Dynamics of Global Atmospheric CO_2 Concentration from 1850 to 2010: a Linear Approximation

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Abstract

Changes in Earth's temperature have significant impacts on the global carbon cycle, yet the quantification of such impacts using linear schemes is traditionally deemed difficult. Here we show that, by incorporating a temperature sensitivity parameter into a simple linear model, we can satisfactorily characterize the timescale-dependent responses of atmospheric CO_2 concentration to temperature changes and carbon emissions and accurately reproduce the history of atmospheric CO_2 between 1850 and 2010. The linear modeling framework allows us to analytically examine the dynamic characteristics of the carbon system and associate them with the response times of the carbon reservoirs and the temperature sensitivity parameter. These results also have important biogeophysical implications that appear to highlight the intensification of the global carbon cycle. On one hand, they indicate that the elevated atmospheric CO_2 concentration enhanced land carbon uptakes at a rate higher than traditionally thought. On the other hand, such enhanced gross carbon uptakes are partially offset by the increases in global surface temperatures, which accelerate the release of carbon from the surface reservoirs into the atmosphere. As a result, the net rate of atmospheric CO_2 sequestration by global land and oceans has slowed by ~30% since 1960s. We believe the linear modeling framework outlined in this paper provides a convenient tool to diagnose the observed atmospheric CO_2 dynamics and monitor their future changes.

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1. Introduction

Anthropogenic CO_2 emissions from fossil-fuel usage and land-use changes have been almost exponentially increasing since the Industrial Revolution (Fig. 1). Their accumulation in the atmosphere appears to be changing Earth's climate (IPCC 2007). The full strength of anthropogenic CO2 emissions for changing the climate has not yet been reached because only 41-45% of the CO_2 emitted between 1850 and 2010 remained in the atmosphere while the rest was sequestered by lands and oceans (Jones and Cox, 2005; Canadell et al. 2007; Raupach et al. 2008; Knorr 2009) (Fig. 1). This largely constant ratio, generally referred to as the "airborne fraction" (denoted as "γ" in this paper), was conventionally used to evaluate the efficiency of global carbon sinks (carbon sequestered by lands and oceans) in assimilating the extra CO_2 from the atmosphere (Jones and Cox, 2005; Canadell et al. 2007). A few recent studies found that the airborne fraction can also be influenced by other factors and thus may not be an ideal indicator for monitoring changes in the carbon sink efficiency (Knorr 2009; Gloor et al. 2010; Frölicher et a. 2013). Nevertheless, the remarkable constancy of the observed airborne fraction provides important hints for us to examine the dynamic characteristics of the carbon cycle (Gloor et al. 2010). In particular, it indicates that the responses of atmospheric CO_2 concentration to the disturbances of anthropogenic CO_2 emissions since 1850s can be properly approximated by a linear dynamic system with largely constant parameters (such a linear system is said to be linear time-invariant or LTI; the linear systems discussed in this paper are assumed to be LTI unless otherwise stated).

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Our reasoning is based on two important observations. First, it is a basic feature of a stable linear system that its responses to an exponentially increasing forcing [e.g., $\exp(\alpha t)$, $\alpha>0$, $t\geq 0$] will approach the input signal with a constant ratio (Naylor and Sell 1982). Second, a nonlinear dynamic system can be linearly approximated around a steady point within a neighborhood in its state space (Khalil 2001). In the case of the global carbon cycle, the existence of such a (quasi) steady state is evident in that the atmospheric CO_2 concentration (and the corresponding global climatology) had been relatively stable for thousands of years before the industrial era (IPCC 2007). Therefore, the observed simple proportional relationship between the atmospheric CO_2 concentrations and the increasing CO_2 emissions suggests that recent changes of the global carbon cycle are still within the linear neighborhood of the system's last steady state.

There is a rich literature on the application of linear methodology to study the global carbon cycle, either to approximate the system's dynamics or diagnose its characteristics (e.g., Oeschger and Heimann 1983; Meier-Raimer and Hasselmann 1987; Enting and Mansbridge 1987; Wigley 1991; Jarvis et al. 2008; Gloor et al. 2010; Joos et al. 1996, 2013). At the heart of some of the most influential methods is the estimation of the system's Impulse Response Function (IRF; or more generally the Green's function), which describes the time-varying responses of atmospheric CO_2 to a pulse of external disturbances, usually anthropogenic carbon emissions. Because the analytical determination of IRFs is difficult for complex systems, they were often obtained by fitting exponential equations to the numerical experiment results with global carbon-cycle

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models or their sub-components (Meier-Raimer and Hasselmann 1987; Joos et al. 1996, 2013). Once the IRF is known, the state of atmospheric CO_2 can be conveniently calculated through linear convolution of the IRF and the records of CO_2 emissions. Results obtained by such linear approaches well agree with the simulations from the corresponding global carbon-cycle models unless the disturbances to the system are too large (Wigley 1991; Li et al. 2009).

Although previous studies mostly use IRFs as convenient tools to substitute the corresponding "parent" models in calculation, the significance of IRFs in diagnosing the dynamic characteristics of the carbon-cycle system cannot be underestimated. The fact that IRFs can be represented by a few exponential functions (Meier-Raimer and Hasselmann 1987) indicates that the dynamic responses of their parent models are largely captured by a few dominant linear modes (Young 1999) – in other words, the fundamental dynamic characteristics of these global carbon-cycle models can be learned from suitable lower-order linear models. For instance, Li et al. (2009) were able to infer the response (*e*-folding) time constants of the major carbon reservoirs in the carbon-cycle model of Lenton (2000) by studying its IRF with a fifth-order linear model.

Extending the line of thoughts from the literature, this study applies lower-order linear models to investigate the dynamic characteristics of the global carbon cycle based on observations. Because the IRF of the real-world system is unknown, we can only treat the global carbon cycle as a "black box" and use the observed forcing-response relationships to constrain our models. Nevertheless, the independence from a parent model also gives

us more freedom to diagnose some important dynamic modes that have been less investigated in previous linear models. In particular, global surface temperature has increased by ~1 ^{o}C since the beginning of the 20th century (Hansen et al. 1999; Brohan et al. 2006). Given the tight coupling between temperature and the carbon cycle (Keeling et al. 1995; Joos et al. 1999, 2001; Lenton 2000; Rafelski et al. 2009), the warming alone may release a large amount of CO_2 from the land and the oceans into the atmosphere, redistributing carbon among these reservoirs. Previous studies have noticed that the effects of temperature on atmospheric CO_2 vary at different time scales, ranging from 1-2 ppm $^{o}C^{-1}$ at the scale of years to 10-20 ppm $^{o}C^{-1}$ over millennium or centuries (see the literature review by Woodwell et al. 1998). However, such effects are traditionally deemed difficult to quantify by simple schemes (Scheffer et al. 2006). Here we show that, by using a simple sensitivity parameter to represent the effect of temperature in our linear model, we can satisfactorily characterize the dynamic responses of atmospheric CO_2 concentration to temperature changes (and carbon emissions) as reported in the literature while accurately reproducing the history of CO_2 in the atmosphere in the past 160 years (see below).

