Climate Diversity in the Habitable Zone due to Varying pN₂ Levels

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Abstract

A large number of studies have responded to the growing body of confirmed terrestrial habitable zone exoplanets by presenting models of various possible climates. However, the impact of the partial pressure of background gases such as N₂ has been poorly-explored, despite the abundance of N₂ in Earth’s atmosphere and the lack of constraints on its typical abundance in terrestrial planet atmospheres. We use PlaSim, a fast 3D climate model, to simulate many hundreds of climates with varying N₂ partial pressures, insolations, and surface characteristics to identify the impact of the background gas partial pressure on the climate. We find that the climate’s response is nonlinear and highly sensitive to the background gas partial pressure. We identify pressure broadening of CO₂ and H₂O absorption lines, amplification of warming or cooling by the water vapor greenhouse positive feedback, heat transport efficiency, and cooling through Rayleigh scattering as the dominant competing mechanisms that determine the equilibrium climate for a given N₂ partial pressure. Finally, we show that different amounts of N₂ should have a significant effect on broadband reflected light observations of terrestrial exoplanets.

Keywords: terrestrial planets, atmospheres, habitability, climate

1. Introduction

The typical starting assumption for climate models of Earth-like planets is either a CO₂-dominated atmosphere like that on Mars and Venus, or an Earth-like atmosphere, with approximately 1 bar of N₂, trace CO₂, and a small amount of water vapor determined by evaporation and precipitation (e.g. Shields et al., 2016; Turbet et al., 2016; Boutle et al., 2017; Wolf, 2017). These are not unreasonable assumptions; our Solar System presents limited opportunities for in-depth studies of habitable zone terrestrial planets with atmospheres. Greenhouse gases like CO₂, rotation rate, eccentricity, and obliquity all have obviously strong effects on the climate (e.g. Manabe and Wetherald, 1975; Williams and Kasting, 1997; Shields et al., 2016; Haqq-Misra et al., 2018), and sophisticated climate models are computationally resource-intensive, limiting the number of models that can be included in sensitivity studies (e.g. Way et al., 2017). CO₂ in particular has known geochemical processes which alter and constrain its atmospheric abundance in relation to other geophysical processes (Walker et al., 1981; Berner, 2004; Pierrehumbert, 2010; Valencia et al., 2018; Nakayama et al., 2019), which makes it an appealing parameter to vary for studies of both habitability and climate evolution (e.g. Kopparapu et al., 2013; Turbet et al., 2016; Paradise and Menou, 2017). While there has been considerable recent progress in understanding N₂ geochemistry (Wordsworth, 2016), N₂ is nonetheless less well-understood, and in addition is very difficult to observe with transit spectroscopy due to a lack of absorption lines at visible or infrared wavelengths (Benneke and Seager, 2012). We focus on N₂ in this study, but this observational problem extends to all spectrally-inactive background gases. Because N₂ is chemically mostly-inert and lacks absorption lines and therefore cannot directly act as a greenhouse gas, its dominant role in the climate is assumed to be as a

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simple background gas pressure that keeps water liquid at Earth-like temperatures, enables heat transport, and causes pressure broadening of the CO$_2$ and H$_2$O absorption lines (Kopparapu et al., 2014; Wordsworth, 2016; Olson et al., 2018; Ramirez, 2018).

However, there has been relatively little work exploring how different N$_2$ partial pressures (pN$_2$) affect the climate, and existing work has not reached a consensus on the overall effect of N$_2$ on the climate. Nakajima et al. (1992) argued that increasing pN$_2$ (or the mass of any spectrally inactive and inert background gas) would likely lead to cooler climates, as the atmospheric lapse rate would trend towards the dry adiabat, resulting in more efficient cooling. Several studies (e.g. Goldblatt et al., 2009; Li et al., 2009; Charnay et al., 2013; Kopparapu et al., 2014; Wolf and Toon, 2014; Chemke et al., 2016) have shown that pressure-broadening of CO$_2$ by N$_2$ should lead to a stronger greenhouse effect, and therefore warming with increased pN$_2$. This might suggest cooler climates at low pN$_2$, but Zahnle and Buick (2016) noted that low surface pressures would result in less effective trapping of water vapor in the the lower atmosphere, as the tropopause is warmer with low surface pressures (Wordsworth and Pierrehumbert, 2014)—increased stratospheric water vapor would then lead to increased total water vapor and a stronger greenhouse effect, suggesting that climates are warmer at low pN$_2$ than at higher pressures. Vladilo et al. (2013) included the effects of heat transport, increased heat capacity, and pressure broadening in an ensemble of one-dimensional energy-balance models, and found that the habitable zone appeared broader at high pressures, due primarily to warming at low insolations and reduced water loss at high insolations. However, they increased pCO$_2$ proportionally with pN$_2$, making it difficult to disentangle warming from N$_2$ and warming from CO$_2$. Chemke and Kaspi (2017) modeled a range of surface pressures with a 3D climate model, but did not include pressure broadening or Rayleigh scattering, examining only the dynamical effects. Most recently, Komacek and Abbot (2019) presented the results of five models computed using ExoCAM, a fully-coupled three-dimensional climate model, with surface pressures ranging from 0.25 to 4 bars. They found a nonlinear response, with warming up to 1 bar, and dramatic cooling at 2 and 4 bars, which they attributed to increased Rayleigh scattering. However, their models did not include CO$_2$, and they note significant disagreement with previous studies (Goldblatt et al., 2009; Li et al., 2009; Charnay et al., 2013; Wolf and Toon, 2014; Chemke et al., 2016) that suggested increased pN$_2$ should cause warming through pressure broadening.

