Habitable Snowballs: Generalizing the Habitable Zone

Adiv Paradise\textsuperscript{a,b,*}, Kristen Menou\textsuperscript{a,b}, Diana Valencia\textsuperscript{a,b}, Christopher Lee\textsuperscript{c}

\textsuperscript{a}Department of Astronomy and Astrophysics, University of Toronto, St. George, Toronto, ON M5S 3H4, Canada
\textsuperscript{b}Centre for Planetary Sciences, Department of Physical and Environmental Sciences, University of Toronto, Scarborough, ON M1C 1A4, Canada
\textsuperscript{c}Department of Physics, University of Toronto, St. George, Toronto, Ontario, M5S 1A7, Canada

Abstract

Habitable planets are commonly imagined to be temperate planets like Earth, with areas of open ocean and warm land. In contrast, planets with colder surfaces and permanent snowball states, where oceans are entirely ice-covered, are believed to be inhospitable. However, we show using a general circulation model that terrestrial habitable zone planets are able to support large unfrozen areas of land even while in a snowball state. These unfrozen regions reach summer temperatures in excess of 10°Celsius and develop their own hydrological cycles. Such conditions permit substantial carbon dioxide weathering, allowing these snowballs to become stable climate states, rather than transient as is commonly assumed. Glaciated planets can thus be habitable, which represents a generalization of the habitable zone concept.

Keywords: terrestrial planets, atmospheres, habitability, astrobiology

1. Introduction

Snowball Climates

Earth-sized exoplanets are often characterized by their orbits relative to their star’s ‘habitable zone’, the region in which surface temperatures can support liquid water. This is usually determined using climate models, with the inner edge as the point where planets transition to inescapably hot runaway greenhouses, such as Venus, and the outer edge as the point where models freeze over even with increased greenhouse gases (Hart, 1979; Kasting et al., 1993). The inner edge and outer edge are usually given in terms of distance from the star or amount of incident light, but can be generalized to other parameters by defining a ‘hot edge’, a ‘soft cold edge’ where the planet fully-glaciates, and a ‘hard cold edge’ where carbon dioxide condenses (e.g. Haqq-Misra et al., 2016; Abbot, 2016; Paradise and Menou, 2017).

Planets beyond the soft cold edge are characterized by oceans completely covered by sea ice, or ‘snowball’ planets. Due to the high albedo of sea ice, these can remain frozen even at high levels of incident sunlight and greenhouse gas concentrations, resulting in a bistability (Budyko, 1969). On fast-rotating planets such as Earth, the coupling between sea ice and surface temperature results in a positive feedback, leading to sharp transitions between snowball and temperate states (Budyko, 1969; Deser et al., 2000). Geological evidence suggests Earth has gone through snowball episodes possibly twice in its early history (Hoffman and Schrag, 2002; Condon et al., 2002; Tajika, 2007). A return to temperate conditions requires global melting, triggered by substantially elevated greenhouse gases or large increases in insolation (Budyko, 1969; Hoffman et al., 1998; Pierrehumbert, 2005).

Snowball episodes are often considered to be problematic for life (e.g. Pierrehumbert, 2005; Lucarini et al., 2013; Haqq-Misra et al., 2016), despite a lack of evidence of decreased biodiversity as a result of Earth’s snowball episodes, and even some evidence of an increase in biodiversity (Corsetti et al., 2006). However, it is not necessarily the case that very low average global temperatures imply sub-freezing conditions everywhere on the planet. Low-dimensional models which deal primarily with global averages may not capture small regions with warmer temperatures. Spiegel et al. (2008) proposed that planets could be partially or temporally habitable (as in, only at certain times of the year), and Linsenmeier et al. (2015) showed using a GCM that this is indeed the case for planets with high obliquity or high eccentricity. In this study, we use an intermediate-complexity GCM to explore the possibility that small-scale melting could occur more generally in Earth-like planets near the snowball deglaciation threshold.

