

1 **Critical role of vertical radiative cooling contrast in**
2 **triggering episodic deluges in small-domain hothouse**
3 **climates**

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8 **Key Points:**

- 9 • Lower-tropospheric radiative heating is unnecessary for the occurrence of episodic
10 deluges
11 • The strong vertical gradient of radiative cooling is a key factor in triggering episodic
12 deluges
13 • The occurrence of episodic deluges needs strong convective inhibition (CIN)

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Abstract

Seeley and Wordsworth (2021) showed that in small-domain cloud-resolving simulations the pattern of precipitation transforms in extremely hot climates (≥ 320 K) from quasi-steady to organized episodic deluges, with outbursts of heavy rain alternating with several dry days. They proposed a mechanism for this transition involving increased water vapor absorption of solar radiation leading to net lower-tropospheric radiative heating. This heating inhibits lower-tropospheric convection and decouples the boundary layer from the upper troposphere during the dry phase, allowing lower-tropospheric moist static energy to build until it discharges, resulting in a deluge. We perform cloud-resolving simulations in polar night and show that the same transition occurs, implying that some revision of their mechanism is necessary. We show that episodic deluges can occur even if the lower-tropospheric radiative heating rate is negative, as long as the magnitude of the upper-tropospheric radiative cooling is about twice as large. We find that in the episodic deluge regime the mean precipitation can be inferred from the atmospheric column energy budget and the period can be predicted from the time for radiation and reevaporation to cool the lower atmosphere.

Plain Language Summary

Precipitation plays an important role in Earth’s climate and habitability, and also influences important weathering processes such as carbonate-silicate cycle. In the distant future, Earth may experience a very hot and wet “hothouse” climate, just like that in the past Archean. Modelling results show that in a hothouse climate, precipitation transforms into an “episodic deluge” pattern, with outbursts of heavy rain alternating with several dry days. In this study, we find that positive lower-tropospheric heating is not the necessary cause for episodic deluges. Instead, vertical radiative cooling contrast is critical in triggering the episodic deluges in small-domain hothouse climates. We also try to understand the underlying mechanism of episodic deluges through CIN and CAPE analyses.

1 Introduction

Earth might have experienced an extremely warm and wet climate, a “hothouse,” in the Archean (Sleep, 2010; Charnay et al., 2017; Steffen et al., 2018), or in the aftermath of a snowball Earth event (Higgins & Schrag, 2003; Hir et al., 2009; R. Pierrehumbert et al., 2011; Hoffman et al., 2017; Yang et al., 2017), and may experience a hothouse climate again in the distant future (Ingersoll, 1969; Kasting et al., 1984; Kidder & Worsley, 2012; Goldblatt et al., 2013; Leconte et al., 2013; Ramirez et al., 2014). Previous work mainly used general circulation models (GCMs), and concluded that there could be a lower-tropospheric temperature inversion and significant increase in upper-tropospheric cloud cover in hothouse climates (Wordsworth & Pierrehumbert, 2013; E. T. Wolf & Toon, 2015; Popp et al., 2016; E. Wolf et al., 2018). Seeley and Wordsworth (2021) moved beyond GCMs by using convective scale cloud-resolving models and found that the precipitation would organize in time into an “episodic deluge” pattern. In this regime, the majority of the grid points have an outburst of heavy rain at the same time (Figure 4d in Seeley and Wordsworth (2021)), followed by several dry days. Seeley and Wordsworth (2021) investigated this in three different cloud-resolving models, Das Atmosphärische Modell (DAM) (Romps, 2008), the System for Atmospheric Modeling (SAM) (Khairoutdinov & Randall, 2003), and the Cloud Model 1 (CM1) (Bryan and Fritsch, 2002), and modified the radiative transfer scheme of DAM in order to be more accurate in hot climates. They verified that the onset of episodic deluges does not depend on the specific model choice. Most of their experiments are in a small domain of $72 \text{ km} \times 72 \text{ km}$, and the episodic deluge is a synchronized behavior between the grid points.

Seeley and Wordsworth (2021) argued that episodic deluges are mainly caused by lower-tropospheric radiative heating: When the climate is warm enough, more lower-tropospheric water vapor increases shortwave absorption, and results in a net positive heating rate. Fig-

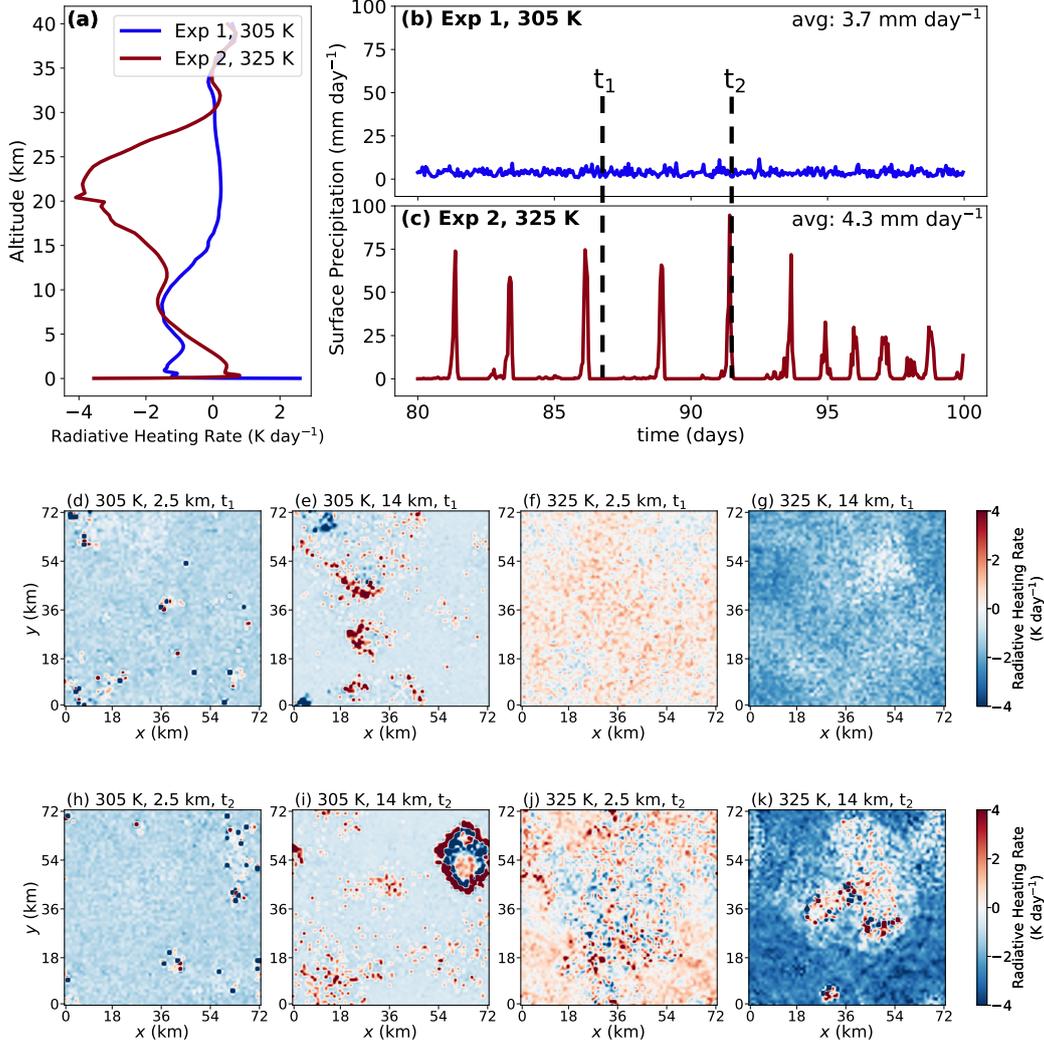


Figure 1. Reproduction of modelling results in Seeley and Wordsworth (2021) using SAM. Panel (a) shows time-averaged radiative heating rate. Panels (b) and (c) show the precipitation pattern in two simulations with the surface temperature fixed at 305 K (Exp 1) and 325 K (Exp 2), respectively. Panels (d)–(g) show the horizontal distribution of the radiative heating rate (longwave plus shortwave) for both cases during a dry spell (t_1). Panels (h)–(k) show the horizontal distributions during a convection (t_2). Each case shows slices at 2.5 km and 14 km. For time variations, see the video version in Supporting Information. Both cases run for 100 days.

64 ure 1 shows our reproduction of the results of Seeley and Wordsworth (2021) using the
 65 cloud-resolving model SAM. When the surface temperature is 305 K (Exp 1), there is net
 66 radiative cooling in the lower troposphere (blue line in Figure 1a), and the precipitation
 67 pattern is quasi-steady (Figure 1b). When the surface temperature is 325 K (Exp 2),
 68 the lower-tropospheric radiative heating rate becomes positive (red line in Figure 1a), and
 69 episodic deluges occur. The full mechanism proposed by Seeley and Wordsworth (2021) to
 70 explain episodic deluges, starting in a dry phase, is as follows: lower-tropospheric radiative
 71 heating inhibits convection in the lower troposphere. Strong radiative cooling in the upper
 72 troposphere leads to condensation and elevated precipitation above an “inhibition layer”.

73 Most droplets of upper-tropospheric precipitation reevaporate in the relatively warm inhi-
 74 bition layer, so very little (or zero) precipitation makes it to the surface. As time goes by,
 75 reevaporation of precipitation cool down the inhibition layer to the point that inhibition is
 76 broken, triggering strong convection. After the convection, inhibition starts again, and the
 77 cycle continues.

78 The heating rate profiles undergo three major changes as the surface temperature in-
 79 creases to 325 K. First, the lower troposphere shifts from cooling to heating. Second, the
 80 vertical gradient of the heating rate profile increases (Figure 1a, see also Figure 2b in Seeley
 81 and Wordsworth (2021)). When the surface temperature is 305 K, the radiative heating rate
 82 is about -1.5 K day^{-1} from the near surface layer to about 10 km, then smoothly transits
 83 to 0 K day^{-1} in the stratosphere (blue line in Figure 1a). When the surface temperature is
 84 325 K, the lower-tropospheric radiative heating rate is about 0.5 K day^{-1} at about 2 km,
 85 but the upper-tropospheric radiative heating rate is about -4 K day^{-1} at 20 km (the red
 86 line in Figure 1a). The increased surface temperature leads to a warmer upper troposphere
 87 and higher water vapor concentration. Both factors intensify longwave cooling in the upper
 88 troposphere. Third, during the dry period, the horizontal distribution of radiative heating
 89 is much more homogenous for the 325 K case, both at the lower troposphere (Figure 1f)
 90 and in the upper troposphere (Figure 1g). Note that the heating rates for both cases are
 91 fairly heterogeneous horizontally during convection (Figure 1h–k). For the video version of
 92 Figure 1, see Supporting Information Video S1.

93 Seeley and Wordsworth (2021) pointed to lower-tropospheric radiative heating as the
 94 primary factor leading to episodic deluges. In this paper, we suggest instead that the vertical
 95 gradient of radiative cooling is a more important factor for the onset of episodic deluges.
 96 This point is emphasized by the fact that episodic deluges occur even when the radiative
 97 heating rate in the lower troposphere is negative (Section 2). More specifically, episodic
 98 deluges require that the magnitude of radiative cooling in the upper-troposphere be about
 99 twice that in the lower troposphere (Section 3.4). The episodic deluge regime the mean
 100 precipitation can be inferred from the atmospheric column energy budget, and the period
 101 can be predicted from the time for radiation and reevaporation to cool the lower troposphere.

