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Shortwave and longwave radiative contributions to global warming under increasing CO2

Aaron Donohoe, Kyle C. Armour, Angeline G. Pendergrass, and David S. Battisti

Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139; Advanced Study Program, National Center for Atmospheric Research, Boulder, CO 80307; and Department of Atmospheric Sciences, University of Washington, Seattle, WA 98195

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In response to increasing concentrations of atmospheric CO2, high-end general circulation models (GCMs) simulate an accumulation of energy at the top of the atmosphere not through a reduction in outgoing longwave radiation (OLR)—as one might expect from greenhouse gas forcing—but through an enhancement of net absorbed solar radiation (ASR). A simple linear radiative feedback framework is used to explain this counterintuitive behavior. It is found that the timescale over which OLR returns to its initial value after a CO2 perturbation depends sensitively on the magnitude of shortwave (SW) feedbacks. If SW feedbacks are sufficiently positive, OLR recovers within merely several decades, and any subsequent global energy accumulation is because of enhanced ASR only. In the GCM mean, this OLR recovery timescale is only 20 yr because of robust SW water vapor and surface albedo feedbacks. However, a large spread in the net SW feedback across models (because of clouds) produces a range of OLR responses; in those few models with a weak SW feedback, OLR takes centuries to recover, and energy accumulation is dominated by reduced OLR. Observational constraints of radiative feedbacks—from satellite radiation and surface temperature data—suggest an OLR recovery timescale of decades or less, consistent with the majority of GCMs. Altogether, these results suggest that, although greenhouse gas forcing predominantly acts to reduce OLR, the resulting global warming is likely caused by enhanced ASR.

Global conservation of energy is a powerful constraint for understanding Earth’s climate and its changes. Variations in atmospheric composition that result in a net positive energy imbalance at the top of atmosphere (TOA) drive global warming, with the world ocean as the primary reservoir for energy accumulation (1). In turn, increasing global surface temperature enhances emission of longwave (LW) radiation to space (the Planck response). A schematic of the global energy budget response to a step change in greenhouse gas (GHG) concentrations is illustrated in Fig. L4: outgoing LW radiation (OLR) initially decreases because of enhanced LW absorption by higher GHG levels; as energy accumulates in the climate system, global temperature rises and OLR increases until the TOA energy balance is restored—when OLR once again balances the net absorbed solar radiation (ASR). In this canonical view of global warming, the net energy accumulation (shaded green area in Fig. L4) is a consequence of decreased OLR driven by GHG forcing. In contrast, consider a hypothetical step change in solar insolation (Fig. 1B): ASR is increased, and energy accumulates until the climate warms sufficiently that OLR balances the ASR perturbation. In this case, the net energy accumulation (shaded red area in Fig. 1) is a consequence of increased ASR and opposed by the increased OLR (hatched green area in Fig. 1).

Is the present global warming caused by reduced OLR (as in Fig. L4) or enhanced ASR (as in Fig. 1B)? Anthropogenic radiative forcing is dominated by LW active constituents, such as CO2 and methane, and shortwave (SW) forcing agents, such as sulfate aerosols, are thought to be acting to reduce ASR compared with their preindustrial levels (2). Reduced OLR, thus, seems the likely cause of the observed global energy accumulation, although the limited length of satellite TOA radiation measurements precludes determination of the relative contributions of ASR and OLR by direct observation. Trenberth and Fasullo (3) considered global energy accumulation within the ensemble of coupled general circulation models (GCMs) participating in phase 3 of the Coupled Model Intercomparison Project (4) (CMIP3). They report that, under the Special Report on Emission Scenarios A1B emissions scenario, wherein increasing radiative forcing is driven principally by increasing GHG concentrations, OLR changes little over the 21st century and global energy accumulation is caused nearly entirely by enhanced ASR—seemingly at odds with the canonical view of global warming by reduced LW emission to space (outlined in Fig. L4).

