Presence or absence of stabilizing Earth system feedbacks on different timescales

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Abstract

The question of how Earth's climate is stabilized on geologic timescales is important for understanding Earth's history, long-term consequences of anthropogenic climate change, and planetary habitability. Here we quantify the typical amplitude of past global temperature fluctuations on timescales from hundreds to tens of millions of years, and use it to assess the presence or absence of long-term stabilizing feedbacks in the climate system. On timescales between 4-400 kyrs, fluctuations fail to grow with timescale, suggesting that stabilizing mechanisms like the hypothesized "weathering feedback" have indeed exerted dominant control in this regime. Fluctuations grow on longer timescales, potentially due to tectonically or biologically driven changes that make weathering act as a climate forcing as well as a feedback. These slower fluctuations show no evidence of being damped, implying that chance may still have played a non-negligible role in maintaining the long-term habitability of Earth.

1 Introduction

The global carbon cycle exerts substantial control over Earth's climate through its influence on the atmospheric CO_2 concentration. CO_2 enters the ocean-atmosphere system due to solid Earth degassing and organic carbon oxidation, and is removed through the chemical weathering of carbonate and silicate rocks and subsequent carbonate burial in ocean sediments, as well as organic carbon burial [1]. Weathering rates increase with temperature and CO_2 concentration: this is hypothesized to lead to a long-term stabilizing feedback [2] in which increases in surface temperatures are countered by drawdown of CO_2, and vice versa. This feedback can help explain the puzzle of Earth's enduring habitability even as stellar luminosity has changed significantly [2, 3]. It also justifies the useful "steady-state" assumption in the study of past carbon-cycle change [4], and is an important foundation for the "habitable zone" concept used in exoplanet research [5].

Understanding such long-term stabilizing feedbacks is also essential for understanding the Earth system's dynamical response to perturbation. A salient example is the case of anthropogenic global climate change [6]. Modeling indicates that the weathering feedback damps perturbations with a characteristic (i.e., e-folding) timescale of about 200–400 kyrs [7, 8]. On timescales of $\sim$10 kyrs, the dynamics of the marine calcium carbonate cycle also play an important role [9, 10]. Because burial rates increase with the deep ocean carbonate ion concentration ($[CO_3^{2-}]$), a feedback emerges which indirectly and partially stabilizes atmospheric CO_2: it has had a relatively fast response timescale since the development of pelagic biogenic calcification in the mid-Mesozoic ($\sim$200 Ma) [11].

The current evidence that Earth's climate is indeed stabilized by long-term carbon-cycle feedbacks is as follows. Paleoclimate data suggest that input and output fluxes of CO_2 into the ocean-atmosphere system have typically been balanced to within a few percent [12, 13]. Together with the actual observation of Earth's apparent enduring habitability [14], this is cited as evidence for stabilizing mechanisms; nevertheless, this line of reasoning can be challenged [15, 16]. Plausible parametrizations of the underlying processes lead to these stabilizing feedbacks emerging in models [7, 8], but this cannot alone confirm the importance of the feedbacks within the real Earth system. Finally, model predictions can be compared with the observed response from individual large climate-carbon cycle perturbations in the geologic past: a recent study focusing on the Paleocene-Eocene Thermal Maximum ($\sim$56 Ma) found an overshoot of the calcium carbonate compensation depth in the aftermath of the event consistent with the weathering feedback [17], although organic matter burial may also have played an important role [18, 19]. Nevertheless, the insight from this approach is limited to those specific intervals of Earth history with large disruption events.

To convincingly assess the role of long-term stabilizing feedbacks in the Earth system, we need evidence that is direct (i.e. rooted in observations of past climate changes), general (i.e. applies continuously throughout geologic time), and that provides good constraints on their dynamics. Here we provide such evidence directly from data of past global temperature fluctuations. We first show how the typical amplitude of these fluctuations provides information about the relative dominance — or lack of dominance — of stabilizing feedbacks on different timescales. We quantify these amplitudes across a vast range of timescales, expanding on previous work by Lovejoy [20], and go beyond this to explain observed scaling regimes in terms of physical and biogeochemical processes. Specifically, the data exhibit a regime between about 4-400 kyrs in which fluctuations fail to grow with timescale, and a longer-timescale regime in which they do. We interpret the former as novel observational confirmation of long-term stabilizing Earth system feedbacks, and link the latter to longer-term tectonic or biological evolution, as well as the potential role of chance in maintaining Earth's observed billion-year habitability.

2 Results

2.1 Simple models of long-term climate variability

Stabilizing feedbacks, in principle, should affect how the typical amplitude of fluctuations within a system changes with timescale [21]. To show how this would work, we take a purposely simplified perspective of the Earth system in which the only variable of interest is globally averaged surface temperature, T. Such simplification is appropriate for a first attempt at extracting information about long-term Earth system feedbacks directly from data of past fluctuations; furthermore, as we will show, it is already sufficient for obtaining useful insight.

Two simple "end-member" scenarios for this simplified view are displayed in Figure 1. Scenario A is the classic established model of climate variability in the absence of stabilizing feedbacks: a random walk [22–24]. This assumes that slowly-evolving components of the Earth system retain an aggregate "memory" of the fast-evolving components that accumulates approximately randomly [22]. In that case, temperature evolution would be described by the following stochastic differential equation (SDE):

$$\frac{dT}{dt} = a\eta(t), \tag{1}$$

where $\eta(t)$ is a Gaussian white noise forcing and a is a constant. In this model, the root-mean-square temperature fluctuation ΔT_{rms} occurring on a timescale Δt is proportional to $\Delta t^{1/2}$ (equivalent to red noise, see Materials and Methods). Many climate time series exhibit such scaling behavior [22–26], and the ability to reproduce it is part of the model's appeal. Throughout this paper we will often refer to the scaling exponent ($1/2$ in this case) as H.