Snowball Stability

On Earth, the planet’s surface temperature is regulated on geological timescales by the carbon-silicate cycle. Carbon dioxide (CO\textsubscript{2}) in the air is dissolved into rainwater and delivered to the surface, where it undergoes a temperature-sensitive weathering reaction with rock to form carbonates. These rocks are eventually recycled into the mantle, where the CO\textsubscript{2} can be regassed into the atmosphere (Berner, 2004; Pierrehumbert, 2010). On planets lacking vascular land plants, the weathering rate is strongly coupled to the surface temperature, so that weathering and outgassing form a negative feedback between the surface temperature and CO\textsubscript{2} level (Walker et al., 1981; Williams and Kasting, 1997; Kump et al., 2000; Pierrehumbert, 2010).

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\*Corresponding author

\textit{Email address: paradise@astro.utoronto.ca (Adiv Paradise)}

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This results in the climate trending towards an equilibrium temperature on geological timescales (Pierrehumbert, 2010). On planets with low outgassing rates, this equilibrium temperature may be below the point at which the climate transitions to snowball states (Pierrehumbert, 2005). Previous studies have proposed that a cessation of weathering due to globally cold conditions could mean that such planets slowly outgas enough CO$_2$ to eventually deglacier, resulting in limit cycles (Menou, 2015; Haqq-Misra et al., 2016; Paradise and Menou, 2017). However, if weathering were able to proceed during snowball episodes as a result of small seasonally-warm regions, it is possible that some planets would become trapped in snowball states. We therefore extend our methods in Paradise and Menou (2017) to measure the weathering rates of such climates, and therefore determine the limits of snowball stability.

2. General Circulation and Weathering Model

PlaSim

We use PlaSim, a 3D general circulation model (GCM) of intermediate complexity to simulate snowball climates. PlaSim uses spectral transforms to solve for vorticity, temperature, divergence, and pressure, and includes a 50-meter mixed-layer slab ocean model, thermodynamic sea ice, a simple soil model, convective and precipitative processes, and a parameterized short- and longwave radiation model (Fraedrich et al., 2005). Physical processes are computed on a grid discretized into latitudes, longitudes, and several vertical levels. PlaSim has been used before to study snowball events on Earth-like planets (e.g. Lucarini et al., 2010; Boschi et al., 2013; Linsenmeier et al., 2015), and is able to reproduce snowball phenomena found by Pierrehumbert (2005) such as increased variability in troposphere lapse rate and tropopause height and weakened overall lapse rates and an on-average lower tropopause. In PlaSim, snowballs are characterized by complete sea ice cover, and deglaciation of a snowball event begins when sea ice starts to melt, triggering an albedo runaway. PlaSim does not account for changes in sea level as the mass of the ice sheets changes. PlaSim’s soil hydrology is described by a bucket model, where each grid cell has a prescribed water capacity. Surface water in excess of this capacity is treated as runoff and advected away according to the local mean topographical slope, forming PlaSim’s river system. Evaporation is driven by both runoff and ground water. We run PlaSim in the T21 configuration, corresponding to a resolution of 5.61° at the equator using 32 latitudes and 64 longitudes.

We assume the same land configuration as modern Earth, along with modern obliquity and eccentricity. As in Paradise and Menou (2017), we sample a range of insolations ranging from 1 S$_{\odot}$ (here assumed to be 1367 W m$^{-2}$) to approximately 0.75 S$_{\odot}$, as well as a range of CO$_2$ partial pressures (hereafter pCO$_2$), using both warm-start initial conditions (temperate conditions similar to modern Earth) and cold-start initial conditions (snowball conditions with a global annual average surface temperature near 220 K, with no above-freezing surface conditions anywhere at any time), increasing our sampling resolution near the snowball transition points. We measure the global weathering rate at each point. The models that equilibrate in snowball conditions just before the deglaciation transition represent the warmest snowballs in our sample, and which we therefore assume to represent the deglaciation threshold.

