The Dependence of the Ice-Albedo Feedback on Atmospheric Properties

P. von Paris,^{1,2,3} F. Selsis,^{1,2} D. Kitzmann,⁴ and H. Rauer^{3,4}

Abstract

Ice-albedo feedback is a potentially important destabilizing effect for the climate of terrestrial planets. It is based on the positive feedback between decreasing surface temperatures, an increase of snow and ice cover, and an associated increase in planetary albedo, which then further decreases surface temperature. A recent study shows that for M stars, the strength of the ice-albedo feedback is reduced due to the strong spectral dependence of stellar radiation and snow/ice albedos; that is, M stars primarily emit in the near IR, where the snow and ice albedo is low, and less in the visible, where the snow/ice albedo is high.

This study investigates the influence of the atmosphere (in terms of surface pressure and atmospheric composition) on this feedback, since an atmosphere was neglected in previous studies. A plane-parallel radiative transfer model was used for the calculation of planetary albedos. We varied CO₂ partial pressures as well as the H₂O, CH₄, and O₃ content in the atmosphere for planets orbiting Sun-like and M type stars.

Results suggest that, for planets around M stars, the ice-albedo effect is significantly reduced, compared to planets around Sun-like stars. Including the effects of an atmosphere further suppresses the sensitivity to the ice-albedo effect. Atmospheric key properties such as surface pressure, but also the abundance of radiative trace gases, can considerably change the strength of the ice-albedo feedback. For dense CO₂ atmospheres of the order of a few to tens of bar, atmospheric rather than surface properties begin to dominate the planetary radiation budget. At high CO₂ pressures, the ice-albedo feedback is strongly reduced for planets around M stars. The presence of trace amounts of H₂O and CH₄ in the atmosphere also weakens the ice-albedo effect for both stellar types considered. For planets around Sun-like stars, O₃ could also lead to a very strong decrease of the ice-albedo feedback at high CO₂ pressures. Key Words: Atmospheric compositions—Extrasolar terrestrial planets—Snowball Earth—Planetary atmospheres—Radiative transfer. *Astrobiology* 13, 899–909.

1. Introduction

THE ICE-ALBEDO FEEDBACK is a very important positive feedback for Earth's climate. In simple words, it describes the possible runaway cooling of the planetary surface. Decreasing surface temperature leads to an increase of snow and ice cover, which increases the surface albedo. Hence, more incoming stellar radiation is reflected back to space, which in turn decreases surface temperature.

A recent work by Joshi and Haberle (2012) connected this ice-albedo feedback to exoplanets and the habitable zone (HZ). The HZ is defined as the region around a star where a planet with a suitable atmosphere could have liquid water on the surface and, hence, the potential for life as we know it (Kasting *et al.*, 1993). Calculations presented by Joshi and

Haberle (2012) showed that the planetary albedos for planets orbiting M stars are much lower than for planets orbiting Sun-like stars because of the strong spectral dependence of both ice and snow albedo and the stellar spectrum. Hence, the ice-albedo feedback is much less important for planets orbiting M stars. Based on these results, Joshi and Haberle (2012) stated that the outer boundary of the HZ is more extended around M stars than around G stars since planetary albedos are much lower.

However, the ice-albedo feedback is not the key factor when determining the extent of the outer boundary of the HZ. It is usually assumed that geochemical cycles such as the carbonate-silicate cycle (*e.g.*, Walker *et al.*, 1981; Abbot *et al.*, 2012) will stabilize the climate (*e.g.*, Kasting *et al.*, 1993; Selsis *et al.*, 2007). The outer boundary of the HZ is then

¹Université Bordeaux, LAB, UMR 5804, F-33270, Floirac, France.

²CNRS, LAB, UMR 5804, F-33270, Floirac, France.

³Institut für Planetenforschung, Deutsches Zentrum für Luft- und Raumfahrt, Berlin, Germany.

⁴Zentrum für Astronomie und Astrophysik, Technische Universität Berlin, Berlin, Germany.

determined by the interplay between an increase of planetary albedo with increasing CO₂ and the atmospheric greenhouse effect provided by CO₂ (Kasting *et al.*, 1993). A higher planetary albedo does not necessarily imply lower surface temperatures, as has been shown by, for example, McKay *et al.* (1989) or Mischna *et al.* (2000). Still, at one point, an increase in planetary albedo will lead to cooler surface temperatures, an effect known for CO₂ as the maximum greenhouse effect (Kasting *et al.*, 1993). Furthermore, atmospheres toward the outer edge of the HZ are assumed to be very dense (of the order of several bar of CO₂, Kasting *et al.*, 1993), implying that the influence of surface albedo on the planetary energy budget is probably rather small (see, *e.g.*, Wordsworth *et al.*, 2011 for 3-D investigations of the outer HZ). Therefore, conclusions on habitability and the extent of the HZ cannot be easily drawn from the investigation of the ice-albedo feedback.

Nevertheless, the ice-albedo feedback is still an important process with regard to the atmospheric and climatic evolution of Earth. It is generally accepted that Earth encountered several global glaciation episodes (the so-called snowball Earth, *e.g.*, Kirschvink, 1992; Hoffman *et al.*, 1998; Hyde *et al.*, 2000; Kasting and Ono, 2006) and, more recently, ice ages with widespread glaciations down to midlatitudes (*e.g.*, Tarasov and Peltier, 1997). Some of these snowball episodes coincide with major events in the evolution of the biosphere, such as the advent of atmospheric O₂ (Goldblatt *et al.*, 2006; Kasting and Howard, 2006) and multicellular organisms (Hyde *et al.*, 2000). Rapid, perhaps geologically or biologically induced, changes in atmospheric composition might have been responsible for triggering these glaciations (*e.g.*, Pavlov *et al.*, 2003).