102 2 Episodic deluges during polar night

103 All the experiments in this study use version 6.11.6 of SAM (Khairoutdinov & Ran-
 104 dall, 2003; Khairoutdinov & Emanuel, 2018), one of the three models used in Seeley and
 105 Wordsworth (2021). The horizontal resolution in each experiment is 2 km with 72 grid
 106 points in each direction. The vertical resolution is 144 grid points within 64 km. CO_2 is set
 107 to 400 ppmv and the experiments do not contain ozone. The time step is 10 seconds, and
 108 the output statistics are hourly averages.

109 We find episodic deluges in polar night simulations, demonstrating that shortwave heat-
 110 ing is not necessary for episodic deluges (Figure 2). We find polar night episodic deluges
 111 both with a fixed sea surface temperature of 330 K (Exp 3) and with a two-meter-deep slab
 112 ocean and a positive ocean heat flux of 230 W m^{-2} (Exp 4), which produces a similar sea
 113 surface temperature. In both cases there is a deluge, or convective, period that lasts several
 114 hours and has a peak precipitation of about 80 mm day^{-1} , ten times the average rate. This
 115 is followed by a dry, inhibition period that lasts several days during which the precipitation
 116 rate is usually below 2 mm day^{-1} .

117 So far, the polar night simulations indicate that shortwave heating is not a necessary
 118 condition for episodic deluges, but we still cannot rule out the necessity of lower-tropospheric
 119 radiative heating, as the heating rate at about 0.5 km is slightly positive (Figures 2a &
 120 b). Random process, transient temperature inversion, or some unknown processes may
 121 cause this longwave heating. To exclude the influence of this longwave heating, we add an
 122 additional experiment, Exp 5, to smooth the near surface layer of the heating rate profile in

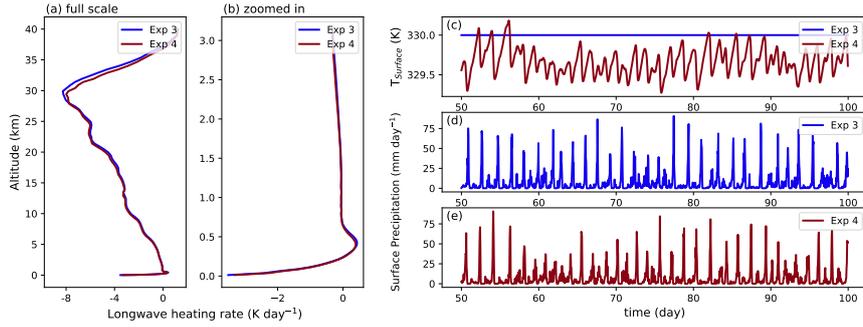


Figure 2. Full scale (a) and zoomed in (b) radiative heating rate profiles in polar night hothouse climate. One experiment fixes the sea surface temperature (SST) at 330 K (Exp 3). The other includes an ocean heat import of 230 W m^{-2} (Exp 4), in order to maintain the SST at around 330 K (c). Episodic deluges occur in both experiments (d and e).

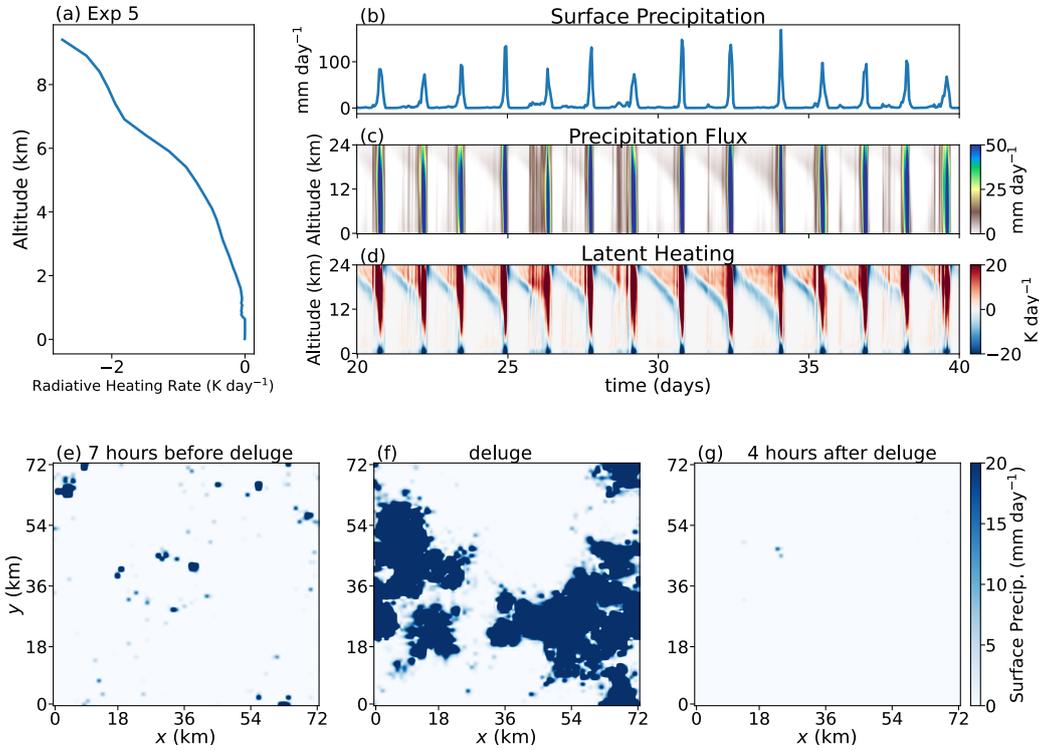


Figure 3. Prescribed radiative heating rate with lower-tropospheric radiative heating removed (a) and simulated precipitation (b) during polar night. Panels (c) and (d) show the altitude-time plot of precipitation flux and latent heating. Panels (e)–(g) show snapshots of surface precipitation at 7 hours before a deluge, during the deluge, and 4 hours after the deluge, respectively. Episodic deluges still exist even if there is no lower-tropospheric radiative heating.

123 the polar night experiment, and set the maximum value to 0 K day^{-1} , as shown in Figure 3a.
 124 Episodic deluges still occur under this radiative heating profile (Figure 3b). Therefore, we
 125 can confirm that lower-tropospheric radiative heating is not necessary for episodic deluges.

126 What is necessary to cause episodic deluges? Let's take a look back to Figure 1a.
 127 Another important feature that changes between the 305 K and 325 K cases is the vertical
 128 gradient in the heating rate: the upper-tropospheric cooling is much stronger in the 325
 129 K case, leading to a larger vertical gradient. In the following sections, we use modelling
 130 experiments to investigate how different heating rate profiles can influence the episodic
 131 deluges.

132 3 Heating Rate Profile Experiments and Results

133 The following experiments are run for 100 days with a fixed sea surface temperature of
 134 325 K. All experiments restart from the 325 K reproduction simulation Exp 2. Considering
 135 the features of the radiative heating rate profile under 325 K, the profile can be roughly
 136 divided into three parts: the lower troposphere, the upper troposphere with strong radiative
 137 cooling, and the stratosphere. We prescribe the radiative heating rate profiles in a three-
 138 layer structure as shown in Figure 4. The heating rate profiles do not evolve with time, so
 139 both longwave and shortwave radiative transfer are turned off in the model. The heating
 140 rate profiles are given by two functions,

$$141 \quad \gamma = \begin{cases} A_{tro} \frac{a^{-z^*} - a^{-z_0}}{a^{-z^*} + a^{z_0}} + B_{tro}, & \text{for the troposphere,} \\ A_{stra} \frac{a^{-z^*} - a^{-35}}{a^{-z^*} + a^{35}} + B_{stra}, & \text{for the stratosphere,} \end{cases} \quad (1)$$

142 where γ is the radiative heating rate in the units of K day^{-1} , z^* is a relative height explained
 143 below, a is a dimensionless number that controls the smoothness of the profiles, and here
 144 $a = 1.5$. Coefficients γ_0 , γ_1 , and z_0 control the width and the central point of the heating
 145 rate profile, as shown in Figure 4. In the troposphere, $z^* = z - z_0$, $A_{tro} = \gamma_0 - \gamma_1$, and
 146 $B_{tro} = \frac{\gamma_0 + \gamma_1}{2}$. In the stratosphere, $z^* = z - 35$, $A_{stra} = \gamma_1$, and $B_{stra} = \frac{\gamma_0 + \gamma_1}{2}$.

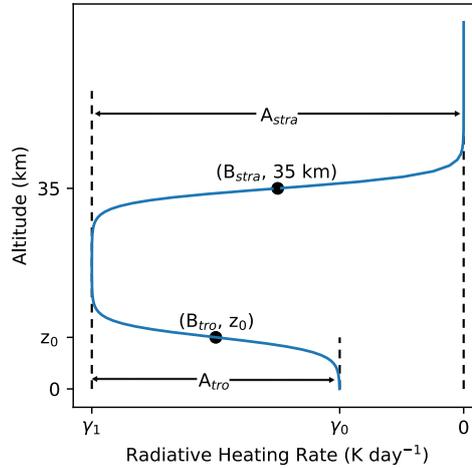


Figure 4. Illustration of the prescribed radiative heating rate profile. The profile can be divided into three parts. The lower troposphere spans from surface to z_0 , with radiative heating rate of γ_0 . The upper troposphere, with radiative heating rate of γ_1 , spans from z_0 to 35 km. The stratosphere spans from 35 km to the model top with radiative heating rate of 0 K day^{-1} . The layers are connected to each other smoothly.

147 In the following sections we conduct 4 groups of experiments (Table 1) focusing on three
 148 factors: experiments in group one (G1) focus on the influence of the lower-tropospheric

149 radiative heating rate γ_0 ; experiments in G2 and G3 focus on the influence of the upper-
 150 tropospheric radiative heating rate γ_1 ; experiments in G4 focus on the height of the inhibition
 151 layer z_0 .

152 3.1 The effect of lower-tropospheric heating rate (γ_0)

153 First, we check the influence of the lower-tropospheric radiative heating rate while
 154 controlling other factors. Figure 5a shows the radiative heating rate profiles in the first
 155 group of experiments (G1). The stratospheric heating rate is 0 K day^{-1} and the upper-
 156 tropospheric heating rate γ_1 is -1.5 K day^{-1} . The inhibition layer height z_0 is 8 km.
 157 The lower-tropospheric heating rates γ_0 are 0.2, 0, -0.2 , -0.5 , -0.8 , and -1.2 K day^{-1} ,
 158 respectively.

