Controls on the Activation and Strength of a High-Latitude Convective Cloud Feedback

DORIAN S. ABBOT

School of Engineering and Applied Sciences, Harvard University, Cambridge, Massachusetts

ELI TZIPERMAN

School of Engineering and Applied Sciences, and Department of Earth and Planetary Science, Harvard University, Cambridge, Massachusetts

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ABSTRACT

Previous work has shown that a convective cloud feedback can greatly increase high-latitude surface temperature upon the removal of sea ice and can keep sea ice from forming throughout polar night. This feedback activates at increased greenhouse gas concentrations. It may help to explain the warm ''equable climates'' of the late Cretaceous and early Paleogene eras (~100 to ~35 million years ago) and may be relevant for future climate under global warming. Here, the factors that determine the critical threshold CO_2 concentration at which this feedback is active and the magnitude of the warming caused by the feedback are analyzed using both a highly idealized model and NCAR's single-column atmospheric model (SCAM) run under Arctic-like conditions. The critical CO_2 is particularly important because it helps to establish the relevance of the feedback for past and future climates.

Both models agree that increased heat flux into the high latitudes at low altitudes generally decreases the critical CO_2. Increases in oceanic heat transport and in solar radiation absorbed during the summer should cause a sharp decrease in the critical CO_2, but the effect of increases in atmospheric heat transport depends on its vertical distribution. It is furthermore found (i) that if the onset of convection produces more clouds and moisture, the critical CO_2 should decrease, and the maximum temperature increase caused by the convective cloud feedback should increase and (ii) that reducing the depth of convection reduces the critical CO_2 but has little effect on the maximum temperature increase caused by the convective cloud feedback. These results should help with interpretation of the strength and onset of the convective cloud feedback as found, for example, in Intergovernmental Panel on Climate Change (IPCC) coupled ocean–atmosphere models with different cloud and convection schemes.

1. Introduction

Cloud feedbacks represent the most important source of uncertainty in the climate system (Cess et al. 1990, 1996; Baker 1997; Murphy et al. 2004; Stainforth et al. 2005; Soden and Held 2006). This motivates the idea that cloud feedbacks might play an important role in explaining past ''equable climates'' and makes understanding clouds important for understanding future climate under increased greenhouse gas levels. Equable climates, which prevailed during the late Cretaceous and early Paleogene (~100 to ~35 million years ago), were characterized by warm high latitudes (e.g., Zachos et al. 2001; Sluijs et al. 2006), particularly during the winter and over continents (e.g., Greenwood and Wing 1995), and tropical temperatures only somewhat higher than modern (e.g., Pearson et al. 2001; Norris et al. 2002; Roche et al. 2006; Tripati et al. 2003). Various mechanisms have been proposed to explain either the relatively cool tropical temperatures or relatively warm polar temperatures, including increased ocean heat transport due to ocean mixing by increased hurricane activity (Emanuel 2002; Korty et al. 2008), the Hadley cell extending nearly to the pole (Farrell 1990), and high-latitude longwave heating due to thick polar stratospheric clouds (Sloan et al. 1992; Sloan and Pollard 1998; Peters and Sloan 2000; Kirk-Davidoff et al. 2002).

Abbot and Tziperman (2008a) proposed a positive feedback on high-latitude temperatures that results

Corresponding author address: Dorian Abbot, Harvard University, 24 Oxford St., Cambridge, MA 02138.
E-mail: abbot@fas.harvard.edu

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from the onset of convective clouds. A related suggestion was also briefly made by Sloan et al. (1999) and Huber and Sloan (1999). In this proposed feedback, an initial warming leads to destabilization of the high-latitude atmosphere to convection, causing convection, which results in convective clouds and increased atmospheric moisture, both of which trap outgoing longwave radiation and lead to further warming.