Scenario B is the same as Scenario A, but also includes a stabilizing feedback with characteristic (i.e. e-folding) timescale τ (also known as an Ornstein-Uhlenbeck process [21]):

$$\frac{dT}{dt} = -\frac{T}{\tau} + a\eta(t), \tag{2}$$

On timescales $\Delta t \ll \tau$, the feedback term is negligible and the root-mean-square fluctuation still scales as $\Delta t^{1/2}$. However, the feedback damps correlations for timescales $\Delta t \gg \tau$, and the root-mean-square fluctuation then scales as $\Delta t^{-1/2}$ (Materials and Methods). Further, aggregating multiple stabilizing feedback processes on different time scales can yield apparent power laws $\Delta T_{\mathrm{rms}} \propto \Delta t^{H}$ for any $-1/2 < H < 1/2$ ([27–29], see also Materials and Methods).

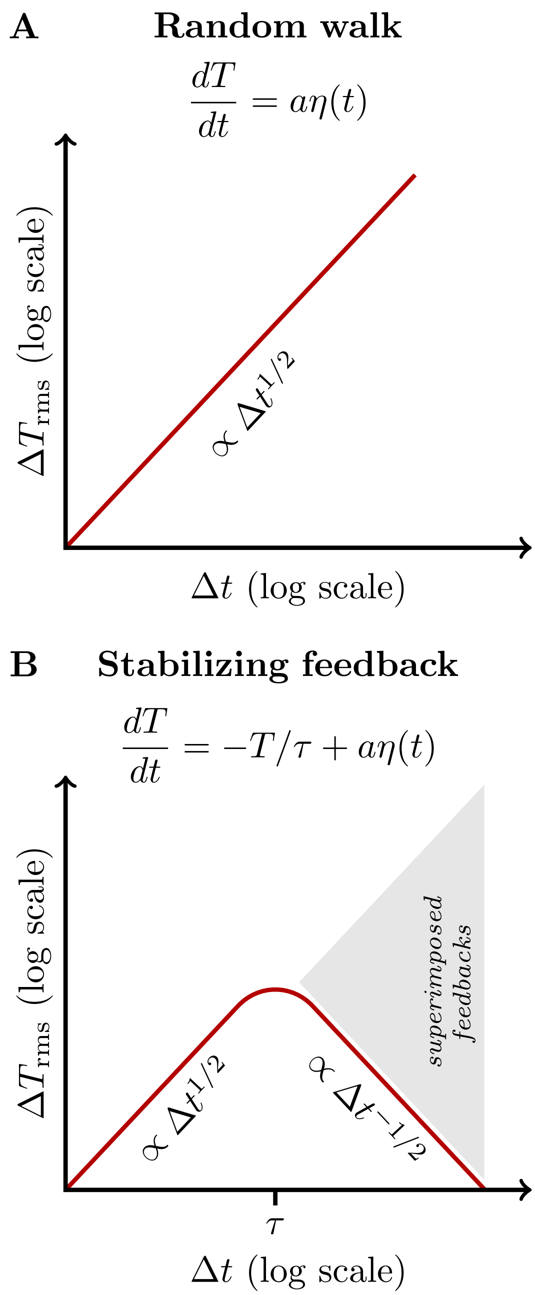

Figure 1: Two "end-member" possibilities for the simplified picture of long-term climate variability discussed in the text. **(A)** A random walk, with no stabilizing feedbacks: here, the root-mean-square temperature variation ΔT_{rms} is proportional to $\Delta t^{1/2}$. **(B)** Incorporating a stabilizing feedback on a timescale τ. Correlations on timescales larger than τ are damped, making the root-mean-square fluctuation scale as $\Delta t^{-1/2}$: i.e. shrink with timescale. Superpositions of multiple such linear feedback processes can yield $\Delta T_{\mathrm{rms}} \propto \Delta t^{H}$ with $-1/2 < H < 1/2$ (Materials and Methods).

The real Earth system is of course much more complicated than this. There are a vast range of processes on a vast range of timescales that are not explicitly accounted for. Nevertheless, as the pioneering work by Hasselmann [22] showed, in complex systems such as Earth's climate, the combined effects of many deterministic processes can be aggregated by the slower components of the system to yield statistics essentially like a random walk (Scenario A above). Thus the $\eta(t)$ in Eqs. (1) and (2) can be considered to already account for many of these processes; the explicit feedback term in Eq. (2) just means that there is a dominant stabilizing feedback on a timescale τ.

Long-term feedbacks in the real Earth system do not necessarily act directly on temperature. For example, of the two mentioned in the Introduction, the silicate weathering feedback responds directly to temperature and the carbonate compensation feedback does not. Nevertheless, if long-term temperature variability is driven at least in part by variability in atmospheric CO_2, any feedback that helps stabilize CO_2 is indirectly helping to stabilize temperature.

A final point needs to be made regarding the possibility of periodic forcings and resonances. On geologic timescales climate is forced by periodic oscillations in Earth's orbital parameters [30, 31]; such forcings, if powerful enough, could be expected to create a peak in fluctuation amplitudes similar to that in Scenario B (Figure 1). The same would be true if the Earth system had an intrinsic tendency to oscillate at a certain timescale. A case study for both would be Plio-Pleistocene glacial variability, and this will be worth addressing once we take a look at the data.

2.2 Observed temperature fluctuations on a range of timescales

We calculate the root-mean-square temperature fluctuation ΔT_{rms} as a function of timescale Δt for five different paleotemperature time series (Materials and Methods). We consider four benthic foraminiferal δ^{18}O records [32–36] and one compilation of isotopic temperatures from Antarctic ice cores [37]: between them, they resolve fluctuations on timescales spanning more than five orders of magnitude. Specifically, "fluctuations" are defined using Haar wavelets [20, 38]. Considering a time series of temperature, $T(t)$, the Haar fluctuation ΔT over a time interval Δt is defined as the difference between the average values of the time series over the first and second halves of the interval; this is described schematically in Figure 2, and discussed further in the Materials and Methods section. We use it because it is simple, accurately measures scaling behavior [38], and is straightforwardly applied to unevenly sampled paleoclimate time series [20]. It also highlights the physically important difference between fluctuations growing with scale ($H > 0$) or shrinking with scale ($H < 0$).