Determining the effect of different N$_2$ partial pressures on the climate is of particular importance for understanding the climate evolution of Earth itself, particularly in the context of the “Faint Young Sun paradox” (Sagan and Mullen, 1972), as a higher N$_2$ partial pressure has been proposed as a partial solution (Goldblatt et al., 2009). However, the actual pN$_2$ during Earth’s early history is a matter of debate. Fossilized raindrops suggest that as late as 2.7 Gya, Earth’s pN$_2$ was less than half its current levels, and that should not represent a significant change from 3–3.5 Gya (Som et al., 2016). However, Johnson and Goldblatt (2017) showed that there was a secular increase in crustal nitrogen during the Precambrian, suggesting that pN$_2$ has been decreasing over time, supporting their finding in Johnson and Goldblatt (2015) that the bulk silicate Earth contains several bars of nitrogen. Emplacing that much nitrogen in the mantle at any time after the hot conditions of the Hadean is difficult (Wordsworth, 2016), though Stüeken et al. (2016) argued that biological burial and release could lead to sharper swings in pN$_2$ than would be permitted by geochemistry alone. A deeper understanding of pN$_2$’s effect on the climate would make it easier to use paleoclimate indicators to help constrain the evolution of Earth’s atmosphere, and to use geological indicators of pN$_2$ to constrain other atmospheric constituents in order to resolve the Faint Young Sun paradox.

In this study, we use PlaSim, a fast 3D climate model, to explore the climates of Earth-sized planets in the habitable zone with pN$_2$ levels ranging from 0.1 to 10 bars. We vary insolation, synchronicity, presence of land, and initial conditions to demonstrate the range of climates permitted by variations in pN$_2$, then perform a series of experiments to thoroughly identify the mechanisms responsible for variations in climate. Finally, we explore the impacts of pN$_2$ on observations of Earth-like exoplanets.

2. Methods

2.1. Climate Model

We use PlaSim, a 3D general circulation model (GCM) designed for Earth-like climates (Fraedrich et al., 2005). PlaSim uses a spectral core to solve the primitive fluid equations for pressure, temperature, divergence, vorticity, and water vapor advection. We use the model in its T21 configuration, in which the atmosphere is discretized into 32 latitudes, 64 longitudes, and 10 vertical levels. The atmosphere is coupled to a surface model, which includes a mixed-layer slab ocean, thermodynamic sea ice, and a soil model on land that allows for variable soil moisture, advection
of excess water to continental margins (the model’s river system), and icy or snowy surfaces. PlaSim’s hydrological cycle is relatively complete, including soil storage, runoff, surface evaporation, storage as atmospheric water vapor, cloud formation, deep and shallow convection, and precipitation as either rain or snow, including re-evaporation of falling precipitation as the water passes through warmer layers. Radiation is computed using a three-band model, with two shortwave bands and one longwave bands. Absorption and transmissivity are computed in each band according to absorber abundances, while clouds are responsible for gray scattering based on cloud fraction and liquid water path. Rayleigh scattering is computed only in the bottom layer. Pressure broadening of each absorbing species is included, and PlaSim assumes that the pressure-broadened effective abundance of each species scales linearly with local pressure (Fraedrich et al., 2005), following Sasamori (1968) and Strong and Plass (1950).

2.2. Modifications to PlaSim

PlaSim does not natively consider varying surface pressure when computing Rayleigh scattering. The scattering transmittance in the bottom layer is computed as direct and diffuse components with prescribed scattering efficiencies (Fraedrich et al., 2005):

\[ T_{\text{direct}} = 1 - \frac{0.219}{1 + 0.816 \mu_0} \]

\[ T_{\text{diffuse}} = 0.856 \]

where \( \mu_0 \) is the cosine of the solar zenith angle. We note that \( T_{\text{diffuse}} = 1 - 0.144 \), and that 0.144 is the actual fit parameter corresponding to PlaSim’s assumed scattering optical depth for diffuse radiation. To adapt this prescription for variable surface pressure, we assume that the scattering optical depth of a column of atmosphere scales mostly linearly with the mass of that column, and therefore surface pressure. According to the Beer-Lambert Law, transmittance is exponentially related to the optical depth \( \tau \) (Ingle and Crouch, 1988):

\[ T = \exp(-\tau) \]

Our modified prescription therefore computes the optical depth, scales it by the column mass relative to that on Earth, and then computes the transmittance:

\[ T' = \exp\left(\frac{p_s g_{e,\oplus}}{p_s g} \ln T_0\right) \]

where \( T_0 \) is the prescribed transmittance for Earth (direct or diffuse), \( p_s \) is the surface pressure, and \( g \) is the surface gravity.