Weathering Model

We implement weathering at each gridpoint following the technique in Paradise and Menou (2017), itself similar to techniques used in Le Hir et al. (2009) and Edson et al. (2012). In Paradise and Menou (2017), however, we parameterized precipitation rates in terms of surface temperature, rather than treating the role of precipitation directly. Here, we make use of the GCM’s grid-level data for quantities such as precipitation to include it directly. Therefore, our weathering parameterization is given as

$$\frac{W_{\odot}}{W_{\oplus}} = \kappa \left( \frac{p_{CO_2}}{p_{CO_2})} \right)^{\beta} \left( \frac{p}{p_{\oplus}} \right)^{0.65} e^{[k_{act}(T_s-288)]}$$

where $W_{\odot}$ is the weathering rate assumed for modern Earth, $p$ is the precipitation rate, $k_{act}$ is related to the chemical activation energy of the weathering reaction and is set to 0.09 (Berner, 1994, 2001), and $T_s$ is the surface temperature. We use 79 cm yr$^{-1}$ for the precipitation on land, $p_{\oplus}$ (Chen et al., 2002; Schneider et al., 2014). Thus, the weathering increases if the amount of CO$_2$ increases, the rate at which it is delivered to the surface increases, or the energy available for the reaction increases. $\beta$ is set to 0.5 (Menou, 2015; Pierrehumbert, 2010; Kump et al., 2000). This parameterization is intended to represent global quantities and is here being evaluated on local scales, where it may not accurately represent small-scale processes and the actual local weathering rate. Therefore, we add a tuning factor $\kappa$, set to 11.9, to account for the deviation from modern Earth’s weathering rate given modern Earth conditions. $\kappa$ is experimentally determined such that at 1367 W m$^{-2}$ and 330 $\mu$bars, $W = W_{\odot}$. The weathering rate is computed at each cell 4 times per day. To compute the global weathering rate, we take an annual average for each cell and then compute an area-weighted average over the entire land surface. Where relevant, we assume that $W_{\odot} = 50$ bar Gyr$^{-1}$ (Gerlach, 2011; Marty and Tolstikhin, 1998). While outgassing rates may change slowly over time as the planet cools (Zhang and Zindler, 2011; Marty and Tolstikhin, 1998), we simply assume a stable outgassing rate in equilibrium with modern Earth weathering, in line with previous work (e.g. Le Hir et al., 2009; Edson et al., 2012; Menou, 2015).

3. Results

We find that snowball climates in the inner habitable zone are consistently characterized not just by complete sea ice coverage, but also by substantial regions of continental land that undergo seasonal melting and attain summer monthly average temperatures in excess of 10°Celsius. Figure 1a illustrates this behavior in a specific Earth-like model with Modern Earth continents and axial tilt, 24 mbars of CO$_2$, and 1300 W m$^{-2}$ insolation. This model in particular has temperate conditions across 35% of its land surface during northern hemispheric summer.
Snowball Weathering at 1300 W m$^{-2}$ and 24 mbars CO$_2$ -- July

Figure 1: Temperate conditions and weathering during snowball conditions with modern Earth continents. Panel (a): Average summer surface temperature, weathering rate, rainfall, and evaporation for a snowball planet at 1300 W m$^{-2}$, or approximately 1.025 AU, with 24 mbars of CO$_2$. This is almost the CO$_2$ level needed to deglaciate at this insolation. Temperate regions, delineated by the black 0 °C isotherm, experience peak temperatures around 25 °C and average summer temperatures above 10 °C. Localized hydrological cycles enable significant weathering. This particular case has approximately 9% modern Earth weathering once spatially and annually averaged, and 35% of the land is temperate in the summer. Panel (b): Annual average surface temperature for a planet at 1300 W m$^{-2}$, with a hypothetical flat supercontinent (delineated in gray) centered on the equator and below the deglaciation threshold. This planet has approximately 2.6 mbar of pCO$_2$ and a global annual average surface temperature of 242 K. However, approximately 41% of the land surface boasts temperate annual average surface temperatures, as delineated by the black 0 °C isotherm. Some snowball planets may therefore have regions with year-round temperate conditions. Snowball climates with modern Earth continents have annual average temperatures below 0 °C everywhere. This hypothetical planet undergoes approximately 7.5% of modern Earth weathering.

