

## Global warming feedbacks on terrestrial carbon uptake under the Intergovernmental Panel on Climate Change (IPCC) emission scenarios

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**Abstract.** A coupled physical-biogeochemical climate model that includes a dynamic global vegetation model and a representation of a coupled atmosphere-ocean general circulation model is driven by the nonintervention emission scenarios recently developed by the Intergovernmental Panel on Climate Change (IPCC). Atmospheric CO<sub>2</sub>, carbon sinks, radiative forcing by greenhouse gases (GHGs) and aerosols, changes in the fields of surface-air temperature, precipitation, cloud cover, ocean thermal expansion, and vegetation structure are projected. Up to 2100, atmospheric CO<sub>2</sub> increases to 540 ppm for the lowest and to 960 ppm for the highest emission scenario analyzed. Sensitivity analyses suggest an uncertainty in these projections of –10 to +30% for a given emission scenario. Radiative forcing is estimated to increase between 3 and 8 W m<sup>–2</sup> between now and 2100. Simulated warmer conditions in North America and Eurasia affect ecosystem structure: boreal trees expand poleward in high latitudes and are partly replaced by temperate trees and grasses at lower latitudes. The consequences for terrestrial carbon storage depend on the assumed sensitivity of climate to radiative forcing, the sensitivity of soil respiration to temperature, and the rate of increase in radiative forcing by both CO<sub>2</sub> and other GHGs. In the most extreme cases, the terrestrial biosphere becomes a source of carbon during the second half of the century. High GHG emissions and high contributions of non-CO<sub>2</sub> agents to radiative forcing favor a transient terrestrial carbon source by enhancing warming and the associated release of soil carbon.

### 1. Introduction

The urgency of climate policy actions depends on how future concentrations of greenhouse gases (GHGs) and global climate change are likely to evolve in the absence of GHG emission control and how global climate change may affect the world's ecosystems and the services they provide. Carbon dioxide (CO<sub>2</sub>) is the most important anthropogenic greenhouse gas (GHG). Currently, only about half of the anthropogenic CO<sub>2</sub> emission stays airborne. The rest is taken up by the terrestrial biosphere and the ocean. Global climate change due to increased GHG concentrations has the potential to reduce these natural CO<sub>2</sub> sinks [e.g., *Cao and Woodward*, 1998; *Sarmiento and Le Quéré*, 1996; *Joos et al.*, 1999b; *Cramer et al.*, 2000] and to affect ecosystem structure [*Smith and Shugart*, 1993] in ways that could enhance or reduce carbon uptake [*Cramer et al.*, 2001; *Smith and Shugart*, 1993; *Cox et al.*, 2000]. However, projections of future atmospheric CO<sub>2</sub> and climate are rendered uncertain by our understanding of how the mechanisms driving oceanic and terrestrial carbon sequestration are influenced by a changing environment.

We investigated possible feedbacks between global climate change and the terrestrial system and their impacts on projections

of future changes in climate and ecosystem structure for the new emission scenarios developed by the writing team of the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emission Scenarios (SRES) [*Nakićenović et al.*, 2000]. The feedbacks involve atmospheric CO<sub>2</sub>, other radiative forcing agents, changes in surface temperature, the hydrological cycle, and the carbon cycle. Rising atmospheric concentrations of CO<sub>2</sub> and other GHGs lead to increased radiative forcing, higher surface-air temperatures, and changes in the hydrological cycle [*Houghton et al.*, 1996]. Such changes may cause increased respiration of the carbon stored in soil and litter owing to higher bacterial activities at higher temperatures [*Lloyd and Taylor*, 1994; *Rustad*, 2000; *Cox et al.*, 2000], reduced net primary production because of excessively high temperatures and/or reduced water availability [*Cramer et al.*, 2001], and dieback of extant forests in response to heat or drought stress [*Cramer et al.*, 2001; *Smith and Shugart*, 1993; *Cox et al.*, 2000], thereby offsetting the carbon uptake stimulated by the increase in atmospheric CO<sub>2</sub> [*Farquhar et al.*, 1980] and, in some regions, by climate change. Although there is still considerable uncertainty about the relative magnitudes of these processes, paleodata clearly indicate that terrestrial carbon storage has varied under different climatic regimes [e.g., *Crowley*, 1995] and that the terrestrial system can show a substantial response to climatic shifts within a few decades [e.g., *MacDonald et al.*, 1993; *Mayle and Cwynier*, 1995; *Birks and Ammann*, 2000].

Most modeling studies to date have only partly addressed the feedback loops described above by prescribing atmospheric CO<sub>2</sub> and climate in dynamic global vegetation models (DGVMs) [*Cramer et al.*, 2001] or in simpler terrestrial carbon models that prescribe the ecosystem distribution as constant in time [*Cao and Woodward*, 1998; *Meyer et al.*, 1999]. These studies show that the

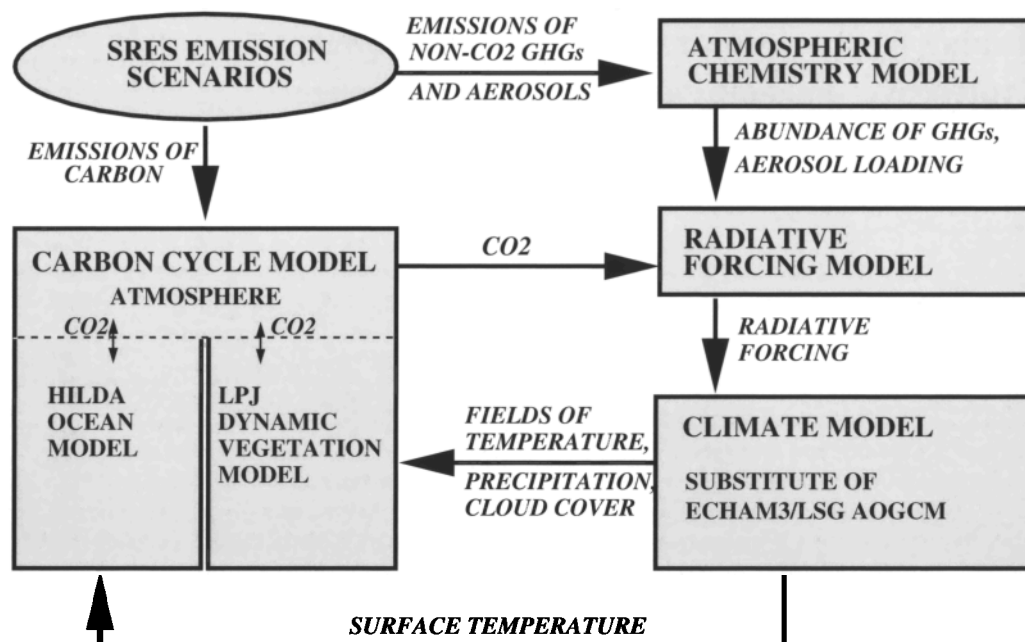
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## BERN CC MODEL