Here, we seek insight into this surprising result. In particular, we examine CO2-only forcing scenarios as simulated by the CMIP5 ensemble of state of the art GCMs (5). Perturbing CO2 alone permits a clean partitioning of radiative forcing and radiative response into their respective SW and LW components and allows an investigation into the relative contributions of reduced OLR and enhanced ASR to global energy accumulation. The CMIP5 multi-GCM mean response to a compounding 1% per year CO2 increase (hereafter, 1% CO2) is shown in Fig. 1D. Although CO2 radiative forcing increases approximately linearly in time for 140 y (dotted lines in Fig. 1D), OLR changes little from its preindustrial value, and global energy accumulation is accomplished nearly entirely by increased ASR, consistent with the multi-GCM mean results in the work Trenberth and Fasullo (3). Perhaps even more striking is the response to an abrupt quadrupling of CO2 (hereafter, 4x CO2), which is shown in Fig. 1C: OLR initially decreases, like in Fig. L4, but recovers to its unperturbed (preindustrial) value within only two decades.

Significance

The greenhouse effect is well-established. Increased concentrations of greenhouse gases, such as CO2, reduce the amount of outgoing longwave radiation (OLR) to space; thus, energy accumulates in the climate system, and the planet warms. However, climate models forced with CO2 reveal that global energy accumulation is, instead, primarily caused by an increase in absorbed solar radiation (ASR). This study resolves this apparent paradox. The solution is in the climate feedbacks that increase ASR with warming—the moistening of the atmosphere and the reduction of snow and sea ice cover. Observations and model simulations suggest that even though global warming is set into motion by greenhouse gases that reduce OLR, it is ultimately sustained by the climate feedbacks that enhance ASR.


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*To whom correspondence should be addressed. Email: thedhoe@mit.edu.

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To interpret these findings, we employ a commonly used linearization of the global TOA energy budget:

\[
d\langle C T_S \rangle/dt = F_{SW} + F_{LW} + (\lambda_{SW} + \lambda_{LW}) T_S,
\]

where \(T_S\) is the global mean surface temperature anomaly, and \(C\) is the time-dependent global heat capacity. Eq. 1 relates the rate of global heat content change to the rate of global TOA energy accumulation, which is given by the sum of SW and LW radiative forcings (\(F_{SW}\) and \(F_{LW}\)) and radiative responses (\(\lambda_{SW} T_S\) and \(\lambda_{LW} T_S\)) (6). Anomalies in OLR and ASR can further be expressed as

\[
\text{ASR} = F_{SW} + \lambda_{SW} T_S
\]

and

\[
-\text{OLR} = F_{LW} + \lambda_{LW} T_S.
\]

The radiative feedbacks (\(\lambda_{SW}\) and \(\lambda_{LW}\)) can be estimated for each GCM by linear regression of ASR and OLR (Fig. 2 C and D) with \(T_S\) (Fig. 2A) over the period after 4× CO\(_2\), wherein radiative forcing is approximately constant (7, 8). Moreover, the LW and SW components of CO\(_2\) forcing (\(F_{LW}\) and \(F_{SW}\)) can be estimated by the \(T_S = 0\) intercept of the regression.\(^1\) Forcing and feedback values for the CMIP5 GCMs (Table S1) are consistent with those estimated by Andrews et al. (10).

As defined by Eq. 1, the effective heat capacity \(C\) (Fig. 2B) is the time-integrated TOA energy accumulation divided by \(T_S\). It has long been recognized that there is no single heat capacity (or characteristic relaxation time) of the climate system (11). Indeed, \(C\) increases with time as heat penetrates below the surface mixed layer and into the ocean interior (12–15). For the CMIP5 GCMs, \(C\) corresponds to an equivalent ocean depth of 50 m in the first decade after 4× CO\(_2\) and increases over time, reaching an equivalent depth of several hundred meters after a century (Fig. 2B). The time evolution of \(C\) together with values of SW and LW feedbacks and forcing permit an iteration of Eq. 1 that accurately reproduces the surface temperature response \(T_S\) of each GCM (Fig. 2A). ASR and OLR predicted by Eq. 2 are in excellent agreement with their respective responses following 4× CO\(_2\) (Fig. 2 C and D) and account for the vast majority (99%) of the variance in \(T_{\text{OLR}}\) across the models. Thus, a simple representation of climate feedbacks (Eqs. 1 and 2) is all that is needed to understand the response of ASR and OLR under GHG forcing.