We also modified PlaSim’s vertical discretization to accommodate variable surface pressures. PlaSim uses a sigma system of vertical discretization, where the vertical coordinate is dimensionless, and its value at a given level is simply the ratio of the current pressure to the surface pressure, such that a vertical coordinate of 1 indicates the surface, and 0 indicates the top of the atmosphere (Fraedrich et al., 2005). PlaSim computes its vertical levels with a fourth-order polynomial that is mostly linear between 0 and 1:

\[ \sigma_{h,k} = \frac{3}{4} \sigma_k + \frac{7}{4} \sigma_k^3 - \frac{3}{2} \sigma_k^4 \]

\[ \sigma'_k = \frac{1}{2}(\sigma_{h,k} + \sigma_{h,k-1}) \]

where \( \sigma_k \) is linearly-spaced between 0 and 1, \( \sigma_{h,k} \) is the coordinate of the interface between model layers \( k \) and \( k + 1 \), and \( \sigma'_k \) is the coordinate of layer \( k \)'s mid-point. This scheme is designed to give slightly higher resolution near the surface and near the tropopause, but because it starts from coordinates evenly-spaced in pressure, increasing the surface pressure results in higher pressures for all model layers. With a surface pressure of 10 bars, this means there is almost 1 bar of atmosphere in the model’s top layer. Particularly in the case of radiative phenomenon which depend on optical depth from the top of the atmosphere, such as ozone absorption and heating, this means important physics
can be lost in the low resolution of the top layer. To mitigate this, we simply rescale the layer interface coordinates so that the top interface always occurs at a prescribed pressure $p_{\text{top}}$, or $p_{\text{top}}/p_s$ in sigma-coordinates. We use 50 mbar for most models, ensuring that the top model layer always spans only the top 50 mbar of the atmosphere. For models with surface pressures less than 1 bar, we linearly scaled this down to 5 mbar at $p_s = 0.1$ bar.

We found in our testing that the model was not particularly sensitive to the choice of $p_{\text{top}}$, though choosing a value too low for a high-pressure model could affect model stability. We also experimented with doubling the vertical resolution to 20 levels, and replacing the vertical discretization scheme entirely with one that was logarithmic in pressure, or linearly-spaced in altitude. Simply using a pinned upper layer had very little effect on our results, and doubling the resolution similarly had little effect. Using a logarithmic scheme resulted in higher peak temperatures in the warmest high-pressure models, but reduced PlaSim’s stability while running and required shorter timesteps. We therefore conclude that our initial approach of simply requiring that the top layer not exceed a certain pressure is sufficient.

To verify the validity of these modifications, we compared PlaSim’s shortwave fluxes to those computed with SBDART (Santa Barbara DISORT Atmospheric Radiative Transfer, Ricchiazzi et al., 1998), which computes plane-parallel radiative transfer as a function of wavelength. We used 18 PlaSim models at 1350 W m$^{-2}$ in an aquaplanet configuration, with surface pressures ranging from 0.1 to 9 bars. We used a single column of output from each model, representing a single instant in time at an equatorial gridpoint during local noon, and used that column as input to SBDART. We ran SBDART on that atmospheric column, ignoring clouds and absorbers such as ozone and water vapor, computing broadband fluxes at each layer from 250 to 750 nm. We then ran PlaSim’s shortwave radiation model on that same column, making the same assumptions, and compared the outgoing shortwave fluxes at the top of the atmosphere. The results of this comparison are shown in Figure 1. Our modified model agrees with SBDART at the top of the atmosphere to within a few percent even with surface pressures up to 10 bars, but we note it does consistently predict more downwelling flux at the surface. It is possible to reduce this disagreement by adjusting the constants in Equation 1 and Equation 2, but this would change PlaSim’s overall output when applied to the standard Earth climate. PlaSim has been used extensively for Earth-like climates (e.g. Fraedrich et al., 2005; Boschi et al., 2013; Holden et al., 2016; Nowajewski et al., 2018; D’Errico et al., 2018), and we do not wish to make modifications that would require significant re-tuning and re-validation on the modern Earth climate. We therefore accept the disagreement at the surface, noting its consistency across surface pressure, as a difference in model parameterization, and take the agreement between PlaSim and SBDART in outgoing shortwave flux and therefore total shortwave energy budget as an indication of the validity of our modifications for the present study.

2.3. Modeling Strategy

To assess the effect of $p_{N_2}$ on the climate, we kept $p_{CO_2}$ constant at 360 µbar, and varied $p_{N_2}$ from 0.1 to 10 bars, and insolation from 1100 to 1550 W m$^{-2}$, computing a grid of 160 models. PlaSim does not differentiate between background gases such as O$_2$ and N$_2$, so here we assume that the entirety of the atmosphere’s background gas fraction is N$_2$, and vary $p_{N_2}$ alone by increasing the surface pressure while reducing the CO$_2$ concentration. Since here we define a ‘background gas’ as having negligible contribution to the energy budget through visible or infrared absorption or emission lines, we do not expect the choice of a different background gas to have much impact on our results beyond changing the atmosphere’s mean molecular weight, with the notable exception that H$_2$-dominated atmospheres are likely to behave very differently due to having a larger thermal scale height than a steam atmosphere (Koll and Cronin, 2019). To isolate the dynamical effects of surface pressure and the radiative effects of N$_2$, we assume a fixed background gas mean molecular weight identical to that in Earth’s atmosphere. We run each model until both the surface and top of the atmosphere have reached a stable energy balance, which typically involves 75–200 years of model time. We run PlaSim on 16-core nodes on our local computing cluster, which takes approximately 45 seconds per model year. We assume Earth’s modern continental distribution and topography for each model in this grid, but we also perform several experiments with aquaplanet configurations (no land; sea surface everywhere) to simplify the model and rule out the role of land surface processes. We assume that the ocean mixed layer thickness is 50 meters in all our models.
Figure 1: Comparison of shortwave fluxes computed using SBDART and PlaSim’s shortwave model with our modified Rayleigh scattering and vertical discretization. We ignore clouds and absorbers such as ozone and water vapor. This comparison uses as input a column of PlaSim output at a single instant in time at an equatorial gridpoint from equilibrated PlaSim models at 1350 W m$^{-2}$, spanning surface pressures of 0.1–9 bars. Our modified PlaSim radiation model shows good agreement with SBDART at the top of the atmosphere for a range of surface pressures.