**Figure 1.** Simplified scheme of the Bern CC model. For scenario calculations, future abundances of greenhouse gases (GHGs) are projected from the emissions of CO<sub>2</sub> and non-CO<sub>2</sub> GHGs. Radiative forcing is calculated from the abundance of CO<sub>2</sub> and non-CO<sub>2</sub> GHGs and emissions of aerosol precursors. The IRF-Eof substitute of the ECHAM3/LSG AOGCM is driven by radiative forcing, and modeled surface fields of temperature, precipitation, and cloud cover are passed to the carbon cycle model. The carbon cycle model includes a well-mixed atmosphere, the HILDA ocean model, and the Lund-Potsdam-Jena (LPJ) Dynamic Global Vegetation Model. Atmospheric CO<sub>2</sub> is projected from carbon emissions by fossil fuel burning and land use changes and the carbon uptake (release) by the ocean and the land biosphere.

terrestrial sink is reduced in simulations considering climate change as compared to simulations without future climatic changes. Recently, an application of a terrestrial model coupled to an atmosphere-ocean general circulation model (AOGCM) has produced the same qualitative result [Cox *et al.*, 2000].

Here a complementary strategy is applied to investigate the climate-land biosphere feedbacks described above under the IPCC SRES nonintervention scenarios. Computationally efficient substitutes of the coupled European Center/Hamburg Model 3 and Large Scale Geostrophic (ECHAM3/LSG) AOGCM [Voss *et al.*, 1998; Meyer *et al.*, 1999; Hooss *et al.*, 2001; Voss and Mikolajewicz, 2001] and of the oceanic carbon cycle [Siegenthaler and Joos, 1992; Joos *et al.*, 1996] are coupled to the Lund-Potsdam-Jena Dynamic Global Vegetation Model (LPJ-DGVM) that simulates vegetation structure, carbon storage, and the water balance here at a resolution of  $3.75^\circ \times 2.5^\circ$  [Cramer *et al.*, 2001; Sitch, 2000]. Adequately designed substitute models consist of a few equations only and require only a very modest amount of CPU time while yielding identical results for selected variables as the parent model [Joos *et al.*, 1996; Huntingford and Cox, 2000]. The use of substitute models combined with the computational efficiency of the LPJ-DGVM allows us to analyze a range of GHG emission scenarios and GHG concentration pathways, to run the model under alternative climate sensitivities, and to allow for alternative hypotheses, e.g., for the relationships between soil respiration and warming, or primary production and atmospheric CO<sub>2</sub>. We are thus able to estimate the uncertainty in future terrestrial carbon storage and the associated uncertainty in projected CO<sub>2</sub>, radiative forcing, and climate change

This paper is organized as follows. In section 2 the coupled model and the SRES scenarios are described. In section 3, projected changes in radiative forcing and climate are presented for six illustrative SRES scenarios. Related changes in carbon storage and ecosystem structure are described. Next, we investigate carbon storage in response to changes in climate or changes in atmospheric CO<sub>2</sub> only and under different growth rates of radiative forcing and atmospheric CO<sub>2</sub>. Before concluding, uncertainties in projected CO<sub>2</sub> and carbon storage and their impact on projected temperatures are estimated. Details of the model can be found in Appendix A.

## 2. Description of Model and Scenarios

### 2.1. Bern Carbon Cycle–Climate Model

The Bern carbon cycle-climate (Bern CC) model consists of a chemistry, radiative forcing, climate, and carbon cycle module (Figure 1). The model components are described in Appendix A and in the literature [Prather *et al.*, 2001; Ramaswamy *et al.*, 2001; Hooss *et al.*, 1999; Voss and Mikolajewicz, 1999; Cubasch *et al.*, 1997; Voss *et al.*, 1998; Joos *et al.*, 1996; Siegenthaler and Joos, 1992; Cramer *et al.*, 2001; Sitch, 2000; Prentice *et al.*, 2000; McGuire *et al.*, 2001]. The main features of the model are summarized below.

For scenario calculations, abundances of non-CO<sub>2</sub> GHGs are prescribed according to observations until 2000 and then calculated from SRES emissions. Calculating the abundance of chemically reactive gases from emissions requires a model that can predict how the lifetimes of these gases are changed by an evolving

atmospheric chemistry. Here we rely on results of a modeling workshop called OxComp where 14 state-of-the-art chemistry transport models were run under a set of emission scenarios. From the results of these simulations, simplified expressions were developed [Prather et al., 2001; M. Prather, personal communication, September 2000] to estimate the evolution of OH and O<sub>3</sub> as a function of pollutant emissions and the impact of changing N<sub>2</sub>O emissions on the lifetime of N<sub>2</sub>O. Simplified expressions, lifetimes, and concentrations at year 2000 and at preindustrial time are given in Appendix A.

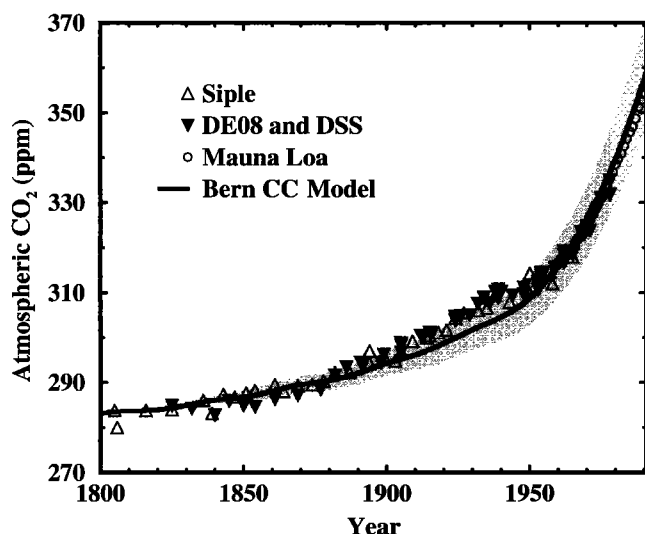
The radiative forcing [Ramaswamy et al., 2001; Shine et al., 1990; Shine et al., 1994; Shine and Forster, 1999] from changes in the abundances of GHGs (CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O, stratospheric and tropospheric O<sub>3</sub>, stratospheric H<sub>2</sub>O due to CH<sub>4</sub> changes, SF<sub>6</sub>, and 28 halocarbons including those controlled by the Montreal Protocol), from direct and indirect effects of sulfate aerosols, and from direct forcing of black and organic carbon is calculated on the basis of simplified expressions [Ramaswamy et al., 2001; Harvey et al., 1996; Myhre et al., 1998].