Insight into the GCM behavior can be gained by considering the values of ASR and OLR required to reach TOA energy balance (equilibrium) with an imposed GHG forcing. If forcing and feedbacks acted only in the LW (as in Fig. 1A), the OLR anomaly would increase from a value of \(-F_{LW} = 0\) after 4× CO\(_2\) (Eq. 2), and global energy accumulation would be driven entirely by reduced OLR. In the multi-GCM mean, however, there is a substantial positive SW feedback of \(\lambda_{SW} = 0.6\) \(\text{W m}^{-2} \text{K}^{-1}\) in addition to the negative LW feedback of \(\lambda_{LW} = -1.7\) \(\text{W m}^{-2} \text{K}^{-1}\) (Fig. 3A). As a result, ASR increases with warming, contributing to global energy accumulation. Moreover, the positive \(\lambda_{SW}\) amplifies the equilibrium temperature response by a gain factor\(^2\) (\(G_{\lambda_{SW}}\)) of \(\sim 1.5\) relative to a system with LW feedbacks only, where

\[
G_{\lambda_{SW}} \equiv 1/(1 + \lambda_{SW}/\lambda_{LW}).
\]

The multi-GCM mean OLR must, therefore, increase by 1.5\(F_{LW}\) after 4× CO\(_2\) (from \(-F_{LW}\) to 0.5\(F_{LW}\)) to reach equilibrium

\(^1\)Radiative forcing by this method includes both the direct radiative forcing by the GHG and the effect of any tropospheric adjustments that occur on timescales of days to weeks (8).

\(^2\)We note that this gain factor differs from the commonly used feedback gain defined as the amplification of the equilibrium temperature response by radiative feedbacks (e.g., water vapor and surface albedo) relative to the response with the Planck feedback only (16, 17).

\*Note that, if OLR remains below its unperturbed value for the entirety of the 150-y simulation, we estimate \(T_{\text{OLR}}\) by linear extrapolation over the final century of the simulations. In this case, \(T_{\text{OLR}}\) would be considered a metric for the GHG forcing ameliorated by the response, because it is possible that OLR may never return to its unperturbed value.
66% of the equilibrium temperature response \( \tau \) is underestimated in these models because we have \( \tau = - \frac{C}{\lambda_{SW} + \lambda_{LW}} \) approaches one, leading to a bimodal forcing, which. There are a few notable exceptions, \( \lambda_{SW} \) provides an upper bound on its value over the first several decades. \( \tau_{cross} \) is far more sensitive to changes in \( \lambda_{SW} \) than \( \lambda_{LW} \) over the parameter space realized in the GCMs (curves in Fig. 3A). From Eq. 4, the \( \sim 66\% \) of the equilibrium temperature change required for OLR to recover to preindustrial values will be achieved at approximately time \( \tau \); that is, \( \tau_{cross} \) is approximately equal to \( \tau \) in the ensemble average. If we take the ensemble mean of \( C \) over the first century of the 4x CO2 simulations as an upper bound on its value over the first several decades (\( C \approx 250 \) m from Fig. 2B), then Eq. 5 provides an upper bound on \( \tau \). For ensemble mean feedback values (Table S1), Eq. 5 gives \( \tau \approx 29 \) y, which is in good agreement with the CMIP5 ensemble mean OLR recovery timescale \( \tau_{cross} = 19 \) y. For all times after \( \tau_{cross} \), energy is lost through enhanced LW emission, and energy accumulation is solely due to enhanced ASR. Thus, the relative contributions of SW and LW anomalies to the total energy accumulation depend directly on the time that it takes for OLR to return to and cross its unperturbed value (\( \tau_{cross} \)). In the multi-GCM mean, OLR takes only two decades to recover, and thus, energy accumulation is due primarily to enhanced ASR.

