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Can thin cirrus clouds in the tropics provide a solution to the faint young Sun paradox?

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[1] In this paper we present radiative-convective simulations to test the idea that tropical cirrus clouds, acting as a negative feedback on climate, can provide a solution to the faint young Sun paradox. We find that global mean surface temperatures above freezing can indeed be found for luminosities larger than about 0.8 (corresponding to ~2.9 Ga and nearly complete tropical cirrus coverage). For luminosities smaller than 0.8, even though global mean surface temperatures are below freezing, tropical mean temperatures are still above freezing, indicating the possibility of a partially ice-free Earth for the Early Archean. We discuss possible mechanisms for the functioning of this negative feedback. While it is feasible for tropical cirrus to completely eliminate the paradox, it is similarly possible for tropical cirrus to reduce the amounts of other greenhouse gases needed for solving the paradox and therefore easing the constraints on CO_2 and CH_4 that appear to be in disagreement with geological evidence.

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1. Introduction

[2] Models for the evolution of the Sun during the main sequence call for a reduced solar luminosity and therefore a reduced Earth's solar constant of about $S = 0.75\ S_0$ around 3.8 Ga (with S_0 the present solar constant ~1353 W/m^2) [*Schwarzschild*, 1958; *Newman and Rood*, 1977]. At the same time, geological evidence shows the presence of a stable ocean and liquid water in the planet at least after 3.9 Ga (and perhaps even earlier [e.g., *Wilde et al.*, 2001; *Pinti*, 2005]). The fact that simple models of the Earth's climate cannot reconcile the reduced luminosity with the presence of liquid water (and the absence of glacial deposits) has become known as the faint young Sun paradox [*Sagan and Mullen*, 1972]. The paradox hinges on the assumption of a constant atmospheric composition or, more precisely, on the assumption of a constant atmospheric greenhouse effect and a constant atmospheric solar reflectivity (both including gases and clouds). Just for illustration purposes, one can use a crude zero-dimensional energy balance for the atmosphere to calculate the mean global surface temperature (T_s) [e.g., *Catling and Kasting*, 2007],

$$T_s = T_g + \left(\frac{(1-A)S}{4\sigma}\right)^{\frac{1}{4}}, \quad (1)$$

where A is the planetary albedo and T_g is a temperature that encapsulates the greenhouse effect of the atmosphere and clouds. For current climate with $A = 0.3$ and $T_g = 34$, $T_s = 288\ K$. According to the standard solar model, the luminosity, and therefore the variation of the solar constant can be approximated by [*Gough*, 1981]

$$S = \frac{S_0}{1 + 0.4t/4.6}, \quad (2)$$

where t is the time in Ga.

[3] Under the assumption of a constant greenhouse effect, the simple zero-dimensional model gives $T_s = 269$ K for a solar luminosity of $S = 0.75\ S_0$, ~ 3.9 Ga. T_s rises above freezing for $S \sim 0.79\ S_0$, which corresponds to 2.9 Ga. It might seem that a much reduced value of A in equation (1) could increase the temperature above freezing. However, the absence of clouds (the main driver of the albedo) would also result in a significant reduction of the greenhouse effect. A first correction to the simple model is to include a water vapor feedback by assuming a constant relative humidity (instead of the implicit assumption of constant specific humidity). By including this positive water vapor feedback in a 1-D radiative convective model one increases the time range of the paradox: a colder surface temperature implies a drier atmosphere and a reduced greenhouse effect. For instance, *Kasting et al.* [1988] found that T_s remains below freezing up until ~2 Ga or $S \sim 0.85\ S_0$. Moreover, *Pierrehumbert* [2010] shows that including an ice-albedo feedback the paradox is even more dramatic and the solution for $S = 0.75\ S_0$ is a snowball Earth with $T_s = 228\ K$ (however, see *Cogley and Henderson-Sellers* [1984] for arguments on a much reduced role for the ice-albedo feedback on the early Earth).

[4] *Sagan and Mullen* [1972] first pointed out the existence of the paradox and suggested that trace amounts of

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NH_3 could solve the paradox. This solution was later found untenable due to the relatively small lifetime of NH_3 to photolysis in an anoxic atmosphere [*Kuhn and Atreya*, 1979]. Most of the solutions to the paradox have relied on changes in T_g produced by either CO_2, CH_4 or both [e.g., *Hart*, 1978; *Owen et al.*, 1979; *Kasting*, 1987; *Kasting et al.*, 1988; *Pavlov et al.*, 2000; *Haqq-Misra et al.*, 2008]. Solutions that involve high CO_2 atmospheric concentrations are particularly appealing given the existence of large reservoirs of carbon in the Earth's mantle and continents (and the relative smallness of the atmospheric and oceanic reservoirs). The temperature dependence of the silicate weathering rate (mainly through the temperature dependence of the precipitation) can act as a negative feedback on climate acting through the CO_2 geological cycle [*Walker et al.*, 1981]. According to this mechanism, climates colder than present are expected to have a higher CO_2 concentrations, compensating to some extent for the reduced solar luminosity.

[5] However, some geological evidence from paleosols and other proxies indicates that CO_2 concentrations must be at least ten times smaller than those required to produce mean surface temperatures above freezing in 1-D radiative-convective models [*Rye et al.*, 1995; *Rollinson*, 2007]). *Zahnle and Sleep* [2002] also argue on the basis of theoretical calculations of the carbon geological cycle, that high CO_2 concentrations are implausible. If the geological evidence is taken at face value, the paradox seems to be unresolved [*Shaw*, 2008]. This realization has prompted even the reconsideration of the relevance of the standard model for solar evolution and therefore the faintness of the early Sun [e.g., *Sackmann and Boothroyd*, 2003]. However, evidence for the standard solar model is strong. In particular, solar analogs appear to show no evidence for the magnitude and time scale of mass loss required to explain an early bright Sun [*Minton and Malhotra*, 2007].

[6] The meridional heat transport can also change under different forcing conditions, potentially providing a stabilizing influence on climate, specially for the onset of snowball solutions [e.g., *Lindzen and Farrell*, 1980]. The moderating influence of meridional heat transport has been discussed in the context of the faint young Sun paradox by *Endal and Schatten* [1982], who proposed a much more effective ocean heat transport in an early Earth with small continents. However, a more effective heat transport would also produce a larger value for the critical insolation for the onset of a snowball Earth. *Gerard et al.* [1990], based on the maximum entropy principle [*Paltridge*, 1978], deduced that the heat transport becomes less efficient for lower solar luminosities and therefore they obtain solutions that are stable to an ice-albedo feedback for the whole evolution of the solar constant.

[7] Besides purely dynamical or radiative mechanisms to account for the moderate temperatures under lower solar luminosity, the rise of life and subsequent changes in atmospheric composition may have played a role in the climate stabilization required to explain the paradox. For instance, the rise of early bacteria could have increased methane fluxes into an early anoxic atmosphere [e.g., *Pavlov et al.*, 2000] providing methane concentrations of about 100 times present concentrations [*Pavlov et al.*, 2003]. The enhancement of the weathering rate due to the rise of life has also been proposed as a negative feedback on climate [*Volk*, 1987; *Schwartzman and Volk*, 1989, 2004] and moreover as a potential self-regulating mechanism for the biosphere [*Lovelock and Whitfield*, 1982].

[8] Water clouds on the other hand, have been only rarely invoked as a possible solution to the paradox, although changes in their composition, height and areal extent can potentially provide large changes in both A and T_g. When studying the effect of greenhouse gases, clouds properties are usually kept constant. The rationale and limitations for the assumption of constant cloud properties are summarized by *Kasting and Catling* [2003]: "If the goal is to determine what is required to create a climate similar to that of today, it is reasonable to assume no change in cloud properties. For model planets that are either much hotter or much colder than present Earth, however, the neglect of cloud feedback may lead to serious error." The matter of how much colder (or hotter) a climate should be so that the effect of cloud feedbacks becomes important has been the subject of some previous studies on the role of clouds in the early Earth climate [*Henderson-Sellers and Cogley*, 1982; *Rossow et al.*, 1982]. In those studies a decrease in cloud liquid water in colder climates is associated with a decrease in planetary albedo large enough to produce mean global surface temperatures above freezing for $S \gtrsim 0.8\ S_0$.

[9] Here, we focus on testing the feasibility of a solution based on changes in the cirrus cloud coverage in the tropics. We are primarily interested whether a plausible change in the coverage of thin cirrus clouds can solve the faint young Sun paradox, regardless of the origin of such a change. We focus on tropical cirrus clouds because contrary to extratropical clouds, in which cloud coverage is mostly related to the relative area of ascent and descent in baroclinic disturbances, the mechanism of formation of cirrus in the tropics appears to be particularly susceptible to a surface temperature dependence. An example of a mechanism that could relate sea surface temperature to thin cirrus cloud coverage is the iris hypothesis proposed by *Lindzen et al.* [2001]. We defer to section 4 the discussion of this particular mechanism.