The model's climate component is an impulse response–empirical orthogonal function (IRF-EOF) substitute [Hooss et al., 1999; Hooss et al., 2001; Voss and Mikolajewicz, 1999; Meyer et al., 1999] of the ECHAM3/LSG AOGCM [Cubasch et al., 1997; Voss et al., 1998], driven by radiative forcing. IRFs for perturbations in surface-air temperature, precipitation, cloud cover, and sea level rise characterize the adjustment time of the climate system to changes in radiative forcing, whereas EOFs describe the spatial patterns of the annual mean perturbations. The substitute was derived from a 850 year simulation with the ECHAM3/LSG model, wherein atmospheric CO<sub>2</sub> was quadrupled in the first 120 years and held constant thereafter [Voss and Mikolajewicz, 1999]. The climate sensitivities of the substitute for a doubling of atmospheric CO<sub>2</sub> corresponding to a change in radiative forcing of 3.7 W m<sup>-2</sup> are 2.5°C (global mean surface-air temperature  $\Delta T_{2x}$ ), 64 mm yr<sup>-1</sup> (global mean precipitation), -0.9% (cloud cover), and 128 cm (sea level) in the standard case. In sensitivity experiments, temperature sensitivities of 0°C (T0; constant climate) and 4.5°C (T45) [Houghton et al., 1996] have been used; the sensitivities of other climate variables were scaled accordingly. This simple scaling tends to overestimate (underestimate) the transient temperature response for climate sensitivities higher (lower) than the ECHAM/LSG sensitivity as uptake of heat by the ocean is not explicitly simulated [Hansen et al., 1984].

The carbon cycle component consists of a well-mixed atmosphere, the High-Latitude Exchange/Interior Diffusion-Advection (HILDA) ocean model [Joos et al., 1996; Siegenthaler and Joos, 1992], and the LPJ-DGVM [Cramer et al., 2001; Stith, 2000; Prentice et al., 2000; McGuire et al., 2001]. The HILDA model and its substitute are used interchangeable and yield identical results. Surface to deep tracer transport in the ocean substitute is described by an IRF. The nonlinearities in air-sea gas exchange and carbon chemistry are captured by separate equations. The effect of sea surface warming on carbonate chemistry is included [Joos et al., 1999b]. The LPJ-DGVM simulates the distribution of nine plant functional types (PFTs) based on bioclimatic limits for plant growth and regeneration and plant-specific parameters that govern plant competition for light and water. The PFTs considered are tropical broad-leaved evergreen trees, tropical broad-leaved rain-green trees, temperate needle-leaved evergreen trees, temperate broad-leaved evergreen trees, temperate broad-leaved summer-green trees, boreal needle-leaved evergreen trees, boreal summer-green trees, C<sub>3</sub> grasses/forbs, and C<sub>4</sub> grasses. Dispersal processes are not explicitly modeled, and an individual PFT can invade new regions if its bioclimatic limits and competition with other PFTs allow establishment. There are six carbon pools per PFT, repre-

senting leaves, sapwood, heartwood, fine roots, aboveground and belowground litter, and two soil carbon pools, which receive input from litter of all PFTs. Photosynthesis is modeled using a form of the Farquhar scheme [Farquhar et al., 1980] with leaf-level optimized nitrogen allocation [Haxeltine and Prentice, 1996] and an empirical convective boundary layer parameterization [Monteith, 1995] to couple the carbon and water cycles. Decomposition rates of soil and litter organic carbon in the standard case depend on soil temperature [Lloyd and Taylor, 1994] and moisture [Foley, 1995]. Fire fluxes are calculated on the basis of litter moisture content, a fuel load threshold, and PFT specific fire resistances. The spatial resolution of the LPJ-DGVM is set to 3.75° × 2.5°. The LPJ-DGVM is spun up for 1000 years under preindustrial CO<sub>2</sub> and a baseline climate that includes interannual variability [Cramer et al., 2001; Leemans and Cramer, 1991]. At year 400, soil carbon pool sizes are calculated analytically from annual litter inputs and annual mean decomposition rates. Spin-up is continued for another 600 years to reach equilibrium. For transient simulations the spatial fields in annual mean perturbations of temperature, cloud cover, and precipitation simulated by the ECHAM3/LSG substitute are added to the baseline climate. Test simulations suggest that the error in global carbon uptake arising from the IRF-EOF approach is 10%. LPJ was forced with the monthly fields of surface temperature, and precipitation was obtained with the ECHAM3/LSG for the "IPCC 1990 Business as Usual" scenario. Alternatively, the climate change pattern was represented by the first EOF of surface temperature and precipitation. Carbon uptake by 2084 is 543 and 607 Gt C for these two simulations. Deviations in local carbon uptake are generally <2 kg m<sup>-2</sup>, and changes in vegetation structure are very similar in both simulations. Atmospheric CO<sub>2</sub> is prescribed until 1999 [Etheridge et al., 1996; Keeling and Whorf, 1994], and the SRES carbon emissions are prescribed afterward. Land use changes are not explicitly considered in the present simulations because land use patterns for the SRES scenarios are not available. Instead, carbon fluxes from land use changes are prescribed externally according to the SRES scenarios. This approach probably tends to overestimate global terrestrial carbon storage as no attempt is made here to correct for the area that is not available for additional carbon storage due to land use.

The performance of the ECHAM3/LSG Model [Cubasch et al., 1997; Voss et al., 1998] has been evaluated in the framework of the Coupled Model Intercomparison Project (CMIP) [Meehl et al., 2000; Covey et al., 2000]. The LPJ-DGVM has been evaluated in combination with atmospheric transport models by comparing model results with the observed seasonal cycle in atmospheric CO<sub>2</sub> [Stith, 2000] and its temporal trend [McGuire et al., 2001] and by comparing the simulated time evolution and interannual variability of terrestrial carbon uptake with reconstructions based on atmospheric, CO<sub>2</sub>,  $\delta^{13}\text{C}$  and O<sub>2</sub>:N<sub>2</sub> data [McGuire et al., 2001]. The parameters of the box diffusion type ocean model were determined such that the model reproduces the observed oceanic distribution of natural and bomb-produced radiocarbon [Siegenthaler and Joos, 1992]; the model was evaluated with oceanic observations of <sup>39</sup>Ar and CFCs, and its parameterization for surface-to-deep transport was tested against the Geophysical Fluid Dynamics Laboratory (GFDL) Princeton ocean general circulation model [Joos et al., 1997]. Model predictions for the decrease in atmospheric  $\delta^{13}\text{C}$  and  $\Delta^{14}\text{C}$  (Suess Effect) as well as the global budget of bomb-produced radiocarbon agree with observations within their error limits when the ocean is coupled to a box-type biosphere [Joos and Bruno, 1998]. Bern CC simulations of CO<sub>2</sub> uptake by ocean and land are within data-based reconstructions [Battle et al., 2000; Keeling et al., 1996; Joos et al., 1999a]. Ocean uptake is 2.1 Gt C yr<sup>-1</sup>, and biospheric uptake is 1.7 Gt C yr<sup>-1</sup> for the period from 1980 to 1999. The combined model in its standard case simulates reasonably well the evolution of atmospheric CO<sub>2</sub>