Over ocean, this feedback should occur preferentially during winter (Abbot and Tziperman 2008b, hereafter AT08b; Abbot et al. 2008, manuscript submitted to *J. Climate,* hereafter AWT) because during summer marine boundary layer clouds block low-level atmospheric solar absorption, so that solar absorption occurs preferentially in the midtroposphere and stabilizes the lower atmosphere.

The convective cloud feedback as outlined in AT08b and AWT is intimately tied to sea ice, which insulates the ocean and prevents convection when it is present, whereas the feedback prevents the formation of sea ice when there is none (AT08b; AWT). Abbot and Tziperman (2008a), however, found that the convective cloud feedback can operate based on atmospheric processes alone. This distinction is important because it underscores the possibility that the convective cloud feedback could lead to further warming even after the complete removal of sea ice, and we will return to it in the discussion (section 4).

The convective cloud feedback allows for multiple equilibria: one solution that is convecting and is warm and another solution that is not convecting and is cold. The purpose of this paper is to determine what parameters control the lowest (critical) CO_2 value at which the warm state can exist and the temperature difference between the two states. The critical CO_2 is important because it determines whether the convective cloud feedback could have been active during periods of equable climate and whether it could be active in a future climate under global warming. The temperature difference between the two states is important because it represents the strength of the convective cloud feedback.

In section 2 we develop a simple two-level atmosphere–surface model that encapsulates the most basic physics that can describe the atmosphere-only convective cloud feedback. We use this model to qualitatively determine the way in which various parameters affect the onset of the feedback and its strength. This analysis should aid interpretation of the convective cloud feedback in more complex models, such as the Intergovernmental Panel on Climate Change (IPCC) coupled GCMs, in which the convective cloud feedback has been shown to be active (AWT).

In section 3 we extend this analysis using the National Center for Atmospheric Research (NCAR) single-column atmospheric model (SCAM). SCAM contains the full cloud, convection, and radiation parameterizations of the NCAR community atmosphere model (CAM), but heat transports into it and velocities acting on it must be prescribed. We show that SCAM's behavior is consistent with that of the two-level model and that the lessons from the simpler model can be used to understand the more complete SCAM.

2. Two-level model

a. Developing the model

In this section we construct a simple two-level model of the atmosphere in which we attempt to capture the simplest system in which the convective cloud feedback can function. Based on previous work (Abbot and Tziperman 2008a; AT08b; AWT), we expect the convective cloud feedback to be active at high latitudes (roughly poleward of 60°) during winter, and we will make assumptions accordingly throughout this section. In this model, the top level represents the free troposphere (200–900 hPa; henceforth the atmosphere) and the lower level (henceforth the surface) represents the combined boundary layer (900–1000 hPa) and surface—for example, a mixed-layer ocean (top 50 m). In effect, we assume that turbulent fluxes tie the surface to the boundary layer so tightly that they behave as one. Energy balance for this model can be written as

$$C_s \frac{dT_\mathrm{s}}{dt} = F_\mathrm{s} - F_\mathrm{c} + \varepsilon\sigma T_\mathrm{a}^4 - \sigma T_\mathrm{s}^4, \tag{1}$$

$$C_a \frac{dT_\mathrm{a}}{dt} = F_\mathrm{a} + F_\mathrm{c} + \varepsilon\sigma\left(T_\mathrm{s}^4 - 2T_\mathrm{a}^4\right). \tag{2}$$

Here, C_s and C_a are the total heat capacities of the surface and atmospheric columns (standard heat capacity multiplied by total column mass), respectively; T_s and T_a are the surface and atmospheric temperatures, respectively; F_s is the heat flux into the surface and boundary layer from solar radiation and by horizontal heat transport, which can be written $F_s = F_o + S(1-\alpha) + F_a^{\mathrm{bl}}$ (where F_o is the meridional ocean heat transport convergence, S is the solar heat flux, α is the albedo, and F_a^{bl} is the atmospheric transport convergence into the boundary layer); F_a is the meridional heat transport convergence into the atmospheric layer; F_c is the convective heat flux from the boundary layer to the free troposphere; ε is the emissivity of the free troposphere; and σ is the Stefan–Boltzmann constant.

