

Most Rocky Sub-Neptunes are Molten: Mapping the Solidification Shoreline for Gas Dwarf Exoplanets

Robb Calder,^{1*} Oliver Shorttle,^{1,2} Harrison Nicholls,^{1,3} Tim Lichtenberg,⁴ Claire Marie Guimond³

¹*Institute of Astronomy, University of Cambridge, Madingley Road, Cambridge CB3 0HA, UK*

²*Department of Earth Sciences, University of Cambridge, Downing Street, Cambridge CB2 3EQ, UK*

³*Atmospheric, Oceanic, and Planetary Physics, Department of Physics, University of Oxford, Oxford OX1 3PU, United Kingdom*

⁴*Kapteyn Astronomical Institute, University of Groningen, P.O. Box 800, 9700 AV Groningen, The Netherlands*

Accepted XXX. Received XXX; in original form XXX

ABSTRACT

Sub-Neptunes are the most common type of detected exoplanet, yet their observed masses and radii are degenerate with several interior structures. One possibility is that sub-Neptunes have silicate/iron interiors and H₂-dominated atmospheres, i.e., they are ‘gas dwarfs’. If gas dwarfs have molten interiors, interactions between their magma oceans and atmospheres will produce distinct observational signatures. These signatures may break the degeneracy in interior structure, while providing insight into their interior processes, history, and population trends. We expect all such planets are born molten, but under what conditions do they remain molten today? We use the coupled interior-climate evolution model, PROTEUS, to estimate the ‘solidification shoreline’: the instellation flux boundary (as a function of stellar T_{eff}) that separates molten gas dwarfs from solidified ones. Our results show that 98% of detected sub-Neptunes occupy a region of parameter space consistent with their having permanent magma oceans, if they are gas dwarfs. While mantle $f\text{O}_2$ and bulk volatile C/H ratio both influence magma ocean lifetimes, planets with oxidising mantles and carbon-rich atmospheres are unlikely to have radii consistent with the sub-Neptune classification. Therefore, most detected sub-Neptunes (if they are gas dwarfs) have permanent magma oceans. This result motivates further research into the interactions between molten interiors and overlying atmospheres, and campaigns to identify unambiguous signatures of these interactions.

Key words: planets and satellites: interiors – planets and satellites: physical evolution – planets and satellites: surfaces – planets and satellites: atmospheres – planets and satellites: composition – planets and satellites: terrestrial planets

1 INTRODUCTION

Exoplanets in the sub-Neptune regime ($1.8 R_{\oplus} < R < 4 R_{\oplus}$; [Fulton et al. 2017](#)) are at the forefront of exoplanet research, with no analogue in our Solar System. The Kepler survey has revealed that they are the most common type of discovered exoplanet ([Howard et al. 2012](#); [Thompson et al. 2018](#); [Hsu et al. 2019](#)), so their study is essential for a unified theory of planetary formation and evolution. With the launch of the JWST and the upcoming launch of Ariel ([Changeat et al. 2025](#)), detailed atmospheric characterisation of sub-Neptunes via transmission spectroscopy is now possible. Recent JWST observations of sub-Neptunes have detected molecules such as H₂O, CO₂ and CH₄ at confidence levels ranging from 3σ to 5σ ([Madhusudhan et al. 2023](#); [Benneke et al. 2024](#); [Holmberg & Madhusudhan 2024](#); [Hu et al. 2025](#)). Furthermore, the upcoming PLATO mission is expected to detect several hundred sub-Neptunes that are ideal for follow up observations with JWST and Ariel ([Rauer et al. 2025](#)). Therefore, the study of sub-Neptunes is one of the most important and timely areas of research in planetary science.

One of the most remarkable features of the exoplanet population is the observed bimodality in its radius-period distribution, known as

the ‘radius valley’. The distribution shows a peak at $\sim 1.5 R_{\oplus}$, which corresponds to super-Earths, and at $\sim 2 R_{\oplus}$, which corresponds to sub-Neptunes ([Van Eylen et al. 2018](#); [Berger et al. 2018, 2020](#)). One hypothesis for the origin of this bimodality is that more massive planets, i.e., sub-Neptunes, are able to retain the H₂-dominated atmospheres they accrete from their protoplanetary disc, whereas less massive planets, i.e., super-Earths, lose most of this atmosphere due to photoevaporation ([Owen & Wu 2013](#); [Owen & Wu 2016](#); [Affolter et al. 2023](#)) and/or core-powered mass loss ([Ginzburg et al. 2018](#); [Gupta & Schlichting 2020](#)). This hypothesis predicts that planets in the sub-Neptune regime are ‘gas dwarfs’: planets with silicate–iron interiors and H₂-dominated atmospheres constituting 0.1–10% of the total planetary mass ([Wolfgang & Lopez 2015](#); [Ginzburg et al. 2016](#); [Tang et al. 2024](#)). Another hypothesis is that the radius valley arises primarily due to planet formation, rather than evolution. In this hypothesis, super-Earths form with a low volatile mass fraction, whereas sub-Neptunes form with a high volatile mass fraction, allowing for larger radii ([Zeng et al. 2019](#); [Venturini et al. 2020](#); [Lee & Connors 2021](#); [Burn et al. 2024](#)). This then implies that planets in the sub-Neptune regime could also be ‘water worlds’, with large interior reservoirs of H₂O ([Luque & Pallé 2022](#)).

It is possible that sub-Neptunes, individually and as a population, represent the outcome of both the atmospheric loss and the planetary

* E-mail: rdc49@cam.ac.uk

formation hypotheses (Bean et al. 2021), resulting in a mixed population of volatile-poor worlds as well as volatile-rich worlds. However, distinguishing between these planet types observationally is challenging, given that the masses and radii we observe for sub-Neptunes are consistent with both volatile-rich and volatile-poor interior structures on a population level (Howe et al. 2014; Rogers et al. 2023; Parc et al. 2024). This problem is made more pressing by the fact that a hypothesised subset of the water world population, ‘hycean worlds’ (Madhusudhan et al. 2021; Rigby & Madhusudhan 2024), have been suggested to have habitable surface conditions. This possibility has been argued with claims of potential biosignature detections in the atmosphere of a candidate hycean world (Madhusudhan et al. 2023, 2025; Pica-Ciamarra et al. 2025). Therefore, a framework for constraining the thermal, compositional and structural make-up of sub-Neptunes is essential not only for a comprehensive understanding of planet formation and evolution, but also for refining the bounds on where habitable conditions may be found in the universe (Lichtenberg et al. 2025).

One method for discriminating between gas dwarfs and water worlds is through observations of their upper-atmospheric chemistry via transmission spectroscopy. This is done by searching for observational signatures of specific molecules in the atmospheres of sub-Neptunes (Constantinou & Madhusudhan 2022; Madhusudhan et al. 2023; Benneke et al. 2024), and comparing these observations with models that relate their upper atmospheric chemistry to their composition, structure and climate state (Shorttle et al. 2024; Cooke & Madhusudhan 2024; Wogan et al. 2024; Rigby et al. 2024). For example, it has been suggested that a lack of NH_3 , alongside a detection of CO_2 and CH_4 , in a sub-Neptune atmosphere is evidence of the presence of a liquid water ocean (Madhusudhan et al. 2021, 2023; Hu et al. 2025), the argument being that the dissolution of NH_3 in a liquid-water ocean is the only explanation for its absence alongside CO_2 and CH_4 . However, it has also been suggested that if a gas dwarf were to instead have a magma ocean in contact with its atmosphere, this would also act as a solvent for N-bearing species, complicating the use of an NH_3 non-detection alone to discriminate between the hycean worlds and gas dwarfs (Shorttle et al. 2024; Glein et al. 2025). This means that there is an observational degeneracy between hycean worlds and gas dwarfs with magma oceans.

As well as being observationally degenerate with other structural scenarios for sub-Neptunes, if gas dwarfs were to have surface magma oceans, these magma oceans could offer valuable insights into the interior composition and population statistics of sub-Neptunes. For example, if an H_2 -dominated atmosphere is in contact with a magma ocean, then it will interact chemically with volatiles dissolved in the magma (Schaefer & Fegley 2017; Kite et al. 2019; Tian & Heng 2024; Horn et al. 2025; Miozzi et al. 2025). Studies have shown that the resulting atmospheric composition will be strongly affected by the oxidation state as well as the volatile content of the magma ocean, potentially allowing us to constrain the interior redox conditions of gas dwarfs via transmission spectroscopy (Shorttle et al. 2024; Rigby et al. 2024; Bower et al. 2025) and emission spectroscopy (Nicholls et al. 2025c). Also, the underabundance of sub-Neptunes with $R > 3 R_\oplus$ could be a consequence of the reduction of sub-Neptune radii due to the dissolution of H_2 in magma oceans (Kite et al. 2020). Given the implications magma oceans have for the climate, habitability and atmospheric compositions of gas dwarfs, it is pertinent to ask: how common are magma oceans on gas dwarfs? While all gas dwarfs are expected to have magma oceans at the point of formation, due to the heat from accretion, a subset of the population may experience surface heating rates sufficient to remain molten permanently, rather than solidifying. Therefore, the key issue

is identifying the fraction of the gas dwarf population that have permanent magma oceans.

Determining the fraction of gas dwarfs with molten surfaces requires accurate modelling of their climate and the heat transport within their silicate interiors. Given that gas dwarfs have atmospheric mass fractions much larger than those expected for Earth-sized planets, the greenhouse effect in these atmospheres is expected to produce surface temperatures far exceeding those of Earth-sized planets. This would lead to molten surfaces at instellation fluxes lower than those required to induce molten surfaces on Earth-sized planets (Lichtenberg et al. 2021; Innes et al. 2023; Rigby et al. 2024; Tang et al. 2024). Volatile exchange between the magma and the atmosphere sets the chemical composition of the atmosphere (Schaefer & Fegley 2017; Kite et al. 2019; Tian & Heng 2024; Shorttle et al. 2024; Rigby et al. 2024), which affects the greenhouse heating of the surface (Lichtenberg et al. 2021; Boer et al. 2025), which in turn affects the surface melt-state. Moreover, even if sub-Neptunes start out in a molten state and solidify over time, this cooling is expected to occur over gigayear timescales (Vazan et al. 2018; Tang et al. 2024). The magnitude of these cooling timescales implies that many sub-Neptunes may be young enough (Sandoval et al. 2021; Bean et al. 2021) such that they still have magma oceans today, even if they will eventually solidify. Therefore, a model coupling the evolution of the rocky interior, the climate, and atmospheric chemistry of sub-Neptunes is required to determine what fraction of the gas dwarf population have surface magma oceans.

Previous studies modelling the behaviour of magma oceans on gas dwarfs have either lacked the required coupling between the interior and the atmosphere, or have not considered their time-dependent evolution (Huang et al. 2022; Shorttle et al. 2024; Rigby et al. 2024; Breza et al. 2025; Nixon et al. 2025). Progress has been made to model the thermal evolution of magma oceans on super-Earths and sub-Neptunes (Vazan et al. 2018; Herath et al. 2024), and some of these models couple the evolution of the rocky interior and the atmosphere (Lehmer & Catling 2017; Kubyskhina et al. 2020; Tang et al. 2024). However, none of these models account for volatile exchange between atmospheres and deep magma oceans, nor do they estimate what fraction of the sub-Neptune population has surface magma oceans. Breza et al. (2025) estimated the fraction of sub-Neptunes that may host surface magma oceans using a static interior structure model. However, as stated in their work, a thorough investigation of this research question requires an evolutionary modelling framework.

In this work, we apply a 1D coupled interior-climate evolution model (Lichtenberg et al. 2021; Nicholls et al. 2024, 2025c) to simulate the thermal, compositional and structural evolution of sub-Neptunes, to determine when a permanent magma ocean is a viable steady state outcome. We consider a parameter space composed of stellar effective temperature, planet instellation flux, mantle oxygen fugacity, volatile C/H ratio and planet mass. In section 2, we discuss the modelling framework, PROTEUS, and the empirical scaling relations used in this study. In section 3, we report the results of our main study, where we determine the instellation flux as a function of stellar effective temperature at which the thermal steady state of a gas dwarf transitions from a permanent magma ocean to mantle solidification; i.e., the ‘solidification shoreline’. We also discuss the evolutionary timescales resulting from the models in this study. In section 4, we show the results of our secondary parameter studies, where we investigate the sensitivity of the location of the solidification shoreline to mantle oxygen fugacity, planet mass and the C/H ratio of the bulk volatile inventory. In section 5, we discuss our results in the context of the population of observed sub-Neptunes and

how the solidification shoreline might be developed as a metric. We present our conclusions in section 6.