A practical factor to decide in developing a diagnostic model for the global carbon cycle is the complexity of the linear tool itself. This may not represent a serious difficulty in the forward model construction and analysis, where well-established mathematical tools are at our disposal (see the example in the Appendix). For the inverse problem of model identification, on the other hand, it is the resolution of available observations that essentially determines the number of independent system parameters that can be reliably

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retrieved. In this study, we decided to demonstrate our analytical framework by a simple two-box model that represents carbon exchanges between the atmosphere and the surface (i.e. land and ocean) reservoirs. This decision is based on multiple considerations besides the constraints of model identification, which include that, for instance, the analysis of a two-box model involves only simple mathematical techniques but render clear physical pictures of the problem under investigation. Though such a "toy" model may sit at the lowest rank on the hierarchy of global carbon-cycle models (Enting 1987), new and important characteristics of the atmospheric CO_2 dynamics can still be learned from it. Furthermore, the use of a simple model by no means implies the compromise of scientific rigor of our findings, which are verified in a generalized linear model framework as described in the Appendix.

Throughout the analysis we also compare the results obtained from the two-box model to those from the more advanced Bern model (Siegenthaler and Joos 1992; Enting et al. 1994; IPCC 1996, 2001). The Bern model couples the atmosphere with a process-based ocean biogeochemical scheme (Siegenthaler and Joos 1992; Shaffer and Sarmiento 1995; Joos et al. 1999) and a multi-component terrestrial biosphere module (Siegenthaler and Oeschger 1987). The original Bern model does not consider the effects of changing global temperatures on terrestrial ecosystem respiration, which however plays an important role in regulating the variability of the global carbon cycle at interannual to multi-decadal time scales (Wang et al. 2013; Rafelski et al. 2009). Therefore, we revised the Bern model to account for temperature's effects on terrestrial ecosystem respiration and recalibrate the model subsequently (see the Appendix for details). The global carbon-

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cycle processes described in the Bern model help us diagnose the biogeophysical mechanisms underlying the characteristics of the atmospheric CO_2 dynamics identified with our simple linear model.

2. Datasets

Annual atmospheric CO_2 concentration data from 1850 to 1960 are based on the ice core CO_2 records from Law Dome, Antarctica (Etheridge et al. 1996) and those between 1960 and 2010 are compiled from NOAA Earth System Research Laboratory (ESRL) (Keeling et al. 1995; Conway et al. 1994). We merged the data following the approach described in Le Quéré et al. (2009) and calculated annual CO_2 growth rate as the first-order difference of the yearly CO_2 concentrations. Long-term records of anthropogenic CO_2 emissions from fossil fuel burning and cement production are compiled by Boden et al. (2011) and those of land-use changes are from Houghton (2003), both downloaded from the Carbon Dioxide Information Analysis Center at Oak Ridge National Laboratory, TN, USA (http://cdiac.ornl.gov). Two sets of monthly surface temperature data are used, including GISTEMP from NASA Goddard Institute for Space Studies (Hansen et al. 1999) and the CRU-NCEP climate dataset (Sitch et al. 2008; Le Quéré et al. 2009), available from 1901 to the present with spatial resolutions of 0.5×0.5 (CRU-NCEP) or 1×1 (GISTEMP) degrees. Monthly time series of temperature are aggregated globally and over the tropics (24^oN-24^oS), and smoothed with a 12-month running window to convert the monthly data to annual values. We calculated temperature anomalies relative to their 1901 to 1920 annual mean and assumed the 20-years mean temperature to be representative of temperature climatologies between 1850 and 1900. This assumption is reasonable as

suggested by analysis of other long-term coarse-resolution temperature datasets (Jones et al. 2003; Brohan et al. 2006).

3. Derivation of the Two-Box Model

This study considers only the "fast" carbon flows between the atmosphere and the surface at time scales within hundreds of years (IPCC 2001). In the two-box approach discussed below the world's land and oceans are treated as one combined carbon reservoir ("box"). A generalized treatment of the surface carbon reservoirs by individuals is presented in the Appendix.

Based on our linearization assumption, we describe the dynamics of the two-box carbon system using the following equations:

$$\dot{A}' = -\alpha_A \cdot A' + \alpha_S \cdot S' + \beta_T \cdot T' + \dot{E}' \qquad (1a)$$

$$\dot{S}' = +\alpha_A \cdot A' - \alpha_S \cdot S' - \beta_T \cdot T' \qquad (1b)$$

where A and S denote carbon storages in the atmosphere and the surface reservoirs, respectively, and E is the accumulated anthropogenic CO_2 emissions since the industrial era. The three variables can be measured by the same unit of parts per million by volume (1 ppm = ~2.13×10^9 metric-ton carbon or GtC). The prime symbol (e.g., "E'") indicates that changes in a variable relative to its preindustrial steady-state level. The preindustrial emissions are assumed negligible so that E'=E. The dot accent (e.g., "$\dot{E}'$") indicates the first-order derivative with regard to time, such that $\dot{E}'$ represents the annual rate of CO_2 emissions (ppm yr^{-1}). The positive constant parameters α_A and α_S (yr^{-1}) describe the decaying rates of corresponding carbon anomalies. Their reciprocals (i.e., $\tau_A = 1/\alpha_A$,

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$\tau_S = 1/\alpha_S$) are often referred to as the response time of the carbon reservoirs (IPCC 2001). T (°C) denotes indices of global (or large-scale) surface temperatures and the coefficient β_T (ppm yr^{-1} °C^{-1}) represents the sensitivity of atmospheric CO_2 growth rate to temperature changes. The term $\beta_T T'$ thus indicates the impacts of temperature increases on the global carbon cycle, which release CO_2 from the surface reservoirs to the atmosphere. In this study we have assumed β_T to be a constant. This assumption is justified later in the paper.

The physical meaning of Eq. 1(a, b) is clear: carbon outflows from the atmosphere, $\alpha_A A'$, are inflows to the surface (e.g., through photosynthesis in green vegetation and the dissolution of CO_2 in the surface water) while carbon outflows from the surface, $\alpha_S S'$, (e.g., through respiration and the outgassing of the dissolved CO_2) are the inflows to the atmosphere. The effects of temperature changes, $\beta_T T'$, revise the relative carbon balance between the atmosphere and the surface reservoirs. Human emissions of CO_2, on the other hand, represent an "external" source of CO_2 to the system by rapidly releasing carbon (e.g., fossil fuel burning) from reservoirs that were formed over millions of years and by permanently altering the structure of land surface carbon pools (e.g., land-cover/land-use changes).

Because mass (carbon) is conserved in the two-box model, Eqs. (1a) and (1b) are not independent. Adding the two equations together leads to

$$\dot{A}' + \dot{S}' = \dot{E}'$$

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$$A' + S' = E'. \quad (1c)$$

Eq. (1c) simply states that the anthropogenically emitted CO_2 either resides in the atmosphere or in the surface reservoirs (i.e., the land and the oceans). Substituting this relationship into Eq. (1a) to replace S', we obtain

$$\dot{A}' + (\alpha_A + \alpha_S) \cdot A' = \beta_T \cdot T' + \alpha_S \cdot E' + \dot{E}'. \quad (2a)$$

Therefore, the dynamics of atmospheric CO_2 represented by the two-box model is determined by an ordinary differential equation of A' under the disturbances of anthropogenic emissions (E' and $\dot{E}'$) and the changing climate (T').

4. Model Determination and Evaluation

We want to determine the parameters of Eq. (2) with observational records of A', T', and E', and evaluate how well the model captures the observed atmospheric CO_2 dynamics. In order to construct a regression model from Eq. (2), we rearrange the equation as follows:

$$\dot{E}' - \dot{A}' = (\alpha_A + \alpha_S) \cdot A' - \alpha_S \cdot E' - \beta_T \cdot T', \quad (2b)$$

where $\dot{E}' - \dot{A}'$ represents the strength of annual carbon sinks. However, because $A' \approx \gamma \cdot E'$ (Fig. 1), where γ is the airborne fraction, the "collinearity" between the two regressors prevents us from determining the coefficients associated with them separately (Chatterjee and Hadi 2006). Indeed, substituting the airborne-fraction relationship into Eq. (2b) leads to

$$\dot{E}' - \dot{A}' = [\alpha_A - (1/\gamma - 1)\alpha_S] \cdot A' - \beta_T \cdot T', \quad (2c)$$

which implies that only a combination of α_A and α_S can be estimated from the observations, and the estimation of β_T also needs some special care (see below).