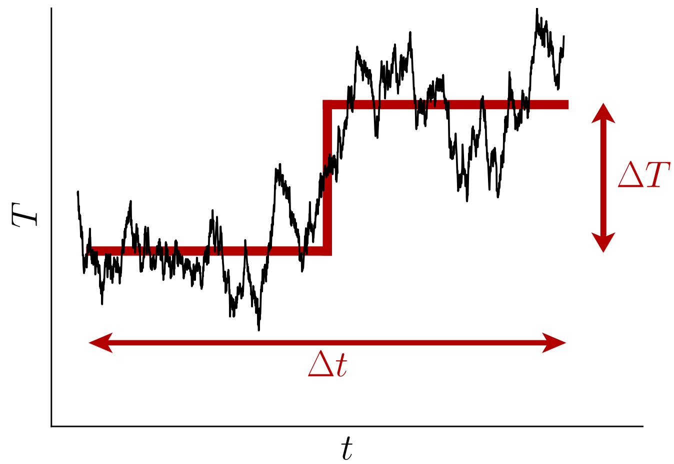

Figure 2: Quantifying the timescale dependence of fluctuation amplitudes using the Haar wavelet. The fluctuation ΔT over an interval Δt is defined as the difference between the average values of the time series over the first and second halves of the interval.

The results of our analysis are shown in Figure 3; some power law scalings (with fixed exponents H) are added as guides for interpretation. A previous analysis by Lovejoy [20] suggested the existence of three regimes that are relevant here: a "climate" regime on timescales below about 80 kyrs in which fluctuations increase with timescale, a "macroclimate" regime in which fluctuations decrease with timescale, and a "megaclimate" regime above about 500 kyrs in which fluctuations increase with timescale again. Our analysis paints a similar picture, but with some key differences.

On timescales shorter than about 4 kyrs and longer than about 400 kyrs, fluctuations increase with timescale: $H \simeq 0.5$, similar to a random walk and consistent with Scenario A. Between 4-400 kyrs, the behavior depends on what interval the data cover. Datasets that contain exclusively Plio-Pleistocene variability (i.e. the last) show a clear peak at a few tens of kyrs and a strongly decreasing regime beyond this; this forms the basis of the regime classification by Lovejoy [20] noted above. However, our analysis reveals that throughout the rest of the Cenozoic these fluctuations consistently obeyed $H \simeq 0$ — that is, their amplitude is essentially timescale-independent. The anomalous Plio-Pleistocene peak and the regime with rapidly decreasing fluctuation amplitudes beyond it likely record the rapid periodic transitions between glacial and interglacial states, rather than evidence regarding stabilizing feedbacks (see the Materials and

Methods for a further discussion).

Following the previous section and Figure 1, the fact that H is much less than 0.5 in this intermediate regime strongly suggests that stabilizing feedbacks have exerted dominant control over Earth's surface temperature on timescales between 4-400 kyrs. We emphasize how remarkable it is that the amplitude of the typical root-mean-square fluctuation in global temperature is essentially constant across two orders of magnitude in timescale! While our analysis cannot conclusively show which feedbacks were responsible, we can make inferences by comparing the timescales to those of various known or hypothesized feedbacks: this is what we will do in the Discussion. To aid this, Figure 3 also shows the approximate timescales of important Earth system feedbacks in this regime, as well as their likely signs (see Materials and Methods for details).