2 METHODS

To simulate the thermal evolution of gas dwarfs, we use the PROTEUS modelling framework (Lichtenberg et al. 2021; Nicholls et al. 2024, 2025c). PROTEUS self-consistently simulates the coupled thermal evolution of a planet’s interior and atmosphere, assuming that the interior starts in a fully molten state. PROTEUS couples the 1D mantle dynamics model SPIDER with the 1D radiative-convective climate model AGNI (Nicholls et al. 2025b), while allowing for volatile exchange between the semi-molten interior and the atmosphere. Volatile partitioning is calculated with the CALLIOPE model, which uses equilibrium gas chemistry and empirical solubility laws to partition volatiles between the magma and the atmosphere (Lichtenberg et al. 2021; Bower et al. 2022; Nicholls et al. 2025c). The simulated evolution terminates when either the planet solidifies or achieves global energetic equilibrium between stellar irradiation, internal heat production, and atmospheric thermal emission to space. This energetic steady state corresponds to the permanent magma ocean scenario. This modelling framework allows us to simulate the thermal evolution of a gas dwarf self-consistently, allowing for either a solidified mantle or global energetic equilibrium with a semi-molten interior as an thermal steady state.

SPIDER (Bower et al. 2018, 2022) numerically evolves the 1D profiles of temperature, specific entropy, melt fraction, and the thermodynamic quantities of a rocky planet’s mantle as a function of time. It accounts for convective heat flux, conduction, phase separation, gravitational settling and radiogenic heat production. The thermal evolution of the core is not explicitly modelled; it is treated as a heat reservoir at the bottom of the mantle with a constant density of $10,738 \text{ kg m}^{-3}$ and a constant heat capacity of $880 \text{ J K}^{-1} \text{ kg}^{-1}$, corresponding to that of a preliminary reference earth model adjusted for compositional differences (Bower et al. 2018). The radius of the core-mantle system is calculated for a given mass, using an equation of state for pure-MgSiO₃ (Bower et al. 2018). We do not account for the effect of dissolved volatiles on the mantle density. The core interior-radius fraction is set to a value equivalent to that of the Earth: 0.55 (Ladders & Fegley 1998). For the study in which we vary the planet mass (section 4.3), we set the core radius fraction to a value of 0.7 to avoid numerical issues with the solver. These issues arise due to high pressures in the deep mantle, where the thermodynamics and equation of state are poorly constrained (Wolf & Bower 2018). The mantle is assumed to begin its evolution with an adiabatic pressure-temperature profile and a surface heat flux of 10^5 W m^{-2} . The initial entropy is set to $3000 \text{ J K}^{-1} \text{ kg}^{-1}$. Long-term evolution of small planets is relatively insensitive to the exact value of initial entropy as long the internal temperature is fully above the mantle liquidus (Lichtenberg et al. 2021; Schaefer et al. 2016). Orbital evolution and tidal heating are not included in our model. If strong tidal heating were expected in this population, then our simulations would represent a lower limit on interior heat generation (Farhat et al. 2025; Nicholls et al. 2025d).

As in previous modelling, we implement a parametrised conductive boundary layer at the mantle-atmosphere interface (Elkins-Tanton 2008; Nicholls et al. 2024). The temperature at the topmost layer of the mantle is used to calculate the temperature at the bottom of the boundary at each time-step, assuming a boundary layer thickness of 1 cm and a thermal conductivity of $2 \text{ W m}^{-1} \text{ K}^{-1}$ (Nicholls et al. 2024). The surface temperature and mantle melt fraction (by

mass) are then used to determine the chemical composition of the atmosphere using the CALLIOPE equilibrium outgassing model (Bower et al. 2022; Nicholls et al. 2024; Boer et al. 2025). In this work, we use CALLIOPE to solve for the partial pressures of eight volatile species: H₂, H₂O, CH₄, CO, CO₂, O₂, NH₃ and N₂. Sulfur chemistry is not included due to its negligible impact on climate in a H₂-dominated atmosphere (Hu et al. 2013; Nicholls et al. 2025d). We determine the partial pressures of the volatile species using empirically-derived solubility laws (O’Neill & Eggers 2002; Ardia et al. 2013; Armstrong et al. 2015; Dasgupta et al. 2022; Gaillard et al. 2022; Sossi et al. 2023) and equilibrium thermochemical coefficients (Stull 1965; Chase Jr et al. 1982). We specify the following parameters as input to the calculation of the composition of the outgassed atmosphere: the total mass of hydrogen in the mantle-atmosphere system (as fraction of the mantle mass), the C/H and N/H ratios of the bulk volatile inventory, and the oxygen fugacity (f_{O_2}) of the near-surface mantle. The oxygen fugacity is quantified in our model in log-units relative to the f_{O_2} of the iron-wüstite buffer, which is assumed to only vary with temperature in our model.

Once the composition of the atmosphere has been determined, the pressure-temperature structure of the atmosphere is calculated using the radiative-convective climate model AGNI (Nicholls et al. 2025c,b) at each time step. AGNI calculates the equilibrium temperature structure of the atmosphere by solving for the pressure-temperature profile that minimises the energy flux lost across each layer of the atmosphere, thereby conserving energy both globally and locally, while permitting a net heat flux to be carried. It solves for the temperature profile using a damped Newton-Raphson optimisation method. Fluxes due to convection, radiative heating, latent heat release from condensation, and surface sensible heat exchange are included. Dry convective fluxes are parametrised using mixing length theory under the Schwarzschild criterion with an asymptotic mixing length (Vitense 1953; Robinson & Marley 2014; Joyce & Tayar 2023). Radiative fluxes are calculated using the established SOCRATES radiative transfer suite (Lacis & Oinas 1991; Edwards & Slingo 1996; Sergeev et al. 2023). We use 48 correlated- k bands fitted to line opacity data from the DACE database (Grimm et al. 2021). Rayleigh scattering and collisionally-induced continuum absorption are also included. We assume the atmosphere is isochemical, in line with previous work (Lichtenberg et al. 2021; Boer et al. 2025; Nicholls et al. 2025d). Real gas equations of state are used to calculate the atmosphere height structure: Haldemann et al. (2020) for water; Saumon et al. (1995) for hydrogen; and the Van der Waals equations of state for the other species. The atmosphere consists of discrete pressure layers from 10^{-5} bar down to the total surface pressure as calculated by the CALLIOPE outgassing code. We do not include the effects of atmospheric escape in this work (although see section 5.4).

In this study, we adopt blackbody stellar emission spectra and therefore we do not account for stellar evolution. While libraries of semi-empirical stellar spectra do exist (e.g., France et al. 2016), they lack the resolution in stellar effective temperature that we require in this study. Libraries of synthetic stellar spectra also exist, but many lack data points in the EUV and X-ray regimes (Husser et al. 2013) and others lack synthetic spectra for M-type stars (Wheeler et al. 2023). In any case, while the exact form of the stellar spectrum would be relevant for photochemistry (Jordan et al. 2025), to first order, it is sufficient for our study to use blackbody stellar spectra. We parametrise the stellar radius, and correspondingly the stellar luminosity, as a function of the stellar effective temperature (T_{eff}) using the empirical mass-radius relation from Demircan & Kahraman (1991) ($R_{\star} \propto M_{\star}^{0.945}$), the linear mass-luminosity relation from Eker et al. (2015) ($L_{\star} \propto M_{\star}^{4.04}$) and the Stefan-Boltzmann law. This also

allows us to calculate the stellar instellation flux (F_{ins}) as a function of T_{eff} and the orbital semi-major axis a_{orb} .

The net flux at the topmost layer of the atmosphere calculated by AGNI is used to determine the flux upper-boundary condition for SPIDER in the next iteration of the evolutionary model. PROTEUS repeats this iterative process until either the global volume melt-fraction of the mantle decreases below 1 wt%, corresponding to solidification, or the net flux leaving the planet decreases below 0.2 W m^{-2} , corresponding to global flux balance and a steady state magma ocean. All the quantities presented in this work, excluding those presented in section 5.1, correspond to the time at which the simulated planet achieves either mantle solidification or net flux balance.

3 THE SOLIDIFICATION SHORELINE

In order to determine the fraction of the gas dwarf population with magma oceans, it is necessary to divide the parameter space relevant for sub-Neptunes into two regions: one in which our model predicts that gas dwarfs will reach a thermal steady state that permits a surface magma ocean, and one in which they are predicted to cool to the point of complete mantle solidification. This concept is analogous to the ‘Habitable Zone’ (Huang 1960; Kasting et al. 1993; Kopparapu et al. 2013; Turbet et al. 2023), which seeks to divide the host star effective temperature-instellation flux parameter space into a region where liquid water is stable on the surface of a planet and a region where liquid water is unstable due to freezing. Another example of a division of a multi-dimensional parameter space into two distinct regions is the ‘Cosmic Shoreline’ (Zahnle & Catling 2017; Lustig-Yaeger et al. 2019; Moran et al. 2023; Chatterjee & Pierrehumbert 2024). The cosmic shoreline is a tool to separate the escape velocity-instellation flux parameter space into a region where planets are expected to have atmospheres and a region where planets are not expected to have atmospheres. Therefore, in a similar vein to studies that have developed the cosmic shoreline as a tool, we seek to determine the ‘solidification shoreline’ that separates gas dwarfs with permanent magma oceans and those with solidified mantles.

To determine the location of the solidification shoreline, we must first choose a parameter space on which the solidification shoreline is defined. Given that the melt fraction of the mantle is strongly dependent on the surface temperature (van Hunen 2015; Monteux et al. 2016), and that the surface temperature correlates with the planet’s equilibrium temperature (Kasting et al. 1993; Del Genio et al. 2019; Herath et al. 2024), it is natural to choose the instellation flux as one of the parameters. Estimates of the habitable zone are strongly dependent on the effective temperature of the host star, given that the bond albedo of the planet depends on the wavelength of the incident starlight (Kasting et al. 1993; Kopparapu et al. 2013). It is unclear how sensitive the thermal steady states of gas dwarfs are to the stellar effective temperature when the instellation flux is held constant.

Tang et al. (2024) show that the envelope mass fraction (EMF) of a gas dwarf has a significant effect on the cooling of the mantle. However, the EMF is more difficult to infer from observations of sub-Neptunes than the host star effective temperature and the instellation flux. It is much easier to place a planet in the instellation flux-stellar effective temperature parameter space without requiring more detailed observation of the system. This allows us to identify planets that merit further investigation with only minimal system information. For this reason, we choose to define the solidification

shoreline in T_{eff} vs. F_{ins} parameter space, while also studying the effect of varying the EMF on the location of the shoreline.

3.1 Determining the Solidification Shoreline as a Function of Envelope Mass Fraction

We perform simulations with PROTEUS across the $T_{\text{eff}}-F_{\text{ins}}$ -EMF parameter space to estimate the location of the solidification shoreline. We sample stellar effective temperatures from 2900 K to 6300 K, covering M, K, G and F stars, similar to Turbet et al. (2023). Given that Tang et al. (2024) investigate the cooling of sub-Neptunes for instellation fluxes spanning several orders of magnitude, and that we are interested in instellation fluxes low enough to result in solidification, we sample instellation fluxes from $10^{-1} F_{\oplus}$ to $10^3 F_{\oplus}$. We adopt a value of 1361 W m^{-2} for $F_{\oplus}$, using the IAU definition of the solar luminosity and the astronomical unit. Given that gas dwarfs are expected to form with EMFs ranging over several orders of magnitude (Wolfgang & Lopez 2015; Ginzburg et al. 2016; Burn et al. 2024; Tang et al. 2024), we perform simulations with EMFs of 0.1%, 0.3% and 1%. While gas dwarfs could form with an EMF as high as 10%, the surface pressures that result from an EMF this high are outside the pressure range for the solubility laws used in this work, therefore we do not consider EMFs higher than 1%.

Since we do not model atmospheric escape, a consideration of the evolutionary endpoint of a given planet, with a given EMF, will underestimate the EMF that the planet will have had in its past. The corollary of this is that by fixing the EMF at its end state, the simulated planet would cool more quickly than if the envelope were initially more massive and allowed to lose mass down to its end state. Atmospheric escape would therefore globally shift the solidification shoreline to lower instellation fluxes, for a given EMF.