[10] Thin cirrus clouds are a ubiquitous feature of the current tropical atmosphere. Recent global data using satellite lidar and radar instruments place the frequency of thin cirrus clouds ($\tau < 3-4$) at ~25% over the tropics (30°S–30°N) [*Sassen et al.*, 2008]. Trajectory studies show that at least two mechanisms explain the formation of cirrus clouds in the tropics; a direct detrainment from convective clouds and also a triggering by gravity waves further away from the original convective region [*Mace et al.*, 2006]. Although cirrus clouds are believed to have a net positive radiative effect, there remains uncertainty on this point [*Liou*, 2005]. Nevertheless, recent satellite estimations of the cloud radiative effect of cirrus clouds [*Choi and Ho*, 2006] seem to confirm the long-held idea that thin cirrus clouds (that is clouds with visible optical depths $\tau \lesssim 10$) have a much larger infrared heating effect than a shortwave cooling, and therefore a strong positive cloud radiative effect.

[11] One-dimensional radiative convective simulations, including at least some representation of cirrus clouds, have already shown the potential of thin cirrus clouds to produce significant surface warming. In the seminal paper by *Manabe and Wetherald* [1967], the addition of a layer of full black cirrus cloud was enough to increase the equilibrium surface

temperature from 280 K to 320 K. Similarly, *Liou and Gebhart* [1982] show radiative-convective equilibrium simulations in which the inclusion of a thin cirrus cloud can increase surface temperatures to ~320 K for total coverage, with the surface temperature being relatively independent of the height of the cloud base. In the next sections, we present results from a simple radiative-convective model in which the tropical thin cirrus cloud coverage (f) is specified.

2. Model Assumptions

[12] The 1-D model is a simple radiative-convective equilibrium model based on the original formulation by *Manabe and Strickler* [1964] and *Manabe and Wetherald* [1967]. The model has 140 levels in pressure from 1000 hPa to 0.04 hPa, following the sigma-level pressure coordinates defined by *Manabe and Wetherald* [1967]. The model is run for 600 days from an initial moist adiabatic atmosphere with surface temperature of 300 K, with time step of 1 day (equilibrium between incoming shortwave and outgoing longwave radiation is reached within less than 1 W/m^2). We use a similar relative humidity profile as in *Manabe and Wetherald* [1967] with a surface relative humidity of 0.8 and a constant stratospheric water vapor mixing ratio of 3×10^{-6}. At each time step we use solar and infrared radiative parameterizations (developed for general circulation models [*Chou and Suarez*, 2002; *Chou et al.*, 2003]) to estimate the radiative heating rates in each vertical layer. A convective adjustment is performed at each time step, so unstable layers are adjusted to a reversible moist adiabat (which at least for the tropics seems to be a very good approximation for the temperature vertical distribution [*Emanuel*, 2007]). In all the runs, unless otherwise noted, the concentration of the radiatively active gases (except for water vapor) is kept fixed and approximately equal to the present atmospheric levels (PAL). That is, CO_2 = 350 ppmv and CH_4 = 1.75 ppmv.

[13] We assess the effect of the coverage of tropical cirrus clouds on surface temperature with some very simple assumptions. The effect of clouds other than thin cirrus (hereafter $\tau < 9$) is not explicitly incorporated, but rather enters as a constant planetary albedo fixed to about 0.2 (this is only part of the planetary albedo, since the radiative parameterization calculates explicitly the scattering by the clear atmosphere and thin cirrus clouds). In this way an incoming solar radiation and a coverage of 0.16 for thin cirrus, will provide a surface temperature close to the observed in the present (298 K for the mean tropical temperature). The incoming solar radiation that provides the current tropical average temperature (~285 W/m^2 after correcting by the solar zenith angle and constant planetary albedo), will serve as a basis for changing the solar constant in the model, mimicking the solar history. The solar zenith angle is kept constant and equal to 60°. The treatment of clouds in the radiative parameterization is explained in detail by *Chou and Suarez* [2002]; *Chou et al.* [2003]. The cloud optical thickness in the visible spectral region (τ_c) is a function of both the effective radius of the cloud particles r_e and the ice water path (IWP) of the cloud, and it is parameterized as

$$\tau_c = \mathrm{IWP}\frac{1.64}{r_e}, \quad (3)$$

where IWP has units of g m^{-2} and r_e has units of μm. The parameterization of the cloud radiative effect in the visible is independent of the solar spectral bands. The value of r_e is calculated according to the empirical regression by *McFarquhar* [2001] as a function of both the local temperature and the value of the cloud water content. The parameterization of the infrared optical depth of the cloud, takes into account the absorption and scattering of radiation by the cloud [*Chou et al.*, 1999]. The extinction coefficient, the single scattering albedo and the asymmetry factor are all dependent on r_e and on the particular spectral band [*Chou et al.*, 2003]. By specifying the thickness of the cloud (here equal to one model vertical layer) and by specifying the cloud water content, both IWP and r_e can be calculated.

[14] In the control case, we specify the value of cloud liquid water content to 7×10^{-4} g/g, so that a cloud with a thickness of 9 hPa results in an IWP ~ 44 g/m^2. The cloud is first located at a fixed level of 200 hPa (we will discuss the effect of relaxing this assumption to make it consistent with the changes in the vertical temperature structure over the range of solar forcings). We use a single cloud as a proxy for the radiative effect of all types of thin cirrus clouds in the tropics. The selection of this particular cloud is not arbitrary, rather it is such that it roughly matches the radiative forcing from observations in current climate as estimated by *Choi and Ho* [2006]. For the control run, the selected cloud provides a longwave cloud radiative effect of +115 W/m^2 and a shortwave cloud radiative effect of −50 W/m^2. These values coincide roughly with the observed values derived by *Choi and Ho* [2006] for both the longwave and the shortwave radiative effect as well as the net positive cloud radiative effect of these clouds, which is about +46 W/m^2 for all clouds with $\tau < 4$.

3. Results

3.1. Single Column Radiative-Convective Simulation

[15] In the first run we explore the behavior of the tropical surface temperature in radiative convective equilibrium for different values of the thin cirrus cloud coverage. Figure 1a shows the results for this tropics-only column. For the current solar insolation S_0 and current cloud coverage $f \sim 0.16$ the surface temperature is ~298 K. An increase in the coverage of this thin cirrus cloud from $f = 0.16$ to $f = 1$ would produce an increase in the surface temperature in the tropics to about 325 K. From Figure 1a, we notice that the mean tropical temperature is above freezing for constant atmospheric conditions (lower gray line), even at solar insolations of about $S \sim 0.81\ S_0$. We note that the usual statement of the faint young Sun paradox is made in terms of mean surface temperature. Therefore a solution is considered as such when the mean surface temperature is above freezing (hereafter, this is what we will consider a solution). A weaker version of the paradox can be envisioned in which temperatures are above freezing for a significantly large area of the planet. One can also envision a stronger version of the paradox in which one takes the absence of evidence of glaciation as an indication of a completely ice-free Earth.

[16] The three black dashed lines in Figures 1a–1d represent three different relative rates of change for the thin cirrus cloud coverage (so a −10%/K rate of change represents a change from 0.16 to 0.176 from 298 K to 297 K. We