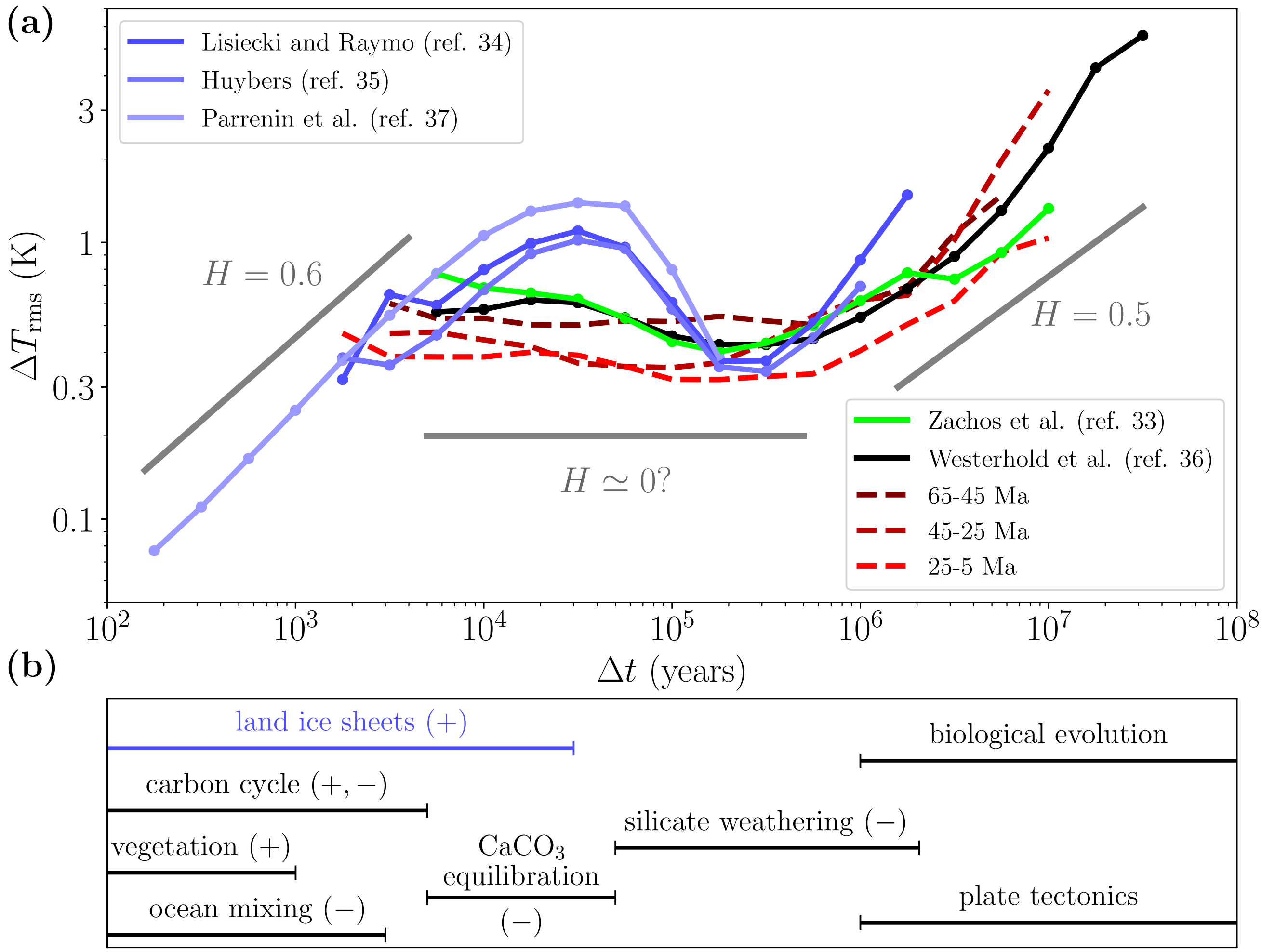

Figure 3: Temperature fluctuations and feedback mechanisms. **(a)** Root-mean-square temperature fluctuations ΔT_{rms} as a function of timescale Δt (Materials and Methods), for five different paleotemperature time series as well as three non-overlapping segments of the data from [36]. Power-law scalings with fixed exponents H are shown as guides for interpretation. On timescales below about 4 kyrs and above about 400 kyrs, fluctuations behave similarly to the random walk ($H \simeq 0.5$, Eq. 1). In contrast, fluctuations do not grow with timescale in the intermediate regime, suggesting that stabilizing feedbacks were indeed dominant here. The peak at $\sim$ 30 kyrs in the Plio-Pleistocene data, and the strongly decreasing regime beyond it, are likely signatures of glacial-interglacial variability. **(b)** Approximate timescales of relevant Earth system processes (see Materials and Methods for details). The symbols + and - indicate positive (destabilizing) and negative (stabilizing) feedbacks, respectively. The land ice sheet feedback is colored blue to emphasize that it is primarily relevant only after the onset of Northern Hemisphere glaciation $\sim$ 3 Ma ago.

2.3 Variability in a system with multiple partial feedbacks

To make clear how multiple feedbacks in a complex system can create a regime with timescale-independent ΔT_{rms} as in Figure 3, and to help develop a more specific interpretation of the three regimes shown in the data, we expand on the stochastic models discussed earlier. Specifically, we consider Earth's surface temperature T to be the sum of multiple stochastic processes, some with stabilizing feedbacks (e.g. Scenario B) and some without (Scenario A). Mathematically, we let

$$\Delta T(t) = \left(\sum_{i}^{n-1} f_i(t) \right) + r(t), \tag{3}$$

where $\dot{f}_i = -f_i/\tau_i + a_i \eta_i(t)$ and $\dot{r}(t) = a_n \eta_n$, and the η_i are independent Gaussian white noise forcings (discussed further in the Materials and Methods). Finally, $a_n < a_i$ for all $i < n$, meaning that variability due to the random walk $r(t)$ grows more slowly than that of the other processes. A key property of this model is that the stabilizing feedbacks have only partial control — in other words, they only stabilize part of the system, and there can still be undamped variability at other scales. The real Earth system shares this property: if it did not, paleoclimate records would exhibit no variability at all on long timescales.

As an example, we choose partial stabilizing feedbacks on timescales of 1, 10, and 100 kyrs (τ_1, τ_2, and τ_3, respectively), numerically simulate Eq. (3) for 200 Myrs, and analyze fluctuations using the same algorithm that we applied to the real data. Results are shown in Figure 4; the general behavior of the observations is well-reproduced. On short timescales ($< \tau_1$) fluctuations grow like a random walk with $H \simeq 0.5$, and then have essentially timescale-independent amplitudes in the regime in which the feedbacks are active. On long timescales ($> \tau_3$), the undamped stochastic variability (reflecting the partial nature of the feedbacks) takes over, and fluctuations again grow like a random walk. Theory predicts that this kind of behavior occurs for a wide range of possible models and parameter values (Materials and Methods): in all cases the position of the intermediate regime is determined by the range of timescales of stabilizing feedbacks.

