Cloud Feedback on Earth’s Long-term Climate Simulated by a Near-global Cloud-permitting Model

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Key Points:

• A near-global cloud-permitting model with a resolution of 10 km × 14 km is employed to test whether there is a long-term cloud feedback.
• The cloud feedback does have a net cooling effect when the Sun becomes brighter and meanwhile the CO₂ concentration decreases.
• These results confirm that cloud feedback is a part of the solution to the faint young Sun problem but its magnitude is relatively small.

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Abstract
The Sun becomes brighter with time, but Earth’s climate is roughly temperate for life during its long-term history; for early Earth, this is known as the Faint Young Sun Problem (FYSP). Besides the carbonate-silicate feedback, recent researches suggest that a long-term cloud feedback may partially solve the FYSP. However, the general circulation models they used cannot resolve convection and clouds explicitly. This study re-investigates the clouds using a near-global cloud-permitting model without cumulus convection parameterization. Our results confirm that a stabilizing shortwave cloud feedback does exist, and its magnitude is \( \approx 6 \text{ W m}^{-2} \) or 14\% of the energy required to offset a 20\% fainter Sun than today, or \( \approx 10 \text{ W m}^{-2} \) or 16\% for a 30\% fainter Sun. When insolation increases and meanwhile CO\(_2\) concentration decreases, low-level clouds increase, acting to stabilize the climate by raising planetary albedo, and vice versa.

Plain Language Summary
The emergence and evolution of life require a relatively stable climate environment. In the solar system, life has been found only on Earth and appeared since about four billion years ago. The underlying mechanisms for maintaining the long-term climate on Earth is an important question but the answer is not completely clear. In this study, we re-investigate a recently proposed mechanism, a stabilizing cloud feedback, using a high-resolution cloud-permitting model in a near-global domain. Our simulations confirm that a stabilizing cloud feedback does exist, but its magnitude is relatively small.

1 Introduction
The luminosity of the Sun increases with time, and the solar constant in the Archean Eon was 20–30\% lower than that today (Gough, 1981). If other climate-controlling factors were the same as present, Earth would have been in a globally ice-covered snowball state during the Archean (Sagan & Mullen, 1972), but much evidence indicates that there was surface liquid water (Feulner, 2012). Hence, there should be some factors compensating for this discrepancy (Sagan & Mullen, 1972; Feulner, 2012; Charnay et al., 2020), and it is one of the most important problems on Earth’s long-term climate (Pierrehumbert, 2010). A number of studies show that biogeochemical feedbacks, such as the carbonate-silicate weathering cycle, are key to stabilizing the climate. The increase of greenhouse gas contents, such as CO\(_2\) and CH\(_4\), is likely responsible for the FYSP (e.g., Wolf and Toon (2013); Byrne and Goldblatt (2014); Le Hir et al. (2014)). Other factors or feedbacks may have also contributed to solving this problem, such as a different atmospheric pressure, a less continent coverage, N\(_2\)-H\(_2\) collision-induced warming, and a stronger tidal heating induced by the closer Moon (Goldblatt et al., 2009; Rosing et al., 2010; Goldblatt & Zahnle, 2011; Wordsworth & Pierrehumbert, 2013; Heller et al., 2021). A recent study of Goldblatt et al. (2021) suggested that a long-term cloud feedback exists and it could also be a part of the solution to the FYSP.