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We may reach the above argument from another perspective. Fig. (1) shows that the anthropogenic CO_2 emissions (E' and $\dot{E}'$) can be approximated by exponential functions. Let $E' = \exp(\alpha_E t)$, and it follows that $\dot{E}' = \alpha_E \cdot \exp(\alpha_E t)$. By applying the airborne fraction relationship ($A' \approx \gamma \cdot E'$ and $\dot{A}' \approx \gamma \cdot \dot{E}'$) and neglecting the influence of temperature ($\beta_T T'$) for now, we obtain

$$\gamma \cdot [\alpha_E + (\alpha_A + \alpha_S)] \cdot \exp(\alpha_E \cdot t) \approx (\alpha_S + \alpha_E) \cdot \exp(\alpha_E \cdot t),$$

or

$$\alpha_A - (1/\gamma - 1) \cdot \alpha_S \approx (1/\gamma - 1) \cdot \alpha_E. \quad (2d)$$

The term on the left-hand side of Eq. (2d) is the same as the (regression) coefficient associated with the A' in Eq. (2c). Eq. (2d) suggests that this coefficient is closely related to the exponential disturbances (α_E) to the system. Because α_E>0 and γ is about 0.41-0.45 (Fig. 1), it is follows that $(1/\gamma - 1)$>1 and $\alpha_A > \alpha_S$ (or $\tau_A < \tau_S$).

The above analysis indicates that additional information is required to resolve α_A and α_S from the regression results of Eq. (2c). One source of such information comes from previous observation-based studies. For instance, by comparing the carbon isotope ratios in wood and in marine material, Revelle and Suess (1957) have long suggested that the response time (τ_A) of atmospheric CO_2 is on the order of 10 years. We also extract information from process-based model studies. Because the initial decaying rate of the IRF of a global carbon-cycle model is mainly determined by α_A (or τ_A; see the proof in the next section and in the Appendix), applying this result to analyze the ensemble IRFs reported in Joos et al. (2013) suggests τ_A to be ~14 years. We choose τ_A to be 12 years

($\alpha_A \approx 0.083$ yr^{-1}) so that the IRF of our linear model closely matches with the Bern model during the initial decaying stage (see the next section). We subsequently estimate τ_S to be ~34 years ($\alpha_S \approx 0.029$ yr^{-1}).

The estimation of the β_T parameter in Eq. (2) requires the choice of a large-scale temperature index that is representative of climate change and closely related with global carbon cycle. Previous studies showed that the land surface air temperature in the tropics (24S-24N) are most strongly coupled with interannual variations in the growth rate of atmospheric CO_2 by a sensitivity (β_T) of ~1.64 ppm yr^{-1} °C^{-1} (Adams and Piovesan 2005; Wang et al. 2013). Here we found that the same temperature-CO_2 coupling also operate at longer time scales. Indeed, because the system is linear, variations in A', T', E', and their derivatives over different time scales must satisfy Eq. (2) separately. Because the interannual variations ("IAV") in the emissions (both $\dot{E}'$and E') and the atmospheric CO_2 concentration (A') are relatively small (Fig. 1), neglecting them in Eq. (2) leads to

$$\dot{A}'_{IAV} \approx \beta_T \cdot T'_{IAV}, \qquad (2e)$$

which is the same linear relationship as previous reported (Wang et al. 2013).

There is another practical reason that we use the β_T estimated from Wang et al. (2013) in this study. Because the long-term increases in global temperature (T') are mainly induced by the growing CO_2 concentrations in the atmosphere (A'), the two variables are indeed significantly correlated ($r \approx 0.9$, with IAV in them removed). Therefore, estimating β_T directly from Eq. (2c) is inevitably subject to the influence of the collinearity between A' and T' (Enting 2010). On the other hand, the short-term variations (i.e., IAV) of global

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temperature are dominated by the natural variability of the climate system (e.g., the El Niño-Southern Oscillations). Therefore, we expect the β_T estimated with Eq. (2e) in Wang et al. (2013) to have less uncertainties.

With the model parameters determined, we use the two-box model to simulate the changes of atmospheric CO_2 concentration between 1850 and 2010 from historical records of temperature and CO_2 emissions (Fig. 2). The simulated results follow the evolution of the observed atmospheric CO_2 concentration to a high degree of accuracy, capturing more than 96% of the variability (i.e., r^2>0.96) of the latter (Fig. 2). The standard deviations (σ) of the differences between simulated values and those measured accurately since 1960 are ~0.9 ppm for the atmospheric CO_2 concentration and ~0.4 ppm for its growth rate, respectively (Fig. 2). These results are highly comparable to those simulated with the revised Bern model (Fig. 2) or other sophisticated climate-carbon models reported in the literature (e.g., Joos et al. 1999; Lenton 2000; Friedlinstein et al. 2006), strongly supporting our argument that the atmospheric CO_2 dynamics in the past one and half centuries can be properly approximated with linear models.

5. Disturbance-Response Functions

We first check the model's responses to an impulse disturbance of anthropogenic CO_2 emissions. Shown in Fig. 3, the initial atmospheric CO2 anomaly decays relatively fast, as 60-70 % of the emitted CO_2 is absorbed by the surface reservoirs within 20 years of the disturbance. However, the rate of carbon assimilation by the land and the oceans significantly slows down in the following decades and eventually becomes neutral as the

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system approaches steady-state. In the end, 15-25% of the simulated CO_2 anomaly will likely stay in the atmosphere for thousands of years (Fig. 3). These results are consistent with the findings from fully coupled climate-carbon models (Cao et al. 2009; Archer et al. 2009; Joos et al. 2013).

The IRF of the linear box models can be analytically characterized. For the two-box model of Eq. (2), when the system approaches a (new) steady state after the disturbance, all the time derivatives ($\dot{E}'$ and $\dot{A}'$) will be zero. Assuming that temperature does not change during the process, we easily obtain the steady state of A' as

$$A' = \frac{\alpha_S}{\alpha_A+\alpha_S} \cdot E' = \frac{\tau_A}{\tau_A+\tau_S} \cdot E', \quad (3a)$$

or more generally

$$\frac{A'}{\tau_A} = \frac{S'}{\tau_S}, \quad (3b)$$

where the mass-conservation relationship represented by Eq. (1d) is used in the derivation. Therefore, the extra CO_2 added to the "fast" carbon cycle by anthropogenic emissions will be partitioned between the atmosphere and the surface corresponding to the response times (τ) of the reservoirs, respectively. Because $\tau_S > \tau_A$ (see Eq. (2d) and the related discussions), a majority of the emitted CO_2 will eventually be absorbed by the surface carbon reservoirs (Fig. 3). In other words, the long-term fate of the CO_2 emitted into the atmosphere is largely determined by the response times of the surface reservoirs (Revelle and Suess 1957).