3. Results

As shown in Figure 2, we find that the effect of increasing pN$_2$ on global temperatures is nonlinear: at low pressures, increasing pN$_2$ results in warmer average temperatures, while at higher pressures, increased pN$_2$ can cool the climate, resulting eventually in a transition to a snowball state, where sea ice extends all the way to the equator (Pierrehumbert, 2005). For planets in snowball states, we find increasing pN$_2$ results in monotonic cooling, rather than the warming trend seen on temperate planets, as shown in the right side of the bottom left plot in Figure 2. The warming associated with pN$_2$ is significant; planets with 4 or 5 bars of N$_2$ can be an average of almost 40 K warmer than planets with 1 bar. We further find that the N$_2$ partial pressure at which average surface temperatures reach their maximum varies as a function of insolation, such that the pN$_2$ required to begin cooling the planet is higher at high insolutions than at low insolutions. Maximum and mean surface temperatures in some parts of the parameter space are high enough to potentially indicate a transition to a runaway greenhouse. While PlaSim has been used before to study this transition (Gomez-Leal et al., 2019), we have not validated PlaSim’s performance in both high-temperature and high-pressure regimes, and therefore refrain from concluding that these models represent the inner edge of the habitable zone. We note however that the observed trends do suggest that the inner edge may move to lower insolutions for intermediate surface pressures, and recommend further study.

We note that a similar nonlinear sensitivity to pN$_2$ was observed in ExoCAM models reported in Komacek and Abbot (2019), but their results included a limited number of models and had no CO$_2$, making it difficult to draw conclusions about the mechanisms responsible for the trend from their models alone. The fact that the same qualitative behavior appears in both GCMs however suggests that this is not model-dependent behavior, and instead representative of underlying physics.

Vladilo et al. (2013) argued that the width of the habitable zone should increase with increasing pN$_2$, using results from a 1D EBM. We find however that the width of the habitable zone is likely to decrease at high pressures. We note two important differences between our work and theirs: Vladilo et al. (2013) assumed constant mixing ratios for non-condensible greenhouse gases such as CO$_2$, such that the non-water greenhouse effect always increases with pressure (beyond that expected from pressure broadening), and their EBM did not include Rayleigh scattering as a source of higher albedo at high pressures, although it did include horizontal heat transport. While we do not attempt to calculate the actual edges of the habitable zone as described in Kopparapu et al. (2013), we note that the climate’s sensitivity to
Fast-Rotating Climates with Varying pN$_2$

Figure 2: Mean, minimum, and maximum surface temperatures, along with sea ice extent, for a grid of 160 PlaSim models of Earth-like planets with varying surface pressures. In varying surface pressure, we only varied the Nitrogen partial pressure (pN$_2$)—pCO$_2$ was held constant. The climate demonstrates nonlinear sensitivity to the amount of background gas, showing significant warming in some regimes and significant cooling in others. This is due to warming by pressure broadening, cooling by Rayleigh scattering, and cooling by heat transport all competing with each other. The stars labeled $P_{\text{max}}$ in the upper-left panel indicate the surface pressures at which the mean temperature is maximized. We note that while temperatures likely exceed the greenhouse runaway limit in some regions of the parameter space, we have not evaluated PlaSim’s performance at that transition and therefore cannot conclusively say that high pN$_2$ could lead to a runaway greenhouse.
insolation is higher at high pressures, resulting in a faster decrease in average temperatures with increasing distance from the Sun. This suggests that with increasing pressure the inner edge should move outwards, while the outer edge moves inwards. A possible explanation for this observation is that in the limit of very efficient heat transport, as is found at high pressures, surface temperature gradients are small. A consequence of this fact is that as insolation or greenhouse forcing changes, the climate is less able to buffer the change in energy budget by moving excess heat away from strongly-irradiated regions or replacing heat lost in cool regions. Additionally, a small horizontal temperature gradient means a small decrease in average temperature results in a large amount of ocean surface falling below the freezing point, so the transition from ice-free to snowball is sharper.

4. Discussion

4.1. Mechanisms of Action on the Climate

Cooling from increased N\textsubscript{2} was expected (Nakajima et al., 1992; Komacek and Abbot, 2019), because increased Rayleigh scattering should increase the top-of-atmosphere albedo, increased heat transport efficiency should mean more efficient heat loss to space, and a drier lapse rate should lead to more efficient cooling of the surface and lower atmosphere. However, because N\textsubscript{2} lacks absorption lines in the thermal infrared, increasing pN\textsubscript{2} was not expected to cause significant warming in our models. To investigate the cause of our observed warming trends, we conducted a number of experiments in which model components were selectively removed. In experiments in which water vapor was removed, we increased the effective solar constant to offset the loss of the water vapor greenhouse, assuming a surface temperature of 280 K for a 1 bar atmosphere, such that $T_{\text{eff}} = [I_0(1 - \alpha)/(4\sigma)]^{1/4}$, where $I_0$ is the insolation, $\alpha$ is albedo, and $\sigma$ is the Stefan-Boltzmann constant. A summary of our experiments and their outcomes is given in Table 1.