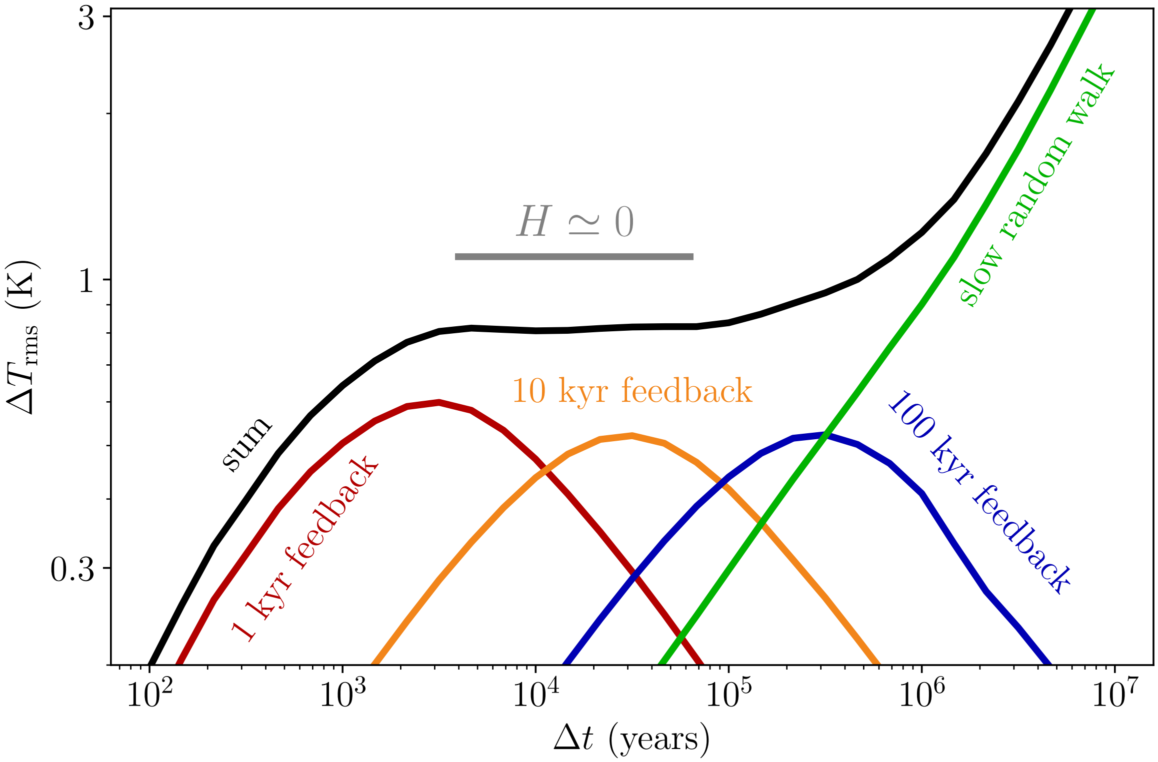

Figure 4: A system with multiple partially stabilizing feedbacks can display the same behavior observed in the data. In our simple conceptual model, Earth's surface temperature T is given by the sum of some stochastic processes with stabilizing feedbacks and some without. Here, we consider feedbacks on timescales of 1, 10, and 100 kyrs, as well as a slow random walk with no feedbacks: results from numerical simulation give remarkable agreement with the observed scaling behavior (Figure 3). Theory predicts similar behavior for a wide range of possible models and parameter values (Materials and Methods). The 'sum' curve is multiplied by a constant for clearer visualization.

3 Discussion

We have calculated the typical amplitude of past global temperature fluctuations on a range of timescales, and have shown that its behavior should reflect the relative dominance or lack of dominance of stabilizing Earth system feedbacks in different timescale regimes. We have identified a regime between about 4-400 kyrs in which fluctuations fail to grow with timescale — consistent with dominant stabilizing feedbacks — and a regime beyond 400 kyrs in which they do — consistent with no dominant stabilizing feedbacks. We now proceed to interpret these observations in light of physical and biogeochemical processes.

3.1 Long-term climate stabilization: confirmed

The identification of the anomalous 4-400 kyr regime is a novel confirmation that stabilizing feedbacks with characteristic timescales in this regime have indeed been a dominant control on Earth's surface temperature. To understand which mechanisms were likely responsible, we can compare this timescale range to the previously proposed timescales for different stabilizing feedbacks.

Of immediate interest is the consistency of this regime with the $\sim$100-kyr timescale proposed for the silicate weathering feedback [7, 8]. We suggest that this is strong observational evidence for the importance of silicate weathering as a climate stabilizer. Through this, it further supports the widely used steady-state assumption [4], existing models of the long-term effects of anthropogenic CO_2 emissions [6, 8], and the idea that the weathering feedback should play a key role in planetary habitability [5].

The fact that the non-growing regime seems to start at timescales as small as 4 kyr suggests that other, shorter-timescale stabilizing feedbacks were also important. One obvious candidate is ocean mixing: the ocean can help damp temperature fluctuations due to its large thermal inertia, and full equilibration is achieved on a timescale of a few kyrs [39]. Another possibility on a $\sim$10 kyr timescale is $CaCO_3$ equilibration [6, 8, 9], which could indirectly stabilize temperature through its effect on atmospheric CO_2. Other feedbacks potentially active at this timescale include vegetation and land ice (see Figure 3); however, these are likely both destabilizing (mathematically positive) feedbacks [40], and as such would not have been responsible for stabilization.

3.2 Beyond stabilization: weathering as a climate forcing?

What is the origin of the increasing fluctuation amplitude beyond 400 kyrs? Following the theory explained above, the random-walk-like growth ($H \simeq 0.5$) should mean that there are are no dominant stabilizing feedbacks in the system on these timescales. Yet, if current thinking is at all accurate, the silicate weathering feedback should still be active on these timescales: it is not inherently timescale-limited. What then is going on?

One possible resolution is the following. Consider Earth's "weathering curve" [41], interpreted here as the dependence of the silicate weathering flux, F_{si}, on Earth's surface temperature. Neglecting changes in organic carbon oxidation or burial, a steady state is established when F_{si} is equal to the volcanic flux F_{volc}

of carbon into the surface environment. Because the weathering curve has a positive slope (weathering increases with temperature), we obtain the familiar stabilizing feedback that tends to drive the system towards a steady state.

Nevertheless, the weathering curve itself may change over time [41], either due to changes in the surface carbon cycle's physical attributes (such as the amount and properties of exposed weatherable rock [13, 42–44]) or in the mechanisms constituting the feedback itself (e.g. due to biological evolution in land plants [1]); see also panel (b) of Figure 3. This will lead to slow "quasistatic" changes in the surface temperature [45], even while the carbon cycle remains in steady state with respect to input and output fluxes. We suggest that it is precisely this class of changes that lead to fluctuations increasing again at the longest timescales.

Figure 5 summarizes this schematically. Imagine that the weathering curve moves upwards, for example due to an increase in weatherability; then, the new steady state will move to a lower surface temperature. On timescales of hundreds of kyrs, the weathering feedback will damp fluctuations towards the steady state. Yet, on longer timescales, weathering will act as a forcing, and the steady state itself will move. The $H \simeq 0.5$ scaling beyond 400 kyrs then suggests that the steady state moves in an undamped way. In other words, while silicate weathering is a stabilizing feedback bringing the system to a steady state, there are no stabilizing feedbacks on the Myr-timescale motion of that steady state itself.

This is of course a highly simplified picture of weathering. We have ignored the effects of changes in organic carbon oxidation and burial, and are considering factors such as CO_2, topography, vegetation types, precipitation and rock types only implicitly (by arguing that they change the weathering curve). Nevertheless, we suggest that the basic reasoning regarding a weathering-established-steady state that moves in an undamped way is likely independent of these details. This could and should be tested using a more detailed carbon-cycle model.

Finally, there is one other possibility that deserves mention: that the increasing fluctuation amplitudes at the longest timescales are due to other destabilizing feedbacks. While there are no obvious candidate mechanisms for such feedbacks on multi-Myr timescales, the data at present cannot rule this out. This would also be very interesting to pursue further.