We explore cases which might yield optimal habitability by considering the more favorable conditions of perpetual equinox (zero obliquity) and a large equatorial land mass. If temperate conditions on land are possible during local summer, a planet with zero obliquity and ample land on the equator might support perpetual temperate conditions on land, and therefore be habitable year-round. We repeat our experiment with a rectangular equatorial supercontinent of roughly the same land area as modern Earth’s continents, as shown in Figure 1b. We confirm that this configuration has persistent temperate conditions on land.

We further find that temperate regions support substantial local weathering, as shown in the upper-right panel of Figure 1a, resulting in moderate global weathering particularly at lower insulations, as shown in Figure 2. This suggests that for planets with outgassing rates below a certain threshold, snowball states may be stable on geological timescales, implying that some fraction of observed terrestrial exoplanets may be in snowball states. We speculate that such planets could be more mature, have a cooler interior, and less tectonic and volcanic activity (Zhang and Zindler, 1993), or they could have formed with a smaller carbon inventory than Earth.

Weathering requires liquid surface water. As shown in Figure 1a, the temperate regions feature their own hydrological cycles, with local evaporation from meltwater and soil water resulting in localized precipitation. The weathering permitted by these regions appears to increase with increasing distance...
into the habitable zone, leveling off beyond \( \sim 1100 \text{ W m}^{-2} \). This is the result of the increased CO\(_2\) partial pressure (pCO\(_2\)), cooler temperate regions, and a significant decrease in temperate area at lower insulations. As shown in Figure 3, as pCO\(_2\) increases throughout the habitable zone, the monthly peak and average temperatures for the warm regions decrease. The temperate land area is relatively constant out to \( \sim 1150 \text{ W m}^{-2} \), and then begins to drop steadily. With less weatherable land area available and increasing pCO\(_2\), the weathering rate levels off.

4. Discussion

Effect of Glaciology

Our results depend on the presence and lower albedo of unfrozen, bare ground, which requires that snowpack be melted away entirely. While PlaSim includes the reflective and thermal properties of prescribed glacial surfaces, PlaSim lacks a dynamic ice sheet model and neglects glacial surface elevation. We explore whether this could affect our results by extending PlaSim with a simplistic toy model to incorporate the effects of snow/ice accumulation and elevation on geological timescales.

We account for dynamic ice sheet growth and collapse by extrapolating the 3-year annual average change in snow depth at each cell, advancing forward in time such that the maximum change in depth is no more than 300 meters of liquid water equivalent. We convert the snow depth in liquid water equivalent to glacier height by assuming an average glacial density of 850 kg m\(^{-3}\) and neglecting spreading and deformation. Noting that spreading and deformation would realistically limit ice sheet height, we arbitrarily limit the height to 3.5 km. We use this new additional elevation to recompute the surface geopotential height. We then allow the GCM to relax to the new conditions. If a cell has managed to continually retain at least 2 meters of snow in liquid water equivalent for a full year or has accumulated 30 meters liquid equivalent of snow, we treat that cell’s surface type as glacial for the following simulation year. It should be stressed that this is not a realistic ice sheet or glacier model, but can be useful for assessing the probability of ice sheet collapse on geological timescales in snowball episodes.

We find that when all land surface in a snowball climate is initially covered by ice sheets that are roughly 1.5 km in height, the ice sheets partially collapse on timescales of thousands to tens of thousands of years due to the lack of snowfall, resulting in large deglaciated temperate regions. This is consistent with more robust investigations of the Marinoan snowball Earth in Benn et al. (2015), in which a 3D ice sheet model (GRISLI) was coupled to a sophisticated GCM (LMDz). However, we find that ice sheets taller than 1.5 km do not collapse, resulting in globally cold conditions. Therefore, our results are sensitive to the glacial history and dynamics of a planet that has entered a snowball state.