**Figure 2.** Observed versus simulated increase in atmospheric  $\text{CO}_2$  during the industrial period. Data are from Antarctic ice cores drilled at Siple [Neftel et al., 1985] and Law Dome (cores DE08 and DSS) [Etheridge et al., 1996] and direct atmospheric observations at Mauna Loa, Hawaii [Keeling and Whorf, 1994]. The Bern CC model is forced by fossil [Marland et al., 1995] and land use carbon emissions [Houghton, 1999] that are available until 1990. The shaded band was obtained by varying the emission data by  $\pm 1$  standard deviation, taken to be 5% for fossil emissions and 31% for land use emissions.

during the industrial period (Figure 2) when driven by historical carbon emissions from fossil fuel burning, cement production, and land use [Marland et al., 1995; Houghton, 1999]. The simulated increase in global average surface temperature since 1900 is  $0.37^\circ\text{C}$ , which is slightly lower than observed ( $0.6 \pm 0.2^\circ\text{C}$ ) [Folland et al., 2001].

## 2.2. SRES Emission Scenarios

The writing team of the Special Report on Emission Scenarios (SRES) of IPCC [Nakićenović et al., 2000] has developed a set of emission scenarios for radiatively important gases up to 2100. There are four scenario “families” comprising a total of 40 scenarios. Each family includes several quantifications of a common narrative. None of the scenarios include any future policies that explicitly address climate change.

Here we analyze the six “illustrative marker scenarios” A1B, A1FI, A1T, A2, B1, and B2 taken to represent each family [Nakićenović et al., 2000]. The A1 family describes a future with low population growth but rapid economic growth and high energy and material demands moderated by rapid technological change. The A1 scenario family develops into three groups that describe alternative directions of technology change in the energy system. The A1FI scenario is representative for a fossil-intensive energy sector; nonfossil energy sources are emphasized in the A1T scenario, whereas nonfossil energy sources and fossil sources are “balanced” in the A1B scenario. The A2 family describes a heterogeneous world with economic development regionally oriented, slow economic growth, and fast population growth. The B1 family describes a convergent world with low population growth as in A1 but with rapid changes in economic structure toward a service and information economy and the introduction of clean technologies. The B2 family describes a world in which the emphasis is on local solutions, with moderate population growth,

intermediate levels of economic development, and less rapid technological change than in A1 or B1.

## 3. Results

### 3.1. Changes in Radiative Forcing and Climate

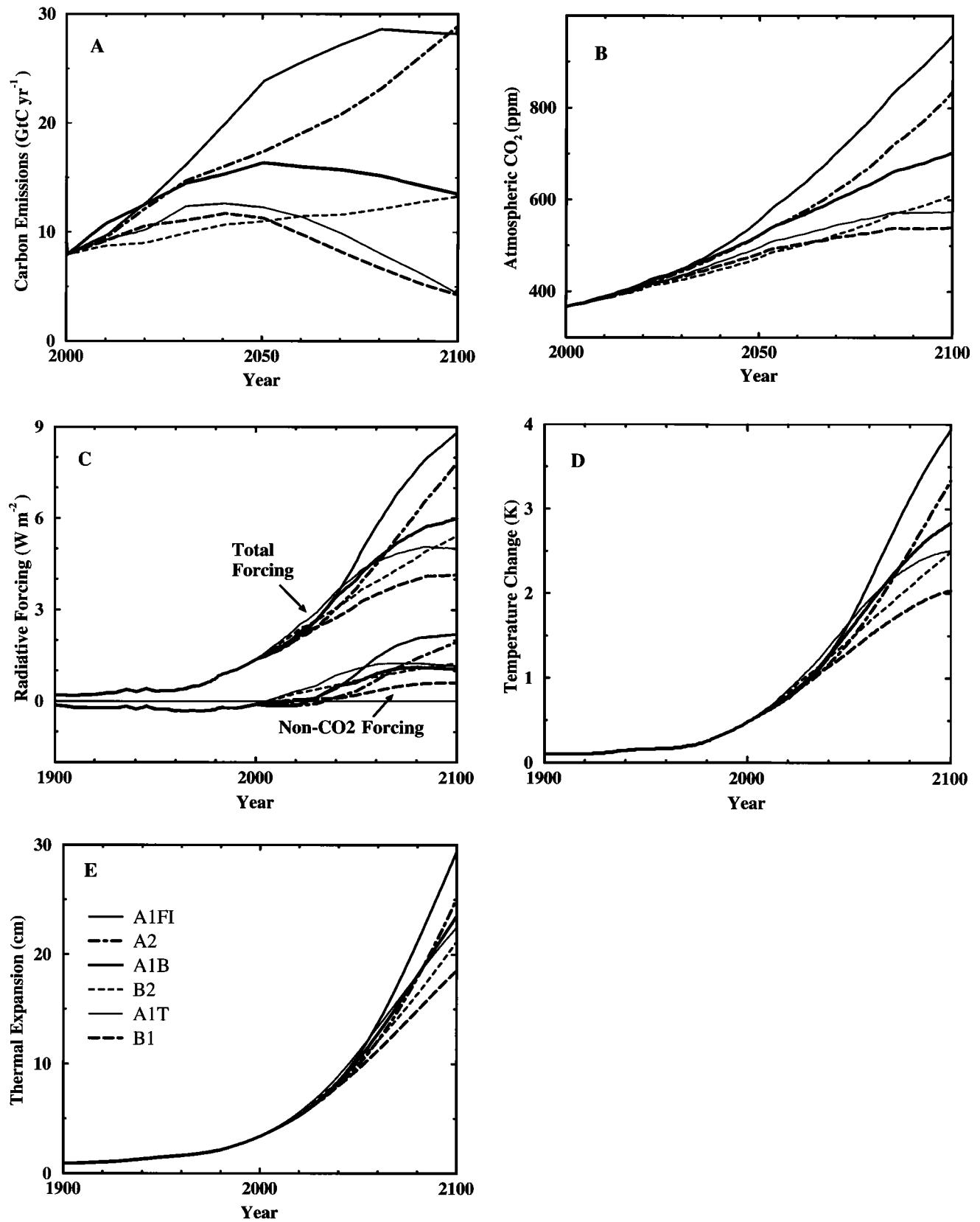
First, we evaluate climatic consequences of the six marker scenarios for the standard model setup. Cumulative carbon emissions between 2000 and 2100 range between 690 and 2455 Gt C for all 40 SRES scenarios. The range of the six analyzed marker scenarios is somewhat narrower: 981 Gt C (B1) to 2107 GtC (A1FI) (Figure 3a). Atmospheric  $\text{CO}_2$  is projected to increase from 367 ppm at the beginning of year 2000 to 540 ppm at year 2100 for the lowest (B1) and to 958 ppm for the highest emission scenario (A1FI) (Figure 3b), corresponding to an increase in radiative forcing of 2.1 and  $5.1 \text{ W m}^{-2}$  respectively.