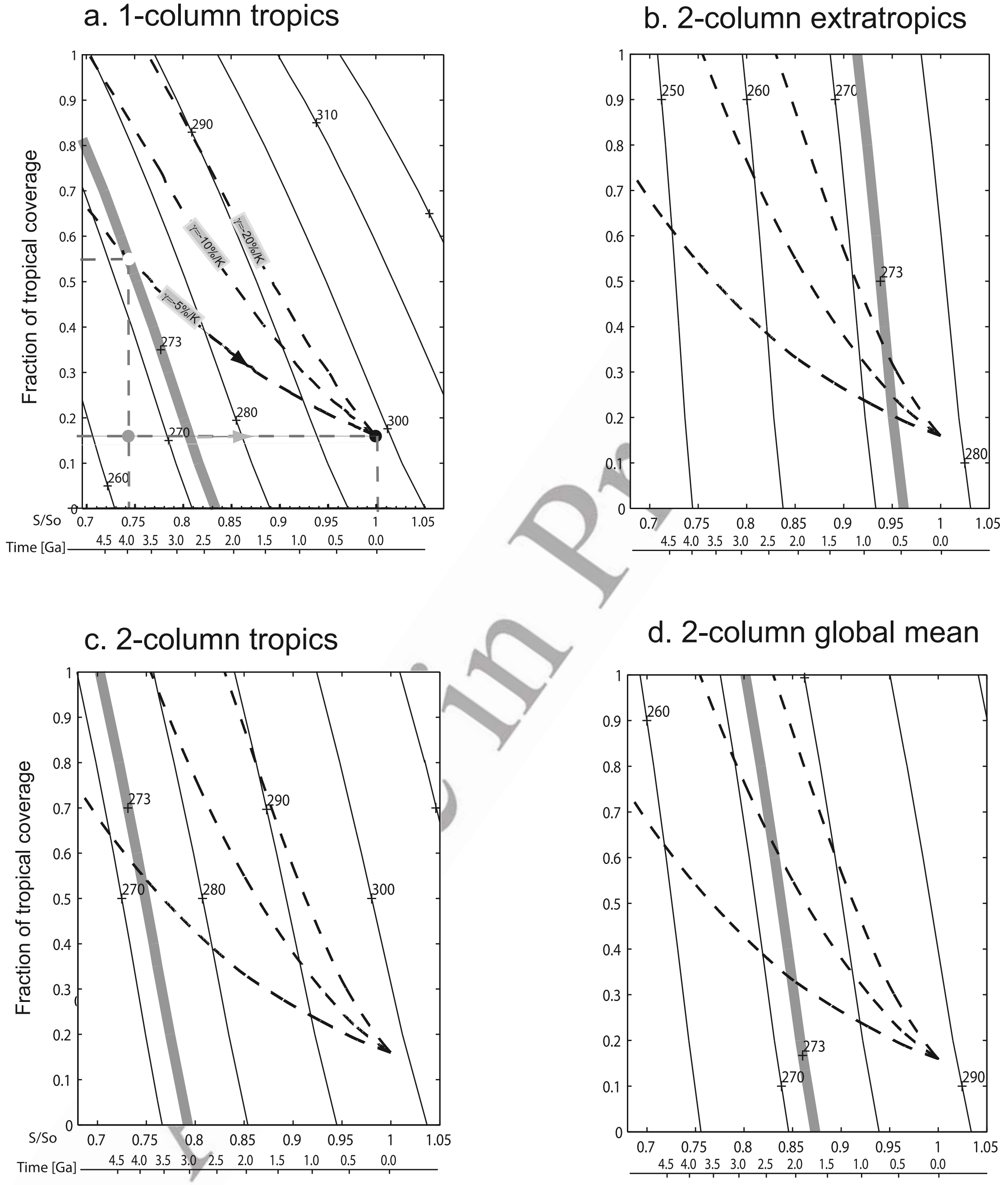

Figure 1. Equilibrium surface temperature corresponding to (a) one-column, tropics-only simulation, (b) extratropical column in the two-column simulation, (c) tropical column in the two-column simulation, and (d) global mean in the two-column simulation. The temperature is indicated by the solid black lines. In Figure 1a a black dot indicates current climate conditions. The white dot indicates the climate surface temperature corresponding to a luminosity of $\sim 0.74 S_0$ and a cloud coverage of 0.55. This climate occurs for a rate of change of -5%/K in the coverage of thin cirrus clouds in the tropics. The two other dashed lines represent rates of change in the cloud coverage of -10%/K and -20%/K as labeled. The grey dot is the equilibrium temperature of a climate with the same luminosity as the white dot but with no cloud feedback. The time scale in the abscissa is calculated according to equation (2).

will denote this rate of change as $\gamma = \frac{1}{f}\frac{\partial f}{\partial T_t}$, where T_t is the mean tropical rate. The rate of change γ represents implicitly the magnitude of the climate feedback associated with increase in thin cirrus clouds. The dashed lines in each of the panels of Figure 1 are for magnitudes of $\gamma = -5$%/K, -10%/K and -20%/K. For this tropics-only case, to sustain surface temperatures above freezing for $S = 0.7\ S_0$, one would need a cirrus coverage of about 0.8. This surface coverage is accomplished with a mere -6%/K change in the cloud coverage.

3.2. Two-Column Radiative-Convective Simulation

[17] Since in the previous simulation we only deal with a tropical column, we cannot test the paradox in its more

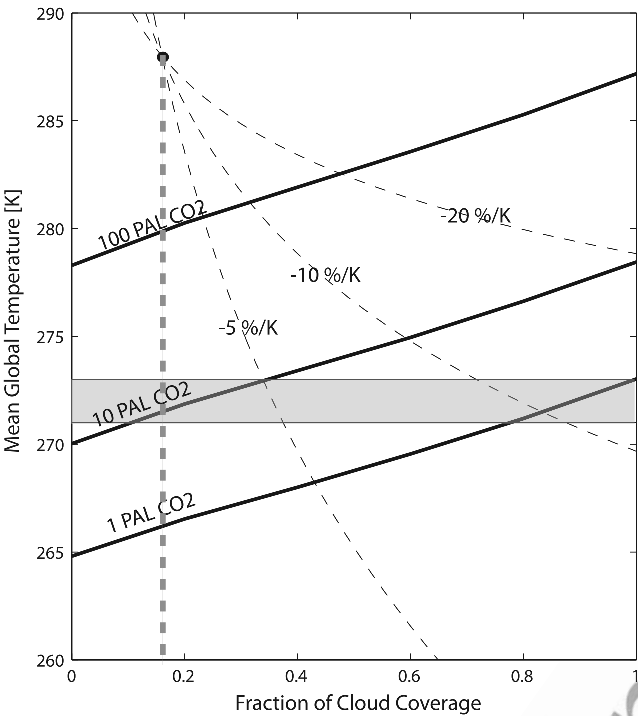

Figure 2. Mean surface temperature corresponding to the two-column radiative convective model for $S = 0.8S_0$. The black solid lines are three different concentrations of CO_2 (PAL stands for present atmospheric level). The dashed lines represent different rates of change in the thin cirrus cloud coverage from the present value of 0.16. The gray horizontal strip is meant to represent a range of temperatures for freezing water between 271 and 273 K.

usual framing, that is, with respect to global mean temperature. Also, since the incoming solar radiation in the single column has been tuned so as to produce the observed current tropical temperature, the heat transported out of the tropical column (implicit in the tuning) decreases in the same proportion as the solar insolation.

[18] We add an extratropical column to the model and we will assume a diffusive heat transport between the two columns, with a constant transport coefficient $K = 3 \times 10^6$ m^2/s over the depth of the model, so that at each time step, the temperature in each layer is calculated as the sum of three tendencies; the radiative heating, the convective adjustment and the meridional transport between the columns.

[19] The results for the two-column simulations are shown in Figures 1b, 1c and 1d. Figure 1c can be directly compared to Figure 1a. We see that assumption of a diffusive transport makes the two-column tropics warmer than the single-column tropics for low cirrus coverages ($f \lesssim 0.45$), and colder for relatively high coverages. Since no change other than the cirrus coverage in the tropical column is made, all change in temperature with cloud coverage in the extratropical column shown in Figure 1b is due to the transport from the tropical column. Figure 1d shows the global mean surface temperature (calculated as simply the average between the surface temperature in the two columns). We see that for constant atmospheric composition (that is following a line of constant $f = 0.16$ in Figure 1d) the global mean surface temperature in our two-column model remains below freezing up until $S = 0.86\ S_0$ giving somewhat warmer temperatures than with previous 1-D radiative-convective simulations (~265 K at $S = 0.8\ S_0$ compared to ~262 K for the same insolation as in work by *Haqq-Misra et al.* [2008]). We are confident that these differences are not due to the peculiarities of the radiative parameterization or to the convective adjustment since our own 1-D tropical simulations with no cloud cover can be used to recover a temperature of about 263 K for $S = 0.8\ S_0$ similar to the ones reported for current atmospheric composition at $S = 0.8\ S_0$ [*Kasting and Catling*, 2003; *Haqq-Misra et al.*, 2008].

[20] The dashed curves in Figure 1d indicate that for some value of γ between -10%/K and -20%/K, there is a solution of the paradox up to $S = 0.8\ S_0$ or for a the range between 2.9 and 1.9 Ga. This solution would imply a total cirrus coverage for the tropics, and a tropical mean temperature of about 285 K. A smaller rate of change of about -7%/K however, can sustain global mean temperatures of only ~261 K for $S = 0.72\ S_0$, although tropical mean surface temperatures in this case would be just above freezing, suggesting that even this moderate rate of change in cloud coverage could explain ice-free conditions for large regions of Earth.