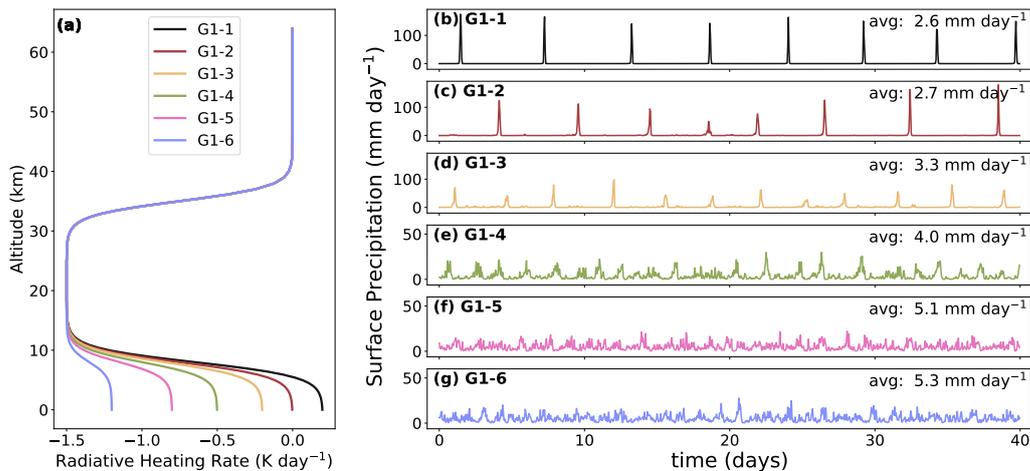


Figure 5. Simulations with fixed upper-tropospheric heating rate and different lower-tropospheric heating rates. Panel (a) shows the prescribed radiative heating rate profiles. The upper-tropospheric heating rate is -1.5 K day^{-1} . The lower-tropospheric heating rates in panels (b) to (g) are 0.2, 0, -0.2 , -0.5 , -0.8 , and -1.2 K day^{-1} , respectively. The inhibition layer height is 8 km.

159 Figures 5b–g show the surface precipitation. Episodic deluges happen even if the lower-
 160 tropospheric radiative heating rate is -0.2 K day^{-1} (Figure 5d), so again lower-tropospheric
 161 radiative heating is not required for episodic deluges. When the lower-tropospheric radiative
 162 heating rate is greater than -0.2 K day^{-1} , precipitation is strongly concentrated in the
 163 deluges (Figures 5b–d). Precipitation during the inhibition period is close to zero and the
 164 convective period is short. As the lower-tropospheric radiative heating rate slowly moves
 165 towards a negative value, the randomness of precipitation gradually increases into a quasi-
 166 steady pattern. When the lower-tropospheric radiative heating rate is less than -0.5 K
 167 day^{-1} , the precipitation pattern is no longer episodic but completely random (Figure 5f &
 168 g).

169 We also carry out another set of experiments with heating rate profiles that are discon-
 170 tinuous at z_0 and at 35 km to exclude the influence of the transition details between the
 171 lower and upper troposphere. The heating rate profile settings are the same as those in G1,
 172 but the transitions are sharp (Figure A1a). The results are similar (Figure A1). The higher
 173 the lower-tropospheric radiative heating rate is, the more episodic the precipitation pattern
 174 is. Episodic deluges can occur even with a negative lower-tropospheric heating rate of -0.2
 175 K day^{-1} .

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3.2 The effect of upper-tropospheric heating rate (γ_1)

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Under what conditions will the negative lower-tropospheric radiative heating trigger episodic deluges? We examine the effect of upper-tropospheric heating rate while keeping the lower-tropospheric heating rate negative in the second group of experiments (G2). Figure 6a shows the radiative heating rate profiles. The stratospheric heating rate is 0 K day^{-1} and the lower-tropospheric heating rate, γ_0 , is -0.2 K day^{-1} . The inhibition layer height z_0 is 8 km. The upper-tropospheric heating rates, γ_1 , are -0.2 , -0.3 , -0.5 , -0.7 , -1.0 , and -1.5 K day^{-1} , respectively.

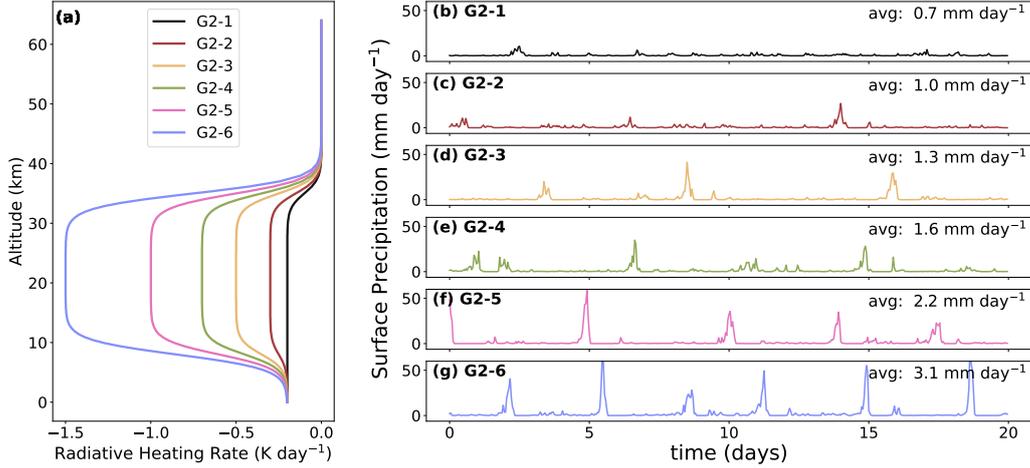


Figure 6. Simulations with fixed lower-tropospheric heating rate and different upper-tropospheric cooling rates. Panel (a) shows the prescribed radiative heating rate profiles. The lower-tropospheric heating rate is -0.2 K day^{-1} . The upper-tropospheric heating rates in panels (b) to (g) are -0.2 , -0.3 , -0.5 , -0.7 , -1.0 , and -1.5 K day^{-1} , respectively. The inhibition layer height is 8 km.

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Figures 6b–g show the surface precipitation. When the lower-tropospheric heating rate is negative, the upper-tropospheric cooling rate needs to be large enough to trigger episodic deluges. In this set of experiments, episodic deluges exist when the upper-tropospheric heating rate is less than -0.5 day^{-1} (Figures 6d–g). The stronger the upper-tropospheric cooling is, the shorter the episodic period is, and the more concentrated the precipitation is in the deluges.

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We also carry out a set of similar experiments with discontinuous heating rate profiles. The radiative heating rates for the lower troposphere, upper troposphere, and stratosphere are exactly the same as in G2, only the transitions between layers are sharp. The results are similar (Figure A2) except for the exact period of episodic deluges. In general, periods with discontinuous heating rate profiles are longer than periods with smoothed profiles.

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Figure 6 suggests that the upper-tropospheric radiative cooling needs to be strong in order to trigger episodic deluges. What if the lower-tropospheric and upper-tropospheric radiative cooling are both strong? Our third group of experiments (G3) set the lower troposphere and upper troposphere to have the same heating rate, that is, $\gamma_0 = \gamma_1$. The upper layer heating rate is 0 K day^{-1} . The lower-tropospheric and upper-tropospheric heating rates are -1 , -2 , -3 , -4 , and -5 K day^{-1} , respectively. The heating rate profiles are similar to those in Figure 2a in Seeley and Wordsworth (2021), but they focused on changing the troposphere height and we focus on changing the tropospheric heating rate. Figure 7 shows the results. Precipitation patterns in G3 are all quasi-steady, showing that

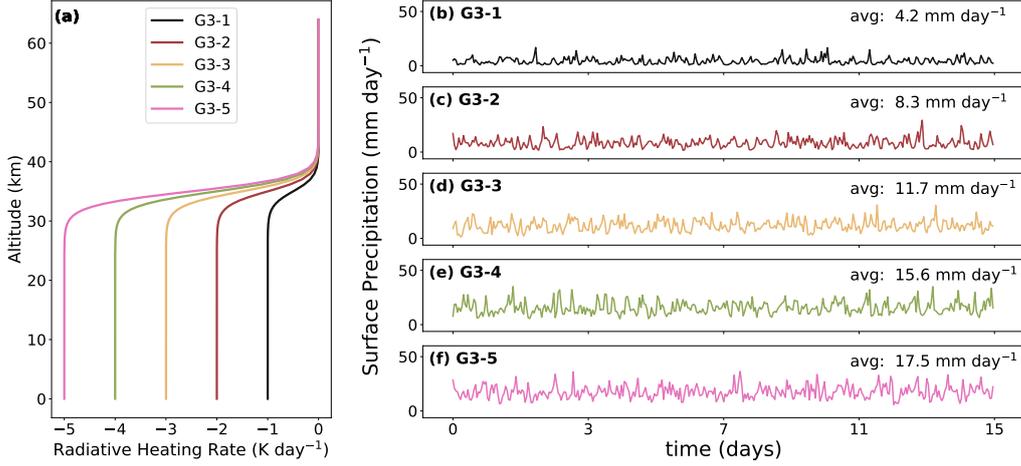


Figure 7. Simulations when lower-tropospheric heating rates and upper-tropospheric cooling rates are the same. Panel (a) shows the prescribed radiative heating rate profiles. The lower-tropospheric heating rates and the upper-tropospheric heating rates in panels (b) to (f) are -1 , -2 , -3 , -4 , and -5 K day⁻¹, respectively.

204 uniform strong radiative cooling is not enough to trigger episodic deluges. This highlights the
 205 fact that the vertical gradient in radiative cooling is more important for episodic deluges
 206 than the magnitude of radiative cooling. We also carry out a set of similar experiments
 207 with discontinuous heating rate profiles. The precipitation patterns are also quasi-steady
 208 (Figure A3).

209 **3.3 The effect of the inhibition layer height (z_0)**

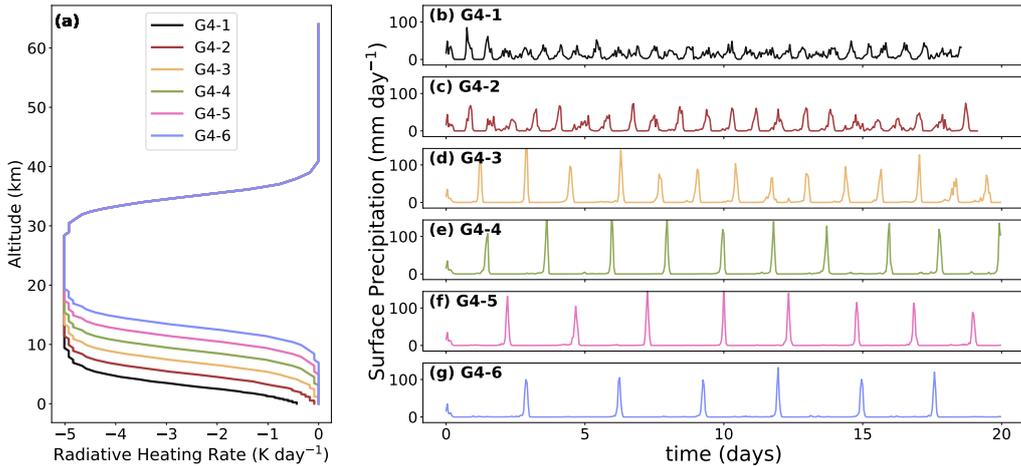


Figure 8. Simulations with fixed lower-tropospheric and upper-tropospheric heating rates but different inhibition layer heights. Panel (a) shows the prescribed radiative heating rate profiles. The lower-tropospheric heating rate is 0 K day⁻¹, and the upper-tropospheric heating rate is -5 K day⁻¹. z_0 in panels (b) to (g) are 3, 5, 7, 9, 11, and 13 km, respectively.

