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## Climate and carbon cycle representation in Integrated Assessment Models

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Carbon cycle, Climate model, Integrated assessment models (IAMs), Impulse response substitute model, pattern scaling

### Abstract

This article presents the concept of a simple, cost-efficient substitute model to simulate the link between carbon emissions, atmospheric CO<sub>2</sub>, radiative forcing, and climate change. The basic idea is to represent the response of complex Earth System models for key variables by a few equations only. The model consists of a series of one-box models each with a single overturning time scale. The latter may be obtained from impulse response functions of the parent model and they characterize the time scales of surface-to-deep transport of carbon and heat in the ocean, and of the flow of carbon through the land biosphere. Non-linearities associated with the marine carbonate chemistry, air-sea fluxes of carbon and heat, land productivity, and radiative forcing by CO<sub>2</sub> and other agents are described separately. A pattern scaling approach is used to simulated regional changes in temperature and other climate variables. The current perturbation of the climate system by anthropogenic carbon emissions is discussed. It is shown that carbon emissions have climatic impacts that are irreversible on human time scales. Substitute models that capture fundamental system behavior are now implemented in many Integrated Assessment Models. On the other hand, the DICE model fails to represent the longevity of the climate perturbation associated with anthropogenic carbon emissions; this implies that analyses based on this model may be biased.

Understanding the coupled climate-carbon cycle and its perturbation by man is of utmost importance in any integrated assessment of future change. Anthropogenic emissions of carbon and other radiative forcing agents force the atmospheric composition, Earth's climate, the land surface and terrestrial biosphere and the chemical state of the ocean towards conditions that have probably not occurred over the past 20 million years (1) and at a speed that is unprecedented at least during the last 20,000 years (2). Today's atmospheric carbon dioxide (CO<sub>2</sub>) concentration of almost 390 ppm is far above the natural range (172 to 300 ppm) of the last 800,000 years (3). Ocean acidification by the uptake of CO<sub>2</sub> is becoming widespread (4) and is imminent in the Arctic (5). CO<sub>2</sub> is projected to rise, surface temperature to warm, and climate to change as carbon emissions continue for a range of business-as-usual as well as for emission mitigation scenarios (6).

One way to improve our quantitative understanding of the intricate coupling between the physical climate system, the carbon and other biogeochemical cycles and the human society is to build integrated assessment models (IAMs). The goal of this article is (i) to highlight critical issues related to the anthropogenic perturbation of climate and the treatment of climate and carbon cycle model components in IAMs and (ii) to describe the representation of the climate-carbon system using an Impulse Response Function approach. First some context for the carbon cycle climate model development is presented. A number of questions arise and are discussed in the next two sections: Which are the characteristics of the anthropogenic CO<sub>2</sub> perturbation? Why is it important to include carbon cycle-climate modules into IAMs? What is the nature of the coupled carbon cycle-climate system? What is the range of space and time scales, the number of processes, linkages, and feedbacks involved? Addressing these questions will allow us to better understand the challenges associated with IAM development and the different model strategies invoked by different groups. Then, an impulse response function and pattern scaling approach is presented in more detail followed by discussion and conclusions.

## **Anthropogenic CO<sub>2</sub> in the climate system**

Several characteristics render CO<sub>2</sub> by far the most important anthropogenic greenhouse gas. The human-caused perturbation in atmospheric CO<sub>2</sub> is two to five orders of magnitude larger than those in the concentrations of other anthropogenic forcing agents such as methane, nitrous oxide, ozone, chlorofluorocarbons or SF<sub>6</sub>. Its contribution to the radiative perturbation of the climate system, measured as radiative forcing, is currently about 56% of that by all anthropogenic greenhouse gases (7) and its share is projected to increase further (8). Most of the projected warming for a range of new reference and mitigation scenarios developed in preparation of the IPCC Fifth Assessment Report is attributable to CO<sub>2</sub>. Most of the warming avoided in mitigation relative to reference scenarios is also attributable to CO<sub>2</sub> (8).

Fossil fuel use is the dominant anthropogenic carbon source (<http://www.globalcarbonproject.org>). Emissions from fossil fuel use (plus emissions from cement production) increased from 6.2 gigatons of carbon (1 GtC=10<sup>12</sup> kg of carbon=3.7

$10^{12}$  kg of CO<sub>2</sub>) per year in 1990 to 8.7 GtC in 2008, a 41% increase from the Kyoto reference year 1990. Land use change is responsible for estimated net emissions of 1.5 GtC per year to the atmosphere, mainly from tropical deforestation. In total about 500 GtC has been released to the atmosphere over the period 1850 to 2007, with contributions of 350 GtC from fossil fuel use and 160 GtC from land use changes.