3.3. Thin Cirrus and Increased Greenhouse Gases

[21] CO_2 alone can provide enough greenhouse effect to overcome the paradox. However, geological evidence seems to point to less CO_2 present in the atmosphere than would be required. For instance, *Rye et al.* [1995] argue on the basis of the absence of siderite that CO_2 concentrations higher than about 10 times the present atmospheric level (10 PAL) at 273 K are unlikely at about 2.8 Ga ($S \sim 0.81\ S_0$). This limit is temperature-dependent and goes up to about 50 PAL at temperatures above 300 K. *Kasting* [1993] quotes levels of CO_2 that are several times higher than the paleosol limit (~50 PAL for reaching $T_s \sim 273$ K for $S = 0.8\ S_0$). The discrepancy between required and estimated CO_2 concentrations is also found in other geological and theoretical evidence [see, e.g., *Rollinson*, 2007, and references therein]. CH_4, with a much longer lifetime in an anoxic atmosphere than in the present atmosphere, could provide an additional greenhouse effect. However, recent calculations by *Haqq-Misra et al.* [2008] show that the required concentrations of CH_4 are larger than previously believed. Also the CH_4 greenhouse effect is limited by the formation of a reflective organic haze when CH_4/CO_2 is higher than ~1.

[22] In this section, we will show calculations with a thin cirrus cloud feedback as the one previously described, operating at the same time as an atmosphere with larger CO_2 concentrations. We perform the same runs as in the control case for three different CO_2 concentrations for $S = 0.8\ S_0$. The longwave parameterization by *Chou et al.* [2002] is deemed appropriate even for concentrations of about 100 times present atmospheric levels of CO_2.

[23] Figure 2 shows the surface temperature for three different CO_2 concentrations at $S = 0.8\ S_0$. We see that for the current climate value of $f = 0.16$ (vertical dashed grey line) and for the present value of CO_2 (1 PAL), the surface temperature is about 266 K. For a constant cloud coverage

Table 1. Value of the Cloud Microphysical and Radiative Properties for the Sensitivity Runs[a]

cwc (10^{-4}g/g)	IWP (g/m^2)	τ	r_e (μm)	LW (W/m^2)	SW (W/m^2)	NET (W/m^2)
7	46	1.3	59	120	−70	50
3.5	23	0.73	52	70	−35	35
28	185	4	75	140	−130	10

[a]Abbreviations: cwc, cloud water content; τ, visible optical depth; and LW, SW, and NET, the cloud radiative forcing in the longwave, shortwave, and net, respectively. For all runs the thickness of the cloud is fixed at ~200 m, and the cloud is located at 200 hPa.

the amount of CO_2 required for mean global temperatures to rise above 273 is about 20 PAL CO_2. We recover here the well-known result that the paradox cannot be solved solely on the basis of a higher concentration of CO_2, without getting a result inconsistent with the paleosol data. If we focus on values of CO_2 allowed by the paleosol constraints, a solution to the paradox can be found with relatively small values for the cloud feedback magnitude. For instance, for 1 PAL CO_2, the tropical coverage required to solve the paradox is about 1. For the case in which $CO_2 \sim 10$ PAL, the paradox can be solved with a tropical coverage of only 0.35 and the magnitude of the cloud rate of change required for providing these cloud coverage is $\gamma \sim -5$%/K. This solution is just barely consistent with the paleosol constraint and of course stronger values of the cloud feedback could solve the paradox for lower levels of CO_2. We stress that both consistency with the paleosol data and global mean temperatures above freezing can be achieved (at least for this particular value of solar insolation) invoking only a small magnitude of the cloud rate of change. We also note that while cirrus coverage is less than full, the effect of further increasing cirrus coverage in the mean temperature is mostly linear with cloud coverage as opposed to the effect of the increase in CO_2 (or other greenhouse gases) in the mean temperature that are only logarithmic.

3.4. Sensitivity to Cloud Water Content

[24] Our results so far, have been obtained with a single cloud with optical depth 1.3. We explore the sensitivity to changes in the cloud water content of the cloud. Table 1 summarizes the cloud properties of the different clouds. The cloud radiative effects were calculated with the runs corresponding to f = 0.2. The clouds with either much larger or much smaller cloud water content than the control case produce smaller net radiative effects. Even though there is a net positive cloud radiative effect for the cwc (cloud water content) = 28 × 10^{-4} run, the cloud radiative effect becomes negative for higher cloud fractions and temperatures decrease with cloud coverage (Figure 3b). For the thinner cloud case, the net radiative effect is smaller but positive and very similar to the control case (Figure 3a). This "optimal" net radiative for the control case coincides with the ordering provided by *Choi and Ho* [2006] with respect to shortwave optical depth; smaller positive radiative effect for thinner clouds and smaller and even negative radiative effects for thicker clouds.

3.5. Sensitivity to the Fixed Height Assumption

[25] We have also tested the possibility that the results are sensitive to the assumption of a fixed height or fixed pressure level cloud. An alternative to specifying the cloud at a constant pressure level is the fixed anvil temperature proposed by *Hartmann and Larson* [2002]. They propose that the level at which radiative cooling decreases substantially is controlled by the distribution of water vapor. At the same time, the total amount of water vapor is a strong function of temperature as a consequence of the Clausius-Clapeyron relation. Therefore, radiative cooling rates in the troposphere are a strong function temperature (as long as water vapor is the main driver of the radiative cooling). The divergence of the radiative cooling would then occur at about the same temperature no matter the surface temperature of the climate considered. Since convective heating balances radiative cooling in the tropical free troposphere,

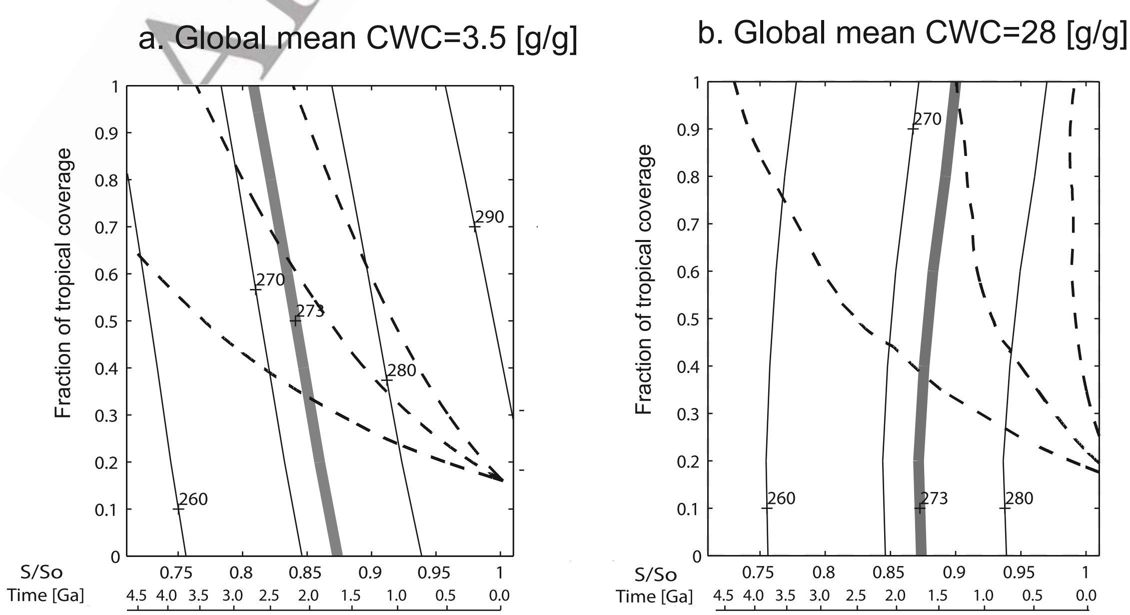

Figure 3. Same as Figure 1d but for clouds with different cloud water content: (a) 3.5 g/g and (b) 28 g/g.

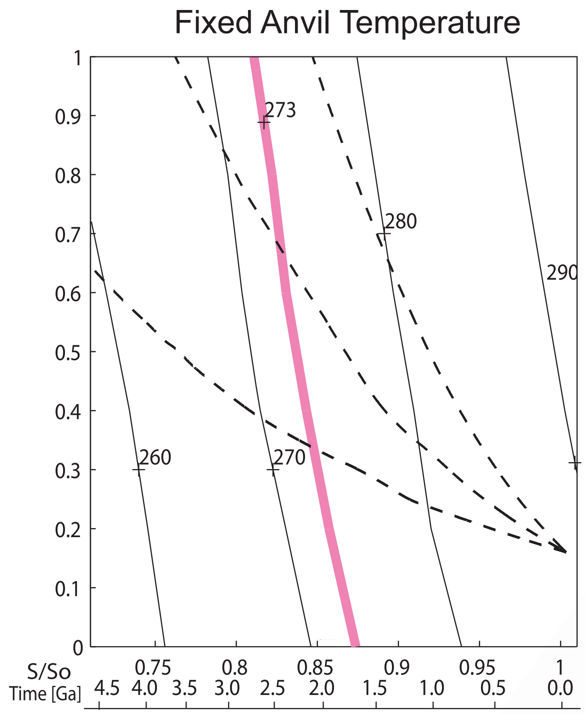

Figure 4. Same as Figure 1d but for a fixed temperature anvil cloud at the 220 K level.

the level at which convection detrains would be strongly constrained to be near a fixed temperature. In Figure 4 we show the results for the global mean surface temperature for the two-column model in the case in which the cloud is located at the 220 K level (this is done iteratively at each time step in the tropical column). The results show that the magnitude of the cloud effect is only modestly reduced. For instance for $S = 0.81\ S_0$, the tropical coverage required for global mean temperatures to be above 273 K in the control case is $f \gtrsim 0.87$. For the fixed anvil temperature case $f \sim 1$.