Therefore, it is of interest for astrobiology and a comprehensive theory of atmospheric evolution to investigate whether planets orbiting other central stars might also be subject to such feedbacks. Joshi and Haberle (2012) calculated planetary albedos based on surface albedos and did not take explicitly the atmosphere into account. However, atmospheric properties such as surface pressure or atmospheric composition have a potentially large influence on the planetary albedo and, hence, the ice-albedo feedback. In this work, we present similar calculations as those of Joshi and Haberle (2012) but with a plane-parallel radiative transfer model that incorporates the radiative effects of an atmosphere. The model is used to calculate planetary albedos of prescribed atmospheric scenarios when varying atmospheric properties (such as atmospheric composition and surface pressure). In a complementary approach, Shields *et al.* (2013) used a combination of 2-D and 3-D models to assess the development of ice and snow cover on planets orbiting M, G, and F stars upon decreasing stellar insolation. However, the considered planetary atmospheres were Earth-like in terms of surface pressure and composition. With a 1-D radiative transfer model, Shields *et al.* (2013) also calculated planetary albedos as a function of CO₂ partial pressure, though without considering additional trace species, such as water, ozone, or methane.

The paper is organized as follows: Section 2 presents the model and scenarios, Section 3 the results, and Section 4 a discussion of additional aspects. We summarize the results and conclude with Section 5.

2. Methods

2.1. Radiative transfer

We used a plane-parallel radiative transfer model to obtain planetary albedos of a wide range of prescribed, fixed atmospheric scenarios. The radiative fluxes important for the planetary albedo (*i.e.*, stellar radiation) are calculated for a spectral range of 0.237–4.545 μm , divided into 38 spectral intervals. Scattering is treated based on a δ -Eddington-2-stream algorithm summarized by Toon *et al.* (1989). The frequency integration is performed by using correlated- k coefficients in up to 4-term exponential sums (Wiscombe and Evans, 1977). Gaseous opacities are taken from the work of Pavlov *et al.* (2000) and Karkoschka (1994). Rayleigh scattering is incorporated with the use of approximations presented by von Paris *et al.* (2010), Vardavas and Carver (1984), and Allen (1973).

The pressure grid in the model atmospheres is determined from the surface pressure p_{surf} up to a pressure of 6.6×10^{-5} bar at the model lid, divided into 52 model layers, approximately spaced equidistantly in log (pressure).

2.2. Model procedure

The planetary albedo A_P is calculated as a Bond albedo in the model, that is,

$$A_P = \frac{F_u}{F_*} = \frac{\int_0^\infty A_{P,\lambda} F_{*,\lambda} d\lambda}{\int_0^\infty F_{*,\lambda} d\lambda} \quad (1)$$

where F_u is the spectrally integrated upwelling shortwave flux at the top of the atmosphere, F_* the total stellar insolation, $A_{P,\lambda}$ the spectral albedo, and $F_{*,\lambda}$ the spectral stellar insolation. As can be seen easily from this formulation, the planetary albedo of a given atmosphere is independent of orbital distance. $F_{*,\lambda}$ was approximated by stellar spectra for the Sun (as a G star proxy) and AD Leo (representing an M type star, with 0.41 solar radii, 0.4–0.5 solar masses, and an effective temperature of ≈ 3400 K, see, *e.g.*, Leggett *et al.*, 1996; Segura *et al.*, 2005). Spectra were taken from Kitzmann *et al.* (2010) and are shown in Fig. 1. The total stellar insolation is scaled to modern-day Earth insolation (*i.e.*, $F_* = 1366 \text{ W m}^{-2}$).

To investigate the influence of different atmospheric and surface properties on planetary albedo, we prescribed atmospheric composition, the surface pressure, and the temperature profile for our calculations. Unless otherwise specified, the temperature profile was fixed at the modern-Earth profile $T_m(z)$ (z altitude), taken from Grenfell *et al.* (2011). The stellar zenith angle was fixed at 60° to simulate globally averaged conditions.

2.3. Parameter variations

We used two different types of surface albedo parameterizations. The first parameterization employed the wavelength-dependent surface albedos A_S of ice and snow, as shown in Fig. 2. We took spectral albedo data presented by Joshi and Haberle (2012), which were originally obtained from measurements in Antarctica (Brandt *et al.*, 2005; Hudson *et al.*, 2006). Note that the albedo remains constant at wavelengths $> 1.5 \mu\text{m}$ (for ice, Brandt *et al.*, 2005) and $> 2.4 \mu\text{m}$ (for snow,

FIG. 1. Stellar spectra of the Sun and AD Leo as used in this work, normalized to equal modern-day Earth insolation. Vertical dashed lines indicate the spectral boundaries of the radiative transfer scheme (0.237–4.545 μm , see Section 2.1). Color graphics available online at www.liebertonline.com/ast

Hudson *et al.*, 2006) since respective measurements were restricted to these spectral ranges.

The second parameterization used a constant (wavelength-independent) surface albedo (values of 0.1, 0.2, 0.7, and 0.9, with 0.1 being close to present Earth's mean surface albedo, Rossow and Schiffer, 1999).

The most important parameter for determining the location of the outer edge of the HZ is the CO_2 partial pressure.