In our baseline simulations, we set the C/H and N/H mass ratios of the bulk volatile inventory in the mantle to 0.32 and 0.09 respectively, equivalent to those corresponding to $100\times$ solar metallicity (Asplund et al. 2021). We use an oxygen fugacity of $\Delta\text{IW} = -3$, and a planet mass of $5M_{\oplus}$. The volatile hydrogen mass fraction, relative to the mass of the mantle, is set to the value needed to achieve the required EMF, assuming that the entire volatile budget is in the atmosphere rather than the melt.

Using the grid of PROTEUS simulations spanning the $T_{\text{eff}}-F_{\text{ins}}$ parameter space, we determine the instellation flux at which the thermal steady state of the planet transitions from a permanent magma ocean to a solidified mantle; i.e., the instellation flux corresponding to the solidification shoreline, F_{shore} , for each stellar effective temperature, for each EMF. We then fit T_{eff} as a function of $\log_{10}(F_{\text{shore}})$ to a second order polynomial using least-squares regression for each EMF,

$$F_{\text{shore}} = a_0 T_{\text{eff}}^2 + a_1 T_{\text{eff}} + a_2. \quad (1)$$

The coefficients for the polynomial fits are shown in table 1. Any further reference to the location of the solidification shoreline refers specifically to these analytical functions.

For a larger EMF, we find that the solidification shoreline is located at lower instellation fluxes (figure 1), implying that sub-Neptunes with a larger EMF are more likely to have a permanent magma ocean for the same instellation flux. This is expected, given that the H_2 continuum opacity scales with the square of pressure, implying that a higher surface pressure results in a higher atmospheric opacity, increasing the surface temperature. These EMFs are not large enough to result in surface pressures above the threshold required to induce solidification (Breza et al. 2025).

The solidification shoreline exists at lower instellation fluxes for

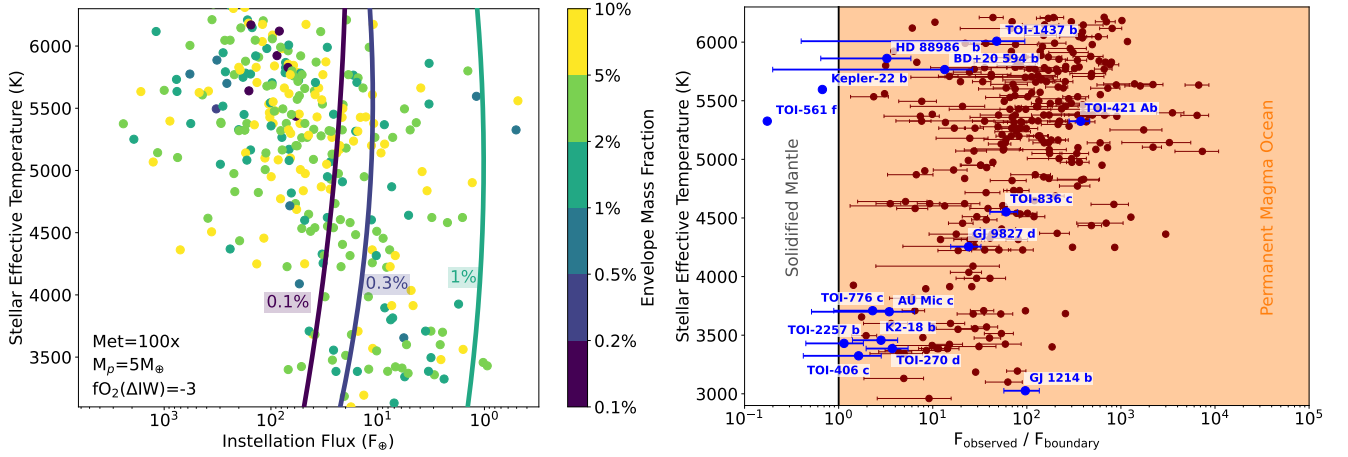


Figure 1. **Left:** Instellation flux at which the thermal steady state of a gas dwarf transitions from a permanent magma ocean to a solidified mantle: the ‘solidification shoreline’. Planets in the region of the parameter space to the left of the shoreline will have permanent magma oceans, and any planets to the right of the shoreline will have solidified mantles. The solidification shoreline is shown for envelope mass fractions of 0.1%, 0.3% and 1%. The positions of all detected sub-Neptunes in this parameter space are shown as scatter points; their colour corresponds to the EMF calculated using equation 2 from [Lopez & Fortney \(2014\)](#). **Right:** Ratio of the observed instellation flux of detected sub-Neptunes (F_{observed}) to the instellation flux required for them to be on their appropriate solidification shoreline (F_{boundary}) as a function of host star effective temperature. Errors in $F_{\text{observed}}/F_{\text{boundary}}$ are propagated from observed masses, host star effective temperature and planet masses. A ratio of $F_{\text{observed}}/F_{\text{boundary}}=1$, corresponding to a planet located on the solidification shoreline, is shown. Sub-Neptunes for which there exist published JWST transmission spectra as well as those in the solidified mantle region are shown in blue, and those without JWST transmission spectra are shown in maroon.

Envelope Mass Fraction (%)	a_0 ($\text{K}^{1/2}$)	a_1 (K)	a_2 (K)
0.1	2.548×10^{-8}	-3.587×10^{-4}	2.557
0.3	5.042×10^{-8}	-5.607×10^{-4}	2.605
1.0	3.909×10^{-8}	-3.951×10^{-4}	1.006

Table 1. Polynomial coefficients for the analytical fit to the instellation flux at the solidification shoreline as a function of stellar effective temperature ($F_{\text{shore}} = a_0 T_{\text{eff}}^2 + a_1 T_{\text{eff}} + a_2$). Coefficients are shown for different envelope mass fractions.

planets around host stars with a higher T_{eff} (the slight increase in F_{shore} with increasing T_{eff} for $T_{\text{eff}} > 5000$ K is a by-product of the polynomial fit to the simulation results), however, this relationship is weak. This insensitivity of the location of the shoreline with respect to T_{eff} contrasts with the strong dependence of the location of the habitable zone on stellar effective temperature ([Huang 1960](#); [Kasting et al. 1993](#); [Kopparapu et al. 2013](#); [Turbet et al. 2023](#)). This is because, unlike studies that estimate the location of the habitable zone, we do not include opacity contributions from clouds, which scatter or absorb incident starlight depending on its wavelength.

The positions of all detected sub-Neptunes (exoplanets with $1.8R_{\oplus} < R < 4R_{\oplus}$) in the $T_{\text{eff}}-F_{\text{ins}}$ parameter space are also shown in figure 1. Data for the planets are taken from the [exoplanet.eu](#) database ([Schneider et al. 2011](#)). The EMFs of the detected sub-Neptunes are estimated using the analytical relationship between EMF, planet radius, planet mass, instellation flux and planet age in [Lopez & Fortney \(2014\)](#). While [Lopez & Fortney \(2014\)](#) state that this analytical law should not be used to predict the EMF for specific planets, we believe that it is a sufficient approximation for our population study. Also, while the equation in [Lopez & Fortney \(2014\)](#) is only valid for pure H/He atmospheres, if it were adapted to account for the carbon- and nitrogen-enriched atmospheres in our simulations, it would only lead to higher predicted EMFs due to larger

atmospheric mean-molecular weights. This increase in EMF would only make our conclusions more robust. We calculate the core radius using equation 2 in their paper, and we neglect the radius of the radiative atmosphere, given that the instellation fluxes corresponding to the solidification shorelines are much lower than the largest instellation fluxes considered in their work. We assume that all planets have an age of 5 Gyr and assume an uncertainty in the age of 3 Gyr.

To quantify the degree to which observed sub-Neptunes are either close to or far away from their appropriate solidification shoreline, we calculate the instellation flux at which the expected thermal steady state transitions from solidified mantle to a permanent magma ocean for each observed sub-Neptune, F_{boundary} . We calculate F_{boundary} for an observed sub-Neptune by predicting the instellation flux it would have at the solidification shoreline corresponding to its estimated EMF based upon the solidification shorelines shown in figure 1, using linear interpolation and extrapolation. The right panel of figure 1 shows the ratio of the observed instellation flux and the instellation flux at the shoreline, $F_{\text{observed}}/F_{\text{boundary}}$, for each observed sub-Neptune. A value greater than 1 for this ratio corresponds to a permanent magma ocean in thermal steady state. We determine the uncertainty for this ratio by propagating the errors for the observed stellar effective temperatures, semi-major axes and the planet masses. We see that, for the majority of the observed sub-Neptunes (98%), $F_{\text{observed}}/F_{\text{boundary}} \gg 1$, indicating that they are very far from their respective solidification shoreline in the $T_{\text{eff}}-F_{\text{ins}}$ parameter space (figure 1, right panel). Only a handful of sub-Neptunes could potentially exist within the ‘solidified mantle’ region of the parameter space, given the uncertainty in their respective flux ratios. Therefore, the majority of detected sub-Neptunes, if they are gas dwarfs with a present EMF equal to or greater than that which they had immediately after their boil-off phase, have permanent magma oceans.

While the EMF, host star effective temperature and instellation flux are the most significant and/or easily observable planetary parameters that determine whether a gas dwarf will solidify, there are

other parameters that could influence its thermal steady state. The location of the solidification shoreline may depend on these secondary parameters; namely the oxygen fugacity of the magma ocean, the C/H ratio of the bulk volatile inventory, and the planet mass. This motivates a study of the effect of varying these secondary parameters on the location of the solidification shoreline, and thus the likelihood of a gas dwarf having a permanent magma ocean.

4 EFFECTS OF SECONDARY PARAMETERS ON THE SOLIDIFICATION SHORELINE

To understand the effect of the oxygen fugacity of the magma ocean, the C/H ratio of the bulk volatile inventory and the planet mass on the location of the solidification shoreline in the $T_{\text{eff}}-F_{\text{ins}}$ parameter space, we perform simulations with PROTEUS in which these parameters are independently varied. In the first and second of these secondary parameter studies, we set the EMF to 0.1%, for computational feasibility. Also, for the first and second studies, we use a stellar effective temperature of 4500 K and an instellation flux of $27 F_{\oplus}$, to ensure that these simulations are located near the solidification shoreline in the $T_{\text{eff}}-F_{\text{ins}}$ parameter space.

To quantify the effect of varying the parameter in question on the location of the solidification shoreline, we record the global melt fraction of the mantle as well as the net atmospheric flux when the planets achieve their thermal steady state: either a solidified mantle or a permanent magma ocean. If varying the parameter results in an increased melt fraction at the end of the thermal evolution for planets with permanent magma oceans, then these planets will require lower instellation fluxes to achieve solidification. Conversely, if varying the parameter increases the net atmospheric flux at the end of the thermal evolution for planets that solidify, then these planets require higher instellation fluxes to achieve flux balance and thus a permanent magma ocean.

4.1 Oxygen Fugacity of the Magma Ocean

The redox state of the mantle exerts significant influence over the speciation of an atmosphere overlaying a magma ocean (Hirschmann 2012; Sossi et al. 2020; Gaillard et al. 2022). Highly reducing mantles result in H_2 dominated atmospheres, with CH_4 and CO as the main secondary constituents, whereas more oxidising mantles result in H_2O - CO_2 dominated atmospheres. Given that the radiation absorption cross-sections of these molecules differ over the wavelength regime in which the host star is most luminous (figure 2), different atmospheric compositions affect the greenhouse heating of the surface induced by the atmosphere, which will affect the thermal evolution of the planet's magma ocean (Nicholls et al. 2024). As a result, the location of the solidification shoreline will change for gas dwarfs with more- or less-oxidising mantles. To investigate the effect of mantle oxidation state on the thermal steady state of gas dwarfs, we vary the oxygen fugacity relative to the iron-wüstite buffer between $\Delta\text{IW} = -5$ and $\Delta\text{IW} = +5$.