What, then, sets the large range of \( \tau_{cross} \) across the CMIP5 GCMs? While a substantial fraction of equilibrium warming is achieved within the first several decades in all GCMs (15, 18)—due to the fast response of the surface components of the climate system (12)—the ASR and OLR responses to warming (and \( \tau_{cross} \)) depend on the SW and LW feedbacks, which vary substantially (Fig. 3A). The dependence of \( \tau_{cross} \) on the feedback parameters can be seen explicitly by solving the linear feedback model for \( \tau_{cross} \) (under the assumption that \( F_{SW} = 0 \)). Substituting Eq. 4 into Eq. 2 and identifying \( \tau = \tau_{cross} \) as the time when OLR = 0 gives \( F_{LW} = F_{LW} G_{\lambda_{SW}}(e^{\tau_{cross}/7} - 1) \), which has the solution

\[
\tau_{cross} = -\tau \ln \left( \frac{1}{1 - \frac{1}{G_{\lambda_{SW}} G_{F_{SW}}}} \right). \tag{6}
\]

Eq. 6 reveals that the OLR recovery time is proportional to (i) the radiative e-folding timescale \( \tau \), which is on the order of several decades, and (ii) a factor \( \ln (1 - 1/G_{\lambda_{SW}}) = \ln (1 - 1/G_{\lambda_{LW}}) \), which is \( \approx 1 \) in the multi-GCM mean but varies by two orders of magnitude across the GCMs. A positive SW feedback amplifies warming, and thus enhances the OLR response and decreases the timescale for OLR recovery. Moreover, \( \tau_{cross} \) is far more sensitive to changes in \( \lambda_{SW} \) than \( \lambda_{LW} \). This result arises from a fundamental asymmetry in the dependence of OLR on \( \lambda_{SW} \) and \( \lambda_{LW} \); a more positive \( \lambda_{SW} \) acts to amplify warming, which enhances OLR and decreases \( \tau_{cross} \); a less negative \( \lambda_{LW} \) similarly acts to amplify warming, which enhances OLR, but it also diminishes the OLR response per degree \( T \) change (Eq. 2), altogether driving only small changes in \( \tau_{cross} \).

Despite its many simplifications, Eq. 6 provides a reasonable estimate of \( \tau_{cross} \) as simulated by the GCMs, explaining 66% of the variance across models (Fig. 3A). In particular, it broadly captures the short OLR recovery time in the CMIP5 models with large and positive \( \lambda_{SW} \) values and the long OLR recovery time in models with a near-zero \( \lambda_{SW} \). There are a few notable exceptions, however, where Eq. 6 predicts a substantially smaller \( \tau_{cross} \) that is not realized. \( \tau_{cross} \) is underestimated in these models because we have not yet accounted for the SW component of CO2 forcing, which is substantial in a few GCMs because of the rapid cloud adjustments that occur on timescales faster than surface temperature changes. Analogous to the SW feedback case discussed above, SW forcing amplifies the equilibrium temperature response by an SW forcing gain factor, \( G_{F_{SW}} \), relative to the system with LW forcing only:

\[
G_{F_{SW}} \equiv 1 + \frac{F_{SW}}{F_{LW}}. \tag{7}
\]

A positive SW forcing amplifies warming, enhancing the OLR response and decreasing \( \tau_{cross} \), whereas a negative SW forcing reduces warming, diminishing the OLR response and increasing \( \tau_{cross} \). Including the effects of SW feedbacks and forcing together gives a simple extension of Eq. 6, wherein the gains are multiplicative (SI Text):

\[
\tau_{cross} = -\tau \ln \left( \frac{1}{1 - \frac{1}{G_{\lambda_{SW}} G_{F_{SW}}}} \right). \tag{8}
\]

In the multi-GCM mean, \( F_{SW} = 0.1 \) and modifies \( \tau_{cross} \) little from that predicted by Eq. 6. However, in some models, \( F_{SW} \) is a substantial fraction of the total CO2 forcing (Fig. 3B), and thus, it has a large impact on \( \tau_{cross} \). With \( F_{SW} \) taken into account, Eq. 8 provides an excellent estimate of \( \tau_{cross} \) as simulated by the GCMs, explaining 78% of the variance across models.