210 The inhibition layer height is the last factor we test. Figure 8a shows the radiative
 211 heating rate profiles in the fourth group of experiments (G4). Both the upper and lower-
 212 tropospheric heating rates γ_0 are 0 K day^{-1} . The upper-tropospheric heating rate γ_1 is -5
 213 K day^{-1} . The inhibition layer heights z_0 are 3, 5, 7, 9, 11, and 13 km, respectively.

214 Most of the cases have episodic deluges, but if the the inhibition layer is not high
 215 enough, for example, 3 km, precipitation is not very episodic (Figure 8b). As the inhibition
 216 height increases from 5 km to 13 km, the period increases from ≈ 1 day to ≈ 3 days. We
 217 will discuss the factors determining the period in section 3.6.

218 We also carry out a set of experiments with discontinuous heating rate profiles (Figure
 219 A4) and the results are almost the same. The discontinuous heating rate profiles, compared
 220 with smoothed profiles, have somewhat higher inhibition heights (Figure A4a vs Figure 8a).
 221 Because of higher inhibition heights, precipitation patterns with discontinuous heating rates
 222 (Figures A4b & c) are more episodic than the precipitation patterns with smoothed heating
 223 rates (Figures 8b & c).

224 3.4 What is the underlying mechanism?

225 The results of the experiments above with prescribed heating rates can be divided
 226 into three types: episodic deluges with positive lower-tropospheric heating rate, episodic
 227 deluges with negative lower-tropospheric heating rate, and quasi-steady precipitation with
 228 negative lower-tropospheric heating rate. According to these three types, we choose three
 229 representative experiments from section 3.1 to study the mechanism of the episodic deluges.
 230 Figure 9 shows the analyses. The left column shows an example of episodic deluges with a
 231 lower-tropospheric heating rate of 0.2 K day^{-1} (Figures 9a & d). The middle column shows
 232 an example of episodic deluges with a negative lower-tropospheric heating rate of -0.2 K
 233 day^{-1} (Figures 9b & e). The right column shows an example of quasi-steady precipitation
 234 with a negative lower-tropospheric heating rate of -0.8 K day^{-1} (Figures 9c & f). The
 235 upper-tropospheric radiative heating rates are all -1.2 K day^{-1} , and the inhibition height
 236 is 8 km.

237 Saturated equivalent potential temperature (θ_{se}) is conserved for a reversible moist
 238 adiabatic process, so θ_{se} is a useful measure of the static stability of saturated atmosphere
 239 (Emanuel et al., 1994). When the lower-tropospheric radiative heating rate is positive
 240 (Figure 9a), the θ_{se} profile has a strong inversion at ≈ 1 km, but has almost no obvious
 241 inversion above 2 km (red line in Figure 9j). The θ_{se} inversion at ≈ 1 km is unlikely to be
 242 important for suppressing the convection. Figures 3c & d suggest that the inhibition layer
 243 reaches up to 10 km (see also Fig. 3 in Seeley and Wordsworth (2021)). Moreover, when
 244 we remove the heating at 1–2 km in the polar night experiment, episodic deluges still exist
 245 (Figure 3). When the lower-tropospheric radiative heating rate is negative (Figures 9b &
 246 c), the θ_{se} profiles of the -0.2 K day^{-1} case and the -0.8 K day^{-1} case (Figure 9k & l)
 247 appear broadly similar, but only the -0.2 K day^{-1} case has episodic deluges (Figure 9e).
 248 Therefore, for episodic deluges, θ_{se} does not appear to yield insight into the location of the
 249 inhibition layer. In what follows, we try to understand the underlying mechanism through
 250 a different angle.

251 The difference in convective available potential energy (CAPE) and convective inhibi-
 252 tion (CIN) at the start and end of the precipitation is large (Figures 9m & p), suggesting
 253 another possible way to understand the episodic deluges. CAPE is defined as

$$254 \text{ CAPE} = R_d \int_{p(\text{LFC})}^{p(\text{EL})} (T_{ve} - T_{vp}) d \ln(p), \quad (2)$$

255 where R_d is the gas constant for dry air, LFC is the level of free convection, EL is the
 256 level of neutral buoyancy, T_{ve} is the virtual temperature of the environment, and T_{vp} is the
 257 virtual temperature of the rising air parcel (Williams & Renno, 1993; Riemann-Campe et
 258 al., 2009). CAPE shows the total energy available for convection (Emanuel et al., 1994),

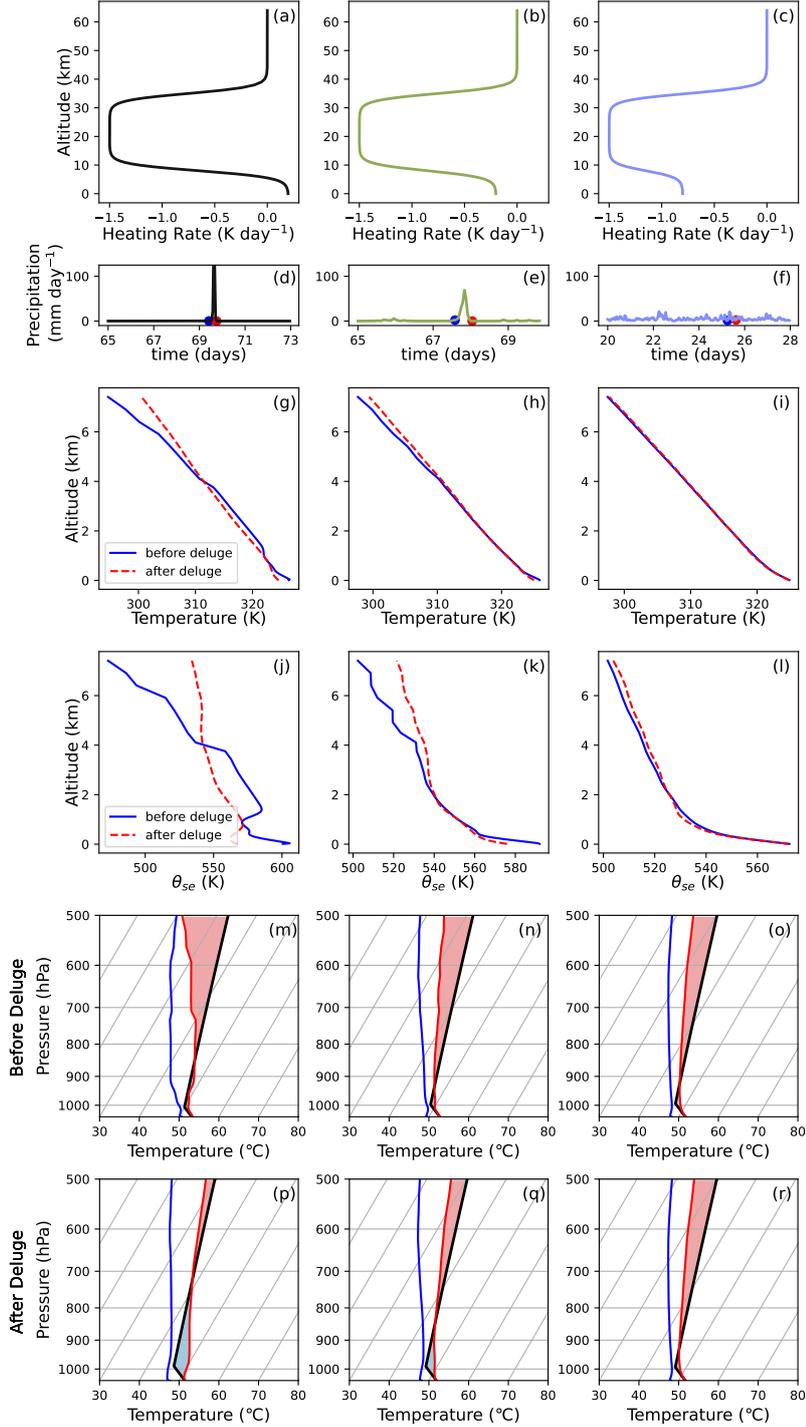


Figure 9. The specified radiative heating rate profiles, precipitation time series, air temperature, saturated equivalent potential temperature, and CAPE and CIN before and after convection. Panels in the left, middle, and right columns correspond to the experiments in Figures 5b, 5d, and 5f, with lower-tropospheric heating rates of 0.2, -0.2 , and -0.8 K day⁻¹, respectively. In the two bottom rows, the blue, black, and red lines are dew point temperature, parcel temperature, and environmental temperature, respectively. Red and blue shades in panels (m)-(o) show the CAPE and CIN before the deluge, and shades in panels (p)-(r) show the CAPE and CIN after the deluge.

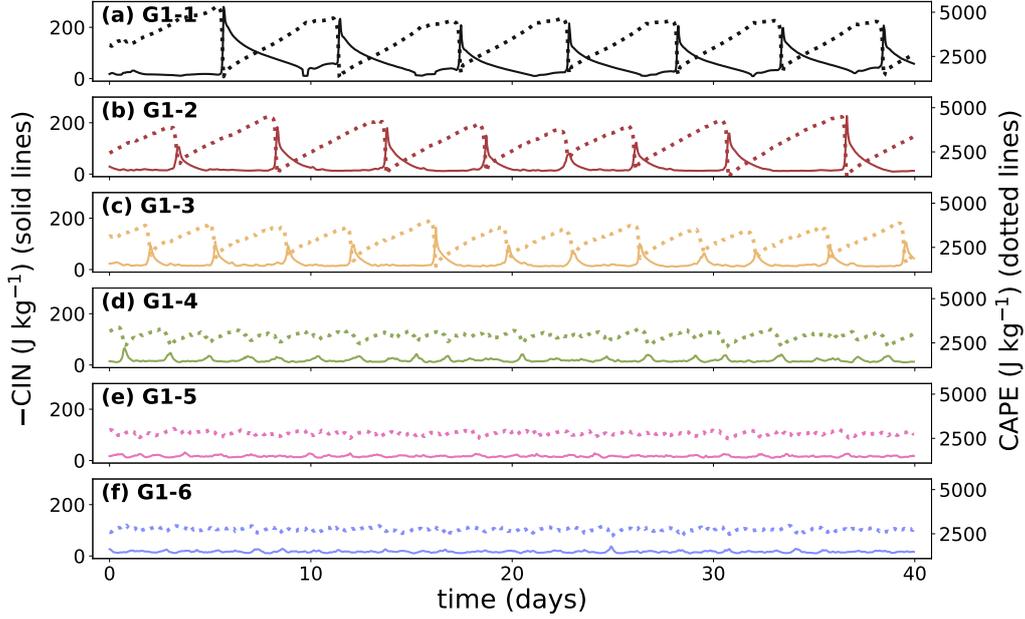


Figure 10. The time series of convective inhibition (CIN) and convective available potential energy (CAPE) for the experiments in Figure 5. We plot the negative value of CIN for clarity.

259 but large CAPE does not guarantee the occurrence of strong convection. To release CAPE,
 260 the air parcel needs to overcome the negative buoyancy from the environment and rise high
 261 enough. CIN measures the intensity of the inhibition energy, and is defined as

$$262 \quad \text{CIN} = R_d \int_{p(\text{SFC})}^{p(\text{LFC})} (T_{ve} - T_{vp}) d \ln(p), \quad (3)$$

263 where SFC is the surface. The value of CIN indicates whether convection will occur
 264 (Colby Jr, 1984; Williams & Renno, 1993; Riemann-Campe et al., 2009), and the value
 265 of CAPE indicates how strong the convection will be if convection occurs (Emanuel et al.,
 266 1994; Rennó & Ingersoll, 1996; Moncrieff & Miller, 1976). Note that CIN is always negative.
 267 The more negative CIN is, the stronger the inhibition is.