More oxidising mantles lead to a larger global melt fraction and lower net atmospheric flux (figure 3) at the end of the thermal evolution of the planet. This result is due to a larger H_2O mixing ratio in the atmosphere (figure 3); the opacity of H_2O over the relevant wavelength regime is much greater than that due to H_2 - H_2 collision-induced absorption (figure 2). The higher opacity results in an increased downward longwave radiative flux at the bottom of the atmosphere (figure 3), implying greater heating of the surface by the atmosphere and an increased melt fraction. However, the increase in

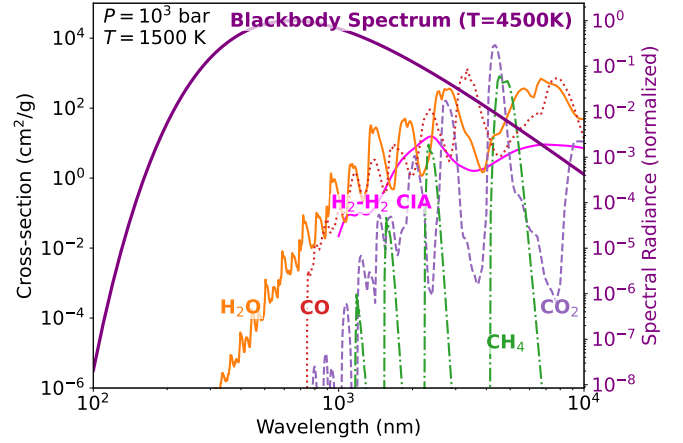


Figure 2. Absorption cross-sections of key species as a function of wavelength for $P = 10^3$ bar and $T = 1500$ K (i.e., the pressure and temperature expected near the base of the atmosphere of a gas dwarf). The cross-section due to H_2 - H_2 collisionally induced absorption is also shown. The blackbody spectrum corresponding to an effective temperature of $T = 4500$ K (i.e., an effective temperature typical of a K-type star) is shown. Absorption cross-sections are taken from the DACE database (Grimm & Heng 2015; Grimm et al. 2021) and cross-sections for H_2 - H_2 collisionally induced absorption are taken from the HITRAN database (Gordon et al. 2022).

melt fraction reaches a maximum at a value of $\Delta\text{IW} = 0$, after which point the melt fraction decreases with increasing oxygen fugacity. This change in behaviour is due to the replacement of CO with CO_2 as the secondary constituent of the atmosphere, which has a smaller absorption cross-section over the relevant wavelength regime than CO (figure 2). The smaller absorption cross-section diminishes the greenhouse effect, leading to a lower global melt fraction.

These results imply that gas dwarfs with more oxidising mantles are more likely to end their thermal evolution with a permanent magma ocean than a solidified mantle, so the solidification shoreline will be pushed to lower instellation fluxes for these planets. This effect only applies, however, to planets with volatile budgets corresponding to $100\times$ solar metallicity. The oxidation state of the mantle could have a different effect on the thermal steady state of gas dwarf under different volatile carbon endowments. In the next section, we investigate the effect of the C/H ratio of the bulk volatile inventory on the location of the solidification shoreline.

4.2 C/H Ratio of the Bulk Volatile Inventory

While we have shown that the redox state of the mantle strongly influences the greenhouse heating of the planetary surface for an H_2 -dominated atmosphere with $100\times$ solar metallicity, gas dwarfs may have atmospheres that are either carbon-rich or carbon-poor compared to this value. The planet's overall carbon endowment at formation may vary due to the metallicity of the host star (Fortney 2012) or the location of the planet in the protoplanetary disc (Li et al. 2021). There is also the possibility of sub-Neptunes with interiors rich in organic molecules; i.e., ‘soot worlds’ (Bergin et al. 2023; Li et al. 2025). Large carbon budgets would result in more volatile carbon outgassing from the magma into the atmosphere. Given the role of molecules such as CH_4 , CO and CO_2 in greenhouse heating of planetary surfaces (Wordsworth & Pierrehumbert 2013), the thermal steady state of gas dwarfs and thus location of the solidification shoreline could depend on the intrinsic carbon budget of the planet.

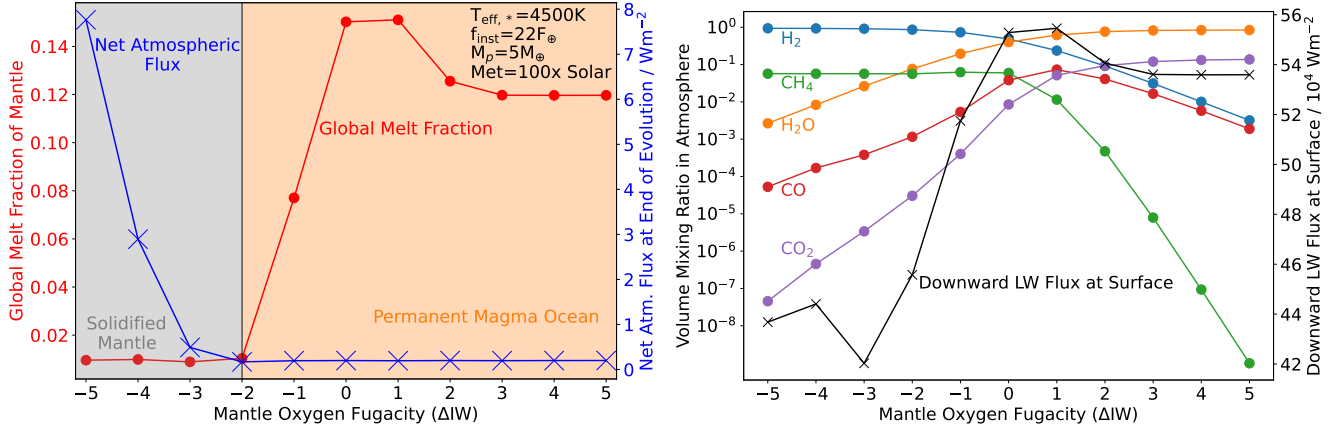


Figure 3. **Left:** Global melt fraction of the mantle (red dots) as well as the net atmospheric flux at the end of the planet’s thermal evolution (blue crosses) as a function of oxygen fugacity of the magma ocean. All simulations with $\Delta IW < -2$ achieve mantle solidification, and all simulations with $\Delta IW > -2$ end their evolution with a permanent magma ocean. The stellar effective temperature, instellation flux, planet mass and metallicity used in these simulations are also shown. The instellation flux was chosen such that the planet was located at the transition from a permanent magma ocean to a solidified mantle in the $T_{eff,*}-F_{inst}$ parameter space. **Right:** Volume mixing ratios of key species in the atmosphere as a function of mantle oxygen fugacity (ΔIW). The downward longwave radiative flux at the base of the atmosphere, which quantifies the rate of surface heating due to the greenhouse effect of the atmosphere, as a function of mantle oxygen fugacity is also shown.

We vary the C/H ratio of the bulk volatile inventory in our model to investigate this dependence.

The volatile C/H ratio of the bulk volatile inventory is varied from 0.1 to 10, encompassing C/H ratios from 100 $\times$ solar metallicity up to the bulk-earth C/H ratio (Wang et al. 2018). We vary the C/H ratio at two fixed values of the oxygen fugacity: $\Delta IW = -4$ and $\Delta IW = +4$, which allows us to explore the speciation of the atmosphere as the C/H ratio is varied for both an oxidising and reducing mantle. We omit nitrogen chemistry in these simulations in order to isolate the effects of carbon chemistry. While this method conserves the total mass of volatile carbon plus hydrogen in the system, it does not conserve the total mass of the atmosphere, given that we do not use a fixed volatile oxygen budget in our simulations. However, the atmospheric masses in these simulations are all consistent to within a factor of 4 (see figures A1 and A2). Therefore, we do not expect these inconsistent atmospheric masses to affect our findings.

When the volatile C/H ratio in the mantle-atmosphere system is increased for a reducing mantle ($\Delta IW = -4$), the net atmospheric flux at the time of solidification increases with increasing carbon volatile fraction (figure 4). This result implies that the planet is further away from the region of the parameter space corresponding to a permanent magma ocean, since the models terminate at mantle solidification, which is not necessarily at radiative equilibrium. If the C/H ratio is increased when $\Delta IW = -4$ for the mantle, we see greater CH_4 volume mixing ratios (figure 5), eventually leading to CH_4 supplanting H_2 as the primary atmospheric constituent. This change in the speciation of the atmosphere leads to a smaller downward longwave radiative flux at the base of the atmosphere (figure 5), which is due to the smaller absorption cross-section of CH_4 relative to that due to the H_2-H_2 continuum (figure 2), thus it also leads to a reduced greenhouse heating rate of the surface.

When the volatile C/H ratio in the mantle-atmosphere system is increased for an oxidising mantle ($\Delta IW = 4$), the global melt fraction of the planet decreases. Similar to the results for a more reducing mantle ($\Delta IW = -4$), an increasing C/H ratio leads to CO_2 replacing H_2O as the primary atmospheric constituent (figure 5). Again, this results in a smaller downward longwave radiative flux at the base of

the atmosphere, and thus a reduced greenhouse heating rate of the surface, given the smaller absorption cross-section of CO_2 compared to H_2O over the relevant wavelength regime (figure 2).

From these simulations, both with oxidising and reducing mantles, we can conclude that gas dwarfs with a larger intrinsic C/H ratio of the bulk volatile inventory are more likely to end their thermal evolution with a solidified mantle due to a diminished greenhouse heating rate of their surfaces. This will result in the solidification shoreline being pushed to higher instellation fluxes.

4.3 Planet Mass

The mass and composition of the atmosphere play a key role in determining the thermal steady state of a gas dwarf, as we have shown in our primary study and our secondary parameter studies. However, the mass of the iron-silicate interior may also play a role, given that previous studies have shown its influence on the thermal evolution of gas dwarfs and partially-molten super-Earths. Herath et al. (2024) show that, for airless planets in the super-Earth regime, a larger planet mass leads to a lower global mantle melt fraction after 10 Gyrs, due to larger pressures in the interior resulting in pressure-induced solidification. The consequences of a larger or smaller silicate mass on the thermal evolution of a gas dwarf are still unclear, however. In this study, we vary the total mass of the planet and investigate its effect on the thermal steady state of gas dwarfs, and thus the location of the solidification shoreline in the $T_{eff,*}-F_{inst}$ parameter space. Note that varying the total planet mass is in effect varying the mass of the silicate interior if the EMF is held constant, given that the volatile inventory and the capacity for the mantle to store volatiles will both increase simultaneously.

We vary the total mass of the planet in our simulation between $2 M_{\oplus}$ and $8 M_{\oplus}$, given that masses larger than $8 M_{\oplus}$ result in deep-mantle pressures outside the range of our $MgSiO_3$ equation of state. We increase the instellation flux to $35 F_{\odot}$ and the EMF to 1%, in order to ensure that the planet is in the ‘permanent magma ocean’ region of the parameter space. We find that the global melt fraction above the middle layer of the mantle, defined by the radial co-ordinate, is

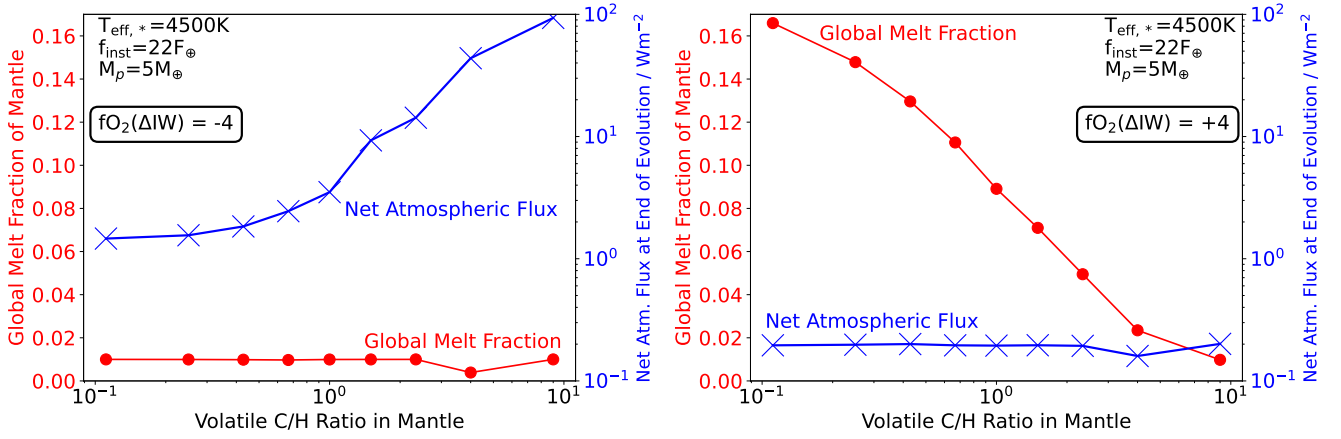


Figure 4. Global melt fraction of the mantle (red dots) as well as the net atmospheric flux at the end of the planet’s thermal evolution (blue crosses) as a function of the C/H ratio of the bulk volatile inventory. Results are shown for simulations with a mantle oxygen fugacity of $\Delta IW = -4$ (left panel) as well as $\Delta IW = 4$ (right panel). All of the simulations corresponding to a mantle oxygen fugacity of $\Delta IW = -4$ end their evolution with a solidified mantle, and all of the simulations corresponding to a mantle oxygen fugacity of $\Delta IW = 4$ end their evolution with a permanent magma ocean. The stellar effective temperature, instillation flux and planet mass used in these simulations are also listed in the upper corners. The instillation flux was chosen such that the planet was located at the transition from a permanent magma ocean to a solidified mantle in the $T_{\text{eff}}-F_{\text{ins}}$ parameter space.