The human-made perturbation in CO<sub>2</sub> and thus in climate is persistent and irreversible on human time scales (9, 10). Figure 1 illustrate the millennial-scale evolution of atmospheric CO<sub>2</sub> and global mean surface temperature for a scenario where all conventional fossil fuel resources are released into the atmosphere and for a scenario where emissions are limited to 1000 GtC, i.e. the cumulative emissions of ambitious mitigation pathways. In both cases, the perturbation in atmospheric CO<sub>2</sub> and global mean surface temperature are irreversible on human time scale. CO<sub>2</sub> released by fossil fuel burning accumulates in the climate system. The perturbation in atmospheric CO<sub>2</sub> is ultimately removed by interactions with the weathering cycles on the time scales of order 100,000 years (11). CO<sub>2</sub> is chemically stable in the atmosphere and not destroyed through reactions with oxidants or deposited on the ground as other agents, instead CO<sub>2</sub> is redistributed among major reservoirs of the Earth system, - the ocean and ocean sediments, vegetation and soils. Consequently, the removal of a CO<sub>2</sub> perturbation from the atmosphere is not dictated by a single removal or reaction time scale, but governed by the cascades of time scales associated with the overturning of carbon within the land biosphere and the ocean and carbon exchange with ocean sediment. It is noted that no scheme is currently operational to remove CO<sub>2</sub> artificially from the atmosphere in suitable amounts, though efforts in this direction are undertaken (12).

The multi-millennial perturbation life time of CO<sub>2</sub> has consequences. For a stabilization of atmosphere CO<sub>2</sub> and its related climatic perturbation it is not sufficient to stabilize emissions, but anthropogenic net emissions (release minus any human-induced removal) have to fall well below current emissions and eventually towards the magnitude of the geological sink strengths of order a few tens of gigatons of carbon (1). For comparison, other quantitatively relevant forcing agents such as aerosols, ozone, methane, nitrous oxide or chlorofluorocarbons, have lifetimes of order days to weeks (tropospheric ozone), a decade (methane), or a century (nitrous oxide, CFC-11, CFC-12). Thus in contrast to CO<sub>2</sub>, atmospheric concentrations of these constituents tend to stabilize if we manage to stabilize related emissions. On the other hand, the climate impacts from non-CO<sub>2</sub> emission reductions will sooner or later be overwhelmed by the climate forcing from rising CO<sub>2</sub> in case we fail to reduce fossil carbon emissions towards low values.

The long perturbation life time of CO<sub>2</sub> (and the long time scales of other climate processes such as ice sheet melting and thermal sea level rise) is at the heart of a fundamental debate in the community assessing the economy of climate change and mitigation options. This debate revolves around the question of how intergenerational equity should be considered. Or phrased differently, what is the appropriate discount rate to compare the costs of investments for emission mitigation and of climate change damage and adaptation? We note that the difference between a discount rate of 1% per year versus a "market-based" discount rate of 5% per year correspond to factor of 49 and  $10^{17}$  when evaluating the costs that CO<sub>2</sub> emission cause in 100 and 1000 years,

respectively. The long time scales involved also implies that conclusions of mitigation and impact analyses that are restricted to the 21<sup>st</sup> century only could potentially be affected by the limited time horizon of the analysis or even be misleading. It is current practice to stop IAM simulations by 2100 or 2200.

## The coupled climate-carbon cycle system

Figure 2 depicts a highly simplified scheme of the coupled climate-carbon cycle-socio-economic system as implemented into various IAMs. Some of the most pertinent couplings are identified by red arrows and carbon fluxes indicated in blue. Fossil fuel use and land use change by humans cause atmospheric CO<sub>2</sub> to increase, which in turn results in a perturbation of the Earth's radiative balance. This radiative forcing by CO<sub>2</sub> and other agents causes climate to change. Climate change in turn affects our society, leading to adaptation and mitigation measures. Climate change also affects the functioning of natural systems for example through changes in currents, temperature and salinity field in the ocean and through changes in the frequency and amount of precipitation, evaporation, and temperature on land.

Each of the individual boxes and arrows shown in Figure 2 represents in reality a multitude of processes and feedbacks that operate on a wide range of time and space scales. For example, we may refer to the atmosphere where climate change is expressed in changes in the space-time variations of temperature or atmospheric humidity or changes in reaction kinetics of atmospheric constituents or the frequency and amount of precipitation or the severity of droughts at the atmosphere-land interface. System time lags, hysteresis behavior, variability around a mean state or between different preferred modes of operation are inherent to such a complex system.

Conservation of mass and energy imposes strong constraints on system behavior. A rich set of observations over the instrumental period as well as extending back into time over many glacial cycles and beyond permit us to confront models with real world data. Instrumental observation (7), model evaluation (13), and paleosciences (14) are all relevant to constrain climate projections. We know that the planet warmed with increasing radiative forcing since the last glacial maximum (20,000 years ago) and that the warming is quantitatively consistent with our understanding of the links between the various climate forcings and the resulting climate change. Uncertainties remain. A way to deal with uncertainties is to focus on probabilities of outcome (15) instead of a relying on a deterministic view. That is not to predict what will exactly happen, but to assess the chance or risk of a certain outcome.

Many of the processes of the climate system might be included into IAMs and several IAM groups incorporate already complex modules to represent certain aspects, e.g. land use, of the Earth System. Increased complexity comes usually in exchange with reduced cost-efficiency and increased demand for computing and human resources. The choice is not only to balance the representation of system complexity versus computational and human resources, but also to which extend to include well understood processes versus more speculative mechanisms.