We performed calculations for 22 CO_2 partial pressures ranging between 0.1 and 30 bar.

Since H_2O has very strong visible and near-IR absorption bands, even low concentrations could have an important impact on calculated albedos. To investigate this further, we performed simulations with the ice and snow surface albedos and atmospheres containing varying amounts of H_2O . H_2O concentrations $c_{\text{H}_2\text{O}}$ as a function of altitude z were set

FIG. 2. Snow and ice spectral albedo. Taken from Joshi and Haberle (2012), their Fig. 1, data from Brandt *et al.* (2005) and Hudson *et al.* (2006). Color graphics available online at www.liebertonline.com/ast

to follow the temperature-dependent saturation vapor pressure $p_{\text{vap,H}_2\text{O}}(T)$; that is, model atmospheres were assumed to be fully saturated:

$$c_{\text{H}_2\text{O}}(z) = \frac{p_{\text{vap,H}_2\text{O}}(T(z))}{p(z)} \quad (2)$$

Atmospheric temperatures $T(z)$ were set to the modern-Earth profile (*i.e.*, $T(z) = T_m(z)$, see Section 2.2) and to a colder profile with $T(z) = T_m(z) - 15$ (corresponding to a surface temperature of 273 K). The colder temperature profile resulted in an approximate decrease in H_2O column amount of about a factor of 5–6.

Methane can build up in the atmosphere if strong surface fluxes exist, as postulated for, say, early Earth (*e.g.*, Pavlov *et al.*, 2003). Planets around M stars favor a buildup of CH_4 because of the weak stellar UV radiation field and the related HO_x chemistry (less OH means more CH_4 ; see, *e.g.*, Segura *et al.*, 2005 or Grenfell *et al.*, 2007). CH_4 could have an important effect on the planetary albedo because of strong near-IR absorption bands. To investigate this, we performed calculations with CH_4 concentrations set to an isoprofile with concentrations of 10^{-6} , 10^{-4} , and 10^{-3} , respectively (again, with the ice and snow surface albedos).

Ozone can build up abiotically in CO_2 -rich or strongly UV-irradiated atmospheres to quite large concentrations (*e.g.*, Selsis *et al.*, 2002; Segura *et al.*, 2007; Domagal-Goldman and Meadows, 2010) and could furthermore be the by-product of an oxygen-producing biosphere. It features strong absorption bands in the UV and visible part of the spectrum where Rayleigh scattering is strongest. Therefore, we performed sensitivity studies where we introduced an artificial ozone profile in the model atmospheres ($1\times$ and $10\times$ modern Earth's concentration following a mean Earth profile from Grenfell *et al.*, 2011, with the ice and snow surface albedos). We note that $10\times$ modern Earth levels most likely represent upper limits for the O_3 profile. However, since we aim at performing sensitivity studies only, we take this case as an end-member case.

All the simulations were done for an AD Leo and a Sun spectrum (see Fig. 1). Table 1 summarizes the scenarios.

Note that we did not include N_2 in our model atmospheres, despite the fact that it is present in substantial amounts in all terrestrial atmospheres of the Solar System (*i.e.*, Venus, Earth, Mars, Titan). The reason is that N_2 does not feature absorption bands or lines in the visible and near IR that are strong enough to affect the albedo, and its Rayleigh scattering coefficient is nearly a factor of 3 smaller than for CO_2 (Vardavas and Carver, 1984).

3. Results

3.1. Atmospheric flux profiles

Figure 3 shows the downward and upward flux profiles in the visible/near-IR spectral domain as a function of pressure for the 0.1 bar and the 20 bar ice case of scenario 1 from Table 1. It is clearly seen that in the case of a 0.1 bar CO_2 atmosphere, the influence of the atmosphere on these radiative fluxes, hence the influence on the planetary albedo, is rather small. The profiles all remain approximately constant throughout the entire atmosphere in this case. In contrast to that, in the 20 bar scenario, the atmospheric influence is very strong.

For the planet around the Sun, the incoming stellar flux is roughly reduced by a factor of 2 (black line, 340 W m^{-2} incoming flux at the top of the atmosphere, $\sim 160 \text{ W m}^{-2}$ reaching the surface). Most of this radiation is reflected back to space due to very efficient Rayleigh scattering (red line). The surface albedo contributes around 70 W m^{-2} to the upward flux at the top of the atmosphere, while the Rayleigh scattering component amounts to 120 W m^{-2} .

For the planet around the M dwarf AD Leo, the influence of the atmosphere in the 20 bar scenario is also very prominent. However, in this case it is primarily due to the absorption of near-IR radiation by CO_2 . The surface and the Rayleigh scattering both contribute with about 50% to the overall outgoing flux, but in absolute terms, the Rayleigh scattering component is reduced to about 35 W m^{-2} (*i.e.*, almost a factor of 4 compared to the Sun). The stellar flux reaching the surface is reduced from 340 W m^{-2} to less than 150 W m^{-2} , which is only slightly less than for the planet around the Sun. Comparing the Rayleigh scattering component with the reduction in downward flux, one finds that approximately 150 W m^{-2} is absorbed in the atmosphere.

It is noteworthy that, although the planetary atmosphere has a huge impact on the flux profiles, in the case of the planet around AD Leo the planetary albedo does not change that much. In the 0.1 bar and the 20 bar scenario, both top-of-atmosphere outgoing flux values are nearly identical.