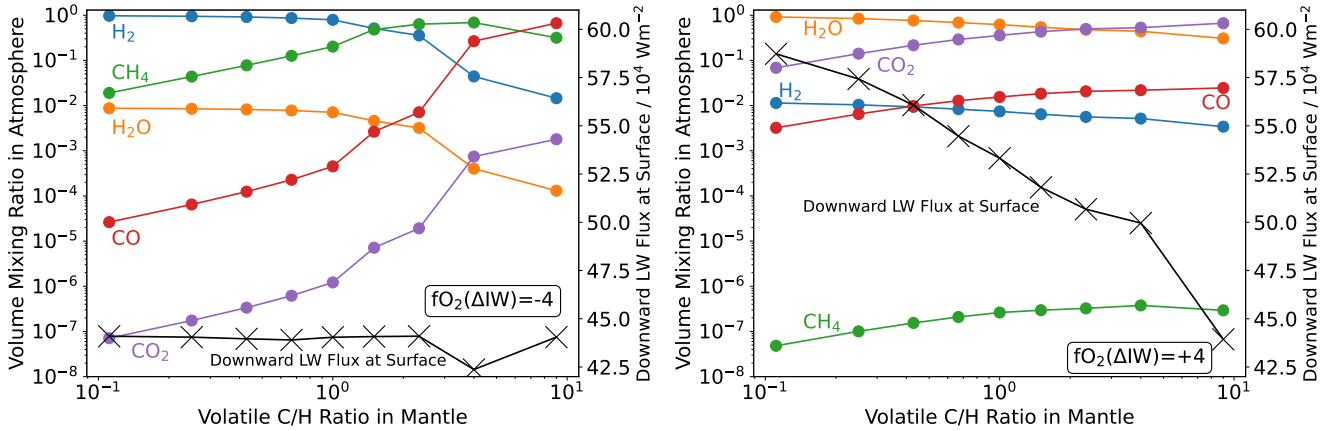


Figure 5. Volume mixing ratios of key species in the atmosphere as a function of the C/H ratio of the bulk volatile inventory. The left and right panels show the results for simulations with mantle oxygen fugacities of $\Delta IW = -4$ and 4 respectively. The downward longwave radiative flux at the base of the atmosphere, which quantifies the rate of surface heating due to the greenhouse effect of the atmosphere, as a function of the C/H ratio of the bulk volatile inventory is also shown.

insensitive to the planet mass (figure 6). While a larger planet mass leads to pressure-induced solidification in the deep mantle (Bower et al. 2018; Herath et al. 2024), the pressures nearer the surface and thus the surface melt fraction are unaffected. Therefore, the likelihood of a gas dwarf possessing a surface magma ocean will be unaffected by its mass, implying that the location of the solidification shoreline is insensitive to planet mass. A caveat to this result is that we do not explore the full mass range expected for sub-Neptunes (i.e., up to and including the mass of Neptune) in this study, therefore we do not capture the full effect of the planet mass on the thermal steady state for gas dwarfs.

5 DISCUSSION

5.1 Sensitivity of the Solidification Shoreline to Evolutionary Timescales

The division of the $T_{\text{eff}}-F_{\text{ins}}$ parameter space for sub-Neptunes into two regions corresponding to permanent magma oceans and solidified mantles shown in figure 1 is only relevant with respect to the thermal steady state of gas dwarfs. The length of their cooling timescales (Vazan et al. 2018; Tang et al. 2024) implies that many of the gas dwarfs in the region of the parameter space corresponding to a solidified mantle may still be cooling, and will thus still have surface magma oceans. This possibility necessitates an investigation into the timescales required to achieve either mantle solidification or a permanent magma ocean.

To study the sensitivity of the location of the solidification shoreline to the age of the planetary system, we show the time taken to

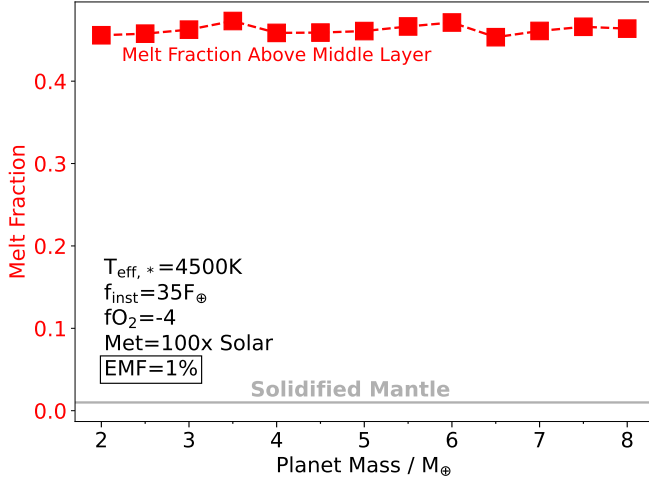


Figure 6. Volume melt fraction in the mantle above the middle layer, defined by the radial co-ordinate, as a function of planet mass. The stellar effective temperature, instellation flux, oxygen fugacity, metallicity and envelope mass fraction used in these simulations are also shown.

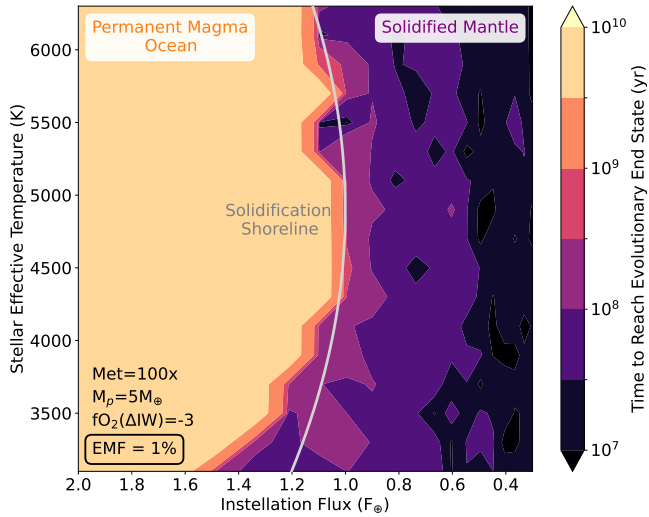


Figure 7. The time taken for gas dwarfs to achieve either a magma ocean steady state or to solidify, as a function of instellation flux and stellar effective temperature. Results are shown for planets modelled with an EMF of 1%. The solidification shoreline is also shown as a white line. The volatile metallicity, planet mass and mantle oxygen fugacity used in these simulations is annotated.

reach a thermal steady state, either a permanent magma ocean or a solidified mantle, for all the simulations in this study with an EMF of 1% (figure 7). For the planets that end their evolution with permanent magma oceans, the timescales required for them to reach a steady state are of order 1 Gyr, an order of magnitude lower than the 10 Gyr cooling timescales reported in Tang et al. (2024). For the planets that end their evolution with solidified mantles, their solidification timescales are of order 100 Myr for $F_{\text{ins}} = 0.8\text{--}1.4 F_{\oplus}$ and of order 10 Myr for lower instellation fluxes: several orders of magnitude shorter than the solidification timescales presented in Tang et al. (2024).

While the timescales presented in our work differ to those reported in Tang et al. (2024), our key result — that most detected sub-

Neptunes will have magma oceans if they are gas dwarfs — still holds regardless of the solidification timescales involved. If the gas dwarfs that are expected to solidify do so over $10^7\text{--}10^8$ yrs, meaning that only the youngest planets will still be solidifying, then most detected sub-Neptunes are still expected to have permanent magma oceans according to the results of our main study. Alternatively, if the gas dwarfs that are expected to solidify do so over Gyr timescales, then there will be a significant fraction of the gas dwarf population with temporary magma oceans in addition to those with permanent magma oceans. In either scenario, the bulk of the detected gas dwarf population is still expected to be molten.

The discrepancy in the gas dwarf solidification timescales presented in this work and in Tang et al. (2024) is a consequence of several key differences in the modelling approaches used in our work and theirs. Tang et al. (2024) treat the envelope in their model as two distinct layers, a radiation-dominated layer and a convection dominated layer, whereas our atmospheric model solves for radiative-convective equilibrium. These different treatments of the atmospheric climate can lead to significantly different cooling timescales (Nicholls et al. 2025a). Furthermore, the timescales shown in Tang et al. (2024) are for gas dwarfs with pure H-He atmospheres, whereas the timescales shown in figure 7 are for planets with atmospheres that are 10% CH_4 by volume. Given the reduced surface heating that results from increasing the CH_4 volume mixing ratio at the expense of the H_2 volume mixing ratio (see section 4.2), gas dwarfs with carbon-enriched atmospheres will have shorter cooling timescales compared to those with pure H-He atmospheres. Finally, the equation of state and material properties for perovskite (i.e., the solid portion of the mantle) used in our work differ from those used in Tang et al. (2024). This will result in different rates of heat transport through the solid material and thus different solidification timescales between the two models.

Regardless of the insensitivity of our conclusions to solidification timescales, it is important to understand the sensitivity of these evolutionary timescales to planetary properties, as their length could affect the atmospheric composition of gas dwarfs, if not the likelihood of them having a permanent magma ocean. Also, if solidification timescales are of order $10^7\text{--}10^8$ yrs, this motivates observational campaigns targeted at young sub-Neptunes (Barat et al. 2025) to study the transition from a magma oceans to a solidified mantle in the sub-Neptune regime. Further work to determine the sensitivity of the location of the solidification shoreline to cooling timescales would help resolve the discrepancy between our solidification timescales and those of Tang et al. (2024), and assist in estimating the melt state of young sub-Neptunes.

5.2 Molten and Solidified Gas Dwarfs in the Mass-Radius Parameter Space

The simulated planets presented in this work, those that have permanent magma oceans as well as those with solidified mantles, span a wide region of the parameter space defined by the stellar effective temperature, instellation flux, EMF, oxygen fugacity, volatile C/H ratio and planet mass. Here, we return to a key question: are these planets still sub-Neptunes, i.e., are their radii consistent with the radius range used to define sub-Neptune exoplanets ($1.8\text{--}4 R_{\oplus}$)?

Studies have shown that the solubility of H_2 in a magma ocean could decrease the radii of molten gas dwarfs, for a given amount of hydrogen, potentially limiting the maximum radii attainable by these planets (Kite et al. 2019). Furthermore, many of the gas dwarfs considered in this work — specifically those in the studies in which the oxygen fugacity and C/H ratio of the bulk volatile inventory is varied — have atmospheres that are not H_2 -dominated. This difference may

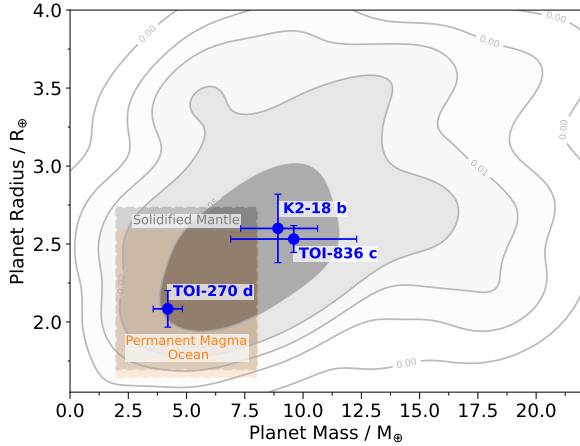


Figure 8. Rectangular boxes showing the range of planetary masses and radii for molten gas dwarfs (orange region) and the solidified gas dwarfs (grey region) modelled in this work. A kernel density contour plot for the mass-radius distribution for all detected sub-Neptunes is also shown. K2-18 b, TOI-270 d and TOI-836 c are also plotted in blue.

result in planetary radii that are below the radius commonly used to define sub-Neptunes. Also, it has been suggested that many solidified planets with Earth-like interiors and H_2 -dominated envelopes are likely to possess thin envelopes due to atmospheric escape, which result in radii that are too small for the planets to be classified as sub-Neptunes (Breza et al. 2025). In this section, we discuss the planetary radii of our simulated gas dwarfs, and how these radii vary with the thermal steady states of these planets as well as their atmospheric compositions.