Two key parameters should be included in any IAM. These are atmospheric CO<sub>2</sub> and global mean temperature. Next, a representation will be described for the cause-effect chain from anthropogenic emission to atmospheric concentrations and radiative forcing to climate change. The idea is to extract essential information contained in complex models by building a cost-efficient substitute that can be applied over broad range of scenarios.

## Simple climate model, impulse response functions, and pattern scaling

The development of Earth System models is evolving and state-of-the art models are generally too expensive for incorporation into IAMs. Fortunately, the response of Earth System models to greenhouse gas emissions may be approximately represented for key variables by building cost-efficient and simple substitute models. First, theoretical concepts for the development of substitute models are presented, followed by some practical considerations. Then, the approach is extended to represent also regional changes for a range of climate variables. It is noted that the equations presented below are specifically tailored to represent more comprehensive models, but they do not describe the physical mechanisms involved. The quality of the substitute model must be evaluated against the results of its comprehensive parent model.

**Theoretical considerations – Impulse response function model:** Impulse response representations have been proven to be particularly useful to represent complex carbon cycle-climate models (16-19). The theoretical justification for the use of impulse response function (IRF) or Green's function is that the dynamic of a linear system is fully characterized by its IRF. Many complex models behave in an approximately linear way and major non-linearities may be singled out and described separately.

IRF's for atmospheric CO<sub>2</sub> are obtained by monitoring the decrease of an atmospheric CO<sub>2</sub> perturbation caused by an initial carbon input at time 0 in a complex model. Figure 1 shows the response to an input of 1000 and 5000 GtC (a uniform atmospheric perturbation of 1 ppm corresponds to 2.123 GtC) into the atmosphere. Normalizing the response to unity at time 0 yields the normalized response function,  $r_a$ , shown in Figure 3. The value of the IRF at any particular time is the fraction of the initially added carbon that is still found in the atmosphere. The atmospheric CO<sub>2</sub> concentration can then be approximately computed as the sum of earlier emissions,  $e$ , at times  $t'$  multiplied by the fraction still remaining airborne after time  $t-t'$ :

$$c_a(t) = \sum_{t_0}^t e(t') \cdot r_a(t-t') \cdot \Delta t' + c_a(t_0) \quad (1)$$

where  $t_0$  refers to the preindustrial steady-state and  $\Delta t'$  is the time step. Non-linearities in the carbonate chemistry of the ocean and terrestrial carbon uptake, as evidenced by comparing the IRFs for different pulse sizes (Figure 3), limit the accuracy of this procedure. Another shortcoming of atmospheric IRFs is that the IRFs are in general different for different tracers.

Next, an alternative to equation (1) is developed that has been shown to represent accurately the relationship between carbon emissions and atmospheric CO<sub>2</sub> as simulated by spatially-resolved, complex models. The problems associated with the non-linear carbonate chemistry and the air-sea coupling can be avoided by using an IRF for the surface ocean layer in combination with appropriate budget equations. The transport of excess CO<sub>2</sub> and other conservative tracers within the ocean is described by a set of linear equations and can therefore be represented by IRFs, at least under the assumption of a constant ocean circulation. The evolution of the surface water concentration in dissolved inorganic carbon,  $c_s$ , can then be computed in analogy to equation 1. The perturbation in  $c_s$  is the sum of earlier net air-to-sea fluxes,  $f_{as}$ , multiplied by the appropriate value of the surface ocean IRF,  $r_s$ :

$$c_s(t) = \frac{1}{A_{oc}h} \sum_{t_0}^t f_{as}(t') \cdot r_s(t-t') \cdot \Delta t' + c_s(t_0) \quad (2)$$

Here,  $r_s$  represents the fraction of earlier input fluxes still present in the well-mixed surface layer with height  $h$ . The net flux of carbon into the ocean is a function of the gas transfer velocity,  $k_g$ , and the partial pressure difference between surface air and seawater:

$$f_{as} = A_{oc} \cdot k_g \cdot \alpha \cdot (\delta pCO_{2,a} - \delta pCO_{2,s}) \quad (3)$$

where  $\delta pCO_2$  represents the perturbation in the CO<sub>2</sub> partial pressure from preindustrial equilibrium and  $A_{oc}$  the ocean surface area and  $\alpha$  is a unit conversion factor. The relationship between the change in surface water pCO<sub>2</sub> and  $c_s$  can be described by a simple analytical expression (see e.g. (16)).

The net uptake of carbon by the land biosphere can be treated using a decay response function,  $r_{b,decay}$ , that describes the fraction of carbon entering the land biosphere by net primary productivity,  $f_{npp}$ , that is released back to the atmosphere in later years. We obtain for the net flux into the land biosphere,  $f_{ab}$ :

$$f_{ab}(t) = \delta f_{npp}(t) - \sum_{t_0}^t \delta f_{npp}(t') \cdot r_{b,decay}(t-t') \cdot \Delta t' \quad (4)$$

The evolution of the terrestrial carbon stock,  $B$ , can be described in analogy to equations (1) and (2) by:

$$B(t) = \sum_{t_0}^t f_{npp}(t') \cdot r_b(t-t') \cdot \Delta t' + B(t_0) \quad (5)$$

, where the IRF for the terrestrial inventory is closely related to the decay response function as further discussed below.