The primary contribution to cooling at higher pN\textsubscript{2} in our experiments was the reduction in net shortwave heating due to increased reflection by Rayleigh scattering. Poleward heat transport does contribute to cooling on planets with Earth-like rotation, but to a lesser extent—its impact on global average temperatures is only apparent when the other major heating and cooling mechanisms are removed. We found that the warming trend in Figure 2 could only be reversed by eliminating the water vapor greenhouse effect, either by removing water vapor entirely or by removing pressure broadening. Without heat transport or Rayleigh scattering, the global average temperature increases monotonically with increasing pN\textsubscript{2}, until moist runaway conditions are reached. The initial heating amplified by the water vapor greenhouse comes from pressure broadening of the CO\textsubscript{2} and H\textsubscript{2}O absorption bands. The effect of pressure broadening on its own, however is small—no more than 5–10 K between 1 and 10 bars of pN\textsubscript{2}. In experiments with
cooling mechanisms such as Rayleigh scattering or heat transport enabled, the heating effect from pressure broadening is sufficiently weak for the overall trend to be colder temperatures with increased pN$_2$—amplification from the water vapor positive feedback is necessary for pressure-broadening to be the dominant mechanism. In the absence of water vapor, heat transport, scattering, or pressure broadening, a slight cooling trend remains—approximately 0.5 K over 1–10 bars. This cooling trend disappears when CO$_2$ is removed as well, suggesting that the vertical distribution of CO$_2$ in taller atmospheres is very slightly more efficient at cooling the surface. The amount of warming we find from the inclusion of water vapor is probably slightly underestimated, since the amount and altitude of stratospheric warming by ozone is prescribed in PlaSim. In reality, because ozone’s major radiative contribution is absorption of ultraviolet incident light, peak absorption should generally occur around the same pressure level, resulting in ozone absorption occurring at higher altitudes on planets with thicker atmospheres (Wilcox et al., 2012). This should in turn result in a higher and colder tropopause (Wordsworth and Pierrehumbert, 2014), allowing a larger volume of water vapor to mix into the troposphere, thus strengthening the water vapor greenhouse effect at higher pN$_2$.

The relationship between these different mechanisms is very apparent when considering the difference between temperate Earth-like planets and snowball planets, where the surface is entirely covered by sea ice (Pierrehumbert, 2005). Due to the high albedo of sea ice, this state is bistable with temperate, mostly ice-free climates across a range of insolations and CO$_2$ levels (Budyko, 1969; Sellers, 1969). We examined the response of both temperate and snowball climates to differences in pN$_2$, as shown in Figure 3, corresponding to a slice through the grid in Figure 2 at 1300 W m$^{-2}$ for both warm- and cold-starts. The temperate models display the strong, nonlinear response shown in Figure 2, with increased heating through pressure broadening amplified by abundant atmospheric water vapor, and eventual cooling by Rayleigh scattering. On the other hand, the snowball models appear much less sensitive to pN$_2$, varying by only 4 K over 10 bars, despite the inclusion of the water vapor greenhouse effect. This low sensitivity is the result of cold temperatures and sea ice imposing limits on evaporation rates, thus limiting the strength of the water vapor positive feedback. The snowball models also appear less sensitive to cooling by Rayleigh scattering because the planet’s overall albedo is already very high, due to the highly-reflective surface. An increase in atmospheric albedo can therefore do little to raise the overall albedo, resulting in much smaller changes to the planet’s energy budget, as shown in Figure 4. In other words, the additional scattered light on a snowball planet is mostly light that would have been reflected anyway, as opposed to a temperate planet or a dry desert planet with a dark surface where that light would have been mostly absorbed. The importance of surface albedo in determining the climate’s sensitivity to pN$_2$ suggests that any planet with a dark surface, i.e. planets without extensive ice or snow cover, will undergo strong cooling from increased Rayleigh scattering at high pN$_2$. Land planets which are not completely dry, such as those in Abe et al. (2011), would exhibit strong cooling from Rayleigh scattering even at snowball-like temperatures, so long as the available water vapor did not produce extensive snow cover.

These results suggest that the effects of changing pN$_2$ can be largely segregated into three different regimes:

1. **High pN$_2$:** Rayleigh scattering dominates, and adding pN$_2$ cools the climate.
2. **Wet Surface:** A large surface water reservoir permits strong evaporation at low pN$_2$. The water vapor greenhouse effect dominates due to pressure broadening, and adding pN$_2$ warms the climate.
3. **Dry Surface:** Atmospheric water vapor is limited by low evaporation. Rayleigh scattering and heat transport are the dominant effects, and increasing pN$_2$ monotonically cools the climate, even at low pN$_2$.

Therefore on planets with a limited water vapor greenhouse, either due to evaporation limited by sea ice or an overall lack of surface water, increasing pN$_2$ results in colder climates, while the opposite effect is found on planets with large surface water inventories and enough insolation to drive large amounts of evaporation. At sufficiently high pN$_2$, other details of the climate matter less, as Rayleigh scattering is strong enough to cause cooling regardless of the planet’s water inventory. We do not expect a more-realistic layer-by-layer treatment of Rayleigh scattering in PlaSim to change this conclusion, as scattering in a high-pN$_2$ atmosphere would simply occur above where water absorption would occur.