First, we investigate the *range* of planetary radii that exist for all of our modeled gas dwarfs with permanent magma oceans as well as all those with solidified mantles (figure 8). There is a notable difference between the maximum planetary radius of the solidified gas dwarfs ($2.73 R_\oplus$) and that of the molten gas dwarfs ($2.58 R_\oplus$). This is in agreement with previous studies that suggest that the dissolution of H_2 in a magma ocean introduces a difference in planetary radii between molten gas dwarfs and solidified gas dwarfs (Kite et al. 2019). Also, we find that solidified gas dwarfs can have radii large enough to exist within the sub-Neptune regime, contrary to the findings of Breza et al. (2025). This is because we consider installation fluxes that are much lower than those considered by Breza et al. (2025). This allows planets with more massive envelopes, and thus larger radii, to cool to the point where they can solidify. Therefore, gas dwarfs too cold to sustain magma oceans can be found in the sub-Neptune regime.

In sections 4.1 and 4.2, we have shown that the location of the solidification shoreline may vary for planets with more oxidising mantles or more carbon-rich volatile inventories. However, as the atmospheres of these planets transition from being H_2 -dominated to being dominated by molecules with a higher mean molecular weight, the scale heights of their atmospheres decrease, thus the radii of the planets decrease. For the planets with the most oxidising or carbon-rich atmospheres, atmospheric scale heights decrease to the point where they may no longer be considered sub-Neptunes. For simulations with an EMF of 0.1%, planets with H_2 atmospheric volume mixing ratios below $\sim 60\%$ have planetary radii that fall below $1.8 R_\oplus$ (figure 9), which Berger et al. (2020) use as the threshold for planets to be classified as sub-Neptunes. For simulations with an

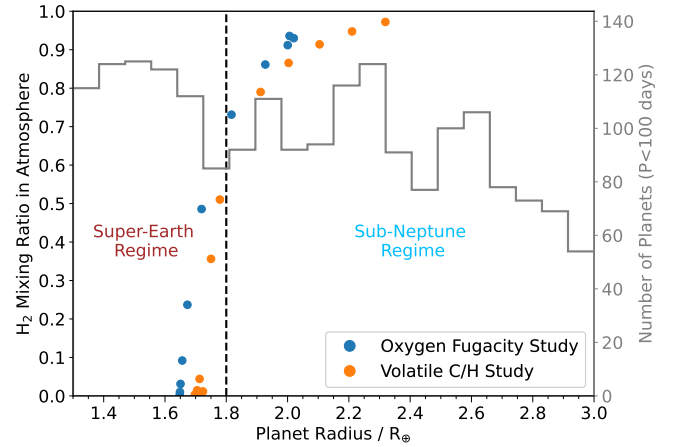


Figure 9. The H_2 mixing ratio as a function of planet radius for the planets from the studies in which the oxygen fugacity (ΔIW) and C/H ratio of the bulk volatile inventory are varied. Blue points correspond to simulations from the study in which the oxygen fugacity is varied, and orange points correspond to simulations in which the C/H ratio of the bulk volatile inventory is varied. Circles and crosses correspond to EMFs of 0.1% and 1% respectively. The distribution of planet radii from the Kepler survey, with orbital periods less than 100 days, are also shown. Planet radii are taken from the exoplanet.eu database (Schneider et al. 2011). The black dashed line denotes the radius at which Berger et al. (2020) define the boundary between the super-Earth and sub-Neptune regime.

EMF of 1%, this H_2 mixing ratio threshold is 20%. Therefore, the simulated planets presented in this work with the most oxidising or carbon-rich atmospheres, which may correspond to solidification shorelines that deviate the most from those presented in figure 1, are not classifiable as sub-Neptunes.

5.3 Sub-Neptunes With Surface Pressures Too Large to Sustain Magma Oceans

One of the main findings in this work is that gas dwarfs with higher envelope mass fractions are more likely to end their thermal evolution with a permanent magma ocean, due to a larger greenhouse heating rate of the surface. However, if the envelope mass fraction increases to a sufficiently high value, the surface pressure will increase to the point where the surface mantle could transition from a molten state to a solid state. Breza et al. (2025) show that, for sub-Neptunes as massive as $16 M_\oplus$ with an atmospheric mean molecular weight of 10 g/mol, an EMF larger than 10% can result in surface pressures large enough to induce a liquid to solid phase change. This implies that the sub-Neptunes shown in figure 1, if their envelopes have a high mean molecular weight and they have an EMF greater than 10%, may be found in the solidified region of the parameter space.

While the existence of such planets is a caveat to our findings, modelling them is outside the scope of this work. The interior component of our model is not capable of simulating planets with masses greater than $8 M_\oplus$ due to the numerical instabilities caused by the large deep-mantle pressures. This motivates experimental work to constrain the behaviour of mantle material under these conditions. Also, Breza et al. (2025) show that surface pressures greater than 100 GPa, well above the 1–10 GPa range for the surface pressures in our simulations, are required to induce a phase change. Such pressures are outside the scope of the solubility laws used in our model and other models (Schlichting & Young 2022; Rigby et al. 2024;

Bower et al. 2025), implying that it is not currently possible to model the solubility of gases in magma at these pressures. Also, at these pressures, hydrogen (Rogers et al. 2025) and H₂O (Ni et al. 2016) will potentially become miscible with the silicate magma. This further complicates the modelling of volatile distribution throughout the planet and may result in depression of the silicate melting curve (Katz et al. 2003; Miozzi et al. 2025).

Finally, it is an open point of discussion as to whether planets with envelope mass fractions as large as 10% and mean molecular weights as high as 10 g/mol should still be classified as gas dwarfs. Benneke et al. (2024) define gas dwarfs as planets with a rocky interior and a low-mean molecular weight atmosphere (<3 g/mol); i.e., a H₂-dominated atmosphere. This definition of a gas dwarf is widely used in the literature surrounding sub-Neptunes (Buchhave et al. 2014; Zeng et al. 2019; Chouqar et al. 2020; Gao et al. 2023; Rigby et al. 2024; Rogers 2025; Bower et al. 2025). Sub-Neptunes with envelope mass fractions and mean molecular weights high enough to prevent a surface magma ocean would be considered ‘metal-rich worlds’ or ‘steam worlds’ in the classification used by Benneke et al. (2024) and Piaulet-Ghorayeb et al. (2024). Nevertheless, the presence and thermal evolution of magma oceans on sub-Neptunes with high mean molecular weight atmospheres is a pertinent research area that requires detailed investigation, primarily because such planets are good targets to observe in transmission spectroscopy with JWST (Glein et al. 2025).

5.4 The Impact of Atmospheric Escape on the Solidification Shoreline

In our modelling framework, we do not account for atmospheric escape through either photo-evaporative or core-powered mass loss. These processes are predicted to significantly erode the atmospheres of close-in sub-Neptunes (Mordasini 2020; Rogers et al. 2023; Tang et al. 2024), reducing the atmospheric mass fractions over the course of their thermal evolution. We have shown that the location of the solidification shoreline is strongly dependent on the atmospheric mass fraction. Therefore, the decrease of the atmospheric mass fraction could reduce the greenhouse heating rate of the surface, which could affect the thermal steady state of a gas dwarf. This would appear to suggest that the lack of atmospheric escape within our modelling framework is a major caveat to our results.

However, while atmospheric escape will affect the thermal evolution of a gas dwarf, it will not affect our finding that most observed sub-Neptunes, if they are gas dwarfs, will be molten. The equation in Lopez & Fortney (2014), as used in this work, predicts the EMF that a sub-Neptune would have currently if it were a gas dwarf, based on its observed radius. If the planet in question were susceptible to atmospheric escape, then its EMF would have been greater earlier in its evolutionary history. The increased EMF would have made it even more likely for the planet to end its thermal evolution with a permanent magma ocean, given the additional surface heating induced by a more massive atmosphere. Similarly, the total volatile inventories in our simulations can be interpreted as the minimum volatile inventory, and thus atmospheric mass fraction, that the simulated planet would have over the course of its evolution. If a real planet had a larger volatile inventory in the past, which had been eroded by atmospheric escape, then the planet would be more likely to have a permanent magma ocean today. Therefore, the solidification shorelines presented in figure 1 can still be used to predict thermal steady states for the detected sub-Neptune population.

5.5 Implications for Sub-Neptunes with Observed JWST Transmission Spectra

Given that 98% of the observed sub-Neptune population lies within the ‘permanent magma ocean’ region of the $T_{\text{eff}}-F_{\text{ins}}$ parameter space, it is germane to ask whether or not observations of sub-Neptunes within this region are consistent with their being gas dwarfs with magma oceans. In figure 1, we show the locations of several sub-Neptunes with published JWST transmission spectra. We consider their observed masses, radii and transmission spectra in the context of our results, and assess their consistency with the molten gas dwarf hypothesis.

Several of these planets, while within the ‘permanent magma ocean’ region of the parameter space in our study, are disfavoured as gas dwarfs. JWST transmission spectra for TOI-421 b (Davenport et al. 2025), GJ 9827 d (Piaulet-Ghorayeb et al. 2024) and GJ 1214 b (Ohno et al. 2025) all favour high mean molecular weight atmospheres for their respective planets. While their surfaces may or may not be molten (cf. Breza et al. 2025), some works may not classify them as gas dwarfs on the basis of these high mean molecular weights; therefore, our results are unlikely to be relevant for these planets. Observations of TOI-836 c (Wallack et al. 2024), however, are less conclusive about the mean molecular weight and the chemistry of the atmosphere. The mass and radius constraints for TOI-836 c are consistent with the mass-radius regime that we have identified for molten gas dwarfs (figure 8), and it lies within the ‘permanent magma ocean’ region of the $T_{\text{eff}}-F_{\text{ins}}$ parameter space (figure 1). In light of this finding, future observations of TOI-836 c could be interpreted in the context of a gas dwarf with a magma ocean, alongside other interior structures.

Different observations of TOI-270 d with transmission spectroscopy have been used to infer different conclusions regarding the mean-molecular weight of its atmosphere. Benneke et al. (2024) report a measurement of $\mu = 5.47^{+1.25}_{-1.14}$, which favours a water-rich, mini-Neptune scenario. However, this result may still be consistent with a molten interior and miscible atmosphere (Glein et al. 2025). Conversely, Holmberg & Madhusudhan (2024) argue that the transmission spectrum reported in their work is consistent with a H₂-dominated, low mean molecular weight atmosphere. The mass and radius constraints for TOI-270 d are consistent with a molten gas dwarf (figure 8), and our results show that TOI-270 d would have a permanent magma ocean if it were a gas dwarf (figure 1). Also, retrievals performed on the transmission spectrum of JWST presented in Holmberg & Madhusudhan (2024) show evidence of CH₄ and CO₂ in the upper atmosphere of TOI-270 d as well as a non-detection of NH₃, which could be explained by the presence of a magma ocean (Shorttle et al. 2024; Rigby et al. 2024). Lastly, Nixon et al. (2025) demonstrate that chemical interactions between a magma ocean and an H₂-rich atmosphere can explain the observational constraints for H₂O and CO₂ for TOI-270 d in Benneke et al. (2024). This leaves the molten gas dwarf hypothesis as a viable explanation for the observations of TOI-270 d.