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The rates at which the atmospheric CO_2 anomaly decays are determined by the solutions (i.e., eigenvalues) to the characteristic equation of the system. For a two-box system like Eq. (2a), the problem is particularly simple because the only non-zero eigenvalue (λ) is

$$\lambda = \alpha_A + \alpha_S, \qquad (4)$$

and the solution of Eq. (2) is therefore

$$A' = \frac{\alpha_A}{\alpha_A+\alpha_S}\exp[-(\alpha_A + \alpha_S)\cdot \mathrm{t}] + \frac{\alpha_S}{\alpha_A+\alpha_S}. \qquad (5a)$$

A helpful observation of Eq. (5) is that, when $t \ll 1/(\alpha_A + \alpha_S)$, the solution can be approximated by

$$A' \approx \frac{\alpha_A}{\alpha_A+\alpha_S}[1 - (\alpha_A + \alpha_S)\cdot t] + \frac{\alpha_S}{\alpha_A+\alpha_S} = 1 - \alpha_A \cdot t \approx \exp(-\alpha_A \cdot t). \qquad (5b)$$

That is, A' initially decays at a maximum rate of α_A as if the capacity of the surface carbon reservoir were unlimited (i.e., α_S=0). This result is also valid for general cases (see the Appendix).

Next we consider the system's responses to disturbances induced by changes in surface temperatures. Unlike anthropogenic CO_2 emissions, changes in temperature do not add additional CO_2 to the "fast" carbon cycle but only re-distribute carbon between the atmosphere and the surface (Eqs. 1a and 1b), and so the system will recover to its initial steady state once the temperature anomaly is removed. However, increases in temperature are persistent under climate-change scenarios. Therefore, we examine the responses of atmospheric CO_2 to a step change in temperature, which is determined from Eq. (2) as:

$$A' = \beta_T/(\alpha_A + \alpha_S)\cdot T'. \qquad (6a)$$

Because $\alpha_A > \alpha_S$, for rough estimates we can also use

$$A' \approx \beta_T/\alpha_A \cdot T'. \quad (6b)$$

Based on the estimated model parameters, therefore, atmospheric CO_2 rises by ~15 ppm for an increase of 1 ^oC in temperature within a few decades (Fig. 3). This long-term temperature sensitivity of atmospheric CO_2 is consistent with the estimate inferred from the (reconstructed) temperature and atmospheric CO_2 records during the Little Ice Age (~20 ppm $^oC^{-1}$; see Woodwell et al. 1998).

The relationships represented by Eqs. (3b), (5b), and (6b) can be generalized to higher-order systems (see the Appendix), providing a convenient way to characterize the models' disturbance-response functions without fully solving the system equations. However, the uncertainties associated with these results – especially the long-term responses of atmospheric CO_2 – need to be emphasized. One key source of the uncertainties is that model's parameters are not fully determined by the observations of the global climate-carbon system. As discussed in Section 4, the estimation of the model parameter α_S depends on the choice of α_A, which is only loosely constrained by the prior knowledge. The analysis in the Appendix indicates that this situation only worsens in higher-order (N-box) systems as the number of system parameters increase at the order of N^2 (also see Joos et al. 1996). It is possible for us to choose another pair of α_A and α_S or a higher-order linear model so that the derived disturbance response functions better approximate those of the Bern model. However, tuning the model in this fashion has only cosmetic effects to the results and does not reduce the associated uncertainties. In addition, in reality the climate system and the global carbon cycle are not independent but tightly coupled. Therefore, a comprehensive assessment of the long-term fate of

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anthropogenic CO_2 emissions in the atmosphere must account for the effects of the associated changes in global temperature, which is beyond the scope of this study.

6. Biogeophysical Implications

The above analysis suggests that the appropriate representation of temperature's effects on the carbon cycle in our linear model helps improve the model's accuracy in approximating the observed dynamics of the atmospheric CO_2 across multiple time scales. To illustrate, we further rearrange Eq. (2c) to obtain

$$\dot{E}' - \dot{A}' + \beta_T \cdot T' = [\alpha_A - (1/\gamma - 1)\alpha_S] \cdot A'. \qquad (7)$$

On the left-hand side of the equation, the term "$\dot{E}' - \dot{A}'$" is usually used to measure the *net* strength of annual global carbon sinks. However, because the warming temperature also releases carbon on from the surface into the atmosphere ($\beta_T \cdot T'$), this extra source of CO_2 has to be absorbed by the global carbon sinks. By accounting for the effects of temperature changes, the term "$\dot{E}' - \dot{A}' + \beta_T \cdot T'$" thus define a *gross* global carbon sinks.

Examining Eq. (7) with the observational data shows that both the net and the gross carbon sinks have been steadily increasing in response to the rising atmospheric CO_2 concentration in the past 160 years, reaching ~2.5 ppm yr^{-1} and ~4.0 ppm yr^{-1} respectively in 2010 (Fig. 4). The gross carbon sinks have a nearly direct linear relationship (with a constant slope ~0.04 yr^{-1}; r=0.98) with the atmospheric CO_2 concentrations throughout the entire data period. In comparison, the relationship between the apparent carbon sinks and the CO_2 concentrations is slightly nonlinear, with its slope

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decreasing from ~0.03 yr^{-1} in 1960 to ~0.02 yr^{-1} in 2010. Therefore, our linear approximation approach would not be able to achieve the same high accuracy if temperature's effects on the carbon cycle were not correctly represented. Note that the slope of these linear relationships (i.e., $\alpha_A - (1/\gamma - 1)\alpha_S$) is sometimes interpreted as the efficiency of surface carbon reservoirs in sequestering annual CO_2 emissions (Gloor et al. 2010; Raupach et al. 2014). Since this coefficient is influenced by the AF factor (γ), it is not an intrinsic characteristic of the carbon-cycle system. Therefore, the "sink efficiency" interpretation of the coefficient is only meaningful when γ is relatively constant. Nevertheless, Fig. 4 shows that although the *gross* carbon-sequestration rates of the surface reservoirs changed little, the *net* "efficiency" of the system has slowed by ~30% in the past five decades. This finding is essentially the same as reported in Raupach et al. (2014) but our analysis emphasizes that this declining carbon sequestration rate mainly reflects the impacts of climate changes on the global carbon cycle.

The biogeophyscial implication of the parameter β_T needs further discussion. Our previous analysis (Wang et al. 2013) suggests that this parameter mainly reflects the temperature sensitivity of respiration of land-surface carbon pools (biomass and soil carbon). This explanation is supported by the simulations of the Bern model in this study, in which terrestrial carbon sinks have much stronger responses to temperature changes than the ocean counterpart (not shown). Furthermore, both our simulations and those from the literature (e.g., Canadell et al. 2007; Le Quéré et al. 2009) indicate that the total carbon storage in the land-surface reservoirs remains largely stable between 1850 and 2010, a necessary condition for β_T to be constant. For instance, because terrestrial carbon

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uptake accounts for 50-60% of the global net sinks in our simulations, the accumulated terrestrial net carbon sinks are about 71-85 ppm in 2010, representing a 7-8% increase in the total terrestrial carbon storage (~1040 ppm as of 1850). At the same time, the accumulated terrestrial carbon losses through land-use changes are about 74 ppm in 2010 based on the dataset of Houghton (2003). These results suggest that the net changes in the total terrestrial biomass and soil carbon are (relatively) small during the past 160 year, providing further justification for our linear modeling approach.

Finally, our analysis suggests that the increasing atmospheric CO_2 concentration must have promoted carbon assimilation by the terrestrial biosphere (Ballantyne et al. 2012), most likely through the CO_2 fertilization effect (Körner and Arnone 1992; Oechel et al. 1994; Long et al. 1991, 2004) and the associated ecological changes (Keenan et al. 2013; Graven et al. 2013). Indeed, because the surface warming rapidly releases a proportion of the assimilated carbon back to the atmosphere (Fig. 4) (Piao et al. 2008; Wang et al. 2013), the increased turnover rate may have obscured the evaluation of the magnitude of the CO_2 fertilization effects, which we found in calibrating the Bern model (see the Appendix). In other words, the gross CO_2 fertilization effect of terrestrial vegetation is likely higher than previously thought (Schimel et al. 2014).