Radiative forcing by non- $\text{CO}_2$  GHGs is  $1.3 \text{ W m}^{-2}$  at 2000 and is projected to be between 1.2 (B1) and  $3.3 \text{ W m}^{-2}$  (A1FI and A2) at year 2100. However, radiative forcing by all non- $\text{CO}_2$  agents is estimated to be close to zero during the twentieth century because the negative forcing by aerosols approximately cancels the positive forcing by non- $\text{CO}_2$  GHGs (Figure 3c) in our model configuration. Projected radiative forcing by all agents is intermediate during the next three decades for the most carbon-intensive scenarios because the effect of high carbon emissions from coal burning is partly masked by concomitant high sulfur emissions (Figure 3c). We emphasize that the magnitude of the aerosol forcing is highly uncertain [Houghton et al., 1996; Shine and Forster, 1999]. However, the relative importance of aerosol-induced cooling decreases under the six marker scenarios because it is assumed that sulfur emission controls are strengthened. Radiative forcing by sulfate aerosols is estimated to be  $-1.2 \text{ W m}^{-2}$  at 2000 and between  $-1.1 \text{ W m}^{-2}$  (A2) and  $-0.4 \text{ W m}^{-2}$  (A1T) at 2100. Changes in radiative forcing by organic and black carbon aerosols are estimated to be small ( $-0.16$  to  $+0.05 \text{ W m}^{-2}$ ). Taken together, we estimate an increase in radiative forcing between 2000 and 2100 of  $0.0 \text{ W m}^{-2}$  (A2) to  $+0.7 \text{ W m}^{-2}$  (B1) from changes in aerosols. This is relatively small compared to the range for total radiative forcing of  $2.8 \text{ W m}^{-2}$  (B1) to  $7.5 \text{ W m}^{-2}$  (A1FI).

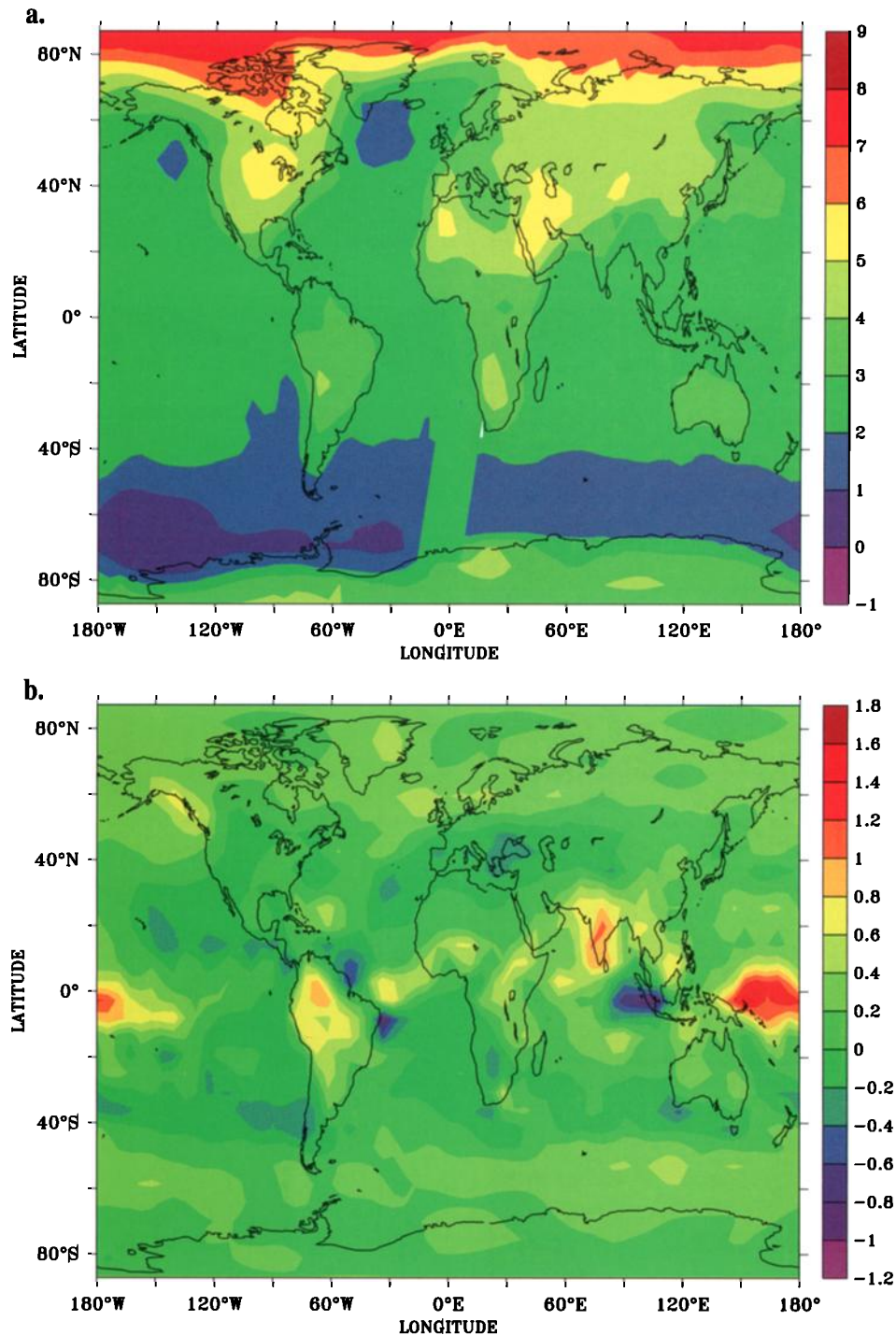
Global average surface-air temperature is modeled to increase by  $0.5^\circ\text{C}$  up to 2000 and a further  $1.6^\circ\text{C}$  (B1) to  $3.5^\circ\text{C}$  (A1FI) up to 2100 (Figure 3d). The simulated temperature increase is largest in high northern latitudes and in continental regions,  $4^\circ$ – $6^\circ\text{C}$  in North America and Eurasia for scenario A1B at year 2100 (Plate 1a). For scenario A1B and year 2100, precipitation changes are between  $-40$  and  $+60 \text{ cm yr}^{-1}$  (Plate 1b). Projected ocean thermal expansion is between 19 and 30 cm (Figure 3e). In conclusion, atmospheric  $\text{CO}_2$ , radiative forcing, temperature, and ocean thermal expansion continue to increase for all six scenarios.