### 4.2. pN$_2$ on Tidally-locked Planets

On planets with Earth-like rotation rates, heat transport is rarely the dominant effect of pN$_2$, as Rayleigh scattering and pressure broadening of trace absorbers both have much stronger impacts on the climate. However, heat transport has a much larger climatic effect on tidally-locked and slow-rotating planets. On these planets, the interplay between
Figure 3: Global mean annual surface temperature and average water column for varying surface pressures, in both warm-start and cold-start (snowball) scenarios. In our cold-start scenarios, the ocean surface is entirely covered in sea ice. Warm-start models display the strongly nonlinear response identified in Figure 2, while snowball models do not. The relatively large availability of water vapor in the temperate models means that pressure broadening of the CO$_2$ and water vapor absorption lines causes a large amount of warming, such that temperatures only decrease with a large amount of Rayleigh scattering at high surface pressures. On snowball planets, however, the limited availability of water vapor results in Pressure broadening causing a small amount of warming at lower surface pressures, but Rayleigh scattering and heat transport quickly dominate, resulting in monotonically-decreasing water vapor abundance. In general, snowball planets are less sensitive to changes in pN$_2$, because the high surface albedo limits the degree to which atmospheric scattering can affect the overall albedo.

Figure 4: Change in shortwave heating due to increased Rayleigh scattering, for temperate and snowball planets at 1300 W m$^{-2}$ and dry land planets at 2000 W m$^{-2}$ with 360 µbars of CO$_2$. The land planets have uniform gray soil albedos of 0.1. The increase in light reflecting off the atmosphere should be the same for all climates (ignoring cloud effects), but the climate sensitivity of snowball planets to Rayleigh scattering is quite a bit lower than that of either temperate or dry land planets with low soil albedos. This is because the overall top-of-atmosphere albedo of a snowball planet is already high, so an increase in reflection from atmospheric scattering primarily reduces how much light would be reflected off the surface, rather than primarily reducing how much light would be absorbed. This results in a smaller change to the energy budget as measured by net shortwave flux at the top of the atmosphere.
Tidally-Locked Climates with Varying pN$_2$

Figure 5: Mean, minimum, and maximum surface temperatures, along with sea ice extent, for a grid of 483 PlaSim models of tidally-locked planets with varying surface pressures. As in Figure 2, the stars labeled $P_{\text{max}}$ in the bottom-left panel indicate the pressure at which the maximum temperature peaks. Only pN$_2$ was varied—pCO$_2$ was held constant. In contrast to Figure 2, on tidally-locked planets pN$_2$ primarily cools the climate at lower insolations, while at higher insolations we see the same nonlinear behavior resulting from competing warming and cooling mechanisms as on Earth-like planets. This difference is because heat transport is much more important on tidally-locked planets, where advection of heat onto the night-side is a major component of the planet’s total energy budget. There appears to be complex structure at high insolations, but as these climates are on the extreme end of what PlaSim can model, more work is required to determine its significance.
these three mechanisms may be more complex. To investigate this, we computed a grid of 483 models with PlaSim in a tidally-locked configuration, with insulations ranging from 400 to 2600 W m\(^{-2}\) and surface pressures ranging from 0.1 to 12 bars. For simplicity, each model was in an aquaplanet surface configuration. PlaSim has been used before to study tidally-locked planets (Menou, 2013; Checlair et al., 2017; Abbot et al., 2018), but we introduced small modifications to ensure that numerical hyper-diffusion timescales would use 24-hour days for unit conversions between days and seconds, rather than simply using the current length of the solar day (which is properly undefined for a tidally-locked planet). While most tidally-locked habitable zone planets orbit low-mass stars with much redder spectra than the Sun, here we assume a solar-like input spectrum. We do not include the effects of a different input spectrum because we wish to isolate the ways in which the geometry and dynamics of tidally-locked planets affect the relative strengths of the mechanisms of action we identified in subsection 4.1. Further work will be required to investigate how a redder spectrum affects the climate’s sensitivity to pN\(_2\). The minimum, mean, and maximum temperatures of our tidally-locked models, along with sea ice fraction, are shown in Figure 5.

On these planets, the climate is very sensitive to the efficiency with which heat can be advected from the dayside to the nightside (Checlair et al., 2017; Haqq-Misra et al., 2018). As pN\(_2\) increases, heat transport efficiency increases, resulting in significant cooling. Combined with increased Rayleigh scattering, warming by pressure broadening only dominates at low pN\(_2\) where heat transport and scattering are weak, and at high insulations where there is high potential for surface evaporation, resulting in very strong amplification through the water vapor greenhouse. The consequence of this difference between high and low insulations is that the surface pressure at which dayside surface temperatures are maximized moves to lower pressures as insolation decreases, as shown in Figure 5. This is consistent with our findings in Figure 2, where we found a similar trend on planets with Earth-like rotation. Here we are using changes in maximum temperature as the metric for the impact on the climate, in contrast to the average surface temperature, which we used in section 3. This is because the geometry of tidally-locked climates and the importance of horizontal heat transport on these planets means that changes in the global average surface temperature may not be particular indicative of trends in the dayside climate. Our finding in section 3 that the habitable zone likely narrows at higher surface pressures seems to hold here as well, with a sharper transition at high pressures from hot to frozen climates than at low surface pressures. In general, we find that the increased importance of heat transport on tidally-locked planets results in increasing pN\(_2\) causing cooling in more of the parameter space than on fast-rotating Earth-like planets at the same pressures and insulations. More work however is necessary to explore the effects of different N\(_2\) partial pressures on the tidally-locked habitable zone, particularly with a redder input spectrum.