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7. Conclusions

We demonstrate in this paper that the observed dynamics of the global atmospheric CO_2 concentration from 1850 to 2010 can be properly approximated as a linear system. In

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particular, we derived a simple box model to describe carbon exchanges between the atmosphere and the surface carbon reservoirs under the disturbances of anthropogenic CO_2 emissions as well as global temperature changes. We show that, with a few appropriately retrieved parameters, the model can successfully simulate the observed changes and variations of the atmospheric CO_2 concentration and its first-order derivative (i.e., CO_2 growth rate) across interannual to multi-decadal time scales. The results are highly comparable to those obtained with more sophisticated models in the literature, confirming that the simple linear model is capable in capturing the main features of atmospheric CO_2 dynamics in the past one and half centuries.

A distinct advantage of our linear modeling framework is that it allows us to analytically, and thus most directly, examine the dynamic characteristics of the (modeled) carbon-cycle system. Our analyses indicate that many of such characteristics are closely associated with the response times of the atmosphere and surface carbon reservoirs. For instance, the initial decaying rate of an impulse of CO_2 emitted into the atmosphere is mainly influenced by the response time of the atmosphere, but the proportion of the extra CO_2 that stays in the atmosphere at long-term time scales is determined by the ratio between the response times of the atmosphere and the surface reservoirs. Unfortunately, the collinearity exhibited by the observed time series of CO_2 emissions and atmospheric CO_2 concentrations has obscured the determination of the response times for individual surface reservoirs, inducing uncertainties of the estimated long-term responses of the global carbon system. In other words, although the steady CO_2 airborne fraction allows

us to represent the carbon-system dynamics with a simple model, it prevents us to resolve further details of the surface carbon reservoirs.

Our model results also have important biogeophysical implications. They highlight that the responses of the global carbon cycle to recent anthropogenic and climatic disturbances are still within the resilience zone of the system, such that annual (gross) terrestrial and ocean carbon sinks linearly increases with the atmospheric CO_2 levels. On one hand, the elevated atmospheric CO_2 concentration must have enhanced land carbon uptakes through the "fertilization" effects and the associated ecological changes. On the other hand, the enhanced gross carbon uptakes are partially offset by the increases in global surface temperatures, which accelerate the release of carbon from the surface reservoirs into the atmosphere. As a result, the "net" efficiency of global land and oceans in sequestering atmospheric CO_2 may have slowed by ~30% since 1960s, although the airborne fraction of CO_2 emissions remains largely constant.

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Finally and importantly, we emphasize that the linear approximation of the global carbon cycle discussed in this paper is conditioned on the pre-industrial (quasi) steady state of the system. The global climate-carbon system is clearly nonlinear beyond this scope (Archer et al. 2009), which can establish different steady states over glacial/interglacial time scales (Sigman and Boyle 2000). A major concern stemming from climate change is that, because the post-industrial anthropogenic disturbances on the global carbon cycle are so strong and rapid, they may abruptly alter the pace at which the natural climate-carbon system evolves and drive the system into a different state at a drastically

accelerated rate (IPCC 2001). Our results clearly indicate that the rising atmospheric CO_2 concentrations and the associated increases in global temperature have significantly intensified the global carbon cycle in the past one and half centuries. Although such intensification of the carbon system seems to be within the linear zone as of now, its resilience may be weakened, or lost, in the future. As the anthropogenic CO_2 emissions continue to increase and the global temperature continues to warm, scientists generally expect surface – in particular, terrestrial – carbon reservoirs to saturate and their CO_2 sequestration efficiency to decrease, such that the responses of the global carbon cycle to the anthropogenic disturbances will eventually deviate from their original path. With this concern regarded, the simple linear model developed in this study may serve as a convenient tool to monitor the early signs when the natural carbon system is pushed away (by anthropogenic disturbances) from its linear zone.

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Appendix

A.1 Calibrations of the Bern carbon-cycle model

The Bern model is a coupled global carbon-cycle box model (Siegenthaler and Joos 1992; Enting et al. 1994) that was used in previous IPCC Assessment Reports to study changes in atmospheric CO_2 concentration under different emission scenarios (IPCC 1996, 2001). It couples the High-Latitude Exchange/Interior Diffusion-Advection (HILDA) ocean biogeochemical model (Siegenthaler and Joos 1992; Shaffer and Sarmiento 1995; Joos et al. 1999) with an atmosphere layer and a multi-component terrestrial biosphere model (Siegenthaler and Oeschger 1987). The HILDA model describes ocean biogeochemical cycling through two well-mixed surface layers in low and high latitudes, a well-mixed deep ocean in the high latitude and a dissipative interior ocean in the low latitude. Ocean tracer transport is represented by four processes: 1) eddy diffusion within the interior ocean (k, 3.2×10^{-5} $m^2\,s^{-1}$); 2) deep upwelling in the interior ocean (w, 2.0×10^{-8} $m\,s^{-1}$), which is balanced by lateral transport between the two surface layers as well as the down-welling in the polar deep ocean; 3) lateral exchange between the interior ocean and the well-mixed polar deep ocean (q, 7.5×10^{-11} s^{-1}); and 4) vertical exchange between the high-latitude surface layer and the deep polar ocean (u, 1.9×10^{-6} m s^{-1}) (Shaffer and Sarmiento 1995). The effective exchange velocity between surface ocean layers and the atmosphere in both low and high latitudes is assumed to be the same (2.32×10^{-5} m s^{-1}) (Shaffer and Sarmiento 1995). Ocean carbonate chemistry (e.g., the Revelle buffer factor) is based on the formulation given by Sarmiento et al. (1992). In addition, we implemented the influence of sea surface temperature on the partial pressure of dissolved CO_2 in seawater with a sensitivity of ~4.3% $^oC^{-1}$ (Gordon and Jones 1973;

Takahashi et al. 1993; Joos et al. 2001). The changes in global mean sea-surface temperature (SST) is approximately 0.8-1.0 ^{o}C from 1850s to 2000s (Rayner et al. 2003; Brohan et al. 2006) slightly lower than that of the tropical land-based air temperature (~1.0 ^{o}C) but with a trend resembling the latter (Rayner et al. 2003; Jones et al. 2003; Hansen et al. 2006). For simplicity, therefore, we used the long-term trend of the tropical land air as a proxy for the corresponding trend in global SST.

The terrestrial biosphere in the Bern model is represented by four carbon compartments (ground vegetation, wood, detritus, and soil) with prescribed turnover rates and allocation ratios. The global net primary production (NPP), the influx to the biosphere, is assumed to be 60 GtC yr^{-1} at the pre-industrial level; and the effect of CO_2 fertilization on NPP (i.e., the β-effect) is described with a logarithmic function with a β parameter of 0.38 (Enting et al. 1994). The original Bern model does not consider the effects of changing global temperatures on terrestrial ecosystem respiration, which have been suggested to play an important role in regulating the variability of the global carbon cycle at interannual to multi-decadal time scales (Wang et al. 2013; Rafelski et al. 2009). Therefore, we implemented temperature's effects on terrestrial ecosystem respiration in the Bern model with an overall sensitivity (Q_{10}) of ~1.5 (Lenton 2000; Davidson and Janssens 2006; Wang et al. 2013). We also changed the pre-industrial CO_2 concentration to 285 ppm in the Bern model to reflect the findings obtained from the observations (Fig. 4 of the main text).