3.2. Planetary spectral albedo

Figure 4 shows the spectral albedos for three different values of the CO_2 pressure of scenario 1 from Table 1. It is clearly seen that an increase of CO_2 increases the spectral albedo in the visible (due to Rayleigh scattering) and decreases the spectral albedo in the near IR (due to enhanced absorption). In general, the behavior of the spectral albedo follows the snow and ice spectral albedo in Fig. 2, that is, a high spectral albedo in the visible and a rather low spectral albedo in the near IR.

An important difference between ice and snow scenarios occurs near $1 \mu\text{m}$ where the calculated albedos differ by a

TABLE 1. PARAMETERS FOR THE MODEL CALCULATIONS PERFORMED

Scenario	A_s	p_{CO_2} (bar)	H_2O	CH_4 vmr	O_3 (PAP)
1	snow, ice (Fig. 2)	0.1, 0.2,..., 1, 2,..., 10,..., 30	0	0	0
2	0.1, 0.2, 0.7, 0.9	0.1, 0.2,..., 1, 2,..., 10,..., 30	0	0	0
3	snow, ice (Fig. 2)	0.1, 0.2,..., 1, 2,..., 10,..., 30	w, c	0	0
4	snow, ice (Fig. 2)	0.1, 0.2,..., 1, 2,..., 10,..., 30	0	10^{-6} , 10^{-4} , 10^{-3}	0
5	snow, ice (Fig. 2)	0.1, 0.2,..., 1, 2,..., 10,..., 30	0	0	1, 10

vmr, volume mixing ratio; PAP, present atmospheric profile; w, warm temperature profile T_m ; c, cold temperature profile $T_m - 15$; see Eq. 2.

FIG. 3. Net upward (F_{up}) and downward (F_{dn}) fluxes in the visible/near-IR for ice surface albedo (scenario 1 from Table 1). Pure CO_2 atmospheres (20 bar solid lines, 0.1 bar dashed lines). Color graphics available online at www.liebertonline.com/ast

factor of about 3. This spectral region is important since a rather large part of stellar radiation is emitted there.

Note that for the visible part of the spectrum, where Rayleigh scattering is important, an increase of CO_2 from 0.1 bar to 10 bar does not significantly affect the spectral albedos for the snow surface case. In contrast, for the ice surface case, the effect is clearly distinguished. This is due to the very high visible albedo of snow (~ 0.9) in this spectral range; hence the surface reflection dominates the spectral albedo in the

visible, independent of atmospheric conditions. The calculated values of the spectral albedos correspond well to values presented by Shields *et al.* (2013).

3.3. Planetary albedo

Figure 5 shows the planetary Bond albedos of scenarios 1 and 2 from Table 1. The calculated values of planetary albedo with a very tenuous CO_2 atmosphere (0.1 bar)

FIG. 4. Spectral albedos for different values of CO_2 pressure and ice and snow surface albedos. Pure CO_2 atmospheres. Color graphics available online at www.liebertonline.com/ast

FIG. 5. Planetary albedo as a function of CO₂ pressure for scenarios 1 and 2 of Table 1. Pure CO₂ atmospheres. Color graphics available online at www.liebertonline.com/ast

approximately confirm the values of Joshi and Haberle (2012), which were done without taking into account atmospheric effects (their Fig. 2).

It is clearly seen that the central star type has a large effect on the planetary albedo. At high CO₂ partial pressures, planetary albedos for the planet around the Sun are 2–3 times higher than for the planet orbiting AD Leo. This is of course due to the different spectral distribution of stellar energy (see Fig. 1). In terms of energy-equivalent orbital distance, this implies that planets around M stars could be 10–20% farther away from their central star and still receive the same net stellar energy input into the atmosphere. This fact is the basis for the claim of Joshi and Haberle (2012) that the HZ around M stars is widened with respect to the HZ around G stars.

As expected, planetary albedos are higher for snow surface albedo than for ice surface albedo. However, with increasing CO₂ partial pressure, the difference becomes noticeably smaller. For very dense, almost entirely optically thick atmospheres, planetary albedos will converge to a value independent of surface albedo.

Interestingly, the amount of CO₂ has a rather weak effect on the planetary albedo when using ice and snow albedos. This was already implied in Figs. 3 and 4. The behavior of the atmospheric spectral albedo is similar to the spectral albedo of ice and snow (see Fig. 2). As before in Section 3.2, calculated values of Bond albedos correspond well to values of Shields *et al.* (2013).

However, for wavelength-independent surface albedo values, the effect of CO₂ on planetary albedo is much stronger, especially for low (Sun) or high (AD Leo) values of surface albedo. This is explained as follows. Scattering alone can only increase the albedo. Thus, in the case of a Sun-like irradiation, for which scattering-only is a fair approximation, the planetary albedo increases with pressure. Therefore, for a

given range of pressure, the increase is more significant for low surface albedos. Absorption alone can only decrease the albedo. Thus, for AD Leo-like irradiation, for which absorption-only is a fair approximation, the planetary albedo decreases with pressure. The effect is more significant if the surface albedo is high. From Fig. 5, it is clear that the scattering-alone and absorption-alone approximations are not perfect. For the Sun and high surface albedos, Fig. 5 shows a slight decrease of planetary albedo when surface pressure increases, which is due to CO₂ absorption in the near IR. For AD Leo and low surface albedos, the planetary albedo increases with pressure due to the scattering that becomes more significant.