The convective heat flux F_c and the free tropospheric emissivity ε depend on whether or not there is convection, which in turn depends on the moist stability. We determine moist stability by comparing the surface moist static energy (M_s)

$$M_s = C_p T_s + L r_s,$$

with the atmospheric saturation moist static energy (M_a^*)

$$M_a^* = C_p T_a + L r_a^* + g z_a,$$

where C_p is the specific heat of air at constant pressure, L is the latent heat of evaporation, r_s is the surface specific humidity, r_a^* is the free tropospheric saturation specific humidity, g is the earth's gravitational constant, and z_a is the height of the atmospheric layer (we specify the pressure of this layer, P_a, and calculate z_a using a scale height of 8 km). We calculate r_s by assuming a constant boundary layer relative humidity, RH. If $M_s < M_a^*$, the model is stable to moist convection and there is no convection; consequently, we set the convective heat flux to zero ($F_c = 0$) and we set the emissivity to a background value [i.e., $\varepsilon = \varepsilon_0$, where ε_0 represents the free tropospheric emissivity in the absence of convection, which should be roughly linear in $\log(CO_2)$ (Sasamori 1968)]. Otherwise, we choose F_c to satisfy the moist stability criticality ($M_s = M_a^*$; see below) and set $\varepsilon = \varepsilon_0 + \Delta\varepsilon$.

Our use of F_c to satisfy the moist stability criticality represents the basic physics of adjustment to a neutrally buoyant profile in a moist atmosphere. Our assumption that the atmospheric emissivity increases from a background emissivity (ε_0) when there is no convection by some offset ($\Delta\varepsilon$) upon the onset of convection represents the advent of radiatively thick convective clouds and the increase in high-altitude moisture; this is how the convective cloud feedback manifests itself in this model. Convective clouds could also affect the model albedo and through it F_s; however, based on previous SCAM and GCM investigations of the seasonality of the convective cloud feedback (AT08b; AWT), we will focus on high-latitude winters when the incoming solar radiation S is small or zero, making such an effect irrelevant.

We can solve for the steady-state solutions of the model by setting the time tendencies of (1) and (2) to zero. First consider the nonconvecting state, in which $F_c = 0$ and $\varepsilon = \varepsilon_0$. We have

$$0 = F_s + \varepsilon_0 \sigma T_{a1}^4 - \sigma T_{s1}^4, \tag{3}$$

$$0 = F_a + \varepsilon_0 \sigma \left(T_{s1}^4 - 2T_{a1}^4\right), \tag{4}$$

where the subscript 1 signifies that this is the nonconvecting solution. We can solve (3) and (4) for the nonconvecting surface and atmospheric temperatures:

$$T_{s1} = \left[\frac{2F_s + F_a}{(2 - \varepsilon_0)\sigma}\right]^{\frac{1}{4}}, \tag{5}$$

$$T_{a1} = \left[\frac{\varepsilon_0 F_s + F_a}{(2 - \varepsilon_0)\varepsilon_0 \sigma}\right]^{\frac{1}{4}}. \tag{6}$$

This solution is valid so long as $M_{s1} \leq M_{a1}^*$.

When the model is convecting, we obtain the equations

$$0 = F_s - F_c + \tilde{\varepsilon}\sigma T_{a2}^4 - \sigma T_{s2}^4, \tag{7}$$

$$0 = F_a + F_c + \tilde{\varepsilon}\sigma\left(T_{s2}^4 - 2T_{a2}^4\right) \quad \text{and} \tag{8}$$

$$C_p T_{s2} + L r_{s2} = C_p T_{a2} + L r_{a2}^* + g z_a, \tag{9}$$

where $\tilde{\varepsilon} \equiv \varepsilon_0 + \Delta\varepsilon$ and the subscript 2 signifies the convecting solution. Equation (9) represents the moist convective criticality $\left(M_{s2} = M_{a2}^*\right)$. Equations (7)–(9) can be solved for T_{s2}, T_{a2}, and F_c. This solution is valid so long as $F_c > 0$.