268 Figure 10 shows the time series of CAPE and CIN for the experiments in G1. When the
 269 precipitation pattern is episodic, CAPE decreases and CIN gets significantly more negative
 270 after the deluge. The strongly negative CIN inhibits convection. During the inhibition
 271 period, CAPE increases while CIN weakens as a function of time. When CIN is too weak to
 272 inhibit convection, convection starts, and the strong CAPE release causes an intense deluge.

273 Table 1 shows the average and the 2- σ control limits of CIN and CAPE for all the
 274 experiments. For the experiments with episodic deluges, lower 2- σ is less than -50 J kg^{-1} .
 275 CAPE does not differ much between episodic or quasi-steady precipitation experiments.
 276 The experiments in group G1 show that episodic deluge cases tend to have strong CIN and
 277 CAPE. The experiments in group G2 show that large CAPE alone is not enough to trigger
 278 episodic deluges. The experiments in group G4 show that when CIN is relatively strong
 279 enough, even if CAPE is a bit small, for example, G4-1 and G4-2, episodic deluges still
 280 exist. The results are consistent with the common view that CIN indicates the strength

Table 1. Convective inhibition (CIN) and convective available potential energy (CAPE) for all the experiments

Group	Configuration*	CIN (-2σ , $+2\sigma$) J kg ⁻¹	CAPE (-2σ , $+2\sigma$) J kg ⁻¹	Precipitation
Exp 1	SST = 305 K	0 (0, 0)	6419 (6687, 0)	quasi-steady
Exp 2	SST = 325 K	-40 (-97, -19)	2991 (3819, 2054)	episodic
Exp 3	polar night, SST = 330 K	-55 (-114, -34)	1687 (2283, 910)	episodic
Exp 4	polar night, slab ocean	-56 (-118, -35)	1647 (2179, 886)	episodic
Exp 5	polar night, modified heating rate	-61 (-143, -35)	1649 (2589, 742)	episodic
G1-1	$\gamma_0 = 0.2$, $\gamma_1 = -1.5$, $z_0 = 8$	-56 (-177, -11)	3443 (4926, 1672)	episodic
G1-2	$\gamma_0 = 0$, $\gamma_1 = -1.5$, $z_0 = 8$	-28 (-115, -11)	3019 (4409, 1414)	episodic
G1-3	$\gamma_0 = -0.2$, $\gamma_1 = -1.5$, $z_0 = 8$	-24 (-87, -12)	2980 (3823, 1890)	episodic
G1-4	$\gamma_0 = -0.5$, $\gamma_1 = -1.5$, $z_0 = 8$	-19 (-38, -11)	2899 (3257, 2392)	undetermined
G1-5	$\gamma_0 = -0.8$, $\gamma_1 = -1.5$, $z_0 = 8$	-16 (-25, -11)	2780 (2970, 2556)	quasi-steady
G1-6	$\gamma_0 = -1.2$, $\gamma_1 = -1.5$, $z_0 = 8$	-16 (-25, -11)	2745 (2921, 2537)	quasi-steady
G2-1	$\gamma_0 = -0.2$, $\gamma_1 = -0.2$, $z_0 = 8$	-8 (-18, -4)	2811 (3015, 2507)	quasi-steady
G2-2	$\gamma_0 = -0.2$, $\gamma_1 = -0.3$, $z_0 = 8$	-15 (-39, -9)	3213 (3593, 2549)	undetermined
G2-3	$\gamma_0 = -0.2$, $\gamma_1 = -0.5$, $z_0 = 8$	-16 (-63, -9)	3033 (3603, 2151))	episodic
G2-4	$\gamma_0 = -0.2$, $\gamma_1 = -0.7$, $z_0 = 8$	-17 (-54, -10)	3025 (3548, 2289)	episodic
G2-5	$\gamma_0 = -0.2$, $\gamma_1 = -1.0$, $z_0 = 8$	-21 (-78, -11)	2856 (3624, 1797)	episodic
G2-6	$\gamma_0 = -0.2$, $\gamma_1 = -1.5$, $z_0 = 8$	-23 (-78, -11)	2940 (3756, 1912)	episodic
G3-1	$\gamma_0 = \gamma_1 = -1$	-17 (-41, -12)	2750 (2916, 2554)	quasi-steady
G3-2	$\gamma_0 = \gamma_1 = -2$	-14 (-34, -10)	2661 (2826, 2523)	quasi-steady
G3-3	$\gamma_0 = \gamma_1 = -3$	-12 (-39, -8)	2634 (2829, 2380)	quasi-steady
G3-4	$\gamma_0 = \gamma_1 = -4$	-10 (-35, -7)	2703 (3033, 2397)	quasi-steady
G3-5	$\gamma_0 = \gamma_1 = -5$	-9 (-34, -6)	2746 (3142, 2324)	quasi-steady
G4-1	$\gamma_0 = 0$, $\gamma_1 = -5$, $z_0 = 3$	-26 (-56, -12)	2768 (3210, 2138)	episodic
G4-2	$\gamma_0 = 0$, $\gamma_1 = -5$, $z_0 = 5$	-33 (-73, -14)	2797 (3489, 2025)	episodic
G4-3	$\gamma_0 = 0$, $\gamma_1 = -5$, $z_0 = 7$	-41 (-118, -15)	2913 (4060, 1579)	episodic
G4-4	$\gamma_0 = 0$, $\gamma_1 = -5$, $z_0 = 9$	-36 (-126, -11)	3079 (4250, 1598)	episodic
G4-5	$\gamma_0 = 0$, $\gamma_1 = -5$, $z_0 = 11$	-35 (-122, -13)	3269 (4348, 1836)	episodic
G4-6	$\gamma_0 = 0$, $\gamma_1 = -5$, $z_0 = 13$	-30 (-111, -13)	3411 (4358, 2024)	episodic

* γ_0 : the lower-tropospheric radiative heating rate in units of K day⁻¹. γ_1 : the upper-tropospheric radiative heating rate in units of K day⁻¹. z_0 : the height of the inhibition layer in units of km.

281 of convective inhibition and CAPE indicates how strong the convection is (Colby Jr, 1984;
282 Emanuel et al., 1994).

283 In short, the onset of episodic deluges depends on two conditions. First, the lower-
284 tropospheric radiative heating rate should be close to zero (can be negative or positive) to
285 maintain an inhibition period. Second, the upper-tropospheric cooling rate should be strong
286 enough to increase the temperature lapse rate and trigger strong convection. Therefore, we
287 use the ratio of the lower-tropospheric heating rate to the upper-tropospheric heating rate
288 as an index of the vertical gradient in radiative cooling (Figure 11). We find that the

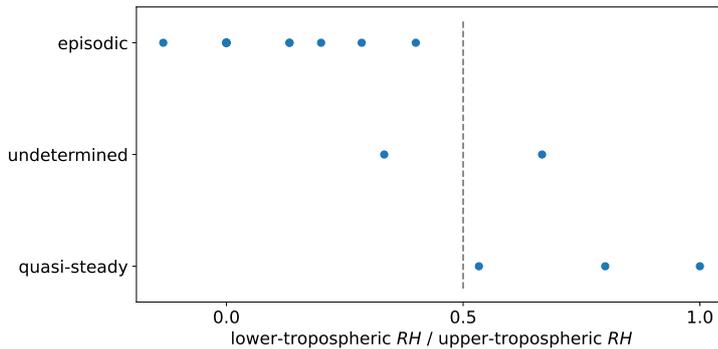


Figure 11. Relative heating rate threshold for episodic deluges. The horizontal axis shows the ratio of the lower-tropospheric radiative heating rate to the upper-tropospheric radiative heating rate (heating rates are positive, cooling rates are negative). Each circle stands for a simulation. Precipitation is episodic when the ratio of lower-tropospheric heating rate to upper-tropospheric heating rate is less than about 0.5.

289 precipitation is episodic when this index is smaller than 0.5, and is non-episodic when this
 290 index is larger than 0.5.

291 Why choose the above index? Because the ratio of the lower-tropospheric heating
 292 rate to the upper-tropospheric heating rate is the key factor for creating a large vertical
 293 temperature gradient. When the lower-tropospheric heating rate is -0.8 K day^{-1} , a -1.5
 294 K day^{-1} upper-tropospheric heating rate is not strong enough compared with the lower-
 295 tropospheric heating rate (Figure 9c). The resulting temperature gradient is too small to
 296 trigger strong convection, and the energy release during convection is not large enough
 297 to start an inhibition period, so CIN and CAPE do not change much before and after
 298 precipitation (Figures 9o & r). The differences in the temperature and θ_{se} profiles before
 299 and after the precipitation are also small (Figures 9i & l). When the lower-tropospheric
 300 heating rate is -0.2 K day^{-1} , a -1.5 K day^{-1} upper-tropospheric heating rate is strong
 301 enough to cause episodic deluges (Figure 9b). The differences in the temperature and θ_{se}
 302 profiles before and after the deluge are obvious (Figures 9h & k). The big difference in
 303 CIN and CAPE before and after the deluge (Figures 9n & q) is necessary to maintain the
 304 oscillation of the deluges.

305 Figure 12 shows the schematic diagram of what happens before and after a deluge.
 306 During the inhibition period, CIN is strongly negative, suppressing convection. The strong
 307 radiative cooling in the upper troposphere “drags” the temperature profile to a greater lapse
 308 rate, and the reevaporation in the lower troposphere also helps with weakening CIN and
 309 building CAPE (step 1). The cooling process triggers strong convection in the majority of
 310 grid points (step 2). The energy release during heavy precipitation adjusts the temperature
 311 profile to have a much smaller gradient and CIN intensifies (step 3). The majority of
 312 grid points do not precipitate under a small vertical temperature gradient and strong CIN,
 313 showing an inhibition period (step 4). Then, the cycle repeats again, with strong radiative
 314 cooling in the upper troposphere increasing the lapse rate again and triggering the next
 315 deluge.

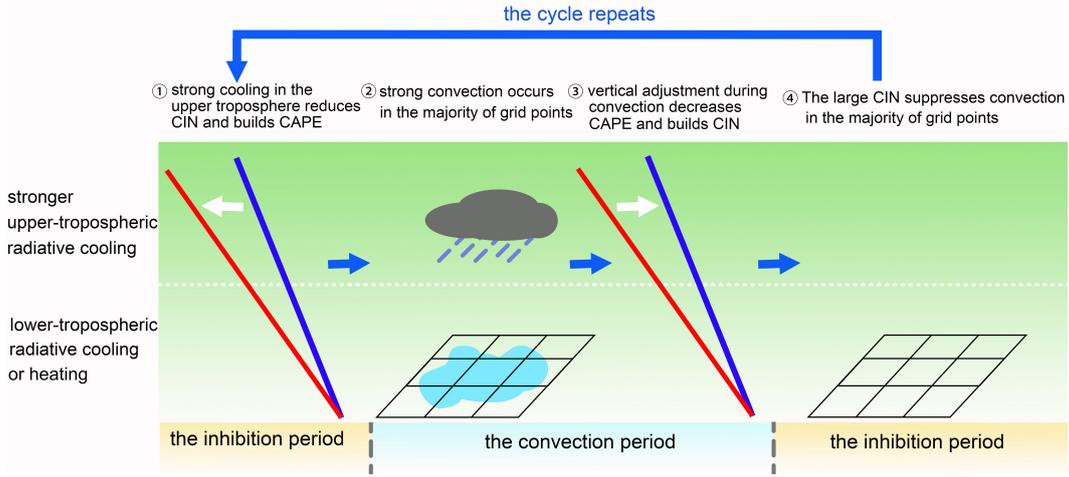


Figure 12. Schematic diagram of the processes before and after the deluge.