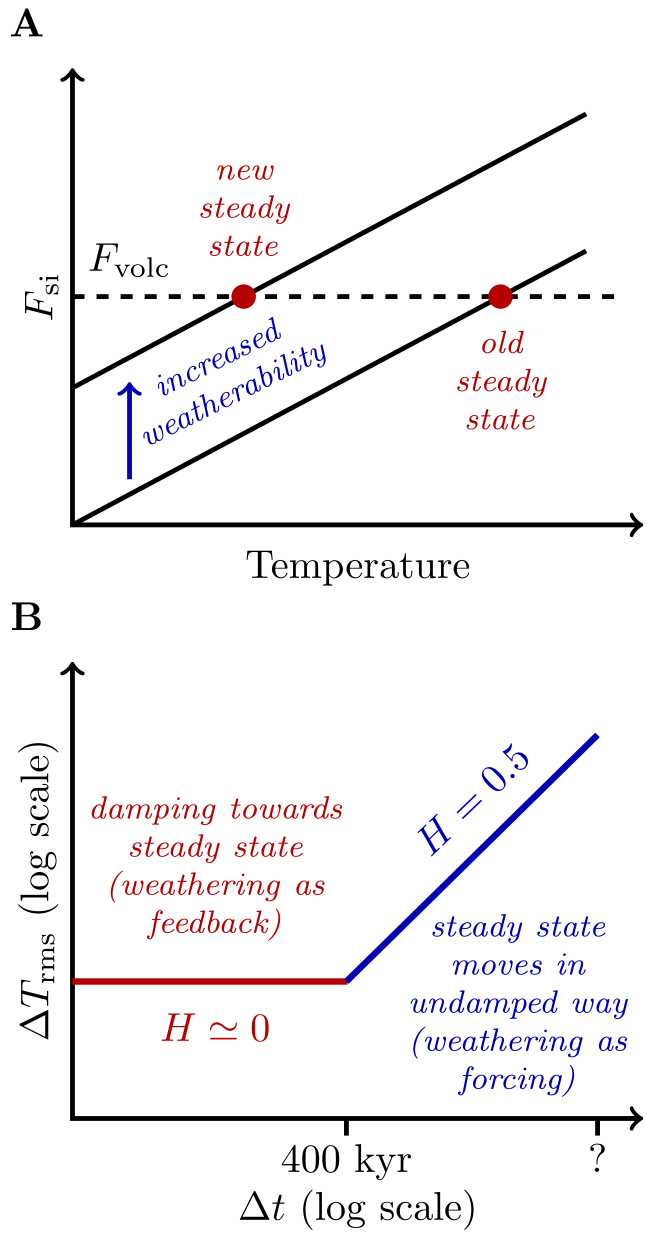

Figure 5: The observation that fluctuation amplitudes increase like a random walk again beyond 400 kyrs, even though the silicate weathering feedback remains active, could potentially be understood as follows. **(A)** Considering Earth's "weathering curve", it is clear how changes such as an increase in weatherability can move the steady state that silicate weathering establishes. **(B)** On timescales below about 400 kyrs, silicate weathering acts as a feedback, driving the system towards a steady state. On longer timescales, the steady state itself moves, and weathering acts as a forcing. There is still damping towards the steady state; the key point is that there is no damping on the motion of the steady state itself.

3.3 Earth's long-term habitability, and the role of chance

The fact that global temperature fluctuations continue to grow like a random walk at the longest timescales has major implications for understanding the long-term habitability of Earth and other Earth-like planets. There is a long-standing debate [14–16] over the extent to which Earth's observed billion-year habitability is a product of stabilization (for example due to the weathering feedback), or a product of chance. The predominant view has been that the weathering feedback is responsible for this long-term habitability, and indeed such stabilization is a key part of the "habitable zone" concept used to search for life on other planets [5, 46].

We have shown that the observations are inconsistent with a dominant stabilizing feedback on the longest timescales, and suggested that those fluctuations arise due to weathering acting as a climate forcing (for example when tectonic processes change the availability of weatherable rocks). Another option is that fluctuations grow on long timescales because of unknown destabilizing feedbacks. In either case, the key question is: Are there any mechanisms in the Earth system that prevent these kinds of fluctuations from eventually driving surface temperature into an uninhabitable regime? If there are none, it would follow that chance may have played a non-negligible role in Earth's continued habitability, and that other Earth-like planets with an active carbonate-silicate cycle and in the conventional "habitable zone" may not necessarily be as accommodating to life over long periods of time as has previously been expected. Obtaining and analyzing well-calibrated, higher-resolution paleotemperature records spanning longer stretches of geologic time, as well as improving our understanding of tectonic evolution and its climatic consequences on timescales of many millions of years [47], should provide further insights.

Materials and Methods

Scaling in time series

It has long been recognized that climate time series on various timescales exhibit self-similar "scaling" behavior [20, 22, 24, 25, 48, 49]. A process $x(t)$ is considered to exhibit self-similarity if

$$x(t) \overset{d}{=} a^{-H} x(at), \tag{4}$$

where H is the self-similarity exponent and $\overset{d}{=}$ denotes equality in terms of probability distribution. For such processes, the power spectrum $S(\omega) \propto \omega^{-\beta}$, where $\beta \simeq 2H + 1$ [50]. When $\beta \simeq 1$ (i.e. $H \simeq 0$), this is the widely-studied "$1/f$ noise" [51]. Observed climate time series often exhibit well-defined timescale regimes in which β and H take on different values [20, 49].

To begin to understand the physical origin of such scaling, a simple null model without feedbacks considers climate fluctuations as a random walk [22]. In the limit of infinitesimal step sizes, this is the Wiener process [21], which has probability distribution

$$p(x,t) = \frac{1}{\sqrt{2\pi(t-t_0)}} \exp\left(-\frac{(x-x_0)^2}{2(t-t_0)}\right). \tag{5}$$

Relating this to Eq. 4, one can show that this process is self-similar with $H = 1/2$ (i.e. $\beta = 2$). In contrast, white noise, which is the long-time limit of Eq. 2, has $\beta = 0$ [21] (i.e. $H = -1/2$).

Superposing multiple processes with stabilizing feedbacks at different timescales can create apparent scaling exponents in the range $-1/2 < H < 1/2$ [27]. For the benefit of the reader the Supplementary Text demonstrates this explicitly. In the literature on "$1/f$ noise", $H = 0$ has also been associated with a continuous log-uniform distribution of feedback timescales [28, 29].