Clouds are one of the key factors that determine planetary climate. For low-level clouds, the cooling effect dominates; but for high-level clouds, the greenhouse warming effect dominates (Hartmann, 2005; Pierrehumbert, 2010). Goldblatt and Zahnle (2011) explored how clouds could resolve the FYSP in one-dimensional (1D) radiative-transfer simulations. They showed that a maximum radiative effect change of 15 W m\(^{-2}\) can be traced to a plausible low-level cloud reduction or high-level cloud increase. Using 3D GCMs, Charnay et al. (2013) and Wolf and Toon (2013) focused on the Archean Eon, and they found that the changes of clouds have a warming effect on the surface, mainly due to the reduction of low-level clouds. Goldblatt et al. (2021) systematically studied the cloud feedback using two GCMs: CAM4 and CAM5. They kept almost the same global-mean surface temperature (\(\Delta T<0.5\)K) by increasing solar constant and meanwhile decreasing CO\(_2\) concentration. Their work showed that shortwave cloud feedback has a stabilizing effect on the long-term
climate, and it can offset 20 W m$^{-2}$ or 40% in percentage of the required forcing when the insolation decreases from 1.0 to 0.8 times modern value. Below, we will show that they have overestimated the strength of the shortwave cloud feedback because the direct effect of varying solar radiation on the magnitude of shortwave cloud radiative effect without any change of cloud properties was not excluded. After subtracting this part, the magnitude of the shortwave cloud feedback in CAM4 is 8.6 W m$^{-2}$ or 18.5% (see Section 3.1).

The horizontal grid sizes of GCMs are always larger than $\approx$100 km, but the sizes of convection and clouds are in the order of $O$($1$–$10$) km or even less, so that empirical convection and cloud parameterizations are necessary for GCMs. However, for GCMs, cloud feedback is the largest uncertainty source, leading to significant inter-model differences (e.g., Soden and Held (2006); Vial et al. (2017)). Therefore, it is important to use higher-resolution models or models with different parameterization schemes to re-examine the long-term cloud feedback proposed in Goldblatt et al. (2021). One method is using a cloud-resolving model that has a resolution of $O$(1) km or smaller, but unfortunately, global-scale cloud-resolving simulations are far beyond present computation resources, especially when long-time integration and multiple experiments are required. Here, we use a cloud-permitting version with a resolution of $\approx$10 km, which is coarser than cloud-resolving models but much finer than GCMs, to re-simulate the clouds and the cloud feedback.

2 Model Descriptions and Experimental Designs

We run a series of simulations with the version 6.11.6 of the System for Atmosphere Modeling (SAM, Khairoutdinov and Randall (2003)). Anelastic momentum equations are used to explicitly resolve non-hydrostatic flows instead of the hydrostatic approximation used in GCMs. Radiation scheme is the same as the Community Atmosphere Model version 3 (CAM3) (Collins et al., 2006), and it is the same as that used in the CAM4 simulations of Goldblatt et al. (2021). There are still parametrizations in SAM that are used to calculate microphysics and sub-grid fluxes. Microphysics parameterization is based on a simple one-moment scheme, and a Smagorinsky-type closure scheme is used to calculate subgrid-scale fluxes (Khairoutdinov & Kogan, 1999; Khairoutdinov & Randall, 2003).

We use an aquaplanet mode, which means there is no land, and the northern and southern hemispheres are symmetric. The horizontal resolution is $\approx$14 km in latitude and $\approx$10 km in longitude, and the domain size corresponds to 60°S to 60°N in latitude and 0° to 180° in longitude. Rigid boundary conditions are used at $y$ boundaries, and periodic boundary conditions are used at $x$ boundaries. There are 38 vertical levels up to $\approx$35 km. The lowest mid-layer level is at 35 m, and the vertical grid spacing is $\approx$80 m between the

Figure 1. (a) Zonal-mean sea surface temperature in fixed SST simulations, (b) solar insolation distribution, and (c) CO$_2$ mixing ratio and insolation in different simulations.
two lowest levels and gradually increases to 1500 m above 10 km. A sponge layer is included above 28 km to reduce the reflection of gravity waves near the model top.

Letting sea surface temperature (SST) be interactive in a near-global domain comes with a huge computational cost, so we prescribe the SST. The spatial pattern of the specified SST is based on modern Earth, but it is zonally uniform and hemispherically symmetric (Figure 1a). We neglect diurnal and seasonal cycles, and set the distribution of solar insolation to be zonally uniform but latitude dependent (Figure 1b). For simplicity, the concentrations of ozone are set to $10^{-7}$ g g$^{-1}$ for all layers, and there is no other trace gas. Besides, cartesian geometry is used in SAM.