The global net primary productivity of a complex model can often be adequately parameterized by analytical functions, e.g. of CO<sub>2</sub> and global mean surface temperature (see e.g. (18)). Finally, we compute the atmospheric concentration from the budget equation:

$$\varepsilon \frac{d}{dt} \delta p CO_{2,a} = e(t) - f_{as,net}(t) - f_{ab,net}(t) \quad (6)$$

$\varepsilon$  is a unit conversion factor; 1  $\mu\text{atm CO}_2$  corresponds to 1 ppm for a surface pressure of 1 atm and in turn to an atmospheric inventory of 2.123 GtC when assuming a uniform perturbation in the mixing ratio (relative to dry air).

The radiative influence of  $\text{CO}_2$  and other agents can be adequately approximated by simplified expressions that have to be shown to represent the output of spatially-resolved radiative transfer models within a few percent (20). The radiative forcing by  $\text{CO}_2$  is:

$$RF_{CO_2} = 5.35 \text{ W m}^{-2} \ln \frac{\delta CO_{2,a} + CO_{2,a}(t_0)}{CO_{2,a}(t_0)} \quad (7)$$

This yields for a doubling of atmospheric  $\text{CO}_2$  a radiative forcing,  $RF_{2 \times CO_2}$ , of  $3.7 \text{ W m}^{-2}$ .

The next step is to compute global mean temperature from the history of radiative forcing by  $\text{CO}_2$  and other agents,  $RF$ . At radiative equilibrium, the global average surface temperature would correspond to an equilibrium temperature,  $\delta T_{eq}$ .  $\delta T_{eq}$  is proportional to the equilibrium climate sensitivity for a nominal doubling of  $\text{CO}_2$ ,  $\Delta T_{2 \times CO_2}$ , and the imposed radiative forcing:

$$\delta T_{eq}(t) = RF(t) \cdot \frac{\Delta T_{2 \times CO_2}}{RF_{2 \times CO_2}} \quad (8)$$

The world ocean has a considerable heat capacity and this thermal inertia leads to a thermal flux into the ocean and a significant delay of surface temperature to a change in radiative forcing. The perturbation heat flux into the ocean,  $fh_{as}$ , is taken to be proportional to the deviation of the actual temperature perturbation,  $\delta T_s$  to the equilibrium temperature perturbation,  $\delta T_{eq}$ , and the radiative forcing:

$$fh_{as}(t) = \frac{RF}{a_{oc}} \cdot \left( \frac{\delta T_{eq} - \delta T}{\delta T_{eq}} \right) \quad (9)$$

$a_{oc}$  is the area fraction of the earth covered the ocean (71%).

The transient response of global average surface air temperature,  $T$ , as simulated by expensive coupled atmosphere-ocean-sea ice models, can then be represented by using the surface ocean response function to account for the downward mixing of heat in the ocean. The perturbation in the surface temperature is given by:

$$\delta T(t) = \frac{c}{h} \sum_{t_0}^t fh_{as}(t') \cdot r_s(t-t') \cdot \Delta t' \quad (10)$$

The constant  $c$  takes into account the heat capacity of sea water and includes factors for unit conversion. Note that the same surface-layer IRF,  $r_s$ , can be applied for heat and carbon.

Equations (2) to (10) together with a few analytical expressions approximate the response of coupled Earth System models to anthropogenic carbon emissions. In other words, the link between emissions, atmospheric concentration, radiative forcing, and climate change can be described by a few equations that can be solved quickly on any personal computer.

**Practical consideration - A box-model substitute:** While it is useful to consider IRFs and the convolution integrals (equations 2, 4 and 10) for theoretical insights, it is often more practicable to remap the convolution integral and the response functions to a multi-box model, thereby avoiding the repeated calculation of the sum in equations 2, 4, and 10. The IRFs of complex models can be fitted by a series of exponentials:

$$r(t) = \sum_i a_i \exp\left(-\frac{t}{\tau_i}\right) \quad (11)$$

The time scales  $\tau_i$  are indicative of the time scales of the perturbation, but they should not be interpreted literally; they are fitting parameters. The sum of all coefficients  $a_i$  equals unity and typically three to four coefficients are sufficient to fit the original IRF.

It can be easily shown that the convolution integrals in combination with equation (11) correspond to a series of independent box models with box  $i$  having an overturning time scale  $\tau_i$  and a share  $a_i$  of the total flux. For example, to compute the concentration of carbon in surface water,  $c_s$ , we solve the following differential equations:

$$A_{oc} h \frac{d}{dt} c_s(t) = \sum_{i=1}^n \frac{d}{dt} N_i(t) \quad (12)$$

where  $N_i$  is the amount of carbon in box  $i$ . and the change of  $N_i$  is the input flux allocated minus the first order decay:

$$\frac{d}{dt} N_i = a_i \cdot A_{oc} \cdot f_{as}(t) - \frac{1}{\tau_i} N_i(t) \quad (13)$$

Analogous box models are readily formulated for the terrestrial carbon uptake and for global mean surface temperature.