### 3.2. Changes in Terrestrial Carbon Storage and Ecosystem Structure

For simplicity, we restrict discussion of the role of terrestrial ecosystem changes to scenario A1B and note that results are qualitatively similar for the other five SRES scenarios. Results of various sensitivity experiments are summarized in Table 1. In the standard experiment T25 ( $\Delta T_{2x} = 2.5^\circ\text{C}$ ), global terrestrial carbon sequestration for scenario A1B is between 2 and  $3 \text{ Gt C yr}^{-1}$  during the 21st century (Figure 4a). Soils and vegetation in tropical areas and high northern latitudes are carbon sinks, whereas soils and vegetation in northern midlatitude regions become carbon sources (Plate 2a). Simulated ecosystem changes are substantial (Plate 2b). Tree coverage by tropical raingreen and evergreen trees is enhanced in the tropics and subtropics, in contrast to Cox et al. [2000], as soil moisture is increased. Boreal trees are partly replaced by temperate deciduous trees or grasses due to warmer



**Figure 3.** (a) Prescribed anthropogenic carbon emissions and projected (b) atmospheric  $\text{CO}_2$ , (c) radiative forcing by all agents and all non- $\text{CO}_2$  agents, (d) change in global average surface temperature, and (e) ocean thermal expansion for the SRES illustrative marker scenarios A1B, A2, B1, B2, A1FI, and A1T. The model's climate sensitivity is set to  $2.5^\circ\text{C}$ .



**Plate 1.** Simulated distribution of (a) annual mean surface warming and (b) precipitation changes since preindustrial time at year 2100 for scenario A1B. Surface warming is in °C; precipitation changes are in mm day<sup>-1</sup>. The model's climate sensitivity is set to 2.5°C.

winters favoring temperate trees, while hot summers reduce the viability of boreal trees. These changes reduce the total forested area by  $1.5 \times 10^{12}$  m<sup>2</sup> between 40° and 60°N. The warming in high northern latitudes leads to a progressive poleward spread of boreal trees in North America and Siberia. Tree coverage north of 60°N is enhanced by  $2.7 \times 10^{12}$  m<sup>2</sup>. These simulated ecosystem changes are comparable with paleodata showing more extensive

distribution of temperate deciduous forests and grasslands in the continental interior, together with a northward shift of the boreal tree line in response to warming of comparable magnitude in northern middle to high latitudes during the mid-Holocene [COHMAP Members, 1988; Prentice *et al.*, 1991]. However, projected future changes cannot be directly compared with Holocene changes as climatic changes progressed much slower during the



**Table 1.** Model Results for Scenario A1B

Simulation	Description	CO <sub>2</sub> in 2100, ppm	Carbon Uptake 2000–2099, Gt C		Mean Surface Warming 1765–2100, °C
			Land	Ocean	
C-slow	CO <sub>2</sub> fertilization capped at year 2000; slow ocean	918	–147	390	3.5
C	CO <sub>2</sub> fertilization capped at year 2000	856	–130	505	3.3
T45	climate sensitivity increased ( $\Delta T_{2\times} = 4.5^\circ\text{C}$ )	805	52	432	5.6
T25	standard model setup ( $\Delta T_{2\times} = 2.5^\circ\text{C}$ )	703	282	418	2.8
T0	no climate change ( $\Delta T_{2\times} = 0.0^\circ\text{C}$ )	632	438	415	0.0
R	soil and litter respiration rates do not depend on global warming	636	470	374	2.6
R-fast	soil and litter respiration rates do not depend on global warming; fast Ocean	617	449	433	2.5

Holocene than the changes projected for the 21st century and because the nature of the forcing was different (i.e., orbital variations as opposed to greenhouse gas variations). Nevertheless, the paleodata indicate that the projected trends of vegetation response are consistent with regional changes in vegetation structure that have occurred in the geologically recent past.

The climate sensitivity of the ECHAM3/LSG model is at the lower end of the current set of coupled AOGCMs. In experiment T45 the climate sensitivity is increased to  $4.5^\circ\text{C}$ . The dieback of boreal trees is then greater, and soil and litter turnover rates are increased. The carbon loss in the boreal and temperate zone becomes larger than the carbon gain in the tropics and high northern latitudes, so that the terrestrial biosphere turns into a carbon source for all six SRES scenarios considered (Figure 4a).

### 3.3. Changes in Carbon Storage in Response to Changes in CO<sub>2</sub> or Climate Only

We present two sensitivity experiments, C0 and T0, to further analyze the pattern of terrestrial carbon sources and sinks. In T0 the model's climate sensitivity is set to zero. Then, CO<sub>2</sub> changes are the only driving force for terrestrial carbon storage. Terrestrial carbon uptake then increases to 4 Gt C yr<sup>–1</sup> in the second half of the century (Figure 4a). The modeled CO<sub>2</sub> fertilization mechanism is especially effective in tropical forests because of their high NPP, high photorespiration rates (which are reduced in high CO<sub>2</sub>), and long timescales for turnover of vegetation carbon (Plate 3a). Ecosystem structure changes are relatively small. Tropical evergreen tree cover is slightly increased, and trees invade arid regions, in part because of increased water use efficiency accompanied by an only modest increase in carbon release by fires and in part because of more effective competition of tree seedlings (with the C<sub>3</sub> photosynthetic pathway) against the dominant C<sub>4</sub> grasses in tropical savannas. The effects of phosphorus limitations and land use pressure which may be large in tropical regions are not considered in the model.