### 4.3. Impact of pN\(_2\) on Observables

Our findings in section 3 and subsection 4.2 suggest that Earth-like climates are highly-sensitive to the amount of background gases present in the atmosphere. However, in the case of N\(_2\), the absence of absorption lines in the visible or infrared makes it difficult to quantify the abundance of background gases on terrestrial planets with transit spectroscopy (Benneke and Seager, 2012). Even if the Rayleigh scattering slope as a function of wavelength could be observed in transit spectroscopy, the presence of a cloud deck could limit efforts to constrain the total thickness of an atmosphere (Kaltenegger et al., 2007; Benneke and Seager, 2012; Barstow et al., 2016). Observations using reflected light can more-easily probe deeper into the atmosphere (e.g. Snellen et al., 2015), so we investigated the possibility that pN\(_2\) might have observable signatures in reflected light.

We used the radiative transfer package SBDART (Ricchiazzi et al., 1998) to compute reflectance and emission spectra for our models. Spatially-resolved images of exoplanets are not currently possible, so to compute the disk-averaged spectrum of each model, we run SBDART on each column of the model’s output that would be visible to an observer for a given snapshot in time, using PlaSim’s output to specify water abundances, cloud fractions, surface type, and air temperature, and specifying the solar zenith angle and observer’s viewing angle according to the latitude and longitude of each model column. In most of our models we assume the observer and Sun are both in the same place in the sky over the planet (corresponding to a planet as seen in secondary eclipse). In addition to this equatorial-view geometry, for our tidally-locked models we also compute spectra for a polar-view geometry, where the system is face-on relative to the viewer, such that only half of the viewer-facing hemisphere is illuminated. We use the MODTRAN-3 input solar spectrum, which has a resolution of 20 cm\(^{-1}\). We assume a solar input spectrum for both Earth-like and tidally-locked models for consistency with PlaSim and to isolate the effect of changing pN\(_2\) from that of changing the input spectrum. Because Rayleigh scattering is responsible for the sky’s distinctive blue color, the effect of increased Rayleigh scattering from higher pN\(_2\) should be apparent in broadband photometry. We
Figure 6: B–V photometric colors of Earth-like and tidally-locked planets in reflected light. The Earth-like models (assuming 24-hour rotation) are all computed assuming equatorial, zenith viewing geometries consistent with the planet being in secondary eclipse. Both equatorial and polar viewing geometries are shown for the tidally-locked planets. These fluxes were computed by running SBDART on each column of PlaSim model output. The Earth-like models include both aquaplanet and Earth-like land configurations together with warm- and cold-starts, while the tidally-locked planets are pure aquaplanets. The Earth-like models are computed at an insolation of 1350 W m$^{-2}$ and the tidally-locked models are at 1400 W m$^{-2}$. We find that across terrestrial planet types, planets with higher pN$_2$ are consistently bluer due to increased Rayleigh scattering.

therefore integrate the resulting high-resolution spectra across B and V bands, 445 ± 47 and 551 ± 44 nm (Binney and Merrifield, 1998), assuming flat box-like filter response functions with 100% transmittance.

We did not compute spectra for every single model reported in this study, as our aim is not to solve the inverse problem in a way that permits robust retrieval of pN$_2$ for any terrestrial habitable zone exoplanet, but rather to identify the impact of pN$_2$ on potential observables. Factors such as cloud coverage, sea ice extent, photochemistry, and the colors of different surface rocks will all likely be sources of confusion in efforts to robustly and uniquely retrieve pN$_2$. We computed spectra for planets lacking any land (aquaplanets) and with modern Earth continents (with Africa centered beneath the observer) at 1350 W m$^{-2}$, as well as cold-start “snowball” versions of both surface configurations, varying pN$_2$ from 0.1 to 10 bars for each type of planet. Snowball models that had land had land surfaces almost entirely covered in snow. Tidally-locked planets do not undergo snowball bifurcations in the way that Earth-like planets do (Checlair et al., 2017), so we cannot perform a similar comparison for those planets. However, tidally-locked planets may appear mostly ice-covered when their systems are observed face-on. Direct imaging campaigns may favor systems with polar viewing geometries due to the potential for longer continuous integration times, and tidally-locked planets at moderate insolutions typically have sea ice fractions above 0.5 due to the cold night-side, with open ocean only found near the center of the day-side (Pierrhumbert, 2011; Heng and Vogt, 2011; Yang et al., 2013, 2014). Combined with projection effects reducing the apparent size of unfrozen areas near the limb of the visible disk, this results in mostly-frozen observer-facing hemispheres. Therefore for our tidally-locked models, we only compute spectra for a slice at 1400 W m$^{-2}$ (to capture a range of sea ice fractions) and instead of warm- and cold-start models, consider equatorial and polar viewing geometries for a single set of models.

The B–V colors of each group of models are shown in Figure 6. Increasing pN$_2$ results in stronger reflectance at short wavelengths compared to longer wavelengths, resulting in bluer planets. Very strong reflectance by surface ice seems to partially counteract this, potentially because scattering works in both directions, such that while incident blue light is more likely to scatter off the atmosphere back into space and therefore less likely to reach the ground, blue light reflectively off the surface is also more likely to scatter off the atmosphere and therefore less likely to reach space. Put another way, the component of reflected light reaching the observer that is reflecting off the atmosphere itself is preferentially blue, but the component that is reflecting off the surface is preferentially red. Because surface ice is
much more reflective than land or ocean, snowball planets therefore appear less blue. Once pN_2 is sufficiently high, however, the atmosphere becomes optically-thick to Rayleigh scattering, surface information is obscured, and both temperate and snowball planets appear similarly blue. We observe a similar effect on the B–V color of tidally-locked planets. The similarity in observable trends observed across both Earth-like and tidally-locked models allows us to identify pN_2 as the direct cause of the change in observables, rather than a climatic effect such as a change in cloud cover.