Finally, we consider the observations of the sub-Neptune K2-18 b in the context of our findings. K2-18 b has been the subject of considerable attention from an observational perspective (Cloutier et al. 2017; Sarkis et al. 2018; Benneke et al. 2019; Madhusudhan et al. 2023, 2025; Hu et al. 2025) and a theoretical perspective (Wogan et al. 2024; Cooke & Madhusudhan 2024; Shorttle et al. 2024; Rigby et al. 2024; Gupta et al. 2025; Liu et al. 2025). Figures 1 and 8 show that the mass, radius and instellation flux constraints for K2-18 b are consistent with a gas dwarf with a permanent magma ocean. Therefore, our results cannot rule out the molten gas dwarf hypothesis

for K2-18 b using these constraints. To further explore this hypothesis for K2-18 b, we must compare self-consistent models of it as a gas dwarf to its observed transmission spectrum. While this has been done previously (Shorttle et al. 2024; Rigby et al. 2024), an evolutionary modelling framework has yet to be used to explore the molten gas dwarf hypothesis for K2-18 b.

5.6 Future Development of the Solidification Shoreline as a Metric

While we have presented the solidification shoreline as a metric for evaluating whether sub-Neptunes are expected to be molten or solid, there is further work to be done in developing this metric. An improvement to our modelling framework would be the inclusion of volatile storage in the solid-phase of the mantle as well as the metallic core, through partitioning into the solid and/or trapping of melt (Tikoo & Elkins-Tanton 2017; Hier-Majumder & Hirschmann 2017; Sim et al. 2024). Incorporating processes such as stellar evolution and orbital migration into an evolutionary modelling framework such as PROTEUS could improve the accuracy of our estimation of the location of the solidification shoreline, given that these processes determine the instellation flux received by the planet. Exploring the full range of planet masses relevant for sub-Neptunes would also refine the robustness of our prediction that most observed gas dwarfs have permanent magma oceans.

Finally, there is the potential for extending the use of the solidification shoreline as a metric for all sub-Neptunes, rather than for gas dwarfs only. This would enable evaluation of the magma ocean hypothesis for sub-Neptunes with high mean molecular weight and/or miscible atmospheres. Such a metric would involve major improvements to our modelling framework, including solubility laws that are valid for the surface pressures expected for these planets and possibly even volatile-silicate miscibility (Rogers et al. 2025). It would also involve accounting for the possibility of large surface pressures preventing the existence of magma oceans within an evolutionary modelling framework (Breza et al. 2025).

6 CONCLUSIONS

The degeneracy in the interior structures of sub-Neptunes necessitates the use of spectroscopy to obtain information about their atmospheric chemistry that can be used to break this degeneracy. If gas dwarfs have surface magma oceans, then the chemical interaction between their magma oceans and their atmospheres could produce chemical signatures that are spectroscopically identifiable. However, it is unclear how common magma oceans are on gas dwarfs. We address this gap by using a 1D coupled climate-interior evolution model to identify the location of the ‘solidification shoreline’: the boundary in instellation flux vs. stellar effective temperature parameter space that separates solidified gas dwarfs from those which maintain permanent magma oceans at steady-state. Our conclusions are as follows:

- 98% of detected sub-Neptunes are in the region of the $T_{\text{eff}}-F_{\text{ins}}$ parameter space corresponding to permanent, steady state magma oceans for gas dwarfs. Therefore, most observed sub-Neptunes, if they are gas dwarfs, have permanent magma oceans. If a sizeable fraction of the sub-Neptune population are gas dwarfs, this motivates theoretical work to understand and predict the consequences of interactions between the magma and the atmospheres of these planets, as well as dedicated observational campaigns to search for chemical signatures of these interactions.

- The most significant parameters that determine the steady state melt fraction of gas dwarfs are the instellation flux and the atmospheric mass fraction. The stellar effective temperature exerts negligible influence on their thermal steady state when the instellation flux is kept constant.

- The C/H ratio of the bulk volatile inventory and the mantle oxygen fugacity exert some influence over the thermal steady state, with higher volatile C/H ratios and more reducing mantles making a solidified mantle a more likely evolutionary outcome. However, gas dwarfs with higher volatile C/H ratios and more oxidising mantles also have higher mean molecular weight atmospheres, giving them radii below the sub-Neptune regime, unless they also have water-rich interiors. While their degree of deep-mantle melting is sensitive to planet mass, the degree of melting near the surface is not; for this reason, the likelihood of gas dwarfs possessing a surface magma ocean will be unaffected by their mass.

- Of the sub-Neptunes for which there are JWST transmission spectra, we find that three planets (TOI-836 c, TOI-270 d, K2-18 b) have masses, radii and instellation fluxes consistent with them being gas dwarfs with permanent magma oceans. Given that low mean molecular weight atmospheres have not been ruled out for these planets, a molten gas dwarf scenario cannot be ruled out for these planets based on their masses, radii and instellation fluxes alone. Future observational constraints on their atmospheric chemistry, combined with self-consistent evolutionary modelling frameworks, are required to further explore the magma ocean hypothesis for these planets.

ACKNOWLEDGMENTS

R.C. thanks the Science and Technology Facilities Council (STFC) for the PhD studentship (grant reference ST/Y509139/1). O.S. acknowledges support from STFC grant UKRI1184. C.M.G. is supported by the STFC (grant reference ST/W000903/1). T.L. was supported by the Branco Weiss Foundation, the Netherlands eScience Center (PROTEUS, NLESC.OEC.2023.017), the Alfred P. Sloan Foundation (AETHER, G202114194), NASA’s Nexus for Exoplanet System Science research coordination network (Alien Earths, 80NSSC21K0593), and the Dutch Research Council NWA-ORC PRELIFE Consortium (NWA.1630.23.013).

DATA AVAILABILITY

The data underlying this article will be shared on reasonable request to the corresponding author. The modelling framework used in this work is fully open-source; there are GitHub pages for PROTEUS (<https://github.com/FormingWorlds/PROTEUS>), AGNI (<https://github.com/nichollsh/AGNI>), SPIDER (<https://github.com/FormingWorlds/spider>), SOCRATES (<https://github.com/nichollsh/SOCRATES>) and CALLIOPE (<https://github.com/FormingWorlds/CALLIOPE>).

REFERENCES

- Affolter L., Mordasini C., Oza A. V., Kubyshkina D., Fossati L., 2023, *A&A*, **676**, A119
- Ardia P., Hirschmann M., Withers A., Stanley B., 2013, *Geochimica et Cosmochimica Acta*, **114**, 52
- Armstrong L. S., Hirschmann M. M., Stanley B. D., Falken E. G., Jacobsen S. D., 2015, *Geochimica et Cosmochimica Acta*, **171**, 283
- Asplund M., Amarsi A. M., Grevesse N., 2021, *A&A*, **653**, A141

- Barat S., et al., 2025, *AJ*, **170**, 165
- Bean J. L., Raymond S. N., Owen J. E., 2021, *Journal of Geophysical Research (Planets)*, **126**, e06639
- Benneke B., et al., 2019, *ApJ*, **887**, L14
- Benneke B., et al., 2024, *arXiv e-prints*, p. [arXiv:2403.03325](#)
- Berger T. A., Huber D., Gaidos E., van Saders J. L., 2018, *ApJ*, **866**, 99
- Berger T. A., Huber D., Gaidos E., van Saders J. L., Weiss L. M., 2020, *AJ*, **160**, 108
- Bergin E. A., Kempton E. M.-R., Hirschmann M., Bastelberger S. T., Teal D. J., Blake G. A., Ciesla F. J., Li J., 2023, *ApJ*, **949**, L17
- Boer I. D., Nicholls H., Lichtenberg T., 2025, *ApJ*, **987**, 172
- Bower D. J., Sanan P., Wolf A. S., 2018, *Physics of the Earth and Planetary Interiors*, **274**, 49
- Bower D. J., Hakim K., Sossi P. A., Sanan P., 2022, *The Planetary Science Journal*, **3**, 93
- Bower D. J., Thompson M. A., Hakim K., Tian M., Sossi P. A., 2025, *arXiv e-prints*, p. [arXiv:2507.00499](#)
- Breza B., Nixon M. C., Kempton E. M. R., 2025, *arXiv e-prints*, p. [arXiv:2509.20429](#)
- Buchhave L. A., et al., 2014, *Nature*, **509**, 593
- Burn R., Mordasini C., Mishra L., Haldemann J., Venturini J., Emsenhuber A., Henning T., 2024, *Nature Astronomy*, **8**, 463
- Changeat Q., et al., 2025, *arXiv e-prints*, p. [arXiv:2509.02657](#)
- Chase Jr M., Curnutt J., Downey Jr J., McDonald R., Syverud A., Valenzuela E., 1982, *Journal of Physical and Chemical Reference Data*, **11**, 695
- Chatterjee R. D., Pierrehumbert R. T., 2024, *arXiv e-prints*, p. [arXiv:2412.05188](#)
- Chouqar J., Benkhaldoun Z., Jabiri A., Lustig-Yaeger J., Soubkiou A., Szentgyorgyi A., 2020, *MNRAS*, **495**, 962
- Cloutier R., et al., 2017, *A&A*, **608**, A35
- Constantinou S., Madhusudhan N., 2022, *MNRAS*, **514**, 2073
- Cooke G. J., Madhusudhan N., 2024, *ApJ*, **977**, 209
- Dasgupta R., Falken E., Pal A., Sun C., 2022, *Geochimica et Cosmochimica Acta*, **336**, 291
- Davenport B., et al., 2025, *ApJ*, **984**, L44
- Del Genio A. D., et al., 2019, *The Astrophysical Journal*, **884**, 75
- Demircan O., Kahraman G., 1991, *Ap&SS*, **181**, 313
- Edwards J. M., Slingo A., 1996, *Quarterly Journal of the Royal Meteorological Society*, **122**, 689
- Eker Z., et al., 2015, *AJ*, **149**, 131
- Elkins-Tanton L. T., 2008, *Earth Planet. Sci. Lett.*, **271**, 181
- Farhat M., Auclair-Desrotour P., Boué G., Lichtenberg T., Laskar J., 2025, *ApJ*, **979**, 133
- Fortney J. J., 2012, *ApJ*, **747**, L27
- France K., et al., 2016, *ApJ*, **820**, 89
- Fulton B. J., et al., 2017, *AJ*, **154**, 109
- Gaillard F., et al., 2022, *Earth and Planetary Science Letters*, **577**, 117255
- Gao P., et al., 2023, *ApJ*, **951**, 96
- Ginzburg S., Schlichting H. E., Sari R., 2016, *ApJ*, **825**, 29
- Ginzburg S., Schlichting H. E., Sari R., 2018, *MNRAS*, **476**, 759
- Glein C. R., Yu X., Luu C. N., 2025, *The Astrophysical Journal*, **985**, 187
- Gordon I. E., et al., 2022, *J. Quant. Spectrosc. Radiative Transfer*, **277**, 107949
- Grimm S. L., Heng K., 2015, *ApJ*, **808**, 182
- Grimm S. L., et al., 2021, *ApJS*, **253**, 30
- Gupta A., Schlichting H. E., 2020, *MNRAS*, **493**, 792
- Gupta A., Stixrude L., Schlichting H. E., 2025, *ApJ*, **982**, L35
- Haldemann J., Alibert Y., Mordasini C., Benz W., 2020, *A&A*, **643**, A105
- Herath M., Boukaré C.-É., Cowan N. B., 2024, *MNRAS*, **535**, 2404
- Hier-Majumder S., Hirschmann M. M., 2017, *Geochemistry, Geophysics, Geosystems*, **18**, 3078
- Hirschmann M. M., 2012, *Earth and Planetary Science Letters*, **341-344**, 48
- Holmberg M., Madhusudhan N., 2024, *A&A*, **683**, L2
- Horn H., Vazan A., Chariton S., Prakapenka V., Shim S.-H., 2025, *Nature*, pp 1–6
- Howard A. W., et al., 2012, *ApJS*, **201**, 15
- Howe A. R., Burrows A., Verne W., 2014, *ApJ*, **787**, 173
- Hsu D. C., Ford E. B., Ragozzine D., Ashby K., 2019, *AJ*, **158**, 109
- Hu R., Seager S., Bains W., 2013, *ApJ*, **769**, 6
- Hu R., et al., 2025, *arXiv e-prints*, p. [arXiv:2507.12622](#)
- Huang S.-S., 1960, *Scientific American*, **202**, 55
- Huang C., Rice D. R., Steffen J. H., 2022, *Monthly Notices of the Royal Astronomical Society*, **513**, 5256
- Husser T. O., Wende-von Berg S., Dreizler S., Homeier D., Reinert A., Barman T., Hauschildt P. H., 2013, *A&A*, **553**, A6
- Innes H., Tsai S.-M., Pierrehumbert R. T., 2023, *ApJ*, **953**, 168
- Jordan S., Shorttle O., Rimmer P. B., 2025, *Science Advances*, **11**, eadp8105
- Joyce M., Tayar J., 2023, *Galaxies*, **11**, 75
- Kasting J. F., Whitmire D. P., Reynolds R. T., 1993, *Icarus*, **101**, 108
- Katz R. F., Spiegelman M., Langmuir C. H., 2003, *Geochemistry, Geophysics, Geosystems*, **4**, 1073
- Kite E. S., Fegley Jr. B., Schaefer L., Ford E. B., 2019, *ApJ*, **887**, L33
- Kite E. S., Fegley Jr. B., Schaefer L., Ford E. B., 2020, *ApJ*, **891**, 111
- Kopparapu R. K., et al., 2013, *ApJ*, **765**, 131
- Kubyskhina D., Vidotto A. A., Fossati L., Farrell E., 2020, *Monthly Notices of the Royal Astronomical Society*, **499**, 77
- Lacis A. A., Oinas V., 1991, *J. Geophys. Res.*, **96**, 9027
- Ladders K., Fegley Bruce J., 1998, *The Planetary Scientist's Companion*. Oxford University Press, doi:10.1093/oso/9780195116946.001.0001, <https://doi.org/10.1093/oso/9780195116946.001.0001>
- Lee E. J., Connors N. J., 2021, *ApJ*, **908**, 32
- Lehmer O. R., Catling D. C., 2017, *The Astrophysical Journal*, **845**, 130
- Li J., Bergin E. A., Blake G. A., Ciesla F. J., Hirschmann M. M., 2021, *Science Advances*, **7**, eabd3632
- Li J., Bergin E. A., Hirschmann M. M., Blake G. A., Ciesla F. J., Kempton E. M. R., 2025, *arXiv e-prints*, p. [arXiv:2508.16781](#)
- Lichtenberg T., Bower D. J., Hammond M., Boukrouche R., Sanan P., Tsai S.-M., Pierrehumbert R. T., 2021, *Journal of Geophysical Research (Planets)*, **126**, e06711
- Lichtenberg T., Shorttle O., Teske J., Kempton E. M.-R., 2025, *Science*, **390**, eads3660
- Liu J., Christie D., Yang J., 2025, *MNRAS*, **541**, 2897
- Lopez E. D., Fortney J. J., 2014, *ApJ*, **792**, 1
- Luque R., Pallé E., 2022, *Science*, **377**, 1211
- Lustig-Yaeger J., Meadows V. S., Lincowski A. P., 2019, *ApJ*, **887**, L11
- Madhusudhan N., Piette A. A. A., Constantinou S., 2021, *ApJ*, **918**, 1
- Madhusudhan N., Sarkar S., Constantinou S., Holmberg M., Piette A. A. A., Moses J. I., 2023, *ApJ*, **956**, L13
- Madhusudhan N., Constantinou S., Holmberg M., Sarkar S., Piette A. A. A., Moses J. I., 2025, *ApJ*, **983**, L40
- Miozzi F., et al., 2025, *arXiv e-prints*, p. [arXiv:2511.01351](#)
- Monteux J., Andraut D., Samuel H., 2016, *Earth and Planetary Science Letters*, **448**, 140
- Moran S. E., et al., 2023, *ApJ*, **948**, L11
- Mordasini C., 2020, *A&A*, **638**, A52
- Ni H., Zhang L., Guo X., 2016, *Science China Earth Sciences*, **59**, 720
- Nicholls H., Lichtenberg T., Bower D. J., Pierrehumbert R., 2024, *Journal of Geophysical Research (Planets)*, **129**, 2024JE008576
- Nicholls H., Lichtenberg T., Chatterjee R. D., Guimond C. M., Postolec E., Pierrehumbert R. T., 2025a, *arXiv e-prints*, p. [arXiv:2507.02656](#)
- Nicholls H., Pierrehumbert R., Lichtenberg T., 2025b, *The Journal of Open Source Software*, **10**, 7726
- Nicholls H., Pierrehumbert R. T., Lichtenberg T., Soucasse L., Smeets S., 2025c, *MNRAS*, **536**, 2957
- Nicholls H., Guimond C. M., Hay H. C. F. C., Chatterjee R. D., Lichtenberg T., Pierrehumbert R. T., 2025d, *MNRAS*, **541**, 2566
- Nixon M. C., et al., 2025, *arXiv e-prints*, p. [arXiv:2510.07367](#)
- O'Neill H. S. C., Eggins S. M., 2002, *Chemical Geology*, **186**, 151
- Ohno K., et al., 2025, *The Astrophysical Journal Letters*, **979**, L7
- Owen J. E., Wu Y., 2013, *ApJ*, **775**, 105
- Owen J. E., Wu Y., 2016, *The Astrophysical Journal*, **817**, 107
- Parc L., Bouchy F., Venturini J., Dorn C., Helled R., 2024, *A&A*, **688**, A59
- Piaulet-Ghorayeb C., et al., 2024, *ApJ*, **974**, L10
- Pica-Ciamarra L., Madhusudhan N., Cooke G. J., Constantinou S., Binet M., 2025, *arXiv e-prints*, p. [arXiv:2505.10539](#)
- Rauer H., et al., 2025, *Experimental Astronomy*, **59**, 26
- Rigby F. E., Madhusudhan N., 2024, *MNRAS*, **529**, 409