Coefficients of IRFs for the ocean and the land biosphere can be found for a range of models in the literature, where also unit conversion factors, constants, and initial conditions are provided (16, 18). It is noted, that there is a close link between the decay response function,  $r_{b,decay}$ , and the response function,  $r_b$ , describing the perturbation in the net primary production and the land carbon inventory, respectively. If  $r_b$  is of the form given by eq. (11), it follows:

$$r_{b,decay}(t) = \sum_i \frac{a_i}{\tau_i} \exp\left(-\frac{t}{\tau_i}\right) \quad (14)$$



**Practical consideration – Long time step:** The equilibration time scale of the surface layer with respect to air-sea carbon fluxes is about 1 year. This is shorter than the typical time step of 10 years applied in many IAMs. Then, it might be adequate to consider the surface layer and the atmosphere as one well mixed box to avoid numerical instabilities. The sum of the anthropogenic emissions and the net flux into the biosphere is immediately redistributed between the atmosphere and the ocean surface layer. Thus, we replace equation (6) by:

$$\frac{d}{dt}(A_{oc} \cdot h \cdot c_s + pCO_{2,a} \cdot \varepsilon) = \frac{dc_s}{dt} \cdot (A_{oc} \cdot h + \gamma \cdot \varepsilon) = e(t) - f_{ab,net}(t) \quad (15)$$

Where  $\gamma$  represents the ratio between the change in the atmospheric partial pressure and in the concentration of dissolved inorganic carbon in the ocean at prevailing conditions:

$$\gamma = \frac{\partial pCO_{2,a}}{\partial c_s} \quad (16)$$

$\gamma$  can be computed from carbonate chemistry models assuming equilibrium in surface ocean partial pressure and atmospheric partial pressure. Such models are available on the web (e.g. <http://cdiac.ornl.gov/oceans/co2rprt.html> or <http://www.ipsl.jussieu.fr/OCMIP/> (follow the links OCMIP3-HOWTO-interannual variability) In practice, the calculation may be done once and the results stored in a look-up table to provide the link between  $\delta c_s$  and  $\delta pCO_{2s}$  and  $\gamma$ ; mean ocean surface temperature, alkalinity, and dissolved inorganic carbon concentration at preindustrial time  $t_0$  are specified for this calculation. Similarly, the amount of carbon stored in box  $i$  may be assumed to be in equilibrium with the input flux, when the overturning time scale  $\tau_i$  of this box is shorter than the time step. The equilibrium concentration follows from equation (13) by setting the time derivative to zero.

**Practical consideration – accounting for additional non-linearities:** IRFs and IRF-based substitute model are able to accurately mimic a fully linear system. Above, we have replaced the concept of an atmospheric IRF model by an ocean surface layer and land biosphere IRF model to include the major carbon-cycle non-linearities associated with the ocean chemistry and with net primary productivity on land. One may extend the approach and account for additional non-linearities. This can be relatively easily achieved in the IRF-based box substitute described above.

For example, the overturning time scales of soil carbon depend on soil temperature and the surface-to-deep mixing in the ocean is expected to slow under global warming. Such effects may be taken into account in an approximate way by making the overturning time scales  $\tau_i$  dependent on global average surface temperature. Meyer et al (18) demonstrated that the response of spatially-resolved and climate sensitive terrestrial models can be accurately described by an IRF/box-substitute model; the allocation factors,  $a_i$ , and the overturning time scales are taken to vary with global mean surface temperature.

It is also well known, that the partial pressure of CO<sub>2</sub> in surface water is increasing with increasing temperature. This non-linearity may be represented in a first order approximation by modifying  $\delta pCO_{2s}$  as follows:

$$\delta pCO_{2,s}(\delta T) = (pCO_{2,0} + \delta pCO_{2,s}(T_0)) \cdot \exp(0.0423 \text{ K}^{-1} \cdot \delta T) - pCO_{2,0} \quad (15)$$

**Regional changes and changes in other variables:** There is an increasing need not only to consider mean climate indicators such as global mean average surface temperature, but also to consider regional information and changes in other climatic variables in IAMs. Regional changes in temperature, precipitation, or any other variables (including extreme event statistics) can be represented by a pattern scaling approach. The basic idea is to describe the spatio-temporal climate change signal from simulations with a complex Earth System model by a time-independent spatial pattern and the temporal evolution of a climate indicator, here global mean surface temperature.

One approach is to apply an empirical orthogonal function analysis (EOF). The perturbation of any variable  $\delta v$  is represented as the superposition of a set of mutually orthogonal spatial patterns,  $EOF_i^v(\mathbf{x})$  and the time-dependent scalar coefficients termed principal components,  $PC_i^v(t)$ :

$$\delta v(x, t) = \sum_i PC_i^v(t) \cdot EOF_i^v(x) \quad (16)$$

Each of the PC-EOF pairs is computed in successive order to explain the maximum possible variance in a climate variable. To make the approach viable in practice, it should be possible to explain most of the changes in  $v$  associated with global warming by the first one or two PC-EOF pairs only and the temporal evolution of these PCs should be a function of global mean surface temperature, at least for the scenario range under consideration. Hooss and coauthors (21) showed that mean temperature and precipitation changes related to global warming can be represented by one EOF-PC pair in their model.