4.4. Caveats

While we believe our results are robust due to agreement with ExoCAM (Komacek and Abbot, 2019) and consistency across a diversity of climates, there are nonetheless a number of model caveats that make any quantitative predictions of specific climates suspect. Perhaps most importantly, PlaSim’s implementation of Rayleigh scattering assumes that all scattering happens at the atmosphere’s bottom layer, which is not physically realistic. The impact on the climate’s energy budget for Earth-like surface pressures and lower is negligible, as the atmosphere’s exponential density profile means most scattering does indeed happen in the lower layers of the atmosphere, below cloud tops, and absorption by the surface is the primary way visible light enters the climate’s energy budget (Pierrehumbert, 2010). However, at higher surface pressures, scattering at higher levels of the atmosphere may become important. In models more realistic than PlaSim, atmospheric absorption of visible and near-infrared light by water vapor and oxygen also contributes to the climate’s energy budget (e.g. Way et al., 2017), and scattering in the middle and upper atmosphere may affect that.

In addition to shortcomings in the scattering parameterization, PlaSim’s ozone abundance is prescribed as a normal distribution peaked at 20 km above the surface. This may be appropriate for Earth’s atmosphere, but in an atmosphere where N_2 and O_2 are well-mixed, the height of peak ozone production will depend on the UV optical depth from the top of the atmosphere, resulting in higher-altitude ozone with higher surface pressures (Wilcox et al., 2012). This should have a large effect on the stratospheric heating profile, possibly resulting in a higher and colder tropopause (Wordsworth and Pierrehumbert, 2014). That may allow for a larger water column and thus additional heating, but more work is needed to assess the effect of ozone layer altitude.

As mentioned in subsection 4.3, we have also not considered the role that photochemistry, hazes, aerosols, and differences in cloud formation efficiency might play in thick N_2 atmospheres. In addition to affecting observables, these factors would likely also affect the climate’s energy budget (Arney et al., 2017; Chen et al., 2018, 2019). Most of these factors are not feasible to study fully with PlaSim, and so would require further work with a more sophisticated model.

Finally, we have assumed a solar-like input spectrum in all of our models. One of the major mechanisms by which N_2 affects the climate is Rayleigh scattering, which is most efficient at short wavelengths (Rybicki and Lightman, 2004). Earth-like planets around lower-mass stars would instead receive most of their light at red and infrared wavelengths, meaning Rayleigh scattering’s contribution to the climate’s energy budget would be lessened considerably. In these climates, we can likely expect cooling through heat transport and warming through pressure broadening to be the dominant ways that N_2 affects the climate, as the reduced importance of Rayleigh scattering at high pN_2 will reduce the impact of pN_2 on shortwave heating.

5. Conclusion

We have used a large ensemble of GCMs in which pN_2 was varied for a range of insolations and initial conditions, for both Earth-like and tidally-locked rotation, and for both aquaplanet and modern Earth land distributions. We found that the climate’s response to pN_2 is strong and nonlinear, with increasing pN_2 leading to warming in some regimes, and cooling in others. We performed a number of experiments to isolate the mechanisms responsible, and conclude that at low and moderate surface pressures on planets with large surface water reservoirs for evaporation, pressure broadening of the CO_2 and H_2O absorption lines leads to significant heating. At high surface pressures and in dry climates, Rayleigh scattering dominates over pressure broadening even in the presence of CO_2, leading to cooling. On planets with very efficient cooling through horizontal heat transport, such as tidally-locked planets, increased heat transport at high surface pressures further leads to cooling, reducing the potential of pressure broadening to cause net warming. Our results are consistent with those reported in Komacek and Abbot (2019), and demonstrate
the relationship between the various seemingly-contradictory mechanisms explored and reported in Nakajima et al. (1992); Goldblatt et al. (2009); Zahle and Buick (2016); Chemke and Kaspi (2017), and Komacek and Abbot (2019).

Finally, we showed that, while pN$_2$ may be difficult to quantify with transit spectroscopy, high pN$_2$ will result in planets appearing ‘bluer’ in reflected light, which may be detectable with broadband reflected light photometry.

The history of Earth’s pN$_2$ is poorly-constrained (Olson et al., 2018), and we have little to no constraints on likely N$_2$ partial pressures on terrestrial exoplanets, particularly those with formation histories and stellar environments different from our own. Even within our own solar system, pN$_2$ varies between terrestrial planets with atmospheres—Venus has over 3 bars of N$_2$ in its atmosphere (Oyama et al., 1980). Our results show however that pN$_2$ is a crucial ingredient in assessing a planet’s climate and habitability. We have determined that pN$_2$ will have an impact on observables accessible through direct imaging and secondary eclipse spectroscopy, but more work is needed to understand its role and evolution on planets with both Earth-like and non-Earth-like geophysical and geochemical properties. Furthermore, while increased pN$_2$ has been proposed as a possible resolution to the ‘faint young Sun paradox’ (Johnson and Goldblatt, 2017), in our sample increased pN$_2$ alone is not able to sufficiently warm an Earth-like planet at Archaean Earth insolation. We have not however fully-explored the parameter space of atmospheric compositions that may have accompanied increased pN$_2$. Theoretical studies of potential climate and habitability of specific exoplanets should therefore include a variety of background gas partial pressures and compositions when assessing the end-members of possible climate scenarios.

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