If a constant value \( \tau \approx 29 \) y is used in Eq. 8, the dependence of \( \tau_{cross} \) on the feedback and forcing gains can be visualized (curves in Fig. 3B). \( \tau_{cross} \) has very steep gradients in the region where the product of \( G_{\lambda_{SW}} \) and \( G_{F_{SW}} \) approaches one, leading to a bimodal distribution of \( \tau_{cross} \) with OLR returning to unperturbed values either over a couple decades or at timescales longer than a century.
Although $G_{\lambda_{SW}}$ and $G_{FSW}$ contribute equally to $t_{\text{cross}}$, $G_{\lambda_{SW}}$ varies by a greater amount than $G_{FSW}$ across the GCMs. Thus, it is SW feedback that most strongly controls the range of $t_{\text{cross}}$ and the relative contributions of OLR and ASR to global energy accumulation. However, in models with a sufficiently negative $F_{SW}$ ($G_{FSW} < 0$), $t_{\text{cross}}$ can be on the order of centuries, even with a large and positive $\lambda_{SW}$ ($G_{\lambda_{SW}} > 0$). In general, OLR recovers on timescales of centuries in models with either weak SW feedbacks or weak (or negative) SW forcing, and OLR recovers on timescales of several decades in models with moderate SW feedbacks and SW forcing. This result can be further seen by varying only $\lambda_{SW}$ and $F_{SW}$ in the linear feedback model (Eq. 1) and setting $\lambda_{LW}$, $F_{LW}$, and $C$ equal to their ensemble mean values. The predicted values of $t_{\text{cross}}$ are in excellent agreement with those simulated by the GCMs (Fig. 4), except for two models with $C$ much larger than the ensemble mean value. Importantly, allowing only $\lambda_{SW}$ and $F_{SW}$ to vary between models is sufficient to capture the clear separation between (i) those models with $t_{\text{cross}}$ on the order of centuries (black circles in Fig. 4A), where global energy accumulation is dominated by reduced OLR, and (ii) those models with $t_{\text{cross}}$ on the order of decades (colored circles in Fig. 4A), where global energy accumulation is dominated by enhanced ASR and opposed by enhanced OLR.

With these insights in mind, we return to the relative roles of ASR and OLR in driving global energy accumulation under the 1% CO$_2$ increase per year scenario, where GHG concentrations increase slowly over time, as in nature, rather than abruptly quadrupling. To quantify the relative roles of enhanced ASR and reduced OLR in transient energy accumulation, we define the SW energy accumulation ratio (SWEAR) to be the ratio of time-integrated energy accumulation via enhanced ASR to the time-integrated net radiative imbalance (ASR – OLR) over the 140 y of the 1% CO$_2$ simulations:

$$\text{SWEAR} = \frac{\int \text{ASR} dt}{\int (\text{ASR} – \text{OLR}) dt}$$  \[9\]

Values of SWEAR vary considerably across the GCMs (Fig. 4B), from near zero (energy accumulated primarily by reduced OLR) to near three (energy accumulated by enhanced ASR and lost by enhanced OLR). SWEAR between 0 and 1 indicates energy accumulation through both enhanced ASR and reduced OLR, whereas SWEAR above 0.5 indicates that ASR contributes more than one-half of global energy accumulation. In the multi-GCM mean, SWEAR is 1.1, indicating that OLR changes little and that net energy accumulation is accomplished entirely by enhanced ASR (Fig. 1D).

This range of GCM behavior under slowly increasing GHG forcing follows directly from the range of OLR recovery timescales $t_{\text{cross}}$ identified above under an abrupt change in GHGs, which, in turn, is set by intermodel differences in SW feedbacks and forcing. Indeed, the linear feedback model (Eqs. 1 and 2 with parameters estimated from 4x CO$_2$ as described above) iterated forward under 1% CO$_2$ captures the multi-GCM ASR and OLR response (dashed lines in Fig. 1D) and their variations across models. The linear feedback model, thus, also captures the inter-GCM variance in SWEAR (95%), where the vast majority (85%) of the inter-GCM variance can be explained by varying $\lambda_{SW}$ and $F_{SW}$ only (with $\lambda_{LW}$, $F_{LW}$, and $C$ set to their ensemble means as above) (Fig. 4B).