We do not include changes in sea level in our toy model, so the planet’s water inventory is not conserved. However, 1.5 km of ice on all land points is of the order of 10% of the total mass of Earth’s oceans, suggesting that extreme land glaciation would either require a small land fraction or very deep oceans. Planets with larger water inventories and shallower topological relief might therefore be more susceptible to building massive ice sheets that threaten habitability.
Soil Hydrology and Erosion

We also explore the possibility that geological processes such as groundwater, runoff, and erosion might affect our results. These results depend on the presence of liquid water through a local water cycle. We therefore examine the impact of PlaSim’s treatment of groundwater. PlaSim uses a bucket hydrology for the soil, where each cell has a ‘bucket’ that can hold a prescribed amount of water, with excess treated as runoff. We vary the water capacity of the soil from zero to infinity and find that while the spatial distribution of rainfall and evaporation changes, the weathering rate varies by less than an order of magnitude, such that our qualitative result stands. This is unsurprising, since on large scales evaporation and reaction rates appear to depend primarily on average temperatures through the Arrhenius and Clausius–Clapeyron equations (Berner, 1994). The spatial distribution of evaporation and precipitation should therefore be a minor effect so long as too much water vapor is not transported away from the warm evaporative regions.

Weathering also requires fresh rock, which is supplied by geological processes such as uplifting, volcanic eruptions, and erosion. If erosion rates during snowball climates are low, the availability of unweathered rock could be a limiting factor (West, 2012; Foley, 2015). We test our results’ sensitivity to this by implementing a supply-limited weathering scheme similar to that proposed in West (2012) and Foley (2015), where the weathering rate asymptotically approaches a supply limit set by the erosion rate. In this parameterization, supply-limited weathering is given as

\[ W_d = W_{\text{max}} \left( 1 - e^{-W/W_{\text{max}}} \right) \]  

where \( W_{\text{max}} \) is the maximum weathering possible given a supply limit, in this case assumed to be set by the erosion rate. \( W \) is the weathering that would happen without a supply limit. Weathering rates therefore follow the unlimited formulation when the supply limit is high, and assume the supply limit when weathering is high or the supply limit is low. We assume a globally homogenous erosion rate \( E \), given in units of length over time. Following (Foley, 2015), we define \( W_{\text{max}} \) as

\[ W_{\text{max}} = f_l E_{\text{crust}} \rho_g \tilde{m}_{\text{CO}_2} \tilde{n}_{\text{crust}} \]  

where \( f_l \) is the land fraction, \( g \) is the surface gravity (9.81 m s\(^{-2}\)), \( E_{\text{crust}} \) is the fraction of Mg, Ca, K, and Na in continental crust (taken to be 0.08), \( \rho_g \) is the density of regolith (taken to be 2500 kg m\(^{-3}\)), \( \tilde{m}_{\text{CO}_2} \) is the molar mass of CO\(_2\) (44 g mol\(^{-1}\)), and \( \tilde{n}_{\text{crust}} \) is the average molar mass of Mg, Ca, K, and Na (32 g mol\(^{-1}\)) (West, 2012; Foley, 2015). To determine the impact of limited erosion, we repeat our models of the snowball deglaciation threshold with this modified weathering scheme for erosion rates ranging from 10 cm yr\(^{-1}\) to 1 m yr\(^{-1}\). We ignore any potential feedbacks between groundwater runoff and erosion rates, and assume therefore that their limiting effects are not additive: changes in soil hydrology do not affect the magnitude of the supply limit, or vice versa.