We calibrated the Bern model so that the model outputs fit the observed atmospheric CO_2 data most favorably. Because no major revisions were made to the ocean carbon cycle module (HILDA), we focused mainly on calibrating the biosphere module. With the

original biosphere model parameters, the simulated atmospheric CO_2 concentrations were found to be distinctly higher than observations, reaching ~411 ppm in 2010. These results are induced because rising temperatures enhance respiration in the model, reducing the net land carbon sinks to an unrealistic ~0.5 ppm yr^{-1} in 2010. To balance the temperature-enhanced respiration, we need to increase the β parameter from 0.38 to 0.64 to incorporate a higher rate of gross biosphere carbon uptake as enhanced by CO_2 fertilization (Long et al. 2004) and the associated ecological changes (Keenan et al. 2013). With the β parameter set at 0.64, the simulated global terrestrial NPP increased by 14% from its pre-industrial level and reached ~69 GtC yr^{-1} in 2010, which qualitatively agrees with recent estimates inferred from the isotope measurements (Welp et al. 2011). As such, the re-calibrated Bern model is able to simulate accurately the observed changes/variations in atmospheric CO_2 concentration and growth rate in the past 160 years (Fig. 2 of the main text). The simulated ocean and land components of global carbon sinks are also consistent with estimates found in previous studies (e.g., Canadell et al. 2007; Le Quéré et al. 2009).

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A.2 Analysis of a General *N*-Box Model

Eq. (1) in the main text can be generalized to describe an arbitrary *N*-component ("*N*-box") carbon-cycle system:

$$\dot{\boldsymbol{s}}' = \boldsymbol{X} \cdot \boldsymbol{s}' + \beta_T T' \cdot \boldsymbol{y} + \dot{E}' \cdot \boldsymbol{z}, \qquad \text{(A1)}$$

where $\boldsymbol{s}'$, $\boldsymbol{y}$, and $\boldsymbol{z}$ represent $N\times1$ vectors, and $\boldsymbol{X}$ is an $N\times N$ matrix. Specifically,

- $\boldsymbol{s'}$ represents all the anomalous carbon state variables (e.g., carbon in atmosphere, land, ocean, interior ocean, etc.). In particular, we assume the first element of $\boldsymbol{s'}$ to be the atmospheric carbon anomalies, that is, $\boldsymbol{s'}_1 = A'$ in Eq. (1) of the main text.
- $\boldsymbol{y}$ describes the distribution weights of the carbon impacts of temperature anomalies ($\beta_T T'$) on different carbon pools. Per the reason explained in the main text, the elements of $\boldsymbol{y}$ are subject to the constraint $\sum_{i=1}^{N} \boldsymbol{y_i} = 0$. Without the loss of generality we set $\boldsymbol{y_1} = 1$, reflecting that positive temperature anomalies release more carbon into the atmosphere.
- $\boldsymbol{z}$ describes the distribution weights of the CO_2 anthropogenically emitted into the system. It is clear that $\boldsymbol{z_1} = 1$ and $\boldsymbol{z_i} = 0$ $(i = 2, \cdots, N)$
- $\boldsymbol{X}$ describes dynamics of and interactions among all the carbon reservoirs. In particular, the diagonal elements of $\boldsymbol{X}$ represent the decaying rates of the carbon reservoirs, i.e., $\boldsymbol{X_{ii}} = -\alpha_i = -1/\tau_i$. The off-diagonal elements $\boldsymbol{X_{ij}} \geq 0$ $(i \neq j)$ represent the rates of carbon flow from the *j*-th reservoir to the *i*-th reservoir.

A2.1 Generalization of Eq. (5b) – the short-term responses of atmospheric CO2 to an impulse disturbance of CO_2 emissions.

It is easy to see that the characteristic equation of Eq. (A1) is given by

$$\det(\boldsymbol{X} - \boldsymbol{\lambda} \cdot \boldsymbol{I}) = 0, \quad \text{(A2)}$$

where $\boldsymbol{\lambda}$ is the vector of eigenvalues, $\boldsymbol{I}$ is the identity matrix, and "det" stands for the determinant of the matrix. By the binomial theorem, the sum of the eigenvalues equals the trace of the state matrix $\boldsymbol{X}$, i.e.,

$$\sum_{i=1}^{N} \boldsymbol{\lambda_i} = tr(\boldsymbol{X}) = \sum_{i=1}^{N} \boldsymbol{X_{ii}} = \sum_{i=1}^{N} -\alpha_i, \quad \text{(A3)}$$

Because of the conservation of mass, the rank of $\boldsymbol{X}$ is N-1 (see explanations in the main text). Therefore, one of the eigenvalues is zero. We denote this zero-valued eigenvalue to be $\boldsymbol{\lambda_N}$. For simplicity of discussion, we also assume that the characteristic equation (A2) does not have multiple roots (i.e., the state matrix $\boldsymbol{X}$ is not degenerated). It is clear that this simplification has little influence on the discussion of the system's short-term responses (i.e., when $t \approx 0$).

The response function of atmospheric CO_2 to a unit impulse of emission disturbances is thus determined by

$$A'(t) = \boldsymbol{s'}_1(t) = \sum_{i=1}^{N-1} \varphi_i \cdot \exp(\lambda_i t) + \omega_1, \quad \text{(A4)}$$

where ω_1 is the steady-state (i.e., long-term) response of atmospheric CO_2 to the disturbance and φ_i are some constant coefficients subject to $\sum_{i=1}^{N-1} \varphi_i + \omega_1 = 1$, so that $A'(0) = 1$. Using the approximation that $\exp(\lambda_i t) = 1 + \lambda_i t$ for $\lambda_i t \approx 0$ in Eq. (A4) and rearranging the items on the right-hand side we arrive at:

$$A'(t) = \boldsymbol{s'}_1(t) \approx 1 + (\sum_{i=1}^{N-1} \varphi_i \lambda_i \cdot t) = \exp(-\alpha_{*,1} \cdot t), \quad \text{(A5)}$$

where $\alpha_{*,1} = \sum_{i=1}^{N-1} \varphi_i \lambda_i$. Therefore, $A'(t)$ initially decays as an exponential function.

We next prove that $\alpha_{*,1}$ is indeed α_1. By similar procedures as above, we can derive the response functions of the i-th $(i = 2, \cdots, N)$ carbon reservoir to be

$$\boldsymbol{s'}_i(t) \approx \omega_i\left[1 - \exp\left(-\alpha_{*,i} \cdot t\right)\right], \quad \text{(A6)}$$

where ω_i is the steady-state (i.e., long-term) response of the specific reservoir. By mass conservation it is apparent that $\sum_{i=1}^{N} \omega_i = 1$. For $t \approx 0$, all the responses of the surface

carbon reservoirs $\boldsymbol{s'}_i(t) \approx 0\ (i = 2, \cdots, N)$. Therefore the first equation in (A1) for the atmospheric CO_2 becomes

$$\dot{A}' = \dot{\boldsymbol{s}}'_1 \approx \boldsymbol{X_{11}} \cdot \boldsymbol{s'}_1 = -\alpha_1 A', \quad \text{(A7a)}$$

which simply means that

$$\alpha_{*,1} = \alpha_1, \text{ and } A'(t) \approx \exp(-\alpha_1 \cdot t), \quad \text{(A7b)}$$

which is the same conclusion stated by Eq. (5b) in the main text (where α_1 is denoted by α_A).

Finally, because the trace of the state matrix $\boldsymbol{X}$ is invariant under unitary transforms, rearranging/re-combining the surface carbon reservoirs will not change the results of Eq. A8. This further proves the generality of Eq. (5b) in the main text.