3.4. Effect of atmospheric composition

Figure 6 shows the effect of increasing H₂O concentrations on the calculated albedo. It is clearly seen that this decreases the planetary albedo. The effect is somewhat stronger for AD Leo than for the Sun. For example, for the ice surface albedo at 1 bar of CO₂, when using the modern-day temperature profile and corresponding H₂O profile (see Eq. 2), the planetary albedo decreases from 0.2 to 0.13 for AD Leo, but only from 0.45 to 0.40 for the Sun. This shows that even small amounts of water can enhance the effect of the stellar type on planetary albedo.

Figure 7 shows the effect of increasing CH₄ concentration on the calculated albedo. Again, the effect is much stronger for AD Leo than for the Sun.

Figure 8 shows the effect of increasing O₃ concentration on the calculated albedo. It has a large impact for the Sun, but its effect is almost negligible for AD Leo, as expected from the spectral distribution of the stellar energy (see Fig. 1). O₃ most strongly absorbs at wavelengths below 0.5 μm where AD Leo does not emit much radiation, compared to the Sun.

FIG. 6. Planetary albedo as a function of CO₂ pressure for snow and ice surface albedos. Influence of H₂O content. Solid lines for ice surface albedo; dotted lines for snow surface albedo. Color graphics available online at www.liebertonline.com/ast

4. Discussion of Additional Aspects

4.1. The effect of clouds

The main effect of clouds, such as in particular H₂O droplet and ice as well as CO₂ ice clouds, in the visible and near IR is the scattering of incident stellar radiation. Due to

their high and continuous opacities they can easily dominate the planetary albedo in this wavelength region.

Additionally, clouds can also limit the impact of the scattering by molecules and the surface on the planetary albedo. Shortwave radiation that is scattered upward by gas molecules or the surface can in turn be scattered down again

FIG. 7. Planetary albedo as a function of CO₂ pressure for snow and ice surface albedos. Influence of CH₄ content. Solid lines for ice surface albedo; dotted lines for snow surface albedo. Color graphics available online at www.liebertonline.com/ast

FIG. 8. Planetary albedo as a function of CO₂ pressure for snow and ice surface albedos. Influence of O₃ content. Solid lines for ice surface albedo; dotted lines for snow surface albedo. Color graphics available online at www.liebertonline.com/ast

by the cloud. Thus, the amount of downwelling shortwave radiation in the lower atmosphere can be higher if a cloud is present compared to the cloud-free cases presented in this study.

The exact impact on the incident stellar radiation is a direct function of the cloud's particle sizes. Particles with sizes near $0.1 \mu\text{m}$ will have a Rayleigh scattering-like effect; that is, they will contribute to the ice-albedo feedback in the same way as the scattering by gas molecules presented in this study. On the other hand, the optical properties of particles with sizes larger than a few micrometers are essentially flat in the visible and near IR (e.g., Kitzmann *et al.*, 2010). Compared to the cloud-free cases, this would lead to a constant offset in the planetary albedo, which is almost independent from the incident stellar radiation (see, e.g., Kitzmann *et al.*, 2011 for H₂O clouds and Kitzmann *et al.*, 2013 for CO₂ clouds).

However, the presence of CO₂, H₂O, CH₄, and O₃ might inhibit cloud formation because of the warming of atmospheric layers by absorption of stellar radiation. Detailed atmospheric modeling is warranted to assess the possibility of cloud formation in the middle and upper atmosphere in the future.

4.2. Land-ocean distribution

On Earth, the mean surface albedo is $A_S = 0.13$ (Rossow and Schiffer, 1999), with the oceans ($A_{\text{ocean}} = 0.05$) being less reflective than the continents. The martian surface albedo is much higher than this ($A_{\text{Mars}} = 0.21$, Kieffer *et al.*, 1977). Also, different types of land (deserts, bare rock, forests, etc.) have different spectral surface albedos. Land plants emerged not earlier than about 600 million years ago on Earth (e.g., Kasting and Howard, 2006), which demonstrates that profound changes in a planet's surface albedo might happen

during its evolution. Therefore, it is expected that the land-ocean ratio and the distribution of continents might have a large influence on the overall surface albedo and, by extension, on the sensitivity of the planet to the ice-albedo feedback.

Since this land-ocean ratio and the continental distribution might also change during the planetary evolution (e.g., Pesonen *et al.*, 2012), the strength of the ice-albedo effect might also change quite a lot over the course of the planetary evolution. An investigation with 2-D or 3-D atmosphere models would be necessary to assess the strength of these effects.

4.3. Albedo measurements

Surface and atmospheric properties of terrestrial exoplanets such as continents or oceans, or ice, snow, and cloud coverage induce photometric variability in the visible–near-IR light curves and affect measured secondary eclipse albedos. Recent studies have suggested that it might be possible to infer such properties from spectrophotometric observations (e.g., Oakley and Cash, 2009; Fujii *et al.*, 2010; Cowan *et al.*, 2011; Livengood *et al.*, 2011). Albedo measurements might be indicative of snowball planets (e.g., Cowan *et al.*, 2011). Such observations are easiest for transiting planets and could be performed near secondary eclipse. For hot giant planets, this has already been done (e.g., Borucki *et al.*, 2009; Snellen *et al.*, 2009; Parviainen *et al.*, 2013).