Our results suggest that an erosion-based supply limit can reduce weathering rates by orders of magnitude in low-erosion regimes, as shown in Figure 4. For perspective, the Antarctic Dry Valleys can experience erosion as low as 25 \( \mu \)m kyr\(^{-1}\) (0.2 bar \( \text{pCO}_2 \) Gyr\(^{-1}\), Schafer et al., 1999), alpine bare bedrock may experience 1 mm kyr\(^{-1}\) (8 bar \( \text{pCO}_2 \) Gyr\(^{-1}\), Small et al., 1997), and average rates globally may be 0.1–1 m kyr\(^{-1}\) (10\(^2\)–10\(^3\) bar \( \text{pCO}_2 \) Gyr\(^{-1}\), Willenbring et al., 2013). We assume Earth’s modern outgassing and weathering rate is 50 bar \( \text{pCO}_2 \) Gyr\(^{-1}\) (Marty and Tolstikhin, 1998; Gerlach, 2011). The weathering in our models is characterized by very strong weathering in localized areas, so even a generous supply limit can reduce overall weathering; however we find this effect is less than an order of magnitude. Erosion is primarily controlled by average local slope rather than precipitation (Blanckenburg, 2005; Willenbring et al., 2013), and glacial activity can dramatically increase erosion (Smith et al., 2007), so the dryness of snowball climates may not reduce erosion below modern rates. However, we note that an older and less tectonically-active planet might have less orogeny and thus less erosion.

Model Caveats

Our results include a number of caveats, primarily related to model limitations. We ignore the possibility of \( \text{CO}_2 \) deposition in cold traps near the poles, which could be a possibility in some snowball scenarios (Turbet et al., 2017). This would limit the \( \text{pCO}_2 \) levels attainable by snowball planets and thus limit the extent of habitable regions. Our results further out in the habitable zone are also less certain than those at higher insolations, as the \( \text{pCO}_2 \) levels associated with those climates are of order 1 bar or higher. As shown in Paradise and Menou...
(2017), PlaSim performs poorly in the regime of high-CO₂ atmospheres, demonstrating significant biases in longwave cooling rates and leaving out relevant processes such as CO₂ cloud condensation (Forget and Pierrehumbert, 1997; Kitzmann, 2017). We also ignore the important role that dust deposition has on snow and ice albedo (Gow and Williamson, 1971) and the fact that such dust deposition might be elevated during a snowball event (Kaiser and Lamy, 2010). Dust would tend to reduce the albedo and thus increase warming. We also neglect seafloor weathering, as its functional form is poorly-constrained (Caldeira, 1995; Abbot et al., 2012).

5. Conclusion

We find that snowball climates feature widespread temperate land areas across a range of parameters, including land distribution, insolation, obliquity, and soil water capacity. These climates sustain low to moderate weathering rates, implying potential for long-term stability. Together with the potential for climate cycles consisting of long snowball events (Menou, 2015; Paradise and Menou, 2017), our results suggest that some fraction of observed terrestrial habitable-zone planets may be snowball planets. This may be particularly important for older planets with lower outgassing as well as younger planets in the outer habitable zone, where snowball planets become likely even at Earth-like outgassing rates.

Together with the results in Abbot et al. (2013) suggesting small areas of open ocean, our results expand the prospects of habitability during snowball events. Along with Linsenmeier et al. (2015), who found that snowball planets with high axial tilt and eccentricity could have seasonally temperate regions, this bolsters the argument in Spiegel et al. (2008) that some terrestrial planets could have fractional habitability, with habitable areas restricted to certain times of the year or certain regions of the planet.

Moreover, it may be possible to observationally distinguish these planets from temperate planets. In addition to higher average albedos, snowball planets have high pCO₂ but low average atmospheric water content, suggesting elevated CO₂/H₂O ratios. Snowballs with fractional habitability may have CO₂/H₂O ratios higher than globally temperate planets but lower than completely-frozen snowballs, due to continued weathering and the presence of limited hydrological cycles.

These results also have implications for the climate cycles explored in e.g., Menou (2015), Abbot (2016), and Paradise and Menou (2017)—the parameter space in which no weathering equilibrium is expected is greatly reduced by our results, and depends much more strongly on factors such as the planet’s glaciology, erosive processes, and topography. We also note that at high insulations, the maximum snowball weathering is greater than the minimum weathering supported by temperate climates described by Abbot (2016) and Paradise and Menou (2017). This suggests that for planets with insulations and outgassing rates in that overlapping region, weathering equilibria are possible for both temperate and snowball states, with the outcome dependent on other factors.

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