3.6. Sensitivity to Water Vapor Feedback

[26] So far we have followed the customary assumption of a constant relative humidity profile. In the context of our 1-D single column tropical model, the assumption of strict relative humidity invariance gives a water vapor feedback factor, $\beta \sim 0.4$. Recent studies suggest that the strong positive water vapor feedback implied by the invariance of relative humidity may be within reasonable agreement with satellite observations [*Dessler et al.*, 2008], even though the vertical profile of relative humidity is not strictly conserved [see also *Sun and Held*, 1996]. *Renno et al.* [1994], for instance, showed in the context of a radiative-convective equilibrium model with an explicit hydrological cycle, that changes in the microphysical parameters that control the conversion of water to precipitation and vapor could produce very different equilibrium climates, with different vertical distributions of relative humidity. Since we do not have an explicit parameterization for water vapor in our model, we specify changes in relative humidity with surface temperature to explore the sensitivity of the results to the water vapor feedback strength.

[27] We vary the relative humidity in the model from the original relative humidity profile according to

$$\mathrm{RH}(500\ \mathrm{hPa}) = \alpha \times (T_s - 288) + \mathrm{RH}_0(500\ \mathrm{hPa}), \qquad (4)$$

where RH_0 is the original relative humidity (based on the *Manabe and Wetherald* [1967] profile). Between 200 hPa and 800 hPa, the humidity profile is interpolated from the original profile to the new value at 500 hPa using a cubic spline. Since we have specified the change in the feedback in terms of a change in relative humidity, the magnitude of the feedback will have a dependence on temperature. We use the model output to calculate the magnitude of the water vapor feedback for each case. Figure 5 shows the temperature dependence of the feedback factor for three different values of $\alpha = -0.015$, 0, +0.015. The feedback factor decreases with temperature for all cases. For the imposed changes in relative humidity, the spread of the water vapor feedback tends to decrease with temperature. This is already an indication that uncertainties in the water vapor feedback factor for current climate will be less consequential in determining the temperature for lower global mean surface temperatures.

[28] Figure 6 shows the mean surface temperature for two-column model as a function of the cloud fraction for $S = 0.8\ S_0$. Global mean temperatures ~273 K, are found at $f \sim 1$. Changing α from -0.015/K to 0.015/K has little effect on the total cirrus cloud cover needed for temperatures above freezing. Figure 6 also hints to the fact that changes in water vapor feedback are more efficient for relatively low cloud coverage, since changes in water vapor in the free troposphere are buffered by the presence of the cloud above (notice the shaded regions in Figure 6 showing the reduced range of variation in f required for a given temperature for low coverage). The two effects, namely the decrease in strength of the feedback with temperature and

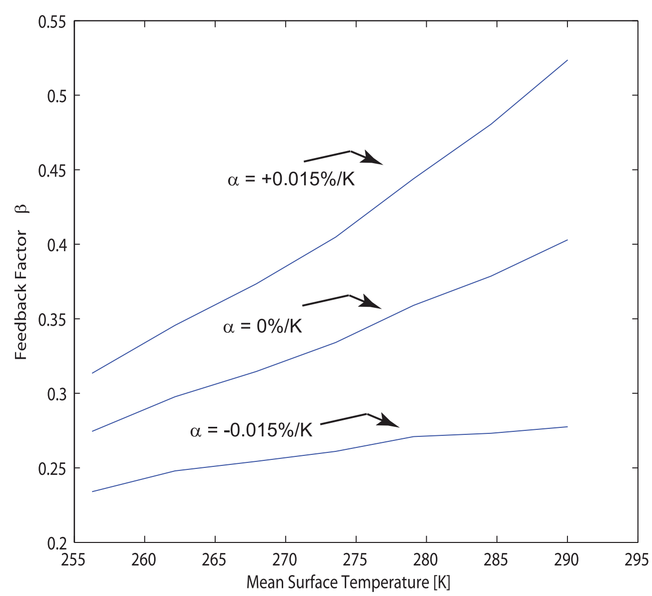

Figure 5. Water vapor feedback factor β as a function of temperature for three different values of the strength of the relative humidity change in equation (4) ($\alpha = -0.015$, 0, and 0.015).

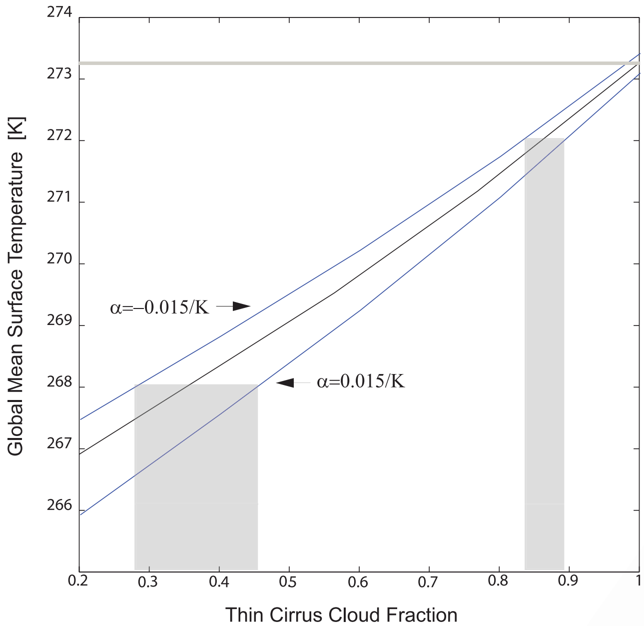

Figure 6. Sensitivity of the results for $S = 0.8S_0$ to the water vapor feedback strength. The two shaded regions show the value of the cloud coverage required to obtain a given global mean temperature (in this case 268 and 272 K).

the decrease in strength of the feedback for large cirrus coverage, suggest that the range of the solution has a low sensitivity to the strength of the water vapor feedback in the context of the present model.

3.7. Sensitivity to the Meridional Heat Flux

[29] We have so far assumed a linear diffusivity law for the heat transport between the tropical and extratropical column. An alternative to the simple linear diffusivity would be to assume a constant temperature difference between the two columns so as to crudely represent a baroclinic adjustment over the different possible climates considered [e.g., *Stone*, 1978]. This is accomplished in the model by allowing the diffusivity coefficient to change while keeping a constant target temperature difference between the two columns (in this case 20 K). In Figure 7 we see the result of this modification. The situation in the global mean is not very different from the constant diffusivity depicted in Figure 1d, so that the main result does not change appreciably; the mean global temperature can be above freezing for luminosities ∼0.81 and full tropical cirrus coverage. However, since in the case of the fixed temperature difference the tropics are colder than in the control case (for instance, the mean tropical temperature is 282 K for $S = 0.81\ S_0$ and $f = 1$ in the fixed meridional temperature case and 285 K in the linear diffusivity case for the same conditions) the values of γ required to accomplish the needed full tropical cirrus coverage are therefore smaller in the fixed meridional temperature case ($\gamma \sim -12\%/K$ compared to $\gamma \sim -15\%/K$ in the control case). By providing warmer extratropical temperatures, this alternative treatment for the meridional heat flux would also delay the onset of solutions unstable to an eventual ice-albedo feedback. Besides the control case and the constant temperature case, we have a third assumption about the meridional heat transport. In the case of a single column tropics depicted in Figure 1 the meridional heat transport is implicit (since the incoming solar radiation is tuned to obtain current tropical temperatures) and reduced by the same fraction as the reduction in incoming solar radiation. In the single column tropical cases the heat transport becomes less effective as the climate cools (similar to the decrease in transport efficiency predicted from maximum entropy considerations [*Gerard et al.*, 1990]). This isolation of the tropics from the extratropics also allows for a more effective functioning of the tropical cirrus clouds in resisting the changes in the solar constant and would provide a more robust "partial" solution to the paradox, with relatively warm oceans in the tropical regions of the planet.

4. Discussion

[30] We have presented simplified radiative-convective equilibrium calculations to investigate the role of thin cirrus clouds in providing a solution for the faint young Sun paradox. In the context of our model, solutions do in fact exist. Tropical thin cirrus clouds can either solve the paradox in the sense of providing *global* mean temperatures above freezing (after ∼2.9 Ga) or in a weaker sense, less than full tropical cirrus coverage can provide *tropical* mean temperatures above freezing for all Earth's existence (in the context of this model). The solutions are characterized by a colder tropical temperature and therefore by thin cirrus clouds acting as a net negative feedback to the solar forcing.