3.5 What determines the mean precipitation?

In a balanced system, the energy lost by the atmosphere should be equal to the energy gained by the atmosphere (Allen & Ingram, 2002; R. T. Pierrehumbert, 2002; O’Gorman et al., 2012; Xiong et al., 2022). In the global mean, the net effect of shortwave heating and longwave cooling on the atmosphere is balanced by the latent heat released in the atmosphere and sensible heat flux from the surface. Therefore, we can calculate the surface precipitation by

$$L\rho_w P = \frac{-c_p}{g} \int_{p_s}^0 (HR_{LW} + HR_{SW}) dp - SH, \quad (4)$$

where L is the latent heat of vaporization of water, ρ_w is the density of liquid water, P is the global mean surface precipitation, c_p is the specific heat capacity ($1004 \text{ J kg}^{-1} \text{ K}^{-1}$), g is the gravitational constant, p_s is the surface pressure, HR_{LW} is the longwave heating rate in the atmosphere, HR_{SW} is the shortwave heating rate in the atmosphere, and SH is the sensible heat flux from the surface. HR_{LW} and HR_{SW} are in the units of K s^{-1} , and SH is in the units of W m^{-2} . The integral starts from the surface to the top of the atmosphere.

Figure 13 shows the calculated theoretical precipitation and the actual model output. Each point stands for a single experiment. Although our experiments have fixed surface temperature, the atmosphere in most of the experiments are energy balanced. The average precipitation of the experiments lie between 1 to 12 mm day^{-1} , and most of the points are scattered along the $y = x$ line, showing good agreement between theoretical and simulated average precipitations.

In the experiment with lower-tropospheric heating rate of -0.2 K day^{-1} , upper-tropospheric heating rate of -1.5 K day^{-1} , and inhibition height of 8 km (Figure 5d), the total radiative heating in the atmosphere is -95.4 W m^{-2} , the sensible heat flux from the surface is -1.8 W m^{-2} , and the total latent heating is 97.2 W m^{-2} . The corresponding average precipitation for this case is 3.3 mm day^{-1} . If the lower-tropospheric heating rate decreases to -0.5 K day^{-1} (Figure 5e), the total radiative heating decreases to -115.8 W m^{-2} , and the average precipitation increases to 4.0 mm day^{-1} .

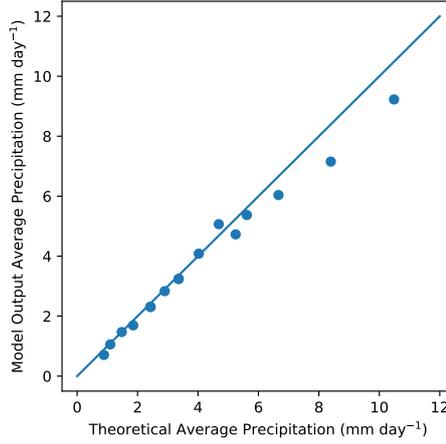


Figure 13. Calculated theoretical precipitation vs actual model output (circles).

3.6 What determines the period of the episodic deluges?

The simulations in sections 3.1-3.3 have different periods. The deluges end rather quickly, so the period of episodic deluges is essentially the time span of the inhibition period. Seeley and Wordsworth (2021) show that precipitation reevaporation cools down the lower troposphere and breaks the inhibition. Therefore, the length of the inhibition phase should be determined by how much time the cooling process of lower troposphere takes. Let's start from the basic temperature equation (Vallis, 2019)

$$\frac{\partial T}{\partial t} = -(u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z})T - \frac{RT}{c_v} \nabla \cdot \vec{v} + \nabla^2(\kappa T) + \frac{J}{c_v}, \quad (5)$$

where R is the molar gas constant ($8.31 \text{ J mol}^{-1} \text{ K}^{-1}$); c_v is the specific heat capacity ($718 \text{ J kg}^{-1} \text{ K}^{-1}$); κ is the thermal diffusivity representing the effect of thermal diffusion and subgrid mixing; J is the external heating source including longwave cooling, shortwave heating, latent heat release through condensation, and latent cooling by reevaporation. Each experiment in this study is in a small domain and without the Coriolis force, so the horizontal temperature gradient is small, and horizontal advection ($u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y}$) is negligible. To calculate the period, we focus on the inhibition periods, so the vertical advection ($w \frac{\partial T}{\partial z}$) and thermodynamic work ($\frac{RT}{c_v} \nabla \cdot \vec{v}$) done by the air parcel in the lower troposphere are also negligible. The influence of temperature diffusion, $\nabla^2(\kappa T)$, is small compared with the timescale of diabatic heating (J). Therefore, diabatic heating alone is the main source of air temperature change. Two factors that can cool down the lower troposphere determine the period: radiative cooling and reevaporation cooling. The equation is given by

$$\Delta t \approx \frac{\Delta T}{-(HR_{LW} + HR_{SW} + HR_e)}, \quad (6)$$

where $-(HR_{LW} + HR_{SW})$ is the radiative cooling rate and $-HR_e$ is the reevaporation cooling rate. The inhibition period starts from the end of a deluge, and ends at the beginning of the next deluge (Figure 14a). All the variables are the time mean and vertically averaged from 2 km to the top of the inhibition layer.

Figure 14d compares the model output period Δt with the calculated theoretical period (Eq. 6). Each circle stands for a single inhibition period. All circles are scattered along the

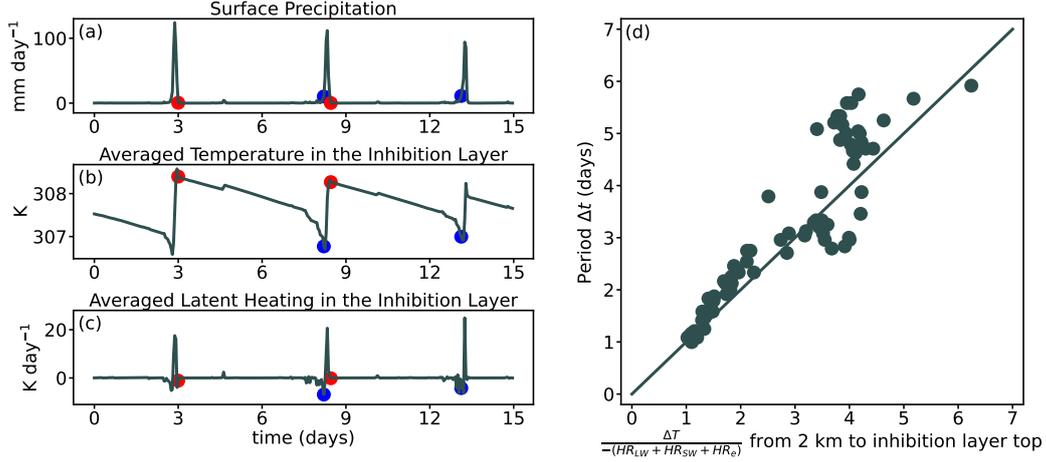


Figure 14. Panels a–c: Surface precipitation (a) and time series of vertically averaged temperature (b) and latent heating (c) in the inhibition layer. An inhibition period starts from the end of previous precipitation (red dot), and ends at the beginning of the next precipitation (blue dot). **Panel d:** The model output period Δt and the calculated theoretical period (Eq. 6). Note that we remove those episodic deluge experiments with unclear boundary between inhibition periods and convection periods, such as experiments G1-4, G2-4, G4-1, and G4-2.

370 $y = x$ line. This result supports the idea that radiative cooling and reevaporation cooling
 371 control the period of episodic deluges.

372 The combined effect of radiative cooling and reevaporation cooling can quantitatively
 373 explain the period differences in Figures 5, 6, and 8. In the experiments G1 with different
 374 lower-tropospheric heating rates (Figure 5), the stronger radiative cooling is, the larger the
 375 denominator in Eq. 6 is, and the quicker the cooling process is. As the lower-tropospheric
 376 radiative heating rate decreases, the inhibition period shortens, until finally the inhibition
 377 period almost vanishes and the precipitation pattern becomes quasi-steady.

378 In the G2 group of experiments, we fix the lower-tropospheric radiative heating rate
 379 at -0.2 K day^{-1} and change the upper-tropospheric cooling rate (Figure 6). A stronger
 380 the upper-tropospheric radiative cooling results in more condensation in the upper tropo-
 381 sphere, so more precipitation droplets reevaporate in the lower troposphere. The increasing
 382 reevaporation cooling leads to shorter period.

383 In the G4 group of experiments with different inhibition layer heights (Figure 8), the
 384 lower-tropospheric radiative cooling rate is zero for all the cases, so reevaporation cooling
 385 decides the periods. Higher inhibition height means weaker total radiative cooling, so less
 386 precipitation from the upper troposphere enters the inhibition layer, and therefore the period
 387 is longer and the inhibition periods are drier.

388 Now we can explain the amplitude differences between the episodic deluge experiments
 389 by considering both mean precipitation and the period of episodic deluges. When controlling
 390 the mean precipitation, longer episode leads to smaller deluge amplitude. In the G1 group
 391 of experiments, when the lower-tropospheric heating rate increases, the period increases,
 392 but the mean precipitation decreases because of less total radiative cooling. The period
 393 increase outweighs the mean precipitation decrease, so the deluge amplitude increases with
 394 a larger lower-tropospheric heating rate (Figures 5b–g). In the G2 group of experiments,
 395 when the upper-tropospheric radiative cooling is stronger, the period becomes shorter, but
 396 the mean precipitation increases because of larger total radiative cooling. Both trends lead

397 to a larger deluge amplitude (Figures 6b–g). In the G4 group of experiments, when the
 398 inhibition layer height increases while controlling other factors, the period becomes longer,
 399 but the mean precipitation decreases because of less total cooling. These two opposite trends
 400 compete with each other, so as the inhibition layer height increases, the deluge amplitude
 401 first increases then decreases (Figures 8b–g).

402 4 Summary and discussion

403 We expand the episodic deluge theory developed by Seeley and Wordsworth (2021).
 404 We show that episodic deluges can occur during polar night, which indicates that shortwave
 405 heating is not a necessary condition for episodic deluges. Moreover, we show that episodic
 406 deluges can occur even if the lower-tropospheric radiative heating rate is negative. We
 407 perform multiple groups of experiments to show that the vertical gradient of the radiative
 408 heating rate profile is an important factor for the onset of episodic deluges. We also discuss
 409 a possible mechanism for the episodic deluges. All of our episodic deluge experiments have
 410 relatively large convective inhibition (CIN), but we cannot explain clearly how the value
 411 of CIN is linked to the inhibition mechanism. How to understand the detailed causes of
 412 the inhibition process? What exactly causes the inhibition? These questions need further
 413 studies.