Fig. 4B shows a clear separation between models with SWEAR $\leq$ 0.5 (OLR-dominated) and models with SWEAR $\geq$ 1 (ASR-dominated). Furthermore, models with SWEAR $\leq$ 0.5 are those with $t_{\text{cross}}$ on the order of centuries (Fig. 4B, black circles), and models with SWEAR $\geq$ 1 are the same as those with $t_{\text{cross}}$ on the order of decades (Fig. 4B, colored circles). This strong dependence of SWEAR on $t_{\text{cross}}$ can be understood by considering the response to 1% CO$_2$ as the superposition of many responses to an instantaneous CO$_2$ forcing, each initiated at a different time. More formally, the time ($t_{\text{ramp}}$) at which OLR returns to its unperturbed value in response to a linear increase in CO$_2$ forcing can be approximated by (SI Text)

$$t_{\text{ramp}} = \frac{\tau}{1 - G_{\lambda_{SW}} G_{FSW}} \equiv \tau \frac{t_{\text{cross}}}{\tau}.$$  \[10\]

For models with $t_{\text{cross}}$ on the order of decades, $t_{\text{ramp}}$ is also on the order of decades, and SWEAR is large. For models with $t_{\text{cross}}$ on the order of a century, $t_{\text{ramp}}$ is on the order of several centuries, and SWEAR is small. Altogether, $t_{\text{cross}}$ explains 85% of the inter-GCM variance in SWEAR.

**Observational Constraints on SW and LW Energy Accumulation**

Global mean surface temperature has increased by about 0.85 K since the pre-industrial period (19) due to a global TOA energy accumulation driven by anthropogenic GHG emissions. Estimates...
of the rate of global heat content change based on ocean temperature measurements indicate that the current TOA energy accumulation is on the order of 0.5–1 W m⁻² (20, 21). Is the observed energy accumulation caused by reduced OLR or enhanced ASR? The limited accuracy and length of continuous satellite measurements of Earth’s radiative budget (22–24) preclude direct determination of anomalies in OLR and ASR. However, the covariance of SW and LW radiation fluxes with global mean surface temperature over the satellite era permits an estimate of \( \lambda_{SW} \) and \( \lambda_{LW} \) (25). Moreover, given the arguments developed above, these feedback parameters can be used to estimate the relative contributions of ASR and OLR anomalies to the present day global energy accumulation.

Murphy et al. (25) estimated \( \lambda_{SW} \) and \( \lambda_{LW} \) using 6 y of data (2000–2005) from the Clouds and the Earth’s Radiant Energy System Energy Balance Filled Project (24). Here, we extend these calculations with the now 14 y (2000–2013) of continuous satellite data and further account for changes in the global radiative forcing of stratospheric aerosols (26) and GHGs (27) over this period (details in SI Text); \( \lambda_{SW} \) and \( \lambda_{LW} \) are calculated from the linear regression of monthly anomalies in forcing-adjusted ASR and OLR on monthly surface temperature anomalies (Fig. S2) from three different datasets: (i) the National Centers for Environment Prediction reanalysis surface air temperature (31), (ii) the Goddard Institute for Space Studies Surface Temperature Analysis (32), and (iii) the adaptation by Cowtan and Way (33) of the Climatic Research Unit of the United Kingdom Met Office’s Hadley Center (34) surface temperature (version 4). The average of all calculations gives \( 4xCO_2 \) increases with warming and that OLR must ultimately become negative. The range of model responses is still dominated by decreased OLR. However, they also suggest that a transition to a regime of global energy accumulation dominated by enhanced ASR could occur with only 0.5 K global warming above present—by the middle of the 21st century if warming trends continue as projected.