Self-similarity in real data can be measured either in real space or in frequency space. In real space, this can be done through a fluctuation function $\Delta x(\Delta t)$ ("structure function" in the field of turbulence [52]). Ideally, this function would obey

$$\langle [\Delta x(\Delta t)]^q \rangle \propto \Delta t^{qH}. \tag{6}$$

One possible choice of fluctuation function is the simple difference

$$\Delta x(t, \Delta t) = x(t + \Delta t) - x(t), \tag{7}$$

but this only accurately reflects scaling behavior (i.e. behaves according to Eq. 7) in the regime $0 < H < 1$ [38]. In this work, we define $\Delta x(\Delta t)$ as the Haar fluctuation: the difference between the time series averaged over the first and second halves of the interval Δt [38, 53]. This accurately reflects scaling behavior in the range $-1 < H < 1$ [38], and is additionally desirable because of its conceptual and computational simplicity. In particular, it is straightforward to measure the scaling behavior for unevenly sampled time series without interpolation. The details of the algorithm we use are described below in ”Algorithm for calculating fluctuation timescale dependence”.

Paleoclimate temperature datasets

We analyze four benthic δ^{18}O datasets: the Cenozoic compilations first introduced by Zachos et al. [32] and updated in ref. [33], the Cenozoic composite record of Westerhold et al. [36], the orbitally-tuned Plio-Pleistocene compilation of Lisiecki and Raymo [34], and the non-orbitally tuned Pleistocene compilation of Huybers [35]. Temperature values are inferred from δ^{18}O following the calibrations proposed by Hansen et al. [54] and slightly redefined by Westerhold et al. [36], which take into account changes in the ice volume contribution to δ^{18}O, and the relationship between deep-ocean and surface temperatures throughout the Cenozoic. This conversion is based on absolute δ^{18}O data and thus the Huybers [35] time series, which has long-term averages removed, has to have a constant offset added to match the others (although the specific value does not affect our results). Deep-ocean temperature T_{do}, following the calibrations discussed above, is given by

$$T_{\mathrm{do}} = \begin{cases} -4\delta^{18}\mathrm{O} + 12, & \text{67-34 Ma} \\ 5 - 8(\delta^{18}\mathrm{O} - 1.75)/3, & \text{34-3.6 Ma} \\ 1 - 4.4(\delta^{18}\mathrm{O} - 3.25)/3, & \text{3.6 Ma-present.} \end{cases} \tag{8}$$

Surface temperature T is then calculated from the deep-ocean temperature according to

$$T = \begin{cases} T_{\mathrm{do}} + 14.15, & \text{67-5.33 Ma} \\ 2.5T_{\mathrm{do}} + 12.15, & \text{5.33-1.81 Ma} \\ 2T_{\mathrm{do}} + 12.25, & \text{1.81 Ma-present.} \end{cases} \tag{9}$$

While the above conversion is used for the sake of accuracy, it is worth noting that the qualitative results remain the same even if a single linear conversion constant relating δ^{18}O and T is used for all datasets.

Finally, we also include in our analysis the ice core temperature compilation of Parrenin et al. [37]. We divide the fluctuations by a factor of 2 to approximately account for high-latitude amplification [20], although again it is important to emphasize that the details of this correction do not affect our conclusions.

Algorithm for calculating fluctuation timescale dependence

We calculate the root-mean-square fluctuation for each paleotemperature time series, ΔT_{rms}, using an interpolation-free algorithm based on that proposed by Lovejoy [20]: all of our code is made freely available at [55] and https://github.com/arnscheidt/stabilizing-earth-system-feedbacks. For an unevenly sampled time series described by two vectors $\mathbf{t} = [t_0, t_1, t_2 ... t_n]$ and $\mathbf{T} = [T_0, T_1, T_2 ... T_n]$, we define the Haar fluctuation at position j of size k:

$$\Delta T(j,k) = \left(\frac{2}{k} \sum_{i=j+k/2}^{j+k-1} T_i \right) - \left(\frac{2}{k} \sum_{i=j}^{j+k/2-1} T_i \right), \tag{10}$$

i.e. the difference between the average value of the second $k/2$ and the first $k/2$ data points. This can be implemented efficiently using cumulative sums.

We consider all even k ranging from 0 to n. Although we could calculate all possible $\Delta T(j,k)$ for the data, $\Delta T(j_1,k)$ is negligibly different from $\Delta T(j_2,k)$ as long as $k \gg |j_1 - j_2|$. Therefore, for each k we calculate $\Delta T(j,k)$ at intervals of ak, where we choose $a = 0.5$: this gives the algorithm $n \log n$ time complexity instead of n^2. Because the interval k is divided in two in terms of indices, but not in terms of time elapsed, we discard any $\Delta x(j,k)$ with $\epsilon < \frac{t_{j+k/2} - t_j}{t_{j+k} - t_j} < 1 - \epsilon$ for some ϵ: the choice of ϵ defines a balance between robustly allowing for unevenly spaced data but ensuring that the extracted $\Delta T(\Delta t)$

remain meaningful. Following Lovejoy [20], we use $\epsilon = 0.25$.

Finally, we calculate $\Delta t = t_{j+k} - t_k$ for each $\Delta T(j,k)$, and average the $\Delta T(\Delta t)^2$ over evenly spaced bins in log-space (4 bins per order of magnitude). The data points in Figure 3 are the bin centers, and the root-mean-square fluctuation $\Delta T_{\rm rms}$ is given by taking the square root of the averaged $\Delta T(\Delta t)^2$. At the extremes (very small or large Δt), there begin to be much fewer data points; we therefore truncate the data where the number of data points per bin are a factor b smaller than the maximum (we use $b = 5$).

For the purposes of Figure 3, we have additionally truncated some of the shortest-timescale fluctuations ($\Delta t < 4$ kyr) from the data of Zachos et al. [33] and Westerhold et al. [36]. They are anomalously large compared to those in the more recent higher-resolution datasets, as well as the results when the 20-Myr segments of the Westerhold et al. [36] data are studied individually. The latter do not consider the time interval from 5 Ma-present, suggesting that this effect arises only due to the data in that interval. Meanwhile, the higher-resolution datasets from this same period (Lisiecki and Raymo [34], Huybers [35], and Parrenin et al. [37]) represent averages over multiple records, while the Zachos et al. [33] and Westerhold et al. [36] data do not. Therefore, we suggest that the anomaly probably reflects contributions to the δ^{18}O signal from sources other than global temperature (regional-scale variability, diagenesis, etc.), which are not of interest here. The 20-Myr segments shown in Figure 3 have not been additionally truncated in this manner, and so none of this affects our conclusions.