We did seven experiments by changing CO$_2$ concentration and solar constant synchronously (black line in Figure 1c). These experiments are respectively labeled as 0.70S$_0$, 0.75S$_0$, 0.80S$_0$, 0.85S$_0$, 0.90S$_0$, 0.95S$_0$, and 1.00S$_0$. The choices of insolation and CO$_2$ concentration are based on Goldblatt et al. (2021), in which a coupled slab ocean was employed and the obtained global-mean surface temperatures are nearly the same although the equator-to-pole surface temperature gradients change somewhat. For each experiment, the last 30 days are used in the following analyses (Figure S1). The details of these simulations are shown in Text S1.

Hourly-mean cloud fraction, precipitation, and precipitable water in the modern Earth simulation (1.00S$_0$) show that the key characteristics, such as the Intertropical Convergence Zone (ITCZ) and mid-latitude baroclinic clouds, can be well simulated in the cloud-permitting framework (Figure S2). However, there are some less realistic features, which were also noted in Bretherton and Khairoutdinov (2015). For example, the differences between continental clouds and oceanic clouds are not simulated because of the employed aquaplanet mode, and neither seasonal cycle nor diurnal cycle is simulated because of the fixed solar radiation and SST. Moreover, as we set the boundaries at 60$^\circ$S(N), a lot of clouds are trapped near the boundary walls; these clouds are mostly advected from lower latitudes and trapped by the walls; if there were not these boundaries, clouds will further flow to higher latitudes and then precipitate to the surface.

Note that energy budgets at the top of the atmosphere (TOA) and the surface are not balanced because we prescribe SSTs. To decrease the energy imbalances of the TOA and surface, we re-do four experiments (0.70S$_0$, 0.80S$_0$, 0.90S$_0$, and 1.00S$_0$) with the solar constants increasing to respectively 0.85, 0.95, 1.05, and 1.15 S$_0$, and meanwhile the CO$_2$ concentrations are unchanged (Figure S3 and red line in Figure 1c). Besides, we also add two slab ocean runs (see Text S1, Figure S4). In these experiments, the energy imbalances at the TOA and surface decrease to be less than 3 W m$^{-2}$ (Figures S3 and S4).

3 Results

3.1 Cloud Feedback

When insolation increases and meanwhile CO$_2$ concentration decreases, global-mean low-level cloud fraction significantly increases, high-level cloud fraction slightly decreases, and middle-level cloud fraction is almost unchanged (Figure 2b). The water paths of clouds at different levels show the same trends as the cloud fractions (Figure 2c). Moreover, the same trends of cloud fractions and cloud water paths can be found in the energy-balanced fixed-SST simulations and also in the slab ocean simulations (Figure 3).

The increase of low-level clouds can reflect more insolation to space, leading to a larger planetary albedo (Figure 2a). For example, comparing the 0.80S$_0$ case and the 1.00S$_0$ case, the planetary albedo increases from 0.36 to 0.38, and the magnitude of shortwave cloud radiative effect increases from 45.1 to 64.0 W m$^{-2}$ (Figure 2d). The change of the shortwave cloud radiative effect is contributed by two parts, one is from the change of the solar radiation without any change of the cloud properties, and the other one is from the change of the cloud
Figure 2. Results under different solar radiations and corresponding CO₂ concentrations. (a) Global-mean planetary albedo in our SAM simulations (black), (b) cloud fractions of low-level (black), middle-level (red), and high-level clouds (blue), (c) same as (b) but for cloud water paths, (d) shortwave (blue) and longwave (red) cloud radiative effects, (e) inversion strength indexes: lower tropospheric stability (LTS; black), estimated inversion strength (EIS; red), and estimated cloud-top entrainment index (ECTEI; blue; see Text S3 in the Supporting Information online), and (f) surface latent (black) and sensible (red) heat fluxes. The low-level cloud fraction in (b) is defined as the fraction of grids with vertically-integrated cloud water path between surface and 700 hPa exceeding 0.02 kg m⁻² (Khairoutdinov & Randall, 2003). Likewise, the middle-level and high-level cloud fractions correspond to the vertically-integrated cloud water path of 700 to 400 hPa and above 400 hPa. In (a), the CAM4 simulation results of Goldblatt et al. (2021) (magenta) are included for comparisons.