We plot the convecting and nonconvecting solutions of the two-level model as a function of ε_0 in Fig. 1. Here we choose $F_a = 100$ W m^{-2}, which is a reasonable high-latitude value (Trenberth and Stepaniak 2003), and $F_s = 250$ W m^{-2}, which we take, for the most part, to represent heat absorbed and stored by the ocean during the summer and released back into the atmosphere during the winter. The simplicity of the model, with only one layer to represent the atmosphere, requires us to choose an unrealistically high F_s (250 W m^{-2}) to obtain the convecting solution; F_s takes much smaller values when we use the more realistic SCAM model (section 3). We take $\Delta\varepsilon = 0.3$ and $P_a = 600$ hPa, representing medium-height convection that produces optically thick clouds.

The nonconvecting solution exists at low values of the clear-sky emissivity, ε_0, but not at higher values (Fig. 1a, solid black line). The convecting solution exists at high ε_0 but disappears for ε_0 below some critical ε_0 which we call ε_c; ε_c is the two-level model analog of the logarithm of the critical CO_2. Below ε_c, the two-level model is no longer warm enough to consistently sustain convection [i.e., (7)–(9) yield $F_c < 0$]. Because the free tropospheric emissivity is increased by $\Delta\varepsilon$ because of the appearance of convective clouds and increased moisture in the convecting solution, the convecting solution has a higher surface temperature than the nonconvecting solution at all ε_0.

The vertical temperature profile of the convecting solution follows the moist lapse rate, whereas the lapse rate of the nonconvecting solution is determined radiatively (Fig. 1d). This causes the nonconvecting surface