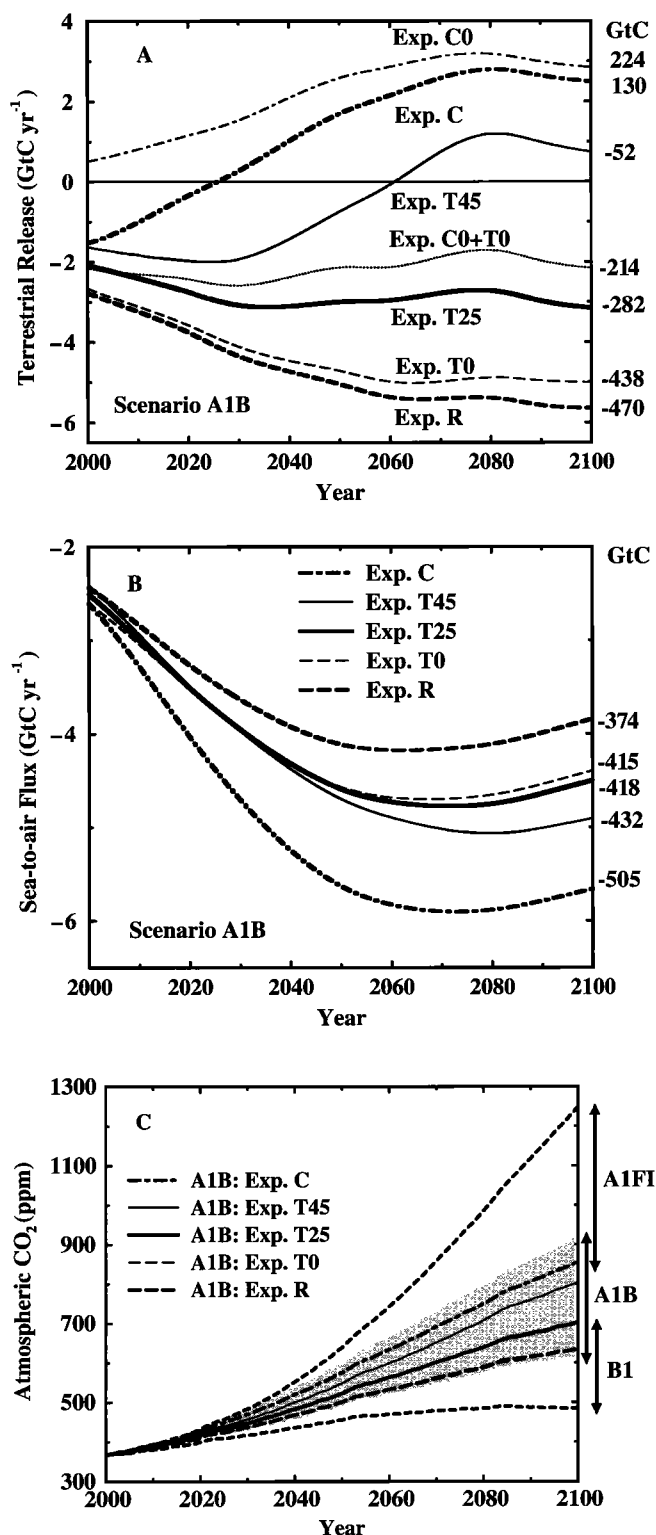
CO<sub>2</sub> fertilization is shut off in experiment C0 by keeping CO<sub>2</sub> in the model's photosynthesis module at preindustrial level, so that only climate change affects terrestrial carbon storage. The terrestrial biosphere then releases carbon, as a consequence of increased soil and litter respiration rates under the simulated warmer and slightly more humid conditions, except in cold climates where growing season length strongly increases (Plate 3b). Forest dieback in limited regions of the northern continental interior leads to a loss of  $>12 \text{ kg C m}^{-2}$  from the boreal zone by 2100. The climate and CO<sub>2</sub> fertilization mechanisms are not additive; their synergistic effects lead to higher carbon storage in T25 than for the sum of C0

and T0. Similarly, the reduction in tree cover in midlatitude regions is smaller, and the stimulation of tree growth in the high north is larger in T25. Experiments C0 and T0 illustrate that global terrestrial carbon uptake or release represents a balance between carbon gain by CO<sub>2</sub> fertilization and vegetation increase mainly in high-latitude and high-altitude regions and carbon loss due to generally increased soil respiration rates, compounded by forest dieback in some boreal regions. This balance may change over time, as illustrated in section 3.4.

### 3.4. Carbon Storage and Growth Rates in Radiative Forcing and CO<sub>2</sub>

We investigated how the magnitude of the terrestrial sink/source may depend on the rate of increase of atmospheric CO<sub>2</sub> and radiative forcing. Atmospheric CO<sub>2</sub> was prescribed to increase exponentially to 1000 ppm within 70 or 210 years and stabilized thereafter (Figure 5a). Non-CO<sub>2</sub> radiative forcing was set to zero or to 25% of the radiative forcing by CO<sub>2</sub> and  $\Delta T_{2\times}$  was set to  $4.5^\circ\text{C}$ . Simulations were continued from the end of the spin-up at model year 1000 until model year 2000 when a new equilibrium is approached. The terrestrial biosphere becomes a CO<sub>2</sub> source for several decades either when atmospheric CO<sub>2</sub> increases rapidly or when non-CO<sub>2</sub> radiative forcing is large (Figure 5b). In this case, the carbon loss due to climate change dominates for a few decades over the carbon gain by CO<sub>2</sub> fertilization. At the new equilibrium, carbon storage has increased by 400 Gt C when non-CO<sub>2</sub> radiative forcing is set to zero but only by 110 Gt C when non-CO<sub>2</sub> radiative forcing is 25% of the radiative forcing by CO<sub>2</sub>. Cumulative ocean uptake until model year 2000 is 2100 and 1990 Gt C for the two cases and the fast atmospheric CO<sub>2</sub> pathway. These findings imply that the emission rate of non-CO<sub>2</sub> GHGs affects the growth rate of atmospheric CO<sub>2</sub>. High rates of GHG emissions and high contributions of non-CO<sub>2</sub> gases to the total radiative forcing could lead to reduced terrestrial carbon storage (or even to a terrestrial carbon source) during this century.

The evolution in simulated carbon storage reflects the different timescales associated with the governing processes. The relevant processes are enhanced productivity and water use efficiency with increasing CO<sub>2</sub> (CO<sub>2</sub> fertilization), loss of soil and litter carbon due to increasing respiration rates, and local dieback of boreal trees and their replacement by temperate trees or grasses under rising temperatures. Figure 5c shows the evolution of primary productivity and the carbon stored globally in vegetation and in soils (including litter) and of global average surface temperature for the experiment where atmospheric CO<sub>2</sub> reaches 1000 ppm within 70 years and non-CO<sub>2</sub> radiative forcing is 25% of the radiative forcing



by CO<sub>2</sub>. The increase in surface temperature lags the increase in atmospheric CO<sub>2</sub> (Figure 5) due to the thermal inertia of the ocean. Primary productivity increases about linearly in the first 80 years of the simulation and stabilizes soon after atmospheric CO<sub>2</sub> has been stabilized. Carbon storage in vegetation increases because of rising productivity during the first 50 years; then, uptake is stalled during the next few decades as the carbon gain by enhanced productivity