properties without any change of solar radiation. The albedo of the 1.00S₀ case is 0.38, so there should be \( \approx 43.6 = \frac{1361.3}{3.87} \times (1.00 - 0.38) \times (1.0 - 0.8) \) W m⁻² less solar radiation absorbed when the solar constant is 80% of the modern Earth. Note that the denominator is 3.87 rather than 4.0; this is because the model uses Cartesian geometry rather than spherical geometry and meanwhile the polar regions are not simulated. When the 1.00S₀ case is chosen as the baseline, the shortwave cloud feedback is \( -45.1 - (-64.0) \times 0.8/1.0 = 6.1 \) W m⁻². In percentage, it is 14.0% \( (= 6.1/43.6) \). Another method for calculating the shortwave cloud feedback is shown in Text S2, and its value is 6.0 W m⁻². For a solar constant of 70% of the modern value, the strength of the shortwave cloud feedback is 10.3 W m⁻² or 15.7%. The magnitude of the shortwave cloud feedback is similar to Wolf and Toon (2013), in which it can contribute \( \approx 9.6 \) W m⁻² or 21% when the Sun is 20% dimmer (Figure 4, Charnay et al. (2020)).

In the CAM4 simulations of Goldblatt et al. (2021), the shortwave cloud radiative effects are \(-33.4\) and \(-52.5\) W m⁻², and the planetary albedos are 0.28 and 0.32, for the 0.80S₀ and 1.00S₀ cases, respectively (Figure S5). There should be 46.5 \( (= 1367.0/4.0 \times (1.00 - 0.32) \times (1.0 - 0.8)) \) W m⁻² less solar radiation absorbed in the 0.80S₀ case. The shortwave cloud feedback equals \(-33.4 - (-52.5) \times 0.8/1.0 = 8.6 \) W m⁻², or 18.5% \( (= 8.6/46.5) \) in percentage. For 0.70S₀, the strength of the shortwave cloud feedback is 10.7 W m⁻² or 15.4% (Figure 4). Note that the estimated magnitudes of the shortwave cloud feedback in Goldblatt et al.
Figure 3. Same as Figure 2 but for the experiments in which the TOA and surface are nearly energy-balanced. The dot markers represent fixed-SST experiments, and the diamond markers represent slab ocean experiments.

(2021) are higher than those shown here, because the effect of varying solar constant on the value of shortwave cloud radiative effect without the changes of clouds was not excluded.

Moreover, the temperature patterns and the trends of cloud-related variables are in line with the CAM4 simulations in Goldblatt et al. (2021) (Figures S5 and S6). However, there are still some differences. For example, the change of high-level cloud fraction in CAM4 is larger than that in SAM (Figures 2b and S5b). Besides, cloud water paths in our simulations are lower than that in CAM4. The high-level cloud water path in SAM decreases from 0.70.S_0 to 1.00.S_0, but CAM4 shows an opposite trend (Figures 2c and S5c). The planetary albedo here is larger (Figure 2a), and the magnitude of the shortwave cloud radiative effect is also larger (Figures 2d and S5d).

Changing insolation and CO\textsubscript{2} concentration synchronously can also lead to changes in atmospheric temperature, atmospheric circulation, and cloud pattern. When we increase the insolation and meanwhile decrease the CO\textsubscript{2} concentration, the lower atmosphere becomes cooler but the upper atmosphere becomes warmer (Figure 5d), the subsidence branch of the Hadley circulation becomes weaker (Figure 5e), and the low-level clouds increase at almost every latitude (Figure 5f). These patterns are similar to that in CAM4 shown in Goldblatt et al. (2021). But, the patterns in our simulations are more symmetric between the northern and southern hemispheres due to the aquaplanet we use.