An alternative is to compute the spatial pattern,  $SP(\mathbf{x})$ , as the difference between the beginning and the end of a simulation and to normalize the field by dividing by the simulated global mean temperature change. Then, the perturbation in a variable is expressed as the product of global mean surface temperature change (equation 10) and the fixed spatial pattern:

$$\delta v(x, \delta T) = \delta T(t) \cdot SP(x) \quad (17)$$

Obviously, it needs to be demonstrated that the pattern scaling is approximately valid over the range and for the variable of interest. This is approximately the case for annual or monthly mean temperature and precipitation.

## Discussion and Conclusion

In the previous paragraph, a box-substitute model approach based on impulse response function in combination with a climate change pattern scaling approach was outlined to describe the link between carbon emissions, atmospheric CO<sub>2</sub>, radiative forcing, global mean surface temperature and regional changes in different variables. The approach is transparent, cost-efficient, and simple, involving a limited set of equations in combination with look-up tables for the carbonate chemistry. The necessary parameters, such as overturning time scales or climate change pattern can be derived and validated from suitable simulations with complex Earth System Models. For specific numerical values, the reader is referred to the literature (16, 18). The model can be readily extended to describe the radiative effects of non-CO<sub>2</sub> agents as outlined elsewhere (19, 22) with a recent update of parameters for non-CO<sub>2</sub> agents (23).

The substitute model approach has limitations. Potential major abrupt and or chaotic regime shifts in ocean circulation or in other climatic components can not be readily represented. Uncertainties in mapping the response of a complex model to its substitute should be put in the context of the uncertainties inherent also in any Earth System model and as illustrated by the range of projections obtained with different models. It has been shown in previous work that substitute models can relatively accurately represent the response of complex models over the current scenario range. Substitute model components developed in Bern have meanwhile been implemented in a range of IAMs including the MESSAGE, and the IMAGE model.

First-order physico-chemical principles and basic knowledge from Earth System science are not yet taken into account in all IAM approaches. The first modules embedded in IAMS of climate change deal with the projection of CO<sub>2</sub> from emissions. Nordhaus (24) developed a substitute for a full carbon cycle model consisting of a single equation and describing the relationship between carbon emissions and atmospheric CO<sub>2</sub> in a simple way. 36% of the carbon emissions into the atmosphere are disappearing instantaneously, while the atmospheric carbon perturbation is removed according to a first order decay with a decay rate,  $\lambda$ , of 1/120 years in his model. Because of its simplicity and convenience, variants of this parameterization have been used in the DICE model of Nordhaus and co-workers as well as in many economic analyses (e.g. (25, 26)). Unfortunately, this parameterization significantly underestimates atmospheric CO<sub>2</sub>, as already evident from the available evidence in the 1980ies (10, 27), and highlighted by (28). It was also shown that this simple parameterization causes a large bias towards high carbon emissions and low abatement levels in economic analyses of optimal carbon emission pathways for a range of utility discount rates (29). Later, Nordhaus and co-workers replaced the parameterization by a three reservoir carbon model fitted to the output of a version of the Bern carbon cycle model (16) and later to the MAGGICC model (30)). Unfortunately, also this current representation does not represent the time scales of the carbon cycle-climate system in a realistic way as shown in Figure 1. Atmospheric CO<sub>2</sub> and climate change is decreasing far too fast and to far too low levels compared to current state-of-the-art-models (9, 31) that have been evaluated with observations.

The number of processes and their coupling through space and time represent a formidable challenge to all natural scientists, including climate and Earth System modelers. This challenge does not become smaller when addressing the linkages with

the socio-economic system as done in IAMs. It is a must to confront results of IAMs with real world data. It is arguably also important for the IAM modeling community to go a step beyond the current approach of linking one particular scenario story line with one particular outcome for example in terms of greenhouse gas emissions (REF SRES or similar) and to progress to probabilistic approaches. Despite difficulties involved how to compare costs over time, the irreversibility associated with anthropogenic carbon emissions and the millennial-scale life time of the climatic perturbations should be given ample attention in any assessment of the benefits of carbon mitigation. Cost-efficient impulse response function/box-substitute models may facilitate work in these directions.