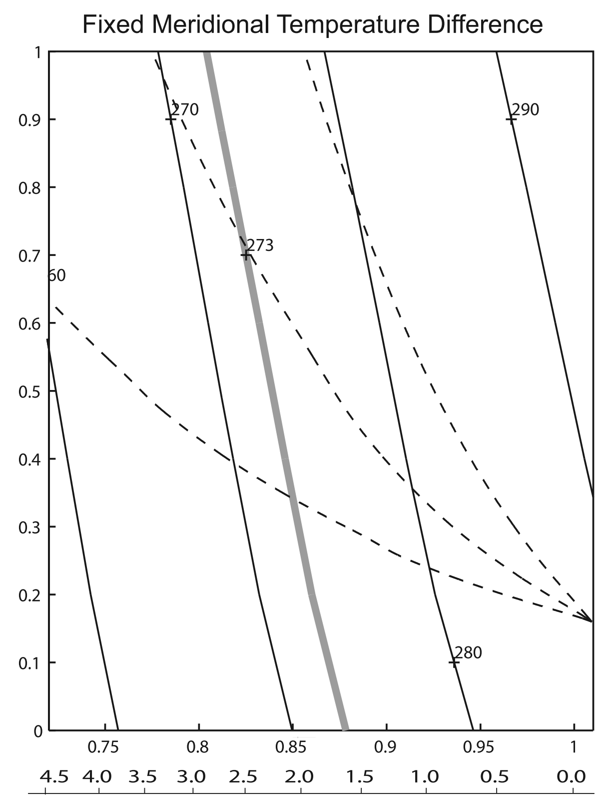

Figure 7. Same as Figure 1d but for a fixed difference in surface temperature between the tropical and the extratropical column.

[31] Given that thin cirrus clouds can indeed solve the paradox, we focus the discussion on the question of the plausibility of these solutions. There is the suggestion that a negative feedback such as the one required might in fact be operating in current climate [*Lindzen et al.*, 2001]. According to this suggestion, called the iris hypothesis, an increase in sea surface temperature (through an increase in the specific humidity of the air that participates in convection) can make precipitation in convective clouds more efficient. In this way less condensate is rained out from deep convective clouds and therefore more condensate is available to be detrained from the top of the cloud to form cirrus clouds. The iris hypothesis is controversial and it would be lengthly to discuss all the arguments here. Apparent confirmation for the iris effect came from the analysis of the OLR trends over the last two decades, showing a strong increase in the OLR compared to a relatively smaller decrease in the shortwave reflectivity in the tropics [*Wielicki et al.*, 2002; *Chen et al.*, 2002]. Using a combination of data sets, *Hatzidimitriou et al.* [2004] traced the OLR increase mainly to a decrease in the upper level cloud coverage and a drying of the upper troposphere. As pointed out by *Chou and Lindzen* [2005] this large increase in OLR was also consistent with a much larger value in the relative change in cloud fraction with temperature than the original −22%/K found by *Lindzen et al.* [2001]. The OLR trends were recently revised down to only about a quarter of the original value [*Wong et al.*, 2006], although the OLR trend continues to be larger than the Planck response expected from an increase in the tropical mean temperature over the same period. Recently, *Lindzen and Choi* [2009] studied variations in the outgoing radiative fluxes with respect to changes in the average tropical temperature in intraseasonal scales. A total negative feedback was deduced from the outgoing longwave response of the tropics. If a strong positive water vapor feedback is realistic [e.g., *Dessler et al.*, 2008], then the combined effect of water vapor feedback and lapse rate feedback must be more than compensated by a strong unknown process acting on modifying the longwave flux. This process cannot be distinguished from the bulk of the longwave response in the analysis by *Lindzen and Choi* [2009], but it most likely resides in the combined behavior of clouds and water vapor in the tropics. This leaves open the possibility that a negative feedback such as the iris is operating in the present climate.

[32] We have assumed so far that the magnitude of the cloud changes with respect to temperature is absolute, that is, they already contain any possible dependence on changes in convective activity that will arise as the incoming radiation at the surface decreases. Theoretical arguments and model simulations both indicate that changes in precipitation with global mean temperature are relatively small (∼2−4%/K [*Held and Soden*, 2006; *O'Gorman and Schneider*, 2008]). A correction to account for the reduction of precipitation or convective activity will indeed be required. One can diagnose from the surface budget, the total convective heating in the model, which, in the tropics has to be equal to the precipitation. The changes in precipitation in the model depend on the magnitude of the feedback itself, given that a stronger feedback would reduce the net incoming solar radiation at the surface more rapidly than in the case of a weaker feedback. This is illustrated in Figure 8 which shows the increase in precipitation with temperature for three different values of the absolute cloud change γ. One can write $\gamma = \gamma' + \frac{1}{P}\frac{\partial P}{\partial T_t}$, so that the relative changes in cloud fraction γ', have to be higher in magnitude than the magnitude of the change γ required to compensate for the decrease of precipitation in a colder climate. Fitting exponential functions to the model-diagnosed precipitation one finds that the quantity $\frac{1}{P}\frac{\partial P}{\partial T_t}$ goes from about 3%/K to 7%/K.

[33] Regarding observed value of γ', different data sets and analyses point to values between −2%/K to −22%/K for current climate (see R. Rondanelli and R. S. Lindzen, Comments on "Variations of tropical upper tropospheric clouds with sea surface temperature and implications for radiative effects," submitted to *Journal of Geophysical Research*, 2010) for a discussion of some of the methodological issues involved). These empirically derived rates of change γ', usually refer to some observable that is a proxy for the thin cirrus clouds rather than the thin cirrus clouds themselves. Nevertheless, the magnitude of these changes is consistent with what is required to solve the paradox (for instance from Figures 1c and 1d, the tropical temperature for $S = 0.8\ S_0$ and $f = 1$ is about 285 K which gives a rate of change of $\gamma \sim -15\%/\mathrm{K}$, $\gamma' \sim -20\%/\mathrm{K}$).

[34] One can ask what happens in the situation in which the tropical atmosphere is already completely covered by cirrus clouds and temperatures continue to decrease. One could expect that if the cloud feedback still operates beyond full coverage, an increase in the cloud water content or in the thickness of the cirrus clouds would ensue. The cloud feedback can only operate until the cloud is thick enough ($\tau \sim 10$) that surface cooling instead of heating is obtained (as in Figure 3b). At the same time, if the cloud cover is thick enough to reflect most of the incoming solar radiation, convection (and therefore the source of the cloud) will shut off. Microphysical effects such as an enhanced precipitation from the cirrus cloud might prevent this from happening. However, without a mechanistic model one cannot go beyond speculation on this point. We only note here that the mechanism such as the one described will have a limit for low temperatures. The availability of water for sustaining a total cirrus coverage does not pose a problem. Even with a weaker hydrological cycle as expected in a colder climate (rainfall rate estimated in ∼2 mm/d for a surface temperature of ∼270 K [*O'Gorman and Schneider*, 2008]) and with a 44 g/m^2 cloud (with an accompanying water vapor layer of 400 g/m^2) and assuming that a typical ice particle dissipates over a day, the detrainment flux required to sustain such a cloud is only about ∼2% of the precipitation rate.

[35] Although the literature about the paradox usually focuses on greenhouse gas solutions [*Kasting and Catling*, 2003; *Shaw*, 2008], solutions based on cloud feedbacks have been put forth in the past. Based on the model developed by *Wang et al.* [1981] in which cloud cover is considered proportional to the convective heating (or total precipitation), *Rossow et al.* [1982] [see also *McGuffie and Henderson-Sellers*, 2005, section 4.6.1] proposed a solution to the paradox based on the negative feedback resulting from a decrease in planetary albedo and a decrease in the cloud water content (and therefore in the visible optical depth) of clouds in a colder climate. Our solution on the