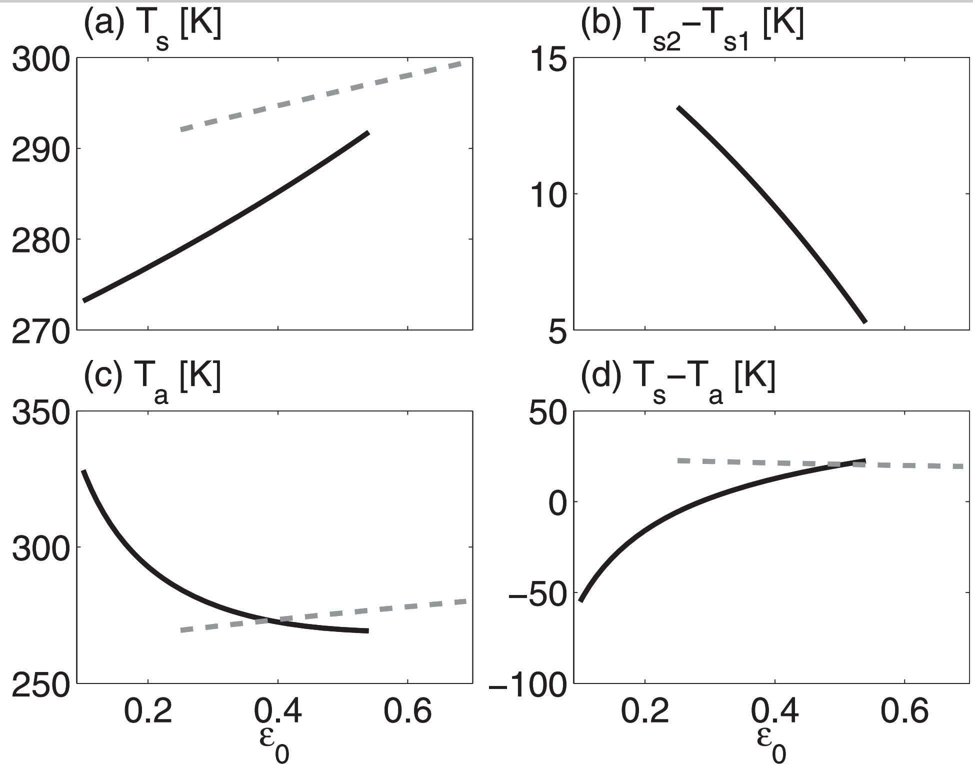

FIG. 1. Solution to the two-level model as a function of ε_0, which is a proxy for $\log(CO_2)$ concentration. (a) Nonconvecting (T_{s1}; black solid line) and convecting (T_{s2}; gray dashed line) solution surface temperature; (b) difference between the surface temperature of the convecting and nonconvecting solutions ($T_{s2} - T_{s1}$); (c) nonconvecting (T_{a1}; black solid line) and convecting (T_{a2}; gray dashed line) solution atmospheric temperature; (d) nonconvecting ($T_{s1} - T_{a1}$; black solid line) and convecting ($T_{s2} - T_{a2}$; gray dashed line) solution lapse rate. Model parameters are $F_s = 250$ W m^{-2}, $F_a = 100$ W m^{-2}, $\Delta\varepsilon = 0.3$, $P_a = 600$ hPa, and RH = 0.85.

temperature to increase much faster with ε_0 than the convecting surface temperature does ($dT_{s1}/d\varepsilon_0 > dT_{s2}/d\varepsilon_0$; Fig. 1a). Consequently, the maximum difference in surface temperature between the convecting and nonconvecting solutions as a function of ε_0, $(T_{s2} - T_{s1})_{\max}$ occurs at the minimum value of ε_0 at which convection is possible ($\varepsilon_0 = \varepsilon_c$).

There is a singularity in the nonconvecting atmospheric temperature (6) as ε_0 approaches zero if F_a, the atmospheric heat transport (AHT), is nonzero. This leads to a negative lapse rate at low ε_0 (Fig. 1d), which to some extent could be a realistic representation of a high-latitude winter inversion; however, the extreme increase of T_{a1} as ε_0 goes to zero is due to the simplicity of the model and is not realistic. In any case, this does not affect the surface temperature (5), which is the quantity in which we are primarily interested.

b. *Using the model to understand the convective cloud feedback*

We now focus on how the model parameters affect ε_c, the lowest ε_0 at which the convecting solution can exist, and $(T_{s2} - T_{s1})_{\max}$, the maximum difference in surface temperature between the convecting and nonconvecting solutions as a function of ε_0. The critical value ε_c is important for two reasons. First, because ε_0 can be thought of as roughly representing $\log(CO_2)$ in this model, ε_c is related to the lowest CO_2 concentration at which the convecting solution can exist, which is critical to whether or not the convecting solution could be realized during an equable climate or future climate with increased greenhouse gases. Second, because $(T_{s2} - T_{s1})_{\max}$ occurs at $\varepsilon_0 = \varepsilon_c$ and $dT_{s1}/d\varepsilon_0 > dT_{s2}/d\varepsilon_0$, decreasing ε_c tends to increase $(T_{s2} - T_{s1})_{\max}$, which is itself important because $(T_{s2} - T_{s1})_{\max}$ represents the strength of the convective cloud feedback. Stated again, the lower the critical CO_2, the larger the maximum temperature increase caused by the convective cloud feedback, all other things being equal.

In Figs. 2 and 3, we show how $(T_{s2} - T_{s1})_{\max}$ and ε_c change as we vary $\Delta\varepsilon$, F_s, F_a, and P_a, which are the important independent model parameters. Here, $\Delta\varepsilon$ represents the increase in optical thickness of the atmosphere associated with clouds and water vapor upon