414 The average precipitation can be understood through atmospheric column energy bud-
 415 get. The period and the deluge amplitude are more complex. We find three factors that
 416 influence the period of episodic deluges: lower-tropospheric radiative heating rate (γ_0), the
 417 upper-tropospheric radiative heating rate (γ_1), and the inhibition layer height (z_0). Gener-
 418 ally, a higher inhibition layer, stronger radiative heating in the inhibition layer, or weaker
 419 radiative cooling in the upper troposphere leads to a shorter period. The period matches the
 420 time for radiation and reevaporation to cool down the lower troposphere. The deluge am-
 421 plitude is decided by the combined effect of the average precipitation and the period. These
 422 two trends compete with each other, so the period does not always vary monotonically with
 423 γ_0 , γ_1 , or z_0 .

424 Another interesting factor is that the heating rate in the 325 K case are nearly hori-
 425 zontally uniform (Figures 1f & g). The homogeneity of radiative heating rates may be
 426 an important factor for episodic deluges, as it causes a spatial coordination among the
 427 grid points, so that heavy precipitation occurs at the same time in the majority of grid
 428 points (Figure 3f). The coordinated dry–convection–dry cycle in the majority of grid points
 429 demonstrates an episodic deluge pattern in a macroscopic view. If the precipitation be-
 430 havior among the grid points are not coordinated, convection occurs here and there at
 431 different times in different grid points, demonstrating a random, or quasi-steady, precipita-
 432 tion pattern in a macroscopic view. Consider, for example, the radiative heating rates in
 433 the reproduction experiments (Figures 1d–g). When the surface temperature is 305 K, both
 434 lower-tropospheric heating and strong upper-tropospheric cooling exist, but not in the ma-
 435 jority of the grid points (Figures 1d & e), so the precipitation pattern is quasi-steady. When
 436 the surface temperature is 325 K, lower-tropospheric heating and strong upper-tropospheric
 437 cooling may not be as strong as in some grid points in the 305 K case during the inhibition
 438 period, but they are more wide spread in over 65% of the grid points (Figures 1f & g), so
 439 the precipitation pattern is episodic. These observations suggest an intimate connection
 440 between convective organization in time and space. Could episodic deluges still exist with
 441 inhomogeneous radiative heating rate? In future studies, we plan to modify the horizontal
 442 distribution of the radiative heating rate and examine how precipitation changes.

443 Here we study episodic deluges in a small domain, and do not include rotation and
 444 the Coriolis force, thus many weather phenomena, such as midlatitude cyclones caused by
 445 baroclinic instability, cannot be considered. In large-scale modelling, with the influence of
 446 Hadley and Walker cells, cyclones and anti-cyclones, large-scale horizontal advection is no
 447 longer negligible, and more factors, such as dynamic lifting, can possibly break inhibition.

448 Can episodic deluges exist in large-scale simulations? How do episodic deluges interact with
449 global circulation and climate? These questions are worthy of further study.

450 Apart from this work, two recent studies, Dagan et al. (2023) and Spaulding-Astudillo
451 and Mitchell (2023), also focus on the temporal variability of precipitation under hothouse
452 climates. Dagan et al. (2023) find that episodic deluges cannot occur on a domain larger
453 than $\mathcal{O}(1000\text{ km})$, and show that the propagation of gravity waves dominates the precipi-
454 tation’s variability in a large domain. Spaulding-Astudillo and Mitchell (2023) show that
455 episodic deluges exist even in 1D single-column radiative-convective model although the
456 period of the episodic deluges is several years, much longer than the several-day period in
457 3D radiative-convective model. They also show that lower-tropospheric radiative heating
458 is not necessary for the onset of the episodic deluges. They propose a mechanism to ex-
459 plain the episodic deluges: “*Emergence occurs when atmospheric instability quantified by the*
460 *convective available potential energy can no longer support the latent heat release of deep,*
461 *entraining convective plumes.*” Whether their proposed mechanism is applicable to our 3D
462 radiative-convective simulations here or not requires further studies.

463 5 Open Research

464 The cloud-resolving model SAM is publicly available at: [http://rossby.msrc.sunysb](http://rossby.msrc.sunysb.edu/~marat/SAM.html)
465 [.edu/~marat/SAM.html](http://rossby.msrc.sunysb.edu/~marat/SAM.html). The data in this study is publicly available at: [https://doi.org/](https://doi.org/10.5281/zenodo.8103889)
466 [10.5281/zenodo.8103889](https://doi.org/10.5281/zenodo.8103889).

467 Acknowledgments

468 We thank Marat F. Khairoutdinov for creating and maintaining SAM. We are grateful to
469 Cheng Li, Da Yang, Daniel D. B. Koll, Feng Ding, Qiu Yang, and Zhihong Tan for the helpful
470 discussions with them. We thank Yixiao Zhang for his help with SAM modelling. Thank
471 Lixiang Gu, Xuelei Wang, and Quxin Cui for their help with CAPE and CIN calculation.
472 J.Y. acknowledges support from the National Natural Science Foundation of China (NSFC)
473 under grants 42275134, 42075046, and 42161144011.

474 References

- 475 Allen, M. R., & Ingram, W. J. (2002). Constraints on future changes in climate and
476 the hydrologic cycle. *Nature*, *419*(6903), 228–232. doi: [https://doi.org/10.1038/](https://doi.org/10.1038/nature01092)
477 [nature01092](https://doi.org/10.1038/nature01092)
- 478 Charnay, B., Hir, G. L., Fluteau, F., Forget, F., & Catling, D. C. (2017). A warm or a cold
479 early Earth? New insights from a 3-D climate-carbon model. *Earth and Planetary*
480 *Science Letters*, *474*, 97–109. doi: [10.1016/j.epsl.2017.06.029](https://doi.org/10.1016/j.epsl.2017.06.029)
- 481 Colby Jr, F. P. (1984). Convective inhibition as a predictor of convection during AVE-
482 SESAME II. *Monthly Weather Review*, *112*(11), 2239–2252. doi: [https://doi.org/](https://doi.org/10.1175/1520-0493(1984)112<2239:CIAAPO>2.0.CO;2)
483 [10.1175/1520-0493\(1984\)112<2239:CIAAPO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<2239:CIAAPO>2.0.CO;2)
- 484 Dagan, G., Seeley, J. T., & Steiger, N. (2023). Convection and convective-organization in
485 hothouse climates. *Authorea Preprints*.
- 486 Emanuel, K. A., et al. (1994). *Atmospheric convection*. Oxford University Press on Demand.
- 487 Goldblatt, C., Robinson, T. D., Zahnle, K. J., & Crisp, D. (2013). Low simulated radiation
488 limit for runaway greenhouse climates. *Nature Geoscience*, *6*(8), 661–667. doi: [10](https://doi.org/10.1038/ngeo1892)
489 [.1038/ngeo1892](https://doi.org/10.1038/ngeo1892)
- 490 Higgins, J. A., & Schrag, D. P. (2003). Aftermath of a snowball Earth. *Geochemistry,*
491 *Geophysics, Geosystems*, *4*(3). doi: [10.1029/2002gc000403](https://doi.org/10.1029/2002gc000403)
- 492 Hir, G. L., Donnadiou, Y., Godd ris, Y., Pierrehumbert, R. T., Halverson, G. P., Macouin,
493 M., . . . Ramstein, G. (2009). The snowball Earth aftermath: Exploring the limits
494 of continental weathering processes. *Earth and Planetary Science Letters*, *277*(3-4),
495 453–463. doi: [10.1016/j.epsl.2008.11.010](https://doi.org/10.1016/j.epsl.2008.11.010)

- 496 Hoffman, P. F., Abbot, D. S., Ashkenazy, Y., Benn, D. I., Brocks, J. J., Cohen, P. A., ...
 497 others (2017). Snowball Earth climate dynamics and Cryogenian geology-geobiology.
 498 *Science Advances*, *3*(11), e1600983.
- 499 Ingersoll, A. P. (1969). The runaway greenhouse: A history of water on Venus. *Journal of*
 500 *the Atmospheric Sciences*, *26*(6), 1191–1198. doi: 10.1175/1520-0469(1969)026<1191:
 501 trgaho>2.0.co;2
- 502 Kasting, J. F., Pollack, J. B., & Ackerman, T. P. (1984). Response of Earth’s atmosphere
 503 to increases in solar flux and implications for loss of water from Venus. *Icarus*, *57*(3),
 504 335–355.
- 505 Khairoutdinov, M. F., & Emanuel, K. (2018). Intraseasonal variability in a cloud-permitting
 506 near-global equatorial aquaplanet model. *Journal of the Atmospheric Sciences*, *75*(12),
 507 4337–4355. doi: 10.1175/jas-d-18-0152.1
- 508 Khairoutdinov, M. F., & Randall, D. A. (2003). Cloud resolving modeling of the ARM
 509 summer 1997 IOP: Model formulation, results, uncertainties, and sensitivities. *Journal*
 510 *of the Atmospheric Sciences*, *60*(4), 607–625. doi: 10.1175/1520-0469(2003)060<0607:
 511 crmota>2.0.co;2
- 512 Kidder, D. L., & Worsley, T. R. (2012). A human-induced hothouse climate? *GSA Today*,
 513 4–11. doi: 10.1130/g131a.1
- 514 Leconte, J., Forget, F., Charnay, B., Wordsworth, R., & Pottier, A. (2013). Increased
 515 insolation threshold for runaway greenhouse processes on Earth-like planets. *Nature*,
 516 *504*(7479), 268–271. doi: 10.1038/nature12827
- 517 Moncrieff, M. W., & Miller, M. J. (1976). The dynamics and simulation of tropical cum-
 518ulonimbus and squall lines. *Quarterly Journal of the Royal Meteorological Society*,
 519 *102*(432), 373–394. doi: <https://doi.org/10.1002/qj.49710243208>
- 520 O’Gorman, P. A., Allan, R. P., Byrne, M. P., & Previdi, M. (2012). Energetic constraints
 521 on precipitation under climate change. *Surveys in geophysics*, *33*(3), 585–608. doi:
 522 <https://doi.org/10.1007/s10712-011-9159-6>
- 523 Pierrehumbert, R., Abbot, D., Voigt, A., & Koll, D. (2011). Climate of the Neopro-
 524terozoic. *Annual Review of Earth and Planetary Sciences*, *39*(1), 417–460. doi:
 525 10.1146/annurev-earth-040809-152447
- 526 Pierrehumbert, R. T. (2002). The hydrologic cycle in deep-time climate problems. *Nature*,
 527 *419*(6903), 191–198. doi: <https://doi.org/10.1038/nature01088>
- 528 Popp, M., Schmidt, H., & Marotzke, J. (2016). Transition to a moist greenhouse with CO₂
 529 and solar forcing. *Nature Communications*, *7*(1). doi: 10.1038/ncomms10627
- 530 Ramirez, R. M., Kopparapu, R. K., Lindner, V., & Kasting, J. F. (2014). Can increased
 531 atmospheric CO₂ levels trigger a runaway greenhouse? *Astrobiology*, *14*(8), 714–731.
 532 doi: 10.1089/ast.2014.1153
- 533 Rennó, N. O., & Ingersoll, A. P. (1996). Natural convection as a heat engine: A theory
 534 for CAPE. *Journal of atmospheric sciences*, *53*(4), 572–585. doi: [https://doi.org/
 535 10.1175/1520-0469\(1996\)053<0572:NCAAHE>2.0.CO;2](https://doi.org/10.1175/1520-0469(1996)053<0572:NCAAHE>2.0.CO;2)
- 536 Riemann-Campe, K., Fraedrich, K., & Lunkeit, F. (2009). Global climatology of convective
 537 available potential energy (CAPE) and convective inhibition (CIN) in ERA-40 reanal-
 538 ysis. *Atmospheric Research*, *93*(1-3), 534–545. doi: 10.1016/j.atmosres.2008.09.037
- 539 Seeley, J. T., & Wordsworth, R. D. (2021). Episodic deluges in simulated hothouse climates.
 540 *Nature*, *599*(7883), 74–79. doi: 10.1038/s41586-021-03919-z
- 541 Sleep, N. H. (2010). The Hadean-Archaean environment. *Cold Spring Harbor Perspectives*
 542 *in Biology*, *2*(6), a002527–a002527. doi: 10.1101/cshperspect.a002527
- 543 Spaulding-Astudillo, F. E., & Mitchell, J. L. (2023). The emergence of relaxation-oscillator
 544 convection on earth and titan. *arXiv preprint arXiv:2306.03219*.
- 545 Steffen, W., Rockström, J., Richardson, K., Lenton, T. M., Folke, C., Liverman, D., ...
 546 others (2018). Trajectories of the Earth system in the Anthropocene. *Proceedings of*
 547 *the National Academy of Sciences*, *115*(33), 8252–8259.
- 548 Vallis, G. (2019). *Essentials of atmospheric and oceanic dynamics*. Cambridge University
 549 Press.
- 550 Williams, E., & Renno, N. (1993). An analysis of the conditional instability of the tropical