Even if they were feasible for terrestrial, potentially habitable planets, it is questionable whether broadband photometric measurements of the reflected light could be useful to constrain the surface albedo and atmospheric characteristics. The contrast C_R between star and planet in reflected light close to secondary eclipse depends on planetary radius R_P , planetary albedo A_P , and orbital distance d :

$$C_R = A_P \cdot f_R = A_P \cdot \left(\frac{R_P}{d}\right)^2 \quad (3)$$

For an Earth around the Sun ($d=1$ AU), this translates into $C_R \sim A_P \cdot 1.8 \times 10^{-9}$ (e.g., Kitzmann *et al.*, 2011). If the planetary radius is known, the only uncertainty is on the deduced planetary albedo. Assuming a photometric precision of about 2×10^{-10} (i.e., $0.1 \times f_R$ in Eq. 3, Oakley and Cash, 2009), the 2σ uncertainty for the obtained planetary albedo A_P translates roughly to ± 0.2 .

As shown above in Section 3, albedo differences for the various scenarios are generally of this order of magnitude. This implies that albedo characterization is very difficult for planets orbiting M stars. Distinguishing between low surface albedos and completely ice-covered planets is unlikely, even if the high albedo due to snow coverage could be excluded reasonably well. Furthermore, it is very difficult to infer the atmospheric properties (i.e., surface pressure, amount of radiative trace gases). For planets orbiting Sun-like stars, it may be possible to distinguish between ice-free and ice-covered planets at low CO_2 pressures. If a high albedo would be measured, however, the albedo solution is no longer unique; that is, it might be an ice-covered planet at low or high atmospheric pressure or even an ice-free planet with a dense CO_2 atmosphere or clouds.

Therefore, the characterization of the atmosphere and surface is not likely to be possible with secondary eclipse observations. To break degeneracies, additional observations are needed. Especially, transmission spectroscopy during primary transit could be used to put constraints on atmospheric scale height and surface pressure (e.g., von Paris *et al.*, 2011; Benneke and Seager, 2012) and thus help interpret measured planetary albedos. These observations would, however, require large telescope facilities to reach sufficient signal-to-noise ratios (e.g., Hedelt *et al.*, 2013).

5. Summary and Conclusions

We have presented detailed calculations of the planetary albedo of terrestrial planets as a function of stellar type and atmospheric composition. Key atmospheric properties such as CO_2 partial pressure and the abundance of radiative trace gases such as H_2O , CH_4 , and O_3 were varied. Furthermore, we considered different types of surface albedo (wavelength-dependent ice or snow albedo, wavelength-independent constant albedo). Based on the planetary albedo, we investigated whether the ice-albedo feedback depends strongly on stellar type and atmospheric composition.

First, our calculations confirm the results of Joshi and Haberle (2012), that is, that for ice- and snow-covered surfaces, the planetary albedos for planets orbiting M stars are much lower than for planets orbiting Sun-like stars (about a factor of 2). This indeed leads to a strong reduction of the ice-albedo effect for planets around M stars.

Second, our results imply that, for dense CO_2 atmospheres, the difference of planetary albedo between ice and ice-free ($A_S=0.1$) cases is strongly reduced (from 0.35 to 0.05 for planets orbiting the Sun). For planets around M stars, the ice-albedo feedback is almost entirely masked at high CO_2 pressures (albedo change less than 0.04).

Third, we have shown that the effect of atmospheric composition may be very important when calculating the

strength of the ice-albedo effect. Even radiative trace species such as H_2O , CH_4 , or O_3 could alter sensitively the planetary albedo. This in turn influences the radiative forcing associated with an ice or snow surface albedo and, hence, could strongly suppress the ice-albedo feedback. The presence of small amounts of H_2O and CH_4 effectively weakens the ice-albedo effect by several percent in planetary albedo for both stellar types considered. In addition, for planets around Sun-like stars, significant amounts of O_3 could also lead to a very strong reduction of the ice-albedo feedback at high CO_2 pressures.

Acknowledgments

P. v. Paris and F. Selsis acknowledge support from the European Research Council (Starting Grant 209622: E3ARThs). This research has been partly supported by the Helmholtz Association through the research alliance “Planetary Evolution and Life.” Valuable discussions with P. Hedelt and J.L. Grenfell are gratefully acknowledged. Helpful, constructive comments by D. Abbot and an anonymous reviewer are also acknowledged.

Author Disclosure Statement

No competing financial interests exist.

Abbreviation

HZ, habitable zone.

References

- Abbot, D., Cowan, N., and Ciesla, F. (2012) Indication of insensitivity of planetary weathering behavior and habitable zone to surface land fraction. *Astrophys J* 756, doi:10.1088/0004-637X/756/2/178.
- Allen, C.W. (1973) *Astrophysical Quantities*, Athlone Press, University of London, London.
- Benneke, B. and Seager, S. (2012) Atmospheric retrieval for super-Earths: uniquely constraining the atmospheric composition with transmission spectroscopy. *Astrophys J* 753, doi:10.1088/0004-637X/753/2/100.
- Borucki, W., Koch, D., Jenkins, J., Sasselov, D., Gilliland, R., Batalha, N., Latham, D., Caldwell, D., Basri, G., Brown, T., Christensen-Dalsgaard, J., Cochran, W., DeVore, E., Dunham, E., Dupree, A., Gautier, T., Geary, J., Gould, A., Howell, S., Kjeldsen, H., Lissauer, J., Marcy, G., Meibom, S., Morrison, D., and Tarter, J. (2009) Kepler’s optical phase curve of the exoplanet HAT-P-7b. *Science* 325, doi:10.1126/science.1178312.
- Brandt, R., Warren, S., Worby, A., and Grenfell, T. (2005) Surface albedo of the Antarctic sea ice zone. *J Clim* 18:3606–3622.
- Cowan, N., Robinson, T., Livengood, T., Deming, D., Agol, E., A’Hearn, M., Charbonneau, D., Lisse, C., Meadows, V., Seager, S., Shields, A., and Welnitz, D. (2011) Rotational variability of Earth’s polar regions: implications for detecting snowball planets. *Astrophys J* 731, doi:10.1088/0004-637X/731/1/76.
- Domagal-Goldman, S. and Meadows, V. (2010) Abiotic buildup of ozone. In *Pathways Towards Habitable Planets*, ASP Conference Series Vol. 430, edited by V. Coudé du Foresto, D.M. Gelino, and I. Ribas, Astronomical Society of the Pacific, San Francisco, p 152.
- Fujii, Y., Kawahara, H., Suto, Y., Taruya, A., Fukuda, S., Nakajima, T., and Turner, E. (2010) Colors of a second Earth: esti-