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## References

1. Prentice, I. C., Farquhar, G. D., Fasham, M. J., Goulden, M. I., Heimann, M., Jaramillo, V. J., Khashgi, H. S., LeQuéré, C., Scholes, R. J. & Wallace, D. W. R. (2001) in *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, eds. Houghton, J. T., Ding, Y., Griggs, D. J., Noguer, M., van der Linden, P. J., Dai, X., Maskell, K. & Johnson, C. A. (Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA), pp. 183-237.
2. Joos, F. & Spahni, R. (2008) Rates of change in natural and anthropogenic radiative forcing over the past 20,000 years. *Proceedings of the National Academy of Sciences of the United States of America* 105: 1425-1430.
3. Lüthi, D., Floch, M. L., Bereiter, B., Blunier, T., Barnola, J. M., Siegenthaler, U., Raynaud, D., Jouzel, J., Fischer, H., Kawamura, K. & Stocker, T. F. (2008) High-resolution carbon dioxide concentration record 650'000-800'000 years before present. *Science* 453: 379-382.
4. Orr, J. C., Fabry, V. J., Aumont, O., Bopp, L., Doney, S. C., Feely, R. M., Gnanadesikan, A., Gruber, N., Ishida, A., Joos, F., Key, R. M., Lindsay, K., Maier-Reimer, E., Matear, R., Monfray, P., Mouchet, A., Najjar, R. G., Plattner, G.-K., Rodgers, K. B., Sabine, C. L., Sarmiento, J. L., Schlitzer, R., Slater, R. D., Totterdell, I. J., Weirig, M.-F., Yamanaka, Y. & Yool, A. (2005) Anthropogenic ocean acidification over the twenty-first century and its impact on calcifying organisms. *Nature* 437: 681-686.
5. Steinacher, M., Joos, F., Frölicher, T. L., Plattner, G.-K. & Doney, S. C. (2009) Imminent ocean acidification in the Arctic projected with the NCAR global coupled carbon cycle-climate model. *Biogeosciences* 6: 515-533.
6. Van Vuuren, D. P., Meinshausen, M., Plattner, G.-K., Joos, F., Strassmann, K. M., Smith, S. J., Wigley, T. M. L., Raper, S. C. B., Riahi, K., de la Chesnaye, F., den Elzen, M., Fujino, J., Jiang, K., Nakicenovic, N., Paltsev, S. & Reilly, J. M. (2008) Temperature increase of 21st century stabilization scenarios. *Proc Natl Acad Sci USA*: in press.

7. Solomon, S., Qin, D., Manning, M., Alley, R., Berntsen, T., Bindoff, N. L., Chen, Z., Chidthaisong, A., Gregory, J., Hegerl, G., Heimann, M., Hewitson, B., Hoskins, B., Joos, F., Jouzel, J., Kattsov, V., Lohmann, U., Matsuno, T., Molina, M., Nicholls, N., Overpeck, J., Raga, G., Ramaswamy, V., Ren, J., Rusticucci, M., Somerville, R., Stocker, T. F., Stouffer, R. J., Whetton, P., Wood, R. A. & Wratt, D. (2007) in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, eds. Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M. & Miller, H. L. (Cambridge University Press, Cambridge United Kingdom and New York, NY, USA), pp. 19-91.
8. Strassmann, K. M., Joos, F. & Plattner, G. K. (2009) CO<sub>2</sub> and non-CO<sub>2</sub> radiative forcing agents in twenty-first century climate change mitigation scenarios. *Climate Dynamics* 33: 737-749.
9. Frölicher, T. L. & Joos, F. (2009) Reversible and irreversible impacts of greenhouse gas emissions in multi-century projections with a comprehensive climate-carbon model. *Climate Dynamics*: in press.
10. Siegenthaler, U. & Oeschger, H. (1978) Predicting future atmospheric carbon dioxide levels. *Science* 199: 388-395.
11. Archer, D., Kheshgi, H. & Maier-Reimer, E. (1999) Dynamics of fossil fuel CO<sub>2</sub> neutralization by marine CaCO<sub>3</sub>. *Global Biogeochemical Cycles* 12: 259-276.
12. IPCC (2005) *IPCC Special Report on Carbon Dioxide Capture and Storage* (Cambridge University Press, Cambridge, UK).
13. Randall, D. A., Wood, R. A., Bony, S., Colman, R., Fichet, T., Fyfe, J., Kattsov, V., Pitman, A., Shukla, J., Srinivasan, J., Stouffer, R. J., Akimasa, S. & Taylor, K. E. (2007) in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, eds. Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M. & Miller, H. L. (Cambridge University Press, Cambridge United Kingdom and New York, NY, USA), pp. 589-662.
14. Jansen, E., Overpeck, J., Briffa, K. R., Duplessy, J.-C., Joos, F., Masson-Delmotte, V., Olago, D., Otto-Bliesner, B., Peltier, W. R., Rahmstorf, S., Ramesh, R., Raynaud, D., Rind, D., Solomina, O., Villalba, R. & Zhang, D. (2007) in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, eds. Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M. & Miller, H. L. (Cambridge University Press, Cambridge United Kingdom and New York, NY, USA), pp. 433-497.
15. Knutti, R., Joos, F., Müller, S. A., Plattner, G. K. & Stocker, T. F. (2005) Probabilistic climate change projections for stabilization profiles. *Geophysical Research Letters* 32: doi:10.1029/2005GL023294.
16. Joos, F., Bruno, M., Fink, R., Siegenthaler, U., Stocker, T. F. & LeQuere, C. (1996) An efficient and accurate representation of complex oceanic and biospheric models of anthropogenic carbon uptake. *Tellus Series B-Chemical and Physical Meteorology* 48: 397-417.
17. Joos, F. & Bruno, M. (1996) Pulse response functions are cost-efficient tools to model the link between carbon emissions, atmospheric CO<sub>2</sub> and global warming. *Phys. Chem. Earth* 21: 471-476.
18. Meyer, R., Joos, F., Esser, G., Heimann, M., Hooss, G., Kohlmaier, G., Sauf, W., Voss, R. & Wittenberg, U. (1999) The substitution of high-resolution terrestrial biosphere models and carbon sequestration in response to changing CO<sub>2</sub> and climate. *Global Biogeochemical Cycles* 13: 785-802.
19. Joos, F., Prentice, I. C., Sitch, S., Meyer, R., Hooss, G., Plattner, G. K., Gerber, S. & Hasselmann, K. (2001) Global warming feedbacks on terrestrial carbon uptake under the Intergovernmental Panel on Climate Change (IPCC) emission scenarios. *Global Biogeochemical Cycles* 15: 891-908.
20. Myhre, G., Highwood, E. J., Shine, K. P. & Stordal, F. (1998) New estimates of radiative forcing due to well mixed greenhouse gases. *Geophysical Research Letters* 25: 2715-2718.