- 551 atmosphere. *Monthly Weather Review*, 121(1), 21–36. doi: 10.1175/1520-0493(1993)
 552 121<0021:aaotci>2.0.co;2
- 553 Wolf, E., Haqq-Misra, J., & Toon, O. (2018). Evaluating climate sensitivity to CO₂ across
 554 Earth’s history. *Journal of Geophysical Research: Atmospheres*, 123(21), 11–861. doi:
 555 10.1029/2018jd029262
- 556 Wolf, E. T., & Toon, O. B. (2015). The evolution of habitable climates under the brightening
 557 sun. *Journal of Geophysical Research: Atmospheres*, 120(12), 5775–5794. doi: 10
 558 .1002/2015jd023302
- 559 Wordsworth, R. D., & Pierrehumbert, R. T. (2013). Water loss from terrestrial planets
 560 with CO₂-rich atmospheres. *The Astrophysical Journal*, 778(2), 154. doi: 10.1088/
 561 0004-637x/778/2/154
- 562 Xiong, J., Yang, J., & Liu, J. (2022). Smaller sensitivity of precipitation to surface tem-
 563 perature under massive atmospheres. *Geophysical Research Letters*, 49(18). doi:
 564 10.1029/2022gl099599
- 565 Yang, J., Jansen, M. F., Macdonald, F. A., & Abbot, D. S. (2017). Persistence of a
 566 freshwater surface ocean after a snowball Earth. *Geology*, 45(7), 615–618.

567 Appendix A Experiment results with discontinuous heating rate profiles

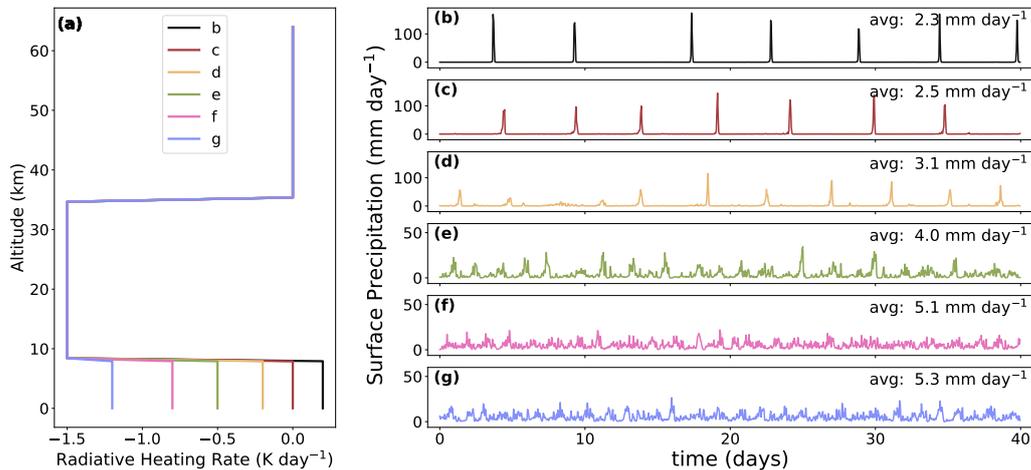


Figure A1. Same as the control group experiments of Figure 5, but the heating rate is discontinuous at the inhibition layer height (z_0) and at the stratosphere (35 km). Panel (a) shows the prescribed radiative heating rate profiles. The upper-tropospheric heating rate is -1.5 K day^{-1} . The lower-tropospheric heating rates in panels (b) to (g) are 0.2, 0, -0.2 , -0.5 , -0.8 , and -1.2 K day^{-1} , respectively. The inhibition layer height is 8 km.

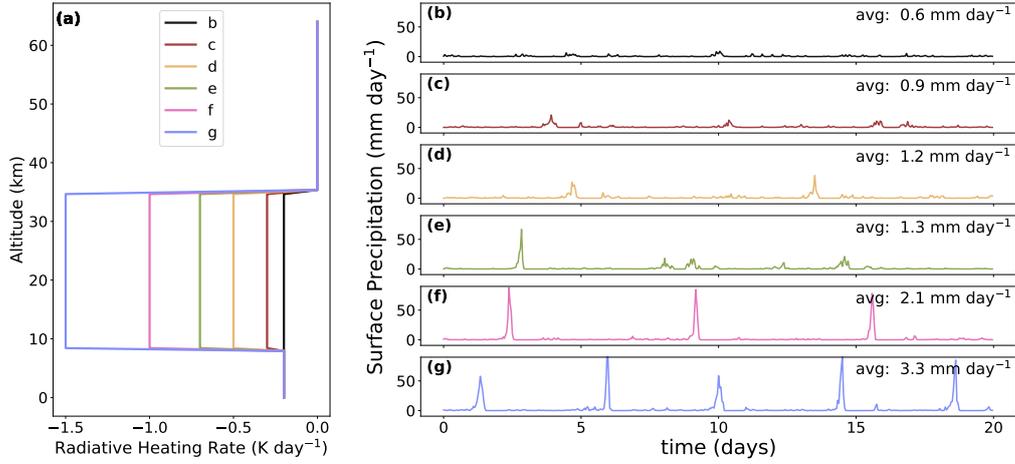


Figure A2. Same as the control group experiments of Figure 6, but the heating rate is discontinuous at z_0 and at 35 km. Panel (a) shows the prescribed radiative heating rate profiles. The lower-tropospheric heating rate is -0.2 K day^{-1} . The upper-tropospheric heating rates in panels (b) to (g) are -0.2 , -0.3 , -0.5 , -0.7 , -1.0 , and -1.5 K day^{-1} , respectively. The inhibition layer height is 8 km.

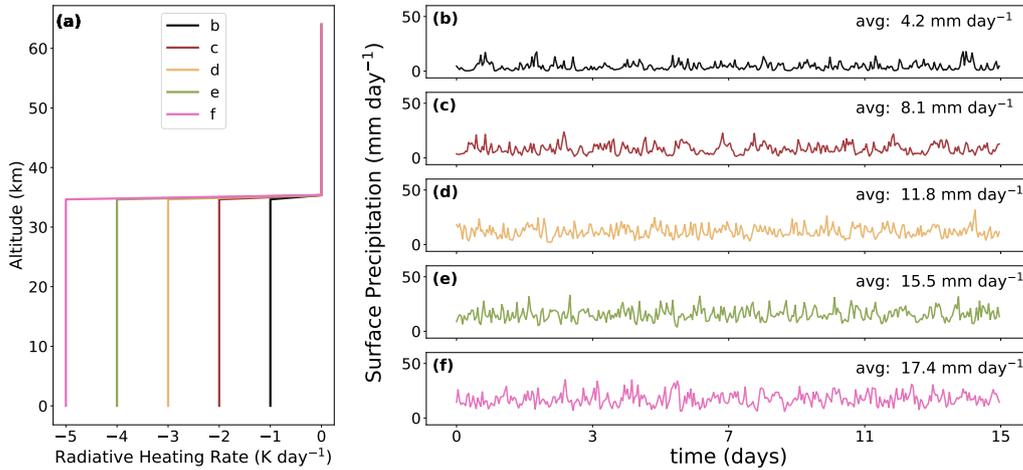


Figure A3. Same as the control group experiments of Figure 7, but the heating rate is discontinuous at z_0 and at 35 km. Panel (a) shows the prescribed radiative heating rate profiles. The lower-tropospheric heating rates and the upper-tropospheric heating rates in panels (b) to (f) are -1 , -2 , -3 , -4 , and -5 K day^{-1} , respectively.

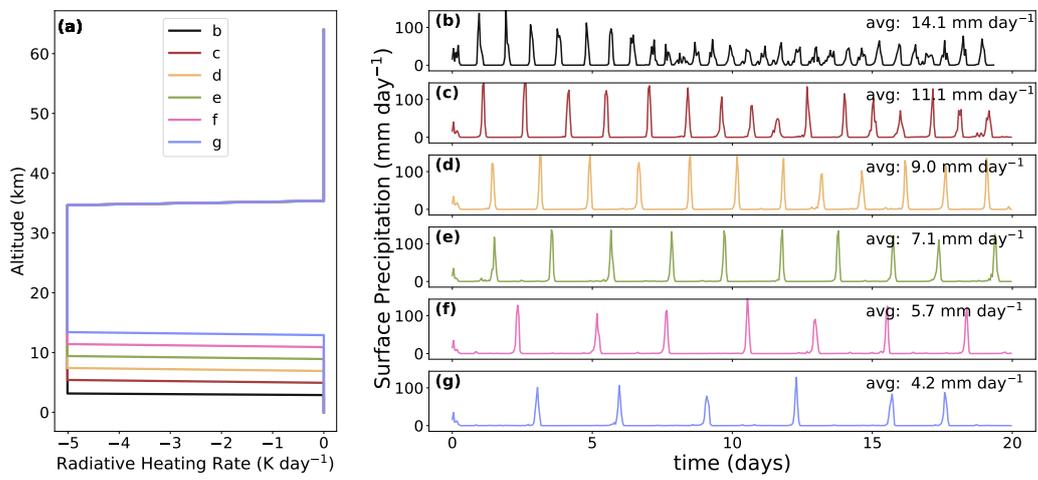


Figure A4. Same as the control group experiments of Figure 8, but the heating rate is discontinuous at z_0 and at 35 km. Panel (a) shows the prescribed radiative heating rate profiles. The lower-tropospheric heating rate is 0 K day^{-1} , and the upper-tropospheric heating rate is -5 K day^{-1} . The inhibition layer heights in panels (b) to (g) are 3, 5, 7, 9, 11, and 13 km, respectively.