- matting the fractional areas of ocean, land, and vegetation of Earth-like exoplanets. *Astrophys J* 715, doi:10.1088/0004-637X/715/2/866.
- Goldblatt, C., Lenton, T., and Watson, A. (2006) Bistability of atmospheric oxygen and the Great Oxidation. *Nature* 443:683–686.
- Grenfell, J.L., Griesmeier, J.-M., Patzer, B., Rauer, H., Segura, A., Stadelmann, A., Stracke, B., Titz, R., and von Paris, P. (2007) Biomarker response to galactic cosmic ray-induced NO_x and the methane greenhouse effect in the atmosphere of an Earth-like planet orbiting an M dwarf star. *Astrobiology* 7:208–221.
- Grenfell, J.L., Gebauer, S., von Paris, P., Godolt, M., Hedelt, P., Patzer, B., Stracke, B., and Rauer, H. (2011) Sensitivity of biomarkers to changes in chemical emissions in the Earth's Proterozoic atmosphere. *Icarus* 211:81–88.
- Hedelt, P., von Paris, P., Godolt, M., Gebauer, S., Grenfell, J.L., Rauer, H., Schreier, G., Selsis, F., and Trautmann, T. (2013) Spectral features of Earth-like planets and their detectability at different orbital distances around F, G, and K-type stars. *Astron Astrophys* 553, doi:10.1051/0004-6361/201117723.
- Hoffman, P., Kaufman, A., Halverson, G., and Schrag, D. (1998) A Neoproterozoic snowball Earth. *Science* 281:1342–1346.
- Hudson, S., Warren, S., Brandt, R., Grenfell, T., and Six, D. (2006) Spectral bidirectional reflectance of Antarctic snow: measurements and parameterization. *J Geophys Res* 111, doi:10.1029/2006JD007290.
- Hyde, W., Crowley, T., Baum, S., and Peltier, W. (2000) Neoproterozoic 'snowball Earth' simulations with a coupled climate/ice-sheet model. *Nature* 405:425–429.
- Joshi, M. and Haberle, R. (2012) Suppression of the water ice and snow albedo feedback on planets orbiting red dwarf stars and the subsequent widening of the habitable zone. *Astrobiology* 12:3–8.
- Karkoschka, E. (1994) Spectrophotometry of the jovian planets and Titan at 300- to 1000-nm wavelength: the methane spectrum. *Icarus* 111:174–192.
- Kasting, J. and Howard, M. (2006) Atmospheric composition and climate on the early Earth. *Philos Trans R Soc Lond B Biol Sci* 361:1733–1742.
- Kasting, J. and Ono, S. (2006) Palaeoclimates: the first two billion years. *Philos Trans R Soc Lond B Biol Sci* 361:917–929.
- Kasting, J., Whitmire, D., and Reynolds, R. (1993) Habitable zones around main sequence stars. *Icarus* 101:108–128.
- Kieffer, H., Martin, T., Peterfreund, A., Jakosky, B., Miner, E., and Palluconi, F. (1977) Thermal and albedo mapping of Mars during the Viking primary mission. *J Geophys Res* 82:4249–4291.
- Kirschvink, J. (1992) Late Proterozoic low-latitude global glaciation: the snowball Earth. In *The Proterozoic Biosphere: A Multidisciplinary Study*, edited by J.W. Schopf and C. Klein, Cambridge University Press, Cambridge, UK, pp 51–52.
- Kitzmann, D., Patzer, A.B.C., von Paris, P., Godolt, M., Stracke, B., Gebauer, S., Grenfell, J.L., and Rauer, H. (2010) Clouds in the atmospheres of extrasolar planets. I. Climatic effects of multi-layered clouds for Earth-like planets and implications for habitable zones. *Astron Astrophys* 511, doi:10.1051/0004-6361/200913491.
- Kitzmann, D., Patzer, A.B.C., von Paris, P., Godolt, M., and Rauer, H. (2011) Clouds in the atmospheres of extrasolar planets. III. Impact of low and high-level clouds on the reflection spectra of Earth-like planets. *Astron Astrophys* 534, doi:10.1051/0004-6361/201117375.
- Kitzmann, D., Patzer, A.B.C., and Rauer, H. (2013) Clouds in the atmospheres of extrasolar planets. IV. On the scattering greenhouse effect of CO₂ ice particles: numerical radiative transfer studies. *Astron Astrophys* 557, 10.1051/0004-6361/201220025.
- Leggett, S., Allard, F., Berriman, G., Dahn, C., and Hauschildt, P. (1996) Infrared spectra of low-mass stars: toward a temperature scale for red dwarfs. *Astrophys J Suppl Ser* 104:117–143.
- Livengood, T., Deming, D., A'Hearn, M., Charbonneau, D., Hewagama, T., Lisse, C., McFadden, L., Meadows, V., Robinson, T., Seager, S., and Wellnitz, D. (2011) Properties of an Earth-like planet orbiting a Sun-like star: Earth observed by the EPOXI Mission. *Astrobiology* 11:907–930.
- McKay, C., Pollack, J., and Courtin, R. (1989) The thermal structure of Titan's atmosphere. *Icarus* 80:23–53.
- Mischna, M., Kasting, J., Pavlov, A., and Freedman, R. (2000) Influence of carbon dioxide clouds on early martian climate. *Icarus* 145:546–554.
- Oakley, P. and Cash, W. (2009) Construction of an Earth model: analysis of exoplanet light curves and mapping the next Earth with the New Worlds Observer. *Astrophys J* 700:1428–1439.
- Parviainen, H., Deeg, H., and Belmonte, J. (2013) Secondary eclipses in the CoRoT light curves. A homogeneous search based on Bayesian model selection. *Astron Astrophys* 550, doi:10.1051/0004-6361/201220081.
- Pavlov, A., Kasting, J., Brown, L., Rages, K., and Freedman, R. (2000) Greenhouse warming by CH₄ in the atmosphere of early Earth. *J Geophys Res* 105:11981–11990.
- Pavlov, A., Hurtgen, M., Kasting, J., and Arthur, M. (2003) Methane-rich Proterozoic atmosphere? *Geology* 31:87–90.
- Pesonen, L., Mertanen, S., and Veikkolainen, T. (2012) Paleomesoproterozoic supercontinents—a paleomagnetic view. *Geophysica* 48:5–47.
- Rossow, W. and Schiffer, R. (1999) Advances in understanding clouds from ISCCP. *Bulletin of the American Meteorological Society* 80:2261–2288.
- Segura, A., Kasting, J., Meadows, V., Cohen, M., Scalo, J., Crisp, D., Butler, R., and Tinetti, G. (2005) Biosignatures from Earth-like planets around M dwarfs. *Astrobiology* 5:706–725.
- Segura, A., Meadows, V., Kasting, J., Crisp, D., and Cohen, M. (2007) Abiotic formation of O₂ and O₃ in high-CO₂ terrestrial atmospheres. *Astron Astrophys* 472:665–679.
- Selsis, F., Despois, D., and Parisot, J.-P. (2002) Signature of life on exoplanets: can Darwin produce false positive detections? *Astron Astrophys* 388:985–1003.
- Selsis, F., Kasting, J., Levrard, B., Paillet, J., Ribas, I., and Delbosse, X. (2007) Habitable planets around the star Gliese 581? *Astron Astrophys* 476:1373–1387.
- Shields, A., Meadows, V., Bitz, C., Pierrehumbert, R., Joshi, M., and Robinson, T. (2013) The effect of host star spectral energy distribution and ice-albedo feedback on the climate of extrasolar planets. *Astrobiology* 13:715–739.
- Snellen, I., de Mooij, E., and Albrecht, S. (2009) The changing phases of extrasolar planet CoRoT-1b. *Nature* 459:543–545.
- Tarasov, L. and Peltier, W. (1997) Terminating the 100 kyr ice age cycle. *J Geophys Res* 102:21665–21694.
- Toon, O., McKay, C., Ackerman, T., and Santhanam, K. (1989) Rapid calculation of radiative heating rates and photodissociation rates in inhomogeneous multiple scattering atmospheres. *J Geophys Res* 94:16287–16301.
- Vardavas, I. and Carver, J. (1984) Solar and terrestrial parameterizations for radiative-convective models. *Planet Space Sci* 32:1307–1325.
- von Paris, P., Gebauer, S., Godolt, M., Grenfell, J.L., Hedelt, P., Kitzmann, D., Patzer, A.B.C., Rauer, H., and Stracke, B. (2010) The extrasolar planet GL 581 d: a potentially habitable planet? *Astron Astrophys* 522, doi:10.1051/0004-6361/201015329.