A2.2 Generalization of Eq. (3b) – the long-term responses of atmospheric CO2 to an impulse disturbance of CO_2 emissions.

At the steady state Eq. (A1) becomes

$$\boldsymbol{X} \cdot \boldsymbol{s}' = 0, \quad \text{(A8)}$$

with $\sum_{i=1}^{N} \boldsymbol{s'}_i = 1$ (mass conservation). This condition indicates that Eq. (A8) has a non-trivial solution $\boldsymbol{s'}^*$, which can be found by well-known procedures of linear algebra. With the solution $\boldsymbol{s'}^*$, by the first row of Eq. (A8) it is clear that

$$A'/\tau_A = \boldsymbol{X_{11}} \boldsymbol{s'}^*_1 = \sum_{i=2}^{N} \boldsymbol{X_{1i}} \boldsymbol{s'}^*_i = S'/\tau_S, \quad \text{(A9)}$$

where $S' = \sum_{i=2}^{N} \boldsymbol{s'}^*_i$, denoting the total responses of all the surface carbon reservoirs; and $\tau_S = S'/\sum_{i=2}^{N} \boldsymbol{X_{1i}} \boldsymbol{s'}^*_i$, representing the "bulk" response time constant of the surface

carbon reservoirs. Eq. (A10) is the same as Eq. (3b) in the main text, that is, the relationship is accurate for general carbon dynamic system if we have complete knowledge of $\boldsymbol{X}$.

Because in reality we do not have accurate information of $\boldsymbol{X}$, the estimates of τ_S (and τ_A) are associated with uncertainties. To illustrate the difficulty of the problem, we consider a particular N-box system that includes only the atmosphere and the global oceans at different layers. Let $\boldsymbol{s'}_1 = A'$ and $\boldsymbol{s'}_i$ $(i = 2, \cdots, N)$ denote different ocean layers with increasing "i" indicating increasing ocean depth. We further assume that each ocean layer only interacts with its neighbors and the atmosphere only interacts with the surface ocean layer. As such, the state matrix $\boldsymbol{X}$ can be written as:

$$\boldsymbol{X} = \begin{bmatrix} -\alpha_1 & \beta_2\alpha_2 & 0 & 0 & \cdots & 0 \\ \alpha_1 & -\alpha_2 & \beta_3\alpha_3 & 0 & \cdots & 0 \\ 0 & (1-\beta_2)\alpha_2 & -\alpha_3 & \beta_4\alpha_4 & \cdots & \vdots \\ 0 & 0 & (1-\beta_3)\alpha_3 & \ddots & \ddots & 0 \\ \vdots & \vdots & \vdots & \ddots & -\alpha_{N-1} & \alpha_N \\ 0 & 0 & \cdots & 0 & (1-\beta_{N-1})\alpha_{N-1} & -\alpha_N \end{bmatrix}, \quad \text{(A10)}$$

where $\alpha_i = 1/\tau_i$ and β_is are constant numbers in the range [0, 1]. Thus $\beta_i$and (1-β_i) reflect the relative weights of the carbon efflux of i-th reservoir to its two neighbors. Note that $\beta_1 = 0$ and $\beta_N = 1$.

Solving the steady-state equation (A8) with the $\boldsymbol{X}$ of Eq. (A10), we obtain the results

$$\boldsymbol{s'^*}_i = \prod_{j=2}^{i} \frac{(1-\beta_{j-1})}{\beta_j} \cdot \frac{\alpha_1 A'}{\alpha_i} = \prod_{j=2}^{i} \frac{(1-\beta_{j-1})}{\beta_j} \cdot \frac{\tau_i A'}{\tau_1}. \quad \text{(A11a)}$$

and by Eq. (A9), we estimate τ_S to be

$$\tau_S = \sum_{i=2}^{N} \left(\prod_{j=2}^{i} \frac{(1-\beta_{j-1})}{\beta_j} \cdot \tau_i \right). \quad \text{(A11b)}$$

Therefore, τ_S is not only a function of τ_i but also of β_j. In particular, the (relative) sensitivity of τ_S to β_2 is

$$(d\tau_S/d\beta_2)/\tau_S = -1/[\beta_2(1-\beta_2)]. \quad \text{(A12)}$$

Because of the characteristic buffering effect of the ocean carbonate chemistry, the anomalous carbon exchange (induced by anthropogenic disturbances) between ocean surface and the atmosphere is much (~10 times) more effective than between ocean surface and deep oceans (Gruber and Sarmiento 2002). This means that the value of β_2 is close to 0.9 or $(d\tau_S/d\beta_2)/\tau_S \approx 10$. Therefore, a 1% uncertainty in β_2 alone could induce 10% uncertainty in τ_S (or S')!

The above example highlights the challenge in estimating τ_S and thus the long-term response of the atmospheric CO_2 to anthropogenic emission disturbances. This problem is particularly emphasized in the main text (Line 10-26, Page 13968). We argued that the problem is mainly induced by the limited observations of the global climate-carbon system such that our knowledge of the state matrix $\boldsymbol{X}$ is incomplete.

A2.3 Generalization of Eq. (6b) – the long-term responses of atmospheric CO_2 to a unit step change of global surface temperature

It is clear that the steady state Eq. (A1) for temperature disturbance is

$$-\boldsymbol{X}\cdot\boldsymbol{s}' = \beta_T T'\cdot\boldsymbol{y}. \quad \text{(A13a)}$$

By the notions developed in Eq. (A9), we can represent the first row of (A13a) as

$$\alpha_A A' - \alpha_S S' = \beta_T T', \quad \text{(A13b)}$$

where $\alpha_A = 1/\tau_A$ and $\alpha_S = 1/\tau_S$.

Because changes in temperature do not induce changes in the total carbon of the system, it means that

$$A' + S' = 0, \text{ or, } A' = -S'. \quad \text{(A14)}$$

Substituting Eq. (A14) in Eq. (A13b) and rearranging the items, we obtain

$$A' = \beta_T/(\alpha_A + \alpha_S) \cdot T' \approx \beta_T/\alpha_A \cdot T', \quad \text{(A15)}$$

for generally $\alpha_A \gg \alpha_S$. This is the same equation as Eq. (6b) in the main text.

Figures

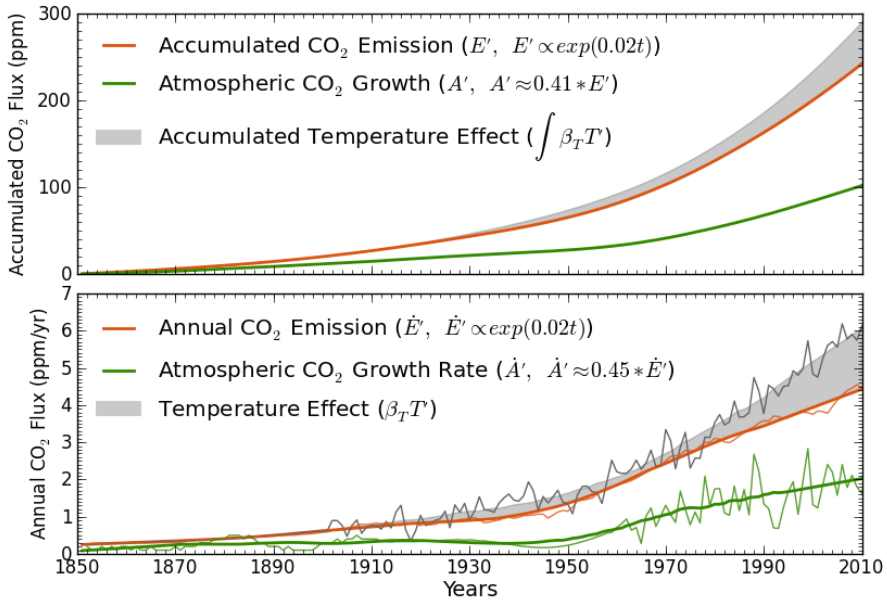

Fig. 1 Time series of global anthropogenic CO_2 emissions (red line), atmospheric CO_2 concentrations (green line), and the anomalous CO_2 fluxes induced by warming surface temperatures (gray shade) between 1850 and 2010. The Top panel indicates the accumulated CO_2 fluxes or the total concentration changes while the Bottom panel shows them at annual steps. The thick and the thin lines indicate long-term and interannual variations of the time series, respectively. The mathematical symbols are the same as in Eq. (1) and explained in the text. In both annual and accumulative cases, CO_2 emissions largely increase as an exponential function of time, while changes in the atmospheric CO_2 concentrations are proportional to the corresponding emissions by a factor about 0.41-0.45.