Anomalous peak in Plio-Pleistocene data

Figure 3 shows that the datasets spanning only the Plio-Pleistocene ($\sim$5 Ma - present) [34, 35, 37] behave somewhat differently in the intermediate regime than the datasets spanning the entire Cenozoic [32, 36]: there is a more pronounced peak at timescales of a few tens of kyrs, and a regime in which fluctuations decrease very rapidly. What is the origin of this difference? A likely solution is that it simply reflects the onset of the Plio-Pleistocene glacial cycles, which feature dramatic transitions between different climate states on timescales of tens of kyrs. This variability would produce an anomalous peak in the averaged fluctuation amplitudes near this timescale, and thus correspondingly steeper slopes on either sides of the peak.

Repeated periodic transitions such as the glacial cycles are a confounding effect when attempting to observe signatures of stabilizing feedbacks in the data. Nevertheless, we know that this kind of behavior is limited to the Plio-Pleistocene: there should be no such confounding effect in other time periods. As shown in Figure 3, the data from the rest of the Cenozoic show the same consistent pattern of fluctuations not growing between timescales of about 4-400 kyr, providing strong evidence of dominant control by stabilizing feedbacks.

Timescales of long-term Earth system feedbacks

The long-term Earth system feedbacks and their timescales shown in Figure 3 are loosely taken after Rohling et al. [56], who made distinctions between annual, decadal, century, millenial, multimillenial, and Myr timescales. We have also included the signs of the feedbacks (i.e. positive/destabilizing or negative/stabilizing) where this is clear (following, e.g. [4, 6, 7, 40, 56]). The timescales are still only intended as approximate; here we offer some additional justification for the more specific values shown. We emphasize that the upper timescale limit for a given feedback does not mean the underlying process is not operating on longer timescales: it simply means that the feedback has reached steady state.

The upper limit of the land ice feedback can be taken approximately as the timescale on which Plio-Pleistocene deglacations occur (10^4 years). More specifically, we can take it as the timescale at which we see the peak in the Plio-Pleistocene temperature fluctuations (i.e. 3×10^4 years.) The upper limit of the vegetation feedback is taken approximately as 10^3 years, based on the observation of strong climate-vegetation correlations on millenial timescales [57]. The upper limit of the stabilizing feedback due to ocean mixing is taken as a few kyrs, based on simulations showing that substantial tracer disequilibrium can persist for at least 2000 years [39]. Finally, we split carbon cycle feedbacks into three categories: those on timescales shorter than 5kyr (which may be stabilizing or destabilizing), $CaCO_3$ equilibration from 5-50kyr [6, 8], and silicate weathering operating between 50 ky and a few million years [4, 7, 8].

Stochastic models of temperature variability

To make concrete the relationship between stabilizing feedbacks and the timescale dependence of fluctuations as shown in Figure 3, we consider stochastic models of long-term temperature variability. Stochastic models have long been employed in the study of glacial cycles [24], but have only recently begun to be applied to deep time climate problems [58–60].

Considering Eq. (3), for $n = 2$ (i.e. one stabilizing feedback), one can show that (Supplementary Text)

$$\Delta T_{\mathrm{rms}} \propto \begin{cases} \Delta t^{1/2}, & \text{if } \Delta t \ll \tau_1 \\ \Delta t^{-1/2}, & \text{if } \tau_1 \ll \Delta t \ll a_1\tau/a_2 \\ \Delta t^{1/2}, & \text{if } \Delta t \gg \tau_s. \end{cases} \tag{11}$$

This already begins to qualitatively reproduce the three regimes seen in the data. For $n > 2$ (i.e. with more stabilizing feedback processes), things become more complicated; yet, it is not hard to obtain behavior like

$$\Delta T_{\mathrm{rms}} \propto \begin{cases} \Delta t^{1/2}, & \text{if } \Delta t \ll \tau_1 \\ \Delta t^{H} \text{ with } -1/2 < H < 1/2, & \text{if } \tau_1 \ll \Delta t \ll \tau_s \\ \Delta t^{1/2}, & \text{if } \Delta t \gg \tau_s. \end{cases} \tag{12}$$

The first crossover timescale τ_1 is that of the fastest feedback, and the slow crossover timescale τ_s is typically determined by the feedback with the slowest timescale together with the amplitude of the slow random walk $r(t)$. Depending on the choice of parameters, it is now possible to reproduce any behavior of the kind seen in Figure 3.

For the example in Fig. 4, $n = 4$, τ_1=1 kyr, τ_2=10 kyr, τ_3=100 kyr, $a_1 = 0.03$ K yr$^{-1/2}$, $a_2 = 0.0085$ K yr$^{-1/2}$, $a_3 = 0.0027$ K yr$^{-1/2}$, $a_4 = 0.0015$ K yr$^{-1/2}$. As can be seen in the figure, the intermediate regime has $H \simeq 0$. The 200-Myr numerical simulation was carried out using an Euler-Maruyama algorithm and the Julia package DifferentialEquations.jl [61].

We emphasize that while we have chosen the noise sources η_i to be uncorrelated for simplicity, our results should in principle be independent of this. For example, the result that multiple stochastic feedback processes can be superimposed to create scaling regimes with $-1/2 < H < 1/2$ holds both for uncorrelated noise sources (Supplementary Text) and for correlated ones [27].

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Author contributions

C.W.A. developed theory, implemented numerical models, and analyzed data all with input from D.H.R.; and C.W.A. and D.H.R. wrote the paper.

Data and code availability

This work generated no new data. Code to replicate the results in this paper is freely available on Zenodo [55] or at https://github.com/arnscheidt/stabilizing-earth-system-feedbacks.

Competing interests

The authors declare no competing interests.

Supplementary information

Materials and Methods

Supplementary Text

References [62].

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