- von Paris, P., Cabrera, J., Godolt, M., Grenfell, J.L., Hedelt, P., Rauer, H., Schreier, F., and Stracke, B. (2011) Spectroscopic characterization of the atmospheres of potentially habitable planets: GL 581 d as a model case study. *Astron Astrophys* 534, doi:10.1051/0004-6361/201117091.
- Walker, J., Hays, P., and Kasting, J. (1981) A negative feedback mechanism for the long-term stabilization of the Earth's surface temperature. *J Geophys Res* 86:9776–9782.
- Wiscombe, W. and Evans, J. (1977) Exponential-sum fitting of radiative transmission functions. *J Comput Phys* 24:416–444.
- Wordsworth, R., Forget, F., Selsis, F., Millour, E., Charnay, B., and Madeleine, J.-B. (2011) Gliese 581d is the first discovered terrestrial-mass exoplanet in the habitable zone. *Astrophys J Letters* 733, doi:10.1088/2041-8205/733/2/L48.

Address correspondence to:

Philip von Paris
Institut für Planetenforschung
Deutsches Zentrum für Luft- und Raumfahrt
Rutherfordstr. 2
12489 Berlin
Germany

E-mail: philip.vonparis@dlr.de

Submitted 5 March 2013

Accepted 27 July 2013