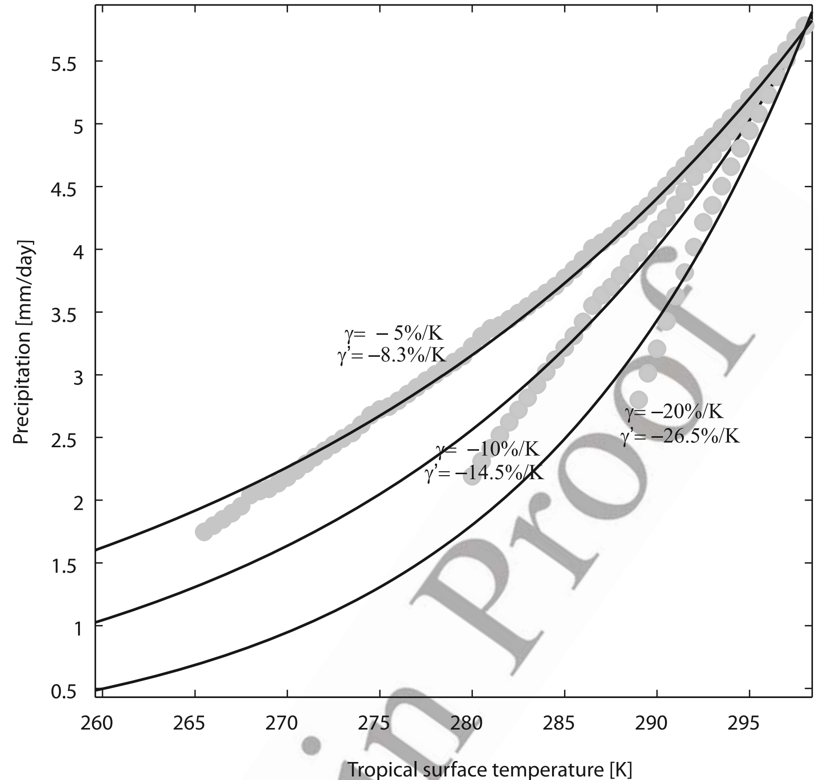

Figure 8. Changes in precipitation diagnosed from the surface balance in the tropical column of the model. The gray dots show the precipitation diagnosed from the model for three values of the magnitude of the feedback $\gamma = 5$, 10, and 20%/K. The black lines are exponential fits to the precipitation curves from which a value of γ' was deduced.

other hand leaves the albedo almost unchanged as it mostly depends on the longwave radiative effect of upper level thin cirrus clouds. The solution by *Rossow et al.* [1982] and our solution are not mutually exclusive. Several cloud feedbacks other than the one resulting from the change in thin cirrus are possible in reality and have been muted in the present model (for instance area coverage and composition of stratocumulus clouds in the subtropics). Despite progress since the time of the writing of the study by *Rossow et al.* [1982], clouds continue to be "the major source of uncertainty" in climate models [e.g., *Schwartz*, 2008]. As in previous studies dealing with clouds and the faint young Sun problem (*Cogley and Henderson-Sellers* [1984] provide references to previous work on this issue) (see also the mechanism proposed by *Shaviv* [2003]), we conclude that a negative cloud feedback can indeed solve the paradox if the Archean climate was somewhat colder than present. (How much colder will also depend on the strength of the feedback.) We have followed the customary assumptions of neglecting the ice-albedo feedback, fixing the relative humidity and muting the effect of clouds to a large extent, we have also assumed a very simplified treatment for the heat transport between tropics and extratropics. None of these assumptions is entirely satisfactory. Given the simplified nature of this radiative-convective model, our study is only exploratory.

[36] Solving the paradox down to a luminosity of $S = 0.8\ S_0$, requires a climate with an equilibrium sensitivity parameter to solar forcing $\lambda = \Delta T_s/\Delta S$ of about 0.29 K/(W m^{-2}). This sensitivity value is certainly smaller than any of the sensitivities to CO_2 forcing in current GCMs [*Intergovernmental Panel on Climate Change*, 2007], but it is within the lower range of estimates made from observations [e.g., *Schwartz*, 2008]. One finds values of $\lambda \sim 0.4$ K/(W m^{-2}) for the 1-D radiative-convective models without clouds (using for instance the results by *Kasting* [1987]); we also found a nearly identical value for λ in our two-column radiative convective model with no cloud feedback. As shown in section 3.3, small changes in the rate of change of cloud coverage can reduce the amount of greenhouse gases needed to reach consistency with the geological evidence. These clouds changes are associated with small changes in the model climate sensitivity (a −5%/K rate of change in the thin cirrus coverage is equivalent to a sensitivity $\lambda \sim 0.37$ K/(W m^{-2})) in the present model).

5. Concluding Remarks

[37] Using simple radiative-convective simulations we have tested the idea that a coverage of tropical cirrus clouds much larger than present could resolve the faint young Sun paradox. We have found that relatively modest cloud changes can indeed provide sufficient cirrus coverage for the mean global temperature to be above freezing for $S \gtrsim 0.8\ S_0$ and for the mean tropical temperature to be above freezing for $S \gtrsim 0.7\ S_0$ without additional greenhouse gases. The model cloud is specified to have similar cloud radiative effect as reported in current climate observations. We tested

the sensitivity of the results to cloud water content, to the assumption of a constant pressure level of detrainment and to a range for the strength of the water vapor feedback. We also looked at two different treatments for the meridional heat transport. We find small sensitivities to all these factors in the present model. Although we describe a very specific cloud negative feedback, our results can be understood in a more general perspective with respect to the faint young Sun paradox; a moderate negative climate feedback can indeed resolve the paradox without resorting to large changes in the greenhouse gas content of the Archean atmosphere.

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References

Catling, D., and J. Kasting (2007), Planetary atmospheres and life, in *Planets and Life: The Emerging Science of Astrobiology*, pp. 91–116, Cambridge Univ. Press, Cambridge, U. K.

Chen, J., B. Carlson, and A. Del Genio (2002), Evidence for strengthening of the tropical general circulation in the 1990s, *Science*, *295*(5556), 838–841.

Choi, Y.-S., and C.-H. Ho (2006), Radiative effect of cirrus with different optical properties over the tropics in MODIS and CERES observations, *Geophys. Res. Lett.*, *33*, L21811, doi:10.1029/2006GL027403.

Chou, M., and R. Lindzen (2005), Comments on "Examination of the decadal tropical mean ERBS nonscanner radiation data for the iris hypothesis," *J. Clim.*, *18*(12), 2123–2127.

Chou, M., and M. Suarez (2002), A Solar Radiation parameterization for atmospheric studies, *NASA Tech. Memo, TM-1999-10460*, vol. 15, 52 pp.

Chou, M., K. Lee, S. Tsay, and Q. Fu (1999), Parameterization for cloud longwave scattering for use in atmospheric models, *J. Clim.*, *12*(1), 159–169.

Chou, M., K. Lee, and P. Yang (2002), Parameterization of shortwave cloud optical properties for a mixture of ice particle habits for use in atmospheric models, *J. Geophys. Res.*, *107*(D21), 4600, doi:10.1029/2002JD002061.

Chou, M. D., M. J. Suarez, X.-Z. Liang, and M. M.-H. Yan (2003), A Thermal Infrared Radiation Parameterization for Atmospheric studies, *NASA Tech. Memo, TM-2001-104606*, vol. 19, 55pp.

Cogley, J., and A. Henderson-Sellers (1984), The origin and earliest state of the Earth's hydrosphere, *Rev. Geophys*, *22*(2), 131–175.

Dessler, A. E., Z. Zhang, and P. Yang (2008), Water-vapor climate feedback inferred from climate fluctuations, 2003–2008, *Geophys. Res. Lett.*, *35*, L20704, doi:10.1029/2008GL035333.

Emanuel, K. A. (2007), Quasi-equilibrium dyamics of the tropical atmosphere, in *The Global Circulation of the Atmosphere*, pp. 186–218, Princeton Univ. Press, Princeton, N. J.

Endal, A., and K. Schatten (1982), The faint young Sun-climate paradox: Continental influences, *J. Geophys. Res.*, *87*, 7295–7302.

Gerard, J., D. Delcourt, and L. Francois (1990), The maximum entropy production principle in climate models: Application to the faint young Sun paradox, *Q. J. R. Meteorol. Soc.*, *116*, 1123–1132, doi:10.1256/smsqj.49505.

Gough, D. (1981), Solar interior structure and luminosity variations, *Sol. Phys.*, *74*(1), 21–34.

Haqq-Misra, J. D., S. D. Domagal-Goldman, P. J. Kasting, and J. F. Kasting (2008), A revised, hazy methane greenhouse for the Archean Earth, *Astrobiology*, *8*(6), 1127–1137, doi:10.1089/ast.2007.0197.

Hart, M. (1978), The evolution of the atmosphere of the Earth, *Icarus*, *33*, 23–39.

Hartmann, D., and K. Larson (2002), An important constraint on tropical cloud–climate feedback, *Geophys. Res. Lett.*, *29*(20), 1951, doi:10.1029/2002GL015835.

Hatzidimitriou, D., I. Vardavas, K. Pavlakis, N. Hatzianastassiou, C. Matsoukas, and E. Drakakis (2004), On the decadal increase in the tropical mean outgoing longwave radiation for the period 1984–2000, *Atmos. Chem. Phys.*, *4*, 1419–1425.

Held, I., and B. Soden (2006), Robust responses of the hydrological cycle to global warming, *J. Clim*, *19*(21), 5686–5699.

Henderson-Sellers, A., and J. G. Cogley (1982), The Earth's early hydrosphere, *Nature*, *298*, 832–835.