- Rigby F. E., et al., 2024, *ApJ*, **975**, 101
 Robinson T. D., Marley M. S., 2014, *ApJ*, **785**, 158
 Rogers J. G., 2025, *MNRAS*, **539**, 2230
 Rogers J. G., Schlichting H. E., Owen J. E., 2023, *ApJ*, **947**, L19
 Rogers J. G., Young E. D., Schlichting H. E., 2025, *arXiv e-prints*, p. [arXiv:2509.13320](#)
 Sandoval A., Contardo G., David T. J., 2021, *Astrophys. J.*, **911**, 117
 Sarkis P., et al., 2018, *AJ*, **155**, 257
 Saumon D., Chabrier G., van Horn H. M., 1995, *ApJS*, **99**, 713
 Schaefer L., Fegley Jr. B., 2017, *ApJ*, **843**, 120
 Schaefer L., Wordsworth R. D., Berta-Thompson Z., Sasselov D., 2016, *Astrophys. J.*, **829**, 63
 Schlichting H. E., Young E. D., 2022, *Planetary Science Journal*, **3**, 127
 Schneider J., Dedieu C., Le Sidaner P., Savalle R., Zolotukhin I., 2011, *A&A*, **532**, A79
 Sergeev D. E., et al., 2023, *Geoscientific Model Development*, **16**, 5601
 Shorttle O., Jordan S., Nicholls H., Lichtenberg T., Bower D. J., 2024, *ApJ*, **962**, L8
 Sim S. J., Hirschmann M. M., Hier-Majumder S., 2024, *Journal of Geophysical Research (Planets)*, **129**, e2024JE008346
 Sossi P. A., Burnham A. D., Badro J., Lanzirotti A., Newville M., O'Neill H. S., 2020, *Science Advances*, **6**, eabd1387
 Sossi P. A., Tollan P. M., Badro J., Bower D. J., 2023, *Earth and Planetary Science Letters*, **601**, 117894
 Stull D. R., 1965, *JANAF Thermochemical Tables*. Vol. 1, Clearinghouse
 Tang Y., Fortney J. J., Nimmo F., Thorngren D., Ohno K., Murray-Clay R., 2024, *arXiv e-prints*, p. [arXiv:2410.21584](#)
 Thompson S. E., et al., 2018, *ApJS*, **235**, 38
 Tian M., Heng K., 2024, *ApJ*, **963**, 157
 Tikoo S. M., Elkins-Tanton L. T., 2017, *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, **375**, 20150394
 Turbet M., et al., 2023, *A&A*, **679**, A126
 Van Eylen V., Agentoft C., Lundkvist M. S., Kjeldsen H., Owen J. E., Fulton B. J., Petigura E., Snellen I., 2018, *MNRAS*, **479**, 4786
 Vazan A., Ormel C. W., Noack L., Dominik C., 2018, *ApJ*, **869**, 163
 Venturini J., Guilera O. M., Haldemann J., Ronco M. P., Mordasini C., 2020, *A&A*, **643**, L1
 Vitense E., 1953, *Z. Astrophys.*, **32**, 135
 Wallack N. L., et al., 2024, *The Astronomical Journal*, **168**, 77
 Wang H. S., Lineweaver C. H., Ireland T. R., 2018, *Icarus*, **299**, 460
 Wheeler A. J., Abruzzo M. W., Casey A. R., Ness M. K., 2023, *AJ*, **165**, 68
 Wogan N. F., Batalha N. E., Zahnle K. J., Krissansen-Totton J., Tsai S.-M., Hu R., 2024, *ApJ*, **963**, L7
 Wolf A. S., Bower D. J., 2018, *Phys. Earth Planet. Inter.*, **278**, 59
 Wolfgang A., Lopez E., 2015, *ApJ*, **806**, 183
 Wordsworth R., Pierrehumbert R., 2013, *Science*, **339**, 64
 Zahnle K. J., Catling D. C., 2017, *ApJ*, **843**, 122
 Zeng L., et al., 2019, *Proceedings of the National Academy of Science*, **116**, 9723
 van Hunen J., 2015, *Geophysical Journal International*, **200**, 1236

This paper has been typeset from a $\text{\LaTeX}$ file prepared by the author.

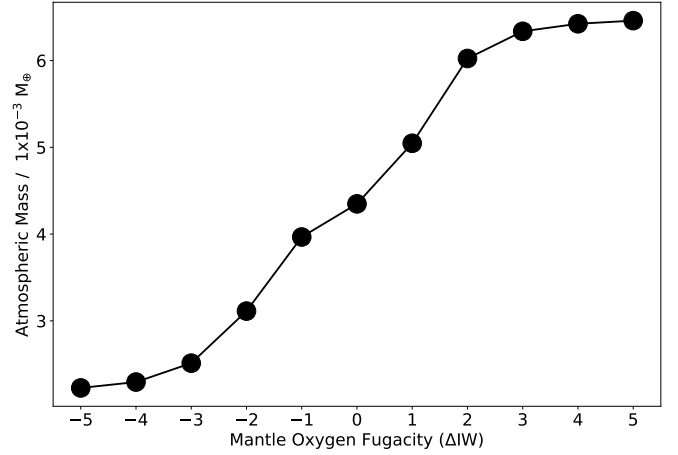


Figure 1. Atmospheric mass as a function of oxygen fugacity (ΔIW) of the mantle for the simulations in which we vary the oxygen fugacity of the magma ocean. The increase in atmospheric mass is due to the transition from an H_2 -dominated atmosphere to an H_2O dominated atmosphere, given the larger mean molecular weight of H_2O compared to H_2 .

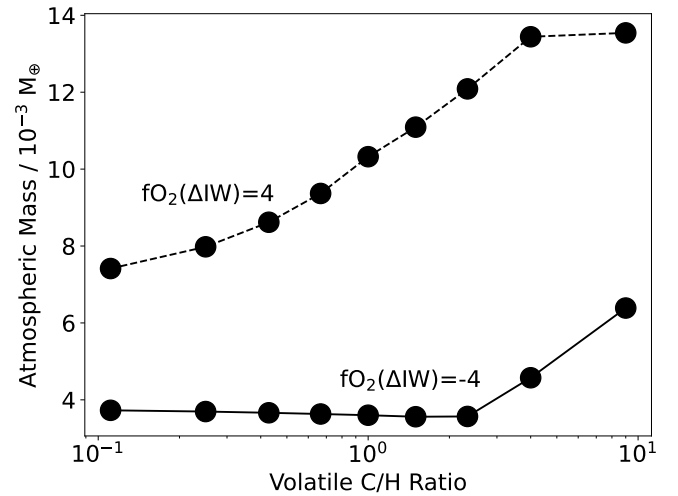


Figure 2. Atmospheric mass as a function of the C/H ratio of the bulk volatile inventory. Results are shown for simulations with a mantle oxygen fugacity (ΔIW) of -4 and 4 as solid and dashed lines respectively.