**Discussion and Conclusions**

We have shown that, in most climate models, the OLR reduction associated with GHG forcing is alleviated within only a few decades and that the subsequent energy accumulation (and thus, global warming) is caused entirely by enhanced ASR. However, in some models, the OLR response is much slower. The range of model behaviors is readily understood in terms of a simple, linear feedback framework: positive SW feedbacks demand that ASR increases with warming and that OLR must ultimately become greater than its unperturbed value to achieve global energy balance with an imposed radiative forcing. The OLR recovery timescale is typically on the order of decades due to the fast response timescale of the surface components of the climate system and the negative LW feedbacks that strongly increase OLR with warming. Observational constraints also suggest an OLR recovery timescale on the order of decades. However, the current global warming rates and the observational estimate of \( \lambda_{SW} \) and \( \lambda_{LW} \) are consistent with previous estimates (ref. 36 and references therein) and values over the first few decades of the CMIP5 simulations (Fig. 2B). Together, these observational estimates of the feedbacks and heat capacity can be used to estimate the Earth’s natural timescale for radiative damping (\( \tau_{\text{ov}} \)) and OLR recovery (\( \tau_{\text{ov}} \)).

We can further estimate the effective global heat capacity \( C \) from observations by regressing global heat content anomalies [from ocean temperature measurements (35)] onto global mean surface temperature anomalies over the period 1970–2013 (Fig. S3), of which reliable ocean observations exist (20). This calculation gives an average value of \( C = 90 \pm 30 \) m of equivalent ocean depth, consistent with previous estimates (ref. 36 and references therein) and values over the first few decades of the CMIP5 simulations (Fig. 2B). Together, these observational estimates of the feedbacks and heat capacity can be used to estimate the Earth’s natural timescale for radiative damping (\( \tau_{\text{ov}} \)) and OLR recovery (\( \tau_{\text{ov}} \)).

**Fig. 4.** (A) Scatterplot of \( \tau_{\text{ov}} \) in the CMIP5 4x CO simulations and those predicted by the linear feedback model (Eq. 8) using the GCM-specific \( \lambda_{SW} \) and \( \lambda_{LW} \) the GCM ensemble average \( \lambda_{SW} \), \( \lambda_{LW} \), and heat capacity. The fill color of each circle indicates the \( \tau_{\text{ov}} \) of each GCM in the 4x CO simulation. The black dashed line is the 1:1 line. (B) The same as in A except for that scatterplot is of the SWW value in the 1% CO increase per year simulations.

We do not account for changes in the radiative forcing of tropospheric aerosols since they have not changed substantially over this time (28–30).

CMIP5 models, although \( \lambda_{SW} \) is at the upper end of the GCM range (Fig. 3A).
global energy imbalance seems to be dominated by reduced OLR because of the substantial SW forcing associated with anthropogenic tropospheric aerosols, which have directly reduced ASR and indirectly reduced OLR by curtailing global warming.

The feedback analysis ignores time dependence (38) and other nonlinearities in climate feedbacks (39). Although both may be important for the details of the response, our results show that the OLR recovery timescale and the relative contributions of ASR and OLR to energy accumulation are largely governed by linear feedbacks (Fig. 4). At times, we simplified the analysis by assuming a constant effective global heat capacity (C) and associated single timescale of temperature response to forcing (r). Although C increases over time (Fig. 2B) and there are, of course, multiple timescales of climate response (12, 15), accounting for these details (e.g., by representing a deep ocean heat capacity) makes no substantive changes to our results and conclusions. Indeed, surface temperature increases quickly after a CO₂ perturbation—much of the equilibrium temperature response is realized within the first few decades in all of the GCMs (Fig. 2A)—and the timescale of OLR recovery is most sensitive to the relative magnitudes of \( \lambda_{SW} \) and \( \lambda_{LW} \). Moreover, when \( C \) becomes small and the resulting heat capacity (governed by linear feedbacks (Fig. 4)) is large and negative could the energy accumulation be dominated by reduced OLR. Instead, observations constrain \( \lambda_{SW} \) to be at the upper end of the CMIP5 range, implying that OLR recovers quickly in response to GHG forcing and that global warming is driven by enhanced ASR.

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