21. Hooss, G., Voss, R., Hasselmann, K., Maier-Reimer, E. & Joos, F. (2001) A nonlinear impulse response model of the coupled carbon cycle-climate system (NICCS). *Climate Dynamics* 18: 189-202.
22. Prather, M., Ehhalt, D., Dentener, F., Derwent, R., Dlugokencky, E., Holland, E., Isaksen, I., Katima, J., Kirchhoff, V., Matson, P., Midgley, P. & Wang, M. (2001) in *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, eds. Houghton, J. T., Ding, Y., Griggs, D. J., Noguer, M., van der Linden, P. J., Dai, X., Maskell, K. & Johnson, C. A. (Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA), pp. 239-287.
23. Forster, P., Ramaswamy, V., Artaxo, P., Bernsten, T., Betts, R., Fahey, D. W., Haywood, J., Lean, J., Lowe, D. C., Myhre, G., Nganga, J., Prinn, R., Raga, G., Schulz, M. & Van Doorland, R. (2007) in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, eds. Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M. & Miller, H. L. (Cambridge United Kingdom and New York, NY, USA, New York, NY, USA), pp. 129-234.
24. Nordhaus, W. D. (1991) To slow or not to slow: The economics of the greenhouse effect. *The Economic Journal* 101: 920-937.
25. Manne, A. S., Mendelsohn, R. & Richels, R. (1995) A model for evaluating regional and global effects of GHG reduction policies. *Energy Policy* 23: 17-34.
26. Yohe, G. W. & Schlesinger, M. E. (1998) Sea-level change: The expected economic costs of protection or abandonment in the United States. *Climatic Change* 337-472.
27. Oeschger, H., Siegenthaler, U., Schotterer, U. & A., G. (1975) A box diffusion model to study the carbon dioxide exchange in nature. *Tellus* 27: 170-192.
28. Schultz, P. A. & Kasting, J. F. (1997) Optimal reductions in CO<sub>2</sub> emissions. *Energy Policy* 25: 491-500.
29. Joos, F., Müller-Fürstenberger, G. & Stephan, G. (1999) Correcting the carbon cycle representation: How important is it for the economics of climate change. *Environmental Modeling and Assessment* 4: 133-140.
30. Nordhaus, W. D. (2007) *A Question of Balance: Economic Modeling of Global Warming* (Yale University Press).
31. Cao, L., Eby, M., Ridgwell, A., Caldeira, K., Archer, D., Ishida, A., Joos, F., Matsumoto, K., Mikolajewicz, U., Mouchet, A., Orr, J. C., Plattner, G. K., Schlitzer, R., Tokos, K., Totterdell, I., Tschumi, T., Yamanaka, Y. & Yool, A. (2009) The role of ocean transport in the uptake of anthropogenic CO<sub>2</sub>. *Biogeosciences* 6: 375-390.

## Figure Captions

Figure 1: Perturbation in atmospheric CO<sub>2</sub> and global mean surface temperature after a pulse-like emission of 5000 GtC (red), corresponding to estimated conventional fossil resources, and of 1000 GtC (blue), corresponding roughly to the cumulative emissions reached under ambitious emission mitigation. The solid line are results from the Bern Carbon Cycle-Climate Model, while the dashed line are results from the DICE model of Nordhaus (30) as available in summer 2009 on the internet. The latter model does misrepresent the redistribution of carbon in the climate system compared to current state-of-the-art climate models (31)

Figure 2: A sketch of the coupled physical climate-carbon cycle–socio-economic system. Blue arrows represent carbon fluxes, red arrow linkages.

Figure 3: Normalized atmospheric impulse response function for an input of 1000 GtC and 5000 GtC into the atmosphere at time 0.

Figure 4: A box-type substitute model of the carbon cycle-climate system based on impulse response functions. Perturbations in the carbon fluxes in and out of the atmosphere are indicated by thick black arrows. Carbon and heat fluxes are allocated to a series of boxes with characteristic time scales for surface-to-deep ocean transport of carbon and heat ( $\tau_i$ ) and of terrestrial carbon overturning ( $\tau_{ib}$ ). The total perturbations in land and surface ocean carbon inventory and in surface temperature are the sums over the corresponding individual perturbations in each box,  $\delta x_i$ .

## Reviewer suggestions

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