Intergovernmental Panel on Climate Change (2007), *Climate Change 2007: The Physical Science Basis*, edited by S. Solomon et al., Cambrige Univ. Press, Cambridge, U. K.

Kasting, J. (1987), Theoretical constraints on oxygen and carbon dioxide concentrations in the Precambrian atmosphere, *Precambrian Res.*, *34*(3–4), 205–229.

Kasting, J. (1993), Earth's early atmosphere, *Science*, *259*(5097), 920–926.

Kasting, J., and D. Catling (2003), Evolution of a habitable planet, *Annu. Rev. Astron. Astrophys.*, *41*, 429–463.

Kasting, J., O. Toon, and J. Pollack (1988), How climate evolved on the terrestrial planets, *Sci. Am.*, *258*(2), 90–97.

Kuhn, W., and S. Atreya (1979), Ammonia photolysis and the greenhouse effect in the primordial atmosphere of the Earth, *Icarus*, *37*, 207–213.

Lindzen, R. S., and Y.-S. Choi (2009), On the determination of climate feedbacks from ERBE data, *Geophys. Res. Lett.*, *36*, L16705, doi:10.1029/2009GL039628.

Lindzen, R., and B. Farrell (1980), The role of polar regions in global climate, and a new parameterization of global heat transport, *Mon. Weather Rev.*, *108*(12), 2064–2079.

Lindzen, R. S., M.-D. Chou, and A. Y. Hou (2001), Does the Earth have an adaptive infrared iris?, *Bull. Am. Meteorol. Soc.*, *82*(3), 417–432.

Liou, K. (2005), Cirrus clouds and climate, in *Yearbook of Science and Technology*, pp. 51–53, McGraw-Hill, New York.

Liou, K., and K. Gebhart (1982), Numerical experiments on the thermal equilibrium temperature in cirrus cloudy atmospheres, *J. Meteorol. Soc. Jpn.*, *60*, 570–582.

Lovelock, J., and M. Whitfield (1982), Life span of the biosphere, *Nature*, *296*, 561–563.

Mace, G., M. Deng, B. Soden, and E. Zipser (2006), Association of tropical cirrus in the 10–15-km layer with deep convective sources: An observational study combining millimeter radar data and satellite-derived trajectories, *J. Atmos. Sci.*, *63*(2), 480–503.

Manabe, S., and R. Strickler (1964), Thermal equilibrium of the atmosphere with a convective adjustment, *J. Atmos. Sci.*, *21*(4), 361–385.

Manabe, S., and R. Wetherald (1967), Thermal equilibrium of the atmosphere with a given distribution of relative humidity, *J. Atmos. Sci.*, *24*(3), 241–259.

McFarquhar, G. (2001), Comments on 'Parametrization of effective sizes of cirrus-cloud particles and its verification against observations' by Zhian Sun and Lawrie Rikus (October B, 1999, 125, 3037–3055), *Q. J. R. Meteorol. Soc.*, *127*, 261–266.

McGuffie, K., and A. Henderson-Sellers (2005), *A Climate Modelling Primer*, John Wiley, Hoboken, N. J.

Minton, D., and R. Malhotra (2007), Assessing the massive young Sun hypothesis to solve the warm young Earth puzzle, *Astrophys. J.*, *660*(2), 1700–1706.

Newman, M., and R. Rood (1977), Implications of solar evolution for the Earth's early atmosphere, *Science*, *198*(4321), 1035–1037.

O'Gorman, P., and T. Schneider (2008), The hydrological cycle over a wide range of climates simulated with an idealized GCM, *J. Clim.*, *22*(21), 5676–5685, doi:10.1175/2009JCLI2701.1.

Owen, T., R. Cess, and V. Ramanathan (1979), Enhanced CO_2 greenhouse to compensate for reduced solar luminosity on early Earth, *Nature*, *277*, 640–642.

Paltridge, G. (1978), The steady-state format of global climate, *Q. J. R. Meteorol. Soc.*, *104*, 927–945.

Pavlov, A., J. Kasting, L. Brown, K. Rages, and R. Freedman (2000), Greenhouse warming by CH_4 in the atmosphere of early Earth, *J. Geophys. Res.*, *105*, 11,981–11,990.

Pavlov, A., M. Hurtgen, J. Kasting, and M. Arthur (2003), Methane-rich Proterozoic atmosphere?, *Geology*, *31*(1), 87–90.

Pierrehumbert, R. T. (2010), *Principles of Planetary Climate*, Cambridge Univ. Press, Cambridge, U. K.

Pinti, D. (2005), The origin and evolution of the oceans, in *Lectures in Astrobiology*, vol. 1, edited by M. Gargaud et al., pp. 83–111, Springer, New York.

Renno, N., K. Emanuel, and P. Stone (1994), Radiative-convective model with an explicit hydrologic cycle: 1. Formulation and sensitivity to model parameters, *J. Geophys. Res.*, *99*, 14,429–14,442, doi:10.1029/94JD00020.

Rollinson, H. (2007), *Early Earth Systems: A Geochemical Approach*, Blackwell, Malden, Mass.

Rossow, W., A. Henderson-Sellers, and S. Weinreich (1982), Cloud feedback: A stabilizing effect for the early Earth?, *Science*, *217*(4566), 1245–1247.

Rye, R., P. Kuo, and H. Holland (1995), Atmospheric carbon dioxide concentrations before 2.2 billion years ago, *Nature*, *378*, 603–605.

Sackmann, I., and A. Boothroyd (2003), Our Sun: V. A bright young Sun consistent with helioseismology and warm temperatures on ancient Earth and Mars, *Astrophys. J.*, *583*(2), 1024–1039.

Sagan, C., and G. Mullen (1972), Earth and Mars: Evolution of atmospheres and surface temperatures, *Science*, *177*(4043), 52–56.

Sassen, K., Z. Wang, and D. Liu (2008), Global distribution of cirrus clouds from CloudSat/Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) measurements, *J. Geophys. Res.*, *113*, D00A12, doi:10.1029/2008JD009972.

Schwartz, S. (2008), Uncertainty in climate sensitivity: Causes, consequences, challenges, *Energy Environ. Sci.*, *1*(4), 430–453.

Schwartzman, D., and T. Volk (1989), Biotic enhancement of weathering and the habitability of Earth, *Nature*, *340*, 457–460.

Schwartzman, D., and T. Volk (2004), Does life drive disequilibrium in the biosphere?, in *Scientists Debate Gaia: The Next Century*, edited by S. H. Schneider et al., pp. 129–135, MIT Press, Cambridge, Mass.

Schwarzschild, M. (1958), *Structure and Evolution of the Stars*, Princeton Univ. Press, Princeton, N. J.

Shaviv, N. (2003), Toward a solution to the early faint Sun paradox: A lower cosmic ray flux from a stronger solar wind, *J. Geophys. Res.*, *108*(A12), 1437, doi:10.1029/2003JA009997.

Shaw, G. H. (2008), Earth's atmosphere—Hadean to early Proterozoic, *Chem. Erde Geochem.*, *68*(3), 235–264, doi:10.1016/j.chemer.2008.05.001.

Stone, P. (1978), Baroclinic adjustment, *J. Atmos. Sci.*, *35*(4), 561–571.

Sun, D., and I. Held (1996), A comparison of modeled and observed relationships between interannual variations of water vapor and temperature, *J. Clim.*, *9*(4), 665–675.

Volk, T. (1987), Feedbacks between weathering and atmospheric CO_2 over the last 100 million years, *Am. J. Sci.*, *287*(8), 763–779.

Walker, J., P. Hays, and J. Kasting (1981), A negative feedback mechanism for the long-term stabilization of the Earth's surface temperature, *J. Geophys. Res.*, *86*, 9776–9782.

Wang, W., W. Rossow, M. Yao, and M. Wolfson (1981), Climate sensitivity of a one-dimensional radiative-convective model with cloud feedback, *J. Atmos. Sci.*, *38*(6), 1167–1178.

Wielicki, B., et al. (2002), Evidence for large decadal variability in the tropical mean radiative energy budget, *Science*, *295*(5556), 841–844.

Wilde, S., J. Valley, W. Peck, and C. Graham (2001), Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago, *Nature*, *409*, 175–178.

Wong, T., B. Wielicki, R. Lee III, G. Smith, K. Bush, and J. Willis (2006), Reexamination of the observed decadal variability of the Earth Radiation Budget using altitude-corrected ERBE/ERBS nonscanner WFOV data, *J. Clim.*, *19*(16), 4028–4040.

Zahnle, K., and N. H. Sleep (2002), Carbon dioxide cycling through the mantle and implications for the climate of ancient Earth, *Geol. Soc. Spec. Publ.*, *199*(1), 231–257, doi:10.1144/GSL.SP.2002.199.01.12.

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