3.2 Mechanisms for Low-level Cloud Feedback

Increasing insolation but meanwhile decreasing CO\textsubscript{2} concentration leads to the change of radiative heating rate, which can cause the change of clouds through several processes. Importantly, the lower troposphere experiences radiative cooling, but the upper troposphere experiences radiative warming (Figures 5h and 5i), and subsequently the lower troposphere becomes cooler but the upper troposphere becomes warmer (Figures 5d and 5g). To examine what factors dominate the changes in the radiative heating rates, we use an off-line radiative transfer model, RRTMG (Mlawer et al., 1997; Clough et al., 2005), to calculate the clear-sky longwave and shortwave heating rates under different conditions (Figure S7). We find that
decreasing CO₂ concentration leads to lower-tropospheric longwave radiative cooling, but the upper-tropospheric longwave heating rate doesn’t change significantly (Figure S7b). In our simulations, the changes of water vapor concentration are small, as well as its radiative effects (Figures S7a, d, and e). For shortwave radiative heating rate, the effect of changing solar constant is greater than that of changing CO₂ or H₂O. When the solar radiation is increased, the clear-sky shortwave heating rate increases at all levels (Figure S7f). In short, from 0.70S₀ to 1.00S₀, the enhanced lower-tropospheric radiative cooling dominated by less CO₂ and upper-tropospheric radiative warming dominated by more insolation make the lower troposphere and the upper troposphere become cooler and warmer, respectively.

Why there are more low-level clouds from 0.70S₀ to 1.00S₀? Following the analyzing methods of previous work (e.g., Bretherton et al. (2013); Bretherton (2015); Qu et al. (2015); Klein et al. (2017); McCoy et al. (2017); Mieslinger et al. (2019)), we find four different processes, including boundary layer inversion, moisture gradient across the inversion, large-scale atmospheric circulation, and sensible heat flux and Bowen ratio. The details are as follows:

**Boundary layer inversion.** The lower troposphere becomes cooler while the upper troposphere becomes warmer from 0.70S₀ to 1.00S₀, so that the strength of the boundary layer inversion increases (Figure 5g). These can also be seen from stability indicators, such as lower tropospheric stability (LTS), estimated inversion strength (EIS), and estimated cloud-top entrainment index (ECTEI) (Slingo (1987); Wood and Bretherton (2006); Kawai et al. (2017); Text S3). All these three indicators show increased inversion strengths at all latitudes (Figures 2e and S8b-d). The stronger inversions can reduce dry air entrainment between the free troposphere and boundary layer and permit more boundary layer clouds (Klein & Hartmann, 1993; Wood & Bretherton, 2006).

**Moisture gradient across the inversion.** The moisture gradient (δq) across the inversion is characterized by the difference in specific humidity between the surface and 700 hPa. Both air temperature and relative humidity (RH) at 700 hPa increase from 0.70S₀ to 1.00S₀, but near-surface temperature and RH nearly do not change (Figures 5g and 5j), which leads to δq gradually decreases (Figure S8e). The smaller specific humidity gradient can lead to less dry entrainment across the inversion, which promotes low-level cloud maintenance (Van der Dussen et al., 2015; Scott et al., 2020; Zhu & Poulsen, 2020).

**Large-scale atmospheric circulation.** The subsidence branch of the Hadley circulation becomes weaker from 0.70S₀ to 1.00S₀ (Figure 5e). The more stable free troposphere (Fig-
Figure 5. Upper two rows: zonal-mean spatial patterns in the 1.00S0 experiment (a–c) and the differences between 1.00S0 and 0.70S0 (d–f). From left to right, the variables are: atmospheric temperature, meridional mass streamfunction, and cloud water. Lower two rows: global-mean profiles of (g) atmosphere temperature, (h)–(i) clear-sky and all-sky radiative heating rate (shortwave plus longwave), (j) relative humidity, (k) cloud fraction, and (l) cloud water. Note that the cloud fraction in (k) is defined as the fraction of grids with cloud water larger than 0.01 g kg$^{-1}$ (Wyant et al., 2009).