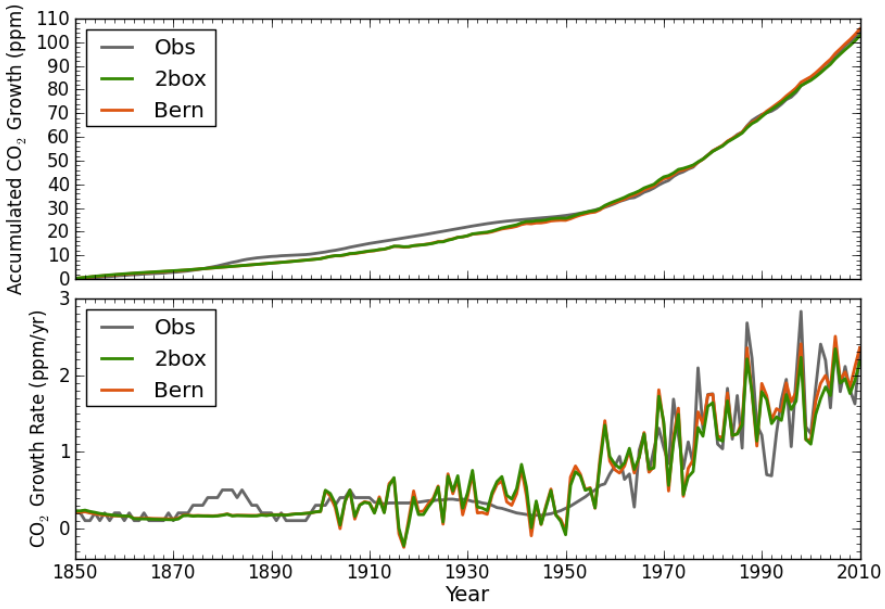

Fig. 2 Simulations of the observed atmospheric CO_2 concentrations (Top Panel) and growth rates (Bottom Panel) from anthropogenic CO_2 emissions and land-surface air-temperature data using the two-box model ("2box") and the revised Bern model ("Bern"). The atmospheric CO_2 concentration in 1850 (i.e., 284.7 ppm) is used as the initial condition for the model integration. Long-term mean temperature before 1901 is assumed to be stable and represented by the 1901-1920 mean. Other model parameters used in these simulations are explained in the main text (the two-box model) or the Appendix (the revised Bern model).

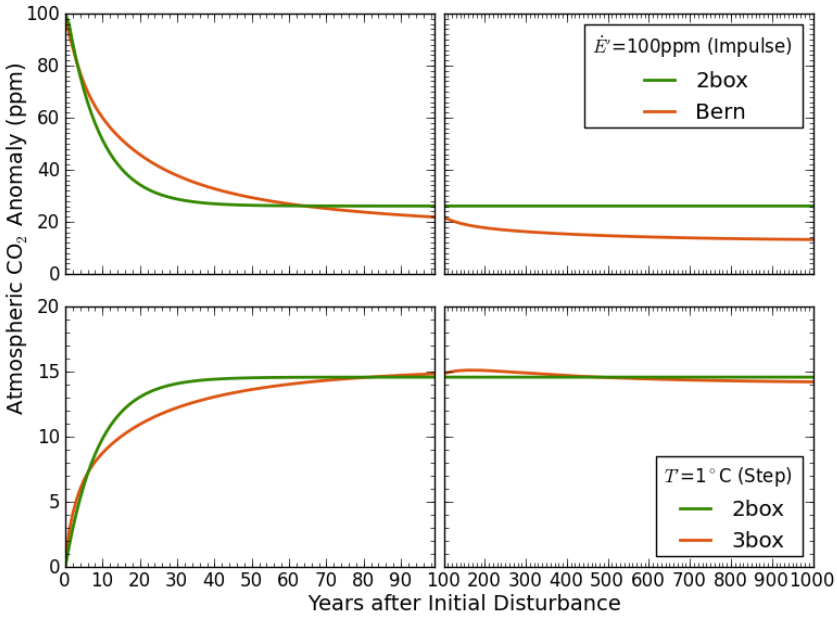

Fig. 3 Disturbance-response functions of the atmospheric CO_2 concentration simulated by the two-box model ("2box") and the revised Bern model ("Bern"). The Top panel shows the responses of atmospheric CO_2 concentration to an impulse increase (of 100 ppm) in anthropogenic CO_2 emissions and the Bottom panel shows the corresponding responses to a step increase (of 1 ^{o}C) in surface temperatures.

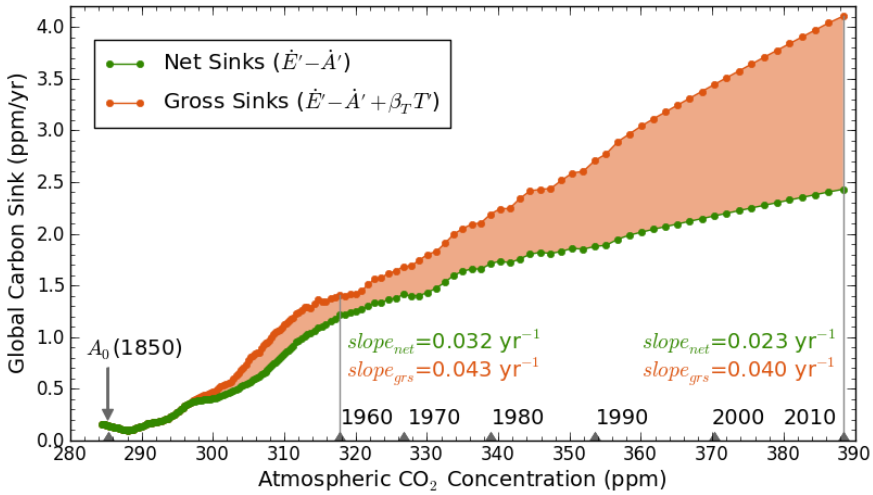

Fig. 4. Global annual carbon sinks (ppm/yr) as a function of atmospheric CO_2 concentration from 1850 to 2010. The green dots indicate the observed "net" carbon sinks and the red dots indicate the "gross" carbon sinks that accounted for the effects of temperature changes (Eq. 7). The differences between the gross and the net carbon sinks (the shaded area) indicate the extra carbon fluxes released into the atmosphere as a result of warming temperatures (Fig. 1). The gray arrow ("A_0") indicates the estimated atmospheric CO_2 level (284.7 ppm) that was stable at pre-industrial CO_2 emission rates and climate conditions. The slopes between the global annual carbon sinks and corresponding changes in atmospheric CO_2 concentration (relative to A_0) generally reflect carbon-sequestration efficiencies of global land and ocean reservoirs.

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