Sensible heat flux and Bowen ratio. Surface sensible heat flux slightly increases and latent heat flux keeps almost constant from 0.70S0 to 1.00S0 (Figures 2f and S8f). The larger temperature gradient near the surface layer (Figure 5d) can contribute to larger sensible heat flux, but water vapor concentration near the surface layer which influences the latent heat flux nearly doesn’t change (Figure S7a). A larger sensible heat flux can increase the buoyancy and supply more heat from the surface to the boundary layer. Meanwhile, Bowen ratio, which is defined as the ratio of surface sensible heat flux to latent heat flux, increases from 0.70S0 to 1.00S0 (Figure S9a). It can increase cloud water flux at cloud base, by influencing the efficiency of the heat cycle. Larger Bowen ratio corresponds to larger ratio of mechanical work to generate convection, supporting more low-level clouds (Figure S9b, Sakradzija and Hohenegger (2017); Mieslinger et al. (2019)).
The mechanisms addressed above can also be found in the four added fixed SST experiments and in the two slab ocean experiments (Figures S10 and S11). This implies that our main conclusions are likely robust.

4 Conclusions and Discussions

In this study, we use a near-global cloud-permitting model to re-investigate the cloud feedback on Earth’s long-term climate. We apply the grid spacing of 10 km × 14 km over a domain of half of the Earth’s equatorial circumference in longitude and 60°S to 60°N in latitude. When solar radiation increases and meanwhile CO₂ concentration decreases, the change of low-level clouds dominates. The underlying mechanisms are analyzed from four different processes (summarized as Figure S12): boundary layer inversion, moisture gradient across the inversion, large-scale atmospheric circulation, and surface heat flux and Bowen ratio. All these four factors contribute to the increase of low-level clouds.

The shortwave cloud feedback in our simulations can contribute ≈6 W m⁻² or 14% of the energy required to offset a fainter Sun that is 20% dimmer than today. For a 30% dimmer Sun, it can contribute ≈10 W m⁻² or 16%. This shortwave cloud feedback is likely to stabilize Earth’s past or future long-term climate, and its impact is limited but unignorable.

Not only shortwave cloud radiative effect but also longwave cloud radiative effect changes in the experiments, but the magnitude of the former is much larger than the latter (Figure 2d). This is due to the fact that the response of low-level cloud water path is greater than that of the high-level cloud water path (Figures 2c, 5l, and S1d–f). Moreover, the concentration of CO₂ can also influence the strength of the longwave cloud radiative effect. When CO₂ concentration is higher, the longwave absorption spectrum is more saturated, which could lead to a weakening of the longwave cloud radiative effect even if there is no change in cloud properties (Wolf & Toon, 2013). This is likely the reason why the cloud fraction and cloud water path of high-level clouds slightly decrease but the longwave cloud radiative effect slightly increases (Figures 2b–d). Unfortunately, the overlaps between CO₂ and cloud absorption spectra are not in model outputs. If these overlaps were considered, the longwave cloud feedback would be smaller than the change of the longwave cloud radiative effect shown in Figure 2d.

There is no continent distribution in the simulations, which would influence the location of ITCZ, the strength of the Hadley circulation, the surface heat fluxes, and other aspects. Future work should examine the role of land-sea distribution as well as use a dynamical ocean to test the effect of ocean dynamics. If continents are taken into account, other additional factors can affect the low-level cloud amount, such as zonal contrast of SST and horizontal energy transport (Norris & Leovy, 1994; Bretherton et al., 2013; Myers & Norris, 2015).

Accuracy resolving the clouds needs a cloud-resolving model with a grid spacing of 4 km or less (Weisman & Klemp, 1997). Therefore, using the horizontal resolution of ≈10 km with no cumulus parameterization here may affect the strength of the cloud feedback. Besides, the conclusions of this study are based on a one-moment microphysics scheme, and different microphysics schemes may lead to different results, which needs to be examined in future work.

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The simulation data in this study are archived at https://doi.org/10.5281/zenodo.6592041. The output data of CAM4 used for comparisons are included in Goldblatt et al. (2021), which can be accessed from https://doi.org/10.20383/101.0308.

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