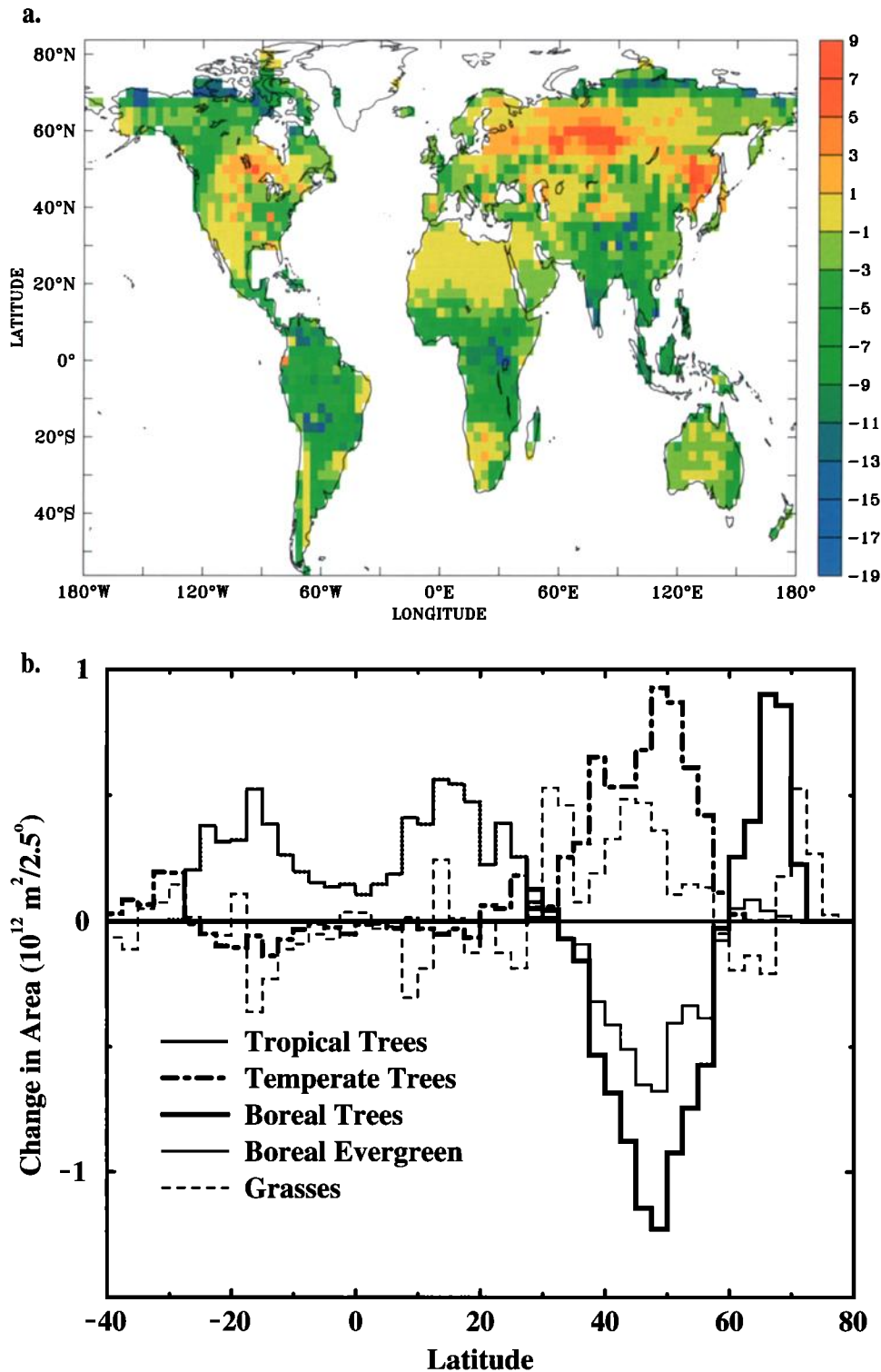
is compensated by carbon loss due to dieback of boreal trees. Temperate trees invade the boreal zone within the following decades, and carbon storage in vegetation increases toward a new equilibrium. Carbon storage in soil increases during the first few decades as input from vegetation outweighs loss by increasing soil respiration rates. At higher temperatures, soil carbon is lost to the atmosphere. The whole terrestrial biosphere acts as a carbon source around the time of CO<sub>2</sub> stabilization (Figure 5), when temperatures and soil respiration reach high levels while soil and litter pools are still large. The biosphere is also a source of carbon during the centuries after model year 1400, as carbon storage in vegetation approaches equilibrium while soil carbon continues to be lost to the atmosphere due to the long adjustment time of the soil reservoir.

### 3.5. Uncertainties in Projected CO<sub>2</sub> and Carbon Storage

We also investigated the consequences of alternative (extreme) model assumptions about terrestrial ecosystem processes. First, the extent of stimulation of carbon storage in natural ecosystems by CO<sub>2</sub> has been a matter of controversy [Häterschwiler *et al.*, 1997; Luo *et al.*, 1999]. There is no proof that the biospheric sink on the global scale is indeed primarily due to CO<sub>2</sub> fertilization, as would be implied by the LPJ-DGVM results. Other process, such as nitrogen fertilization [Schindler and Bayley, 1993; Townsend *et al.*, 1996], climate variations [McGuire *et al.*, 2001; Dai and Fung, 1993], and forest regrowth [Dixon *et al.*, 1994; Caspersen *et al.*, 2000], might in principle be responsible for part or most of the terrestrial sink. If this were the case, then it would be unlikely that primary productivity would increase as a function of future CO<sub>2</sub> concentrations. A simple alternative hypothesis is that primary productivity remains close to its present level. In experiment C, CO<sub>2</sub> fertilization is capped after year 2000; this is achieved by keeping CO<sub>2</sub> constant at the concentration of year 2000 in the photosynthesis routine. Modeled carbon uptake by CO<sub>2</sub> fertilization until year 2000 is thus taken as a surrogate for an unknown process (in order to balance the present global budget), and it is assumed that this unknown sink process has just reached saturation. Then, the terrestrial biosphere turns into a source within the next two decades and around 130 Gt C are released by the end of the century (Figure 4a). Second, there is conflicting evidence on the temperature dependence of soil and litter respiration [Trumbore *et al.*, 1996; Giardina and Ryan, 2000; Jarvis and Linder, 2000; Rustad, 2000] over multiannual timescales. A recent synthesis [Rustad, 2000] of results of 32

**Figure 4.** Sensitivity tests on (a) projected terrestrial carbon release, (b) sea-to-air flux, and (c) atmospheric CO<sub>2</sub> under scenario A1B. Cumulative carbon release for 2000–2100 is given in Gt C at the right; negative numbers denote terrestrial uptake in Figure 4a. In T0, T25, and T45 the climate sensitivity  $\Delta T_{2\times}$  of the ECHAM3/LSG substitute is varied between 0° (constant climate), 2.5° (standard case), and 4.5°C. In R, soil respiration does not depend on global warming. In C0, CO<sub>2</sub> fertilization is shut off, and in C, CO<sub>2</sub> fertilization is capped after year 2000.  $\Delta T_{2\times}$  is 2.5°C in experiments C0, C, and R. The terrestrial model setup of experiment C is combined with a slow overturning version of the ocean model, and the terrestrial setup of experiment R is combined with a fast overturning ocean to yield upper and lower estimates in projected CO<sub>2</sub> (shaded area in Figure 4c). The upper limit (R-fast) for scenario A1FI and the lower limit (R-slow) for scenario B1 is shown by thick dashed lines. The arrows show the ranges (R-fast to C-slow) in projected CO<sub>2</sub> for scenarios A1B, B1, and A1FI at year 2100. Interannual variability in terrestrial release and in sea-to-air fluxes has been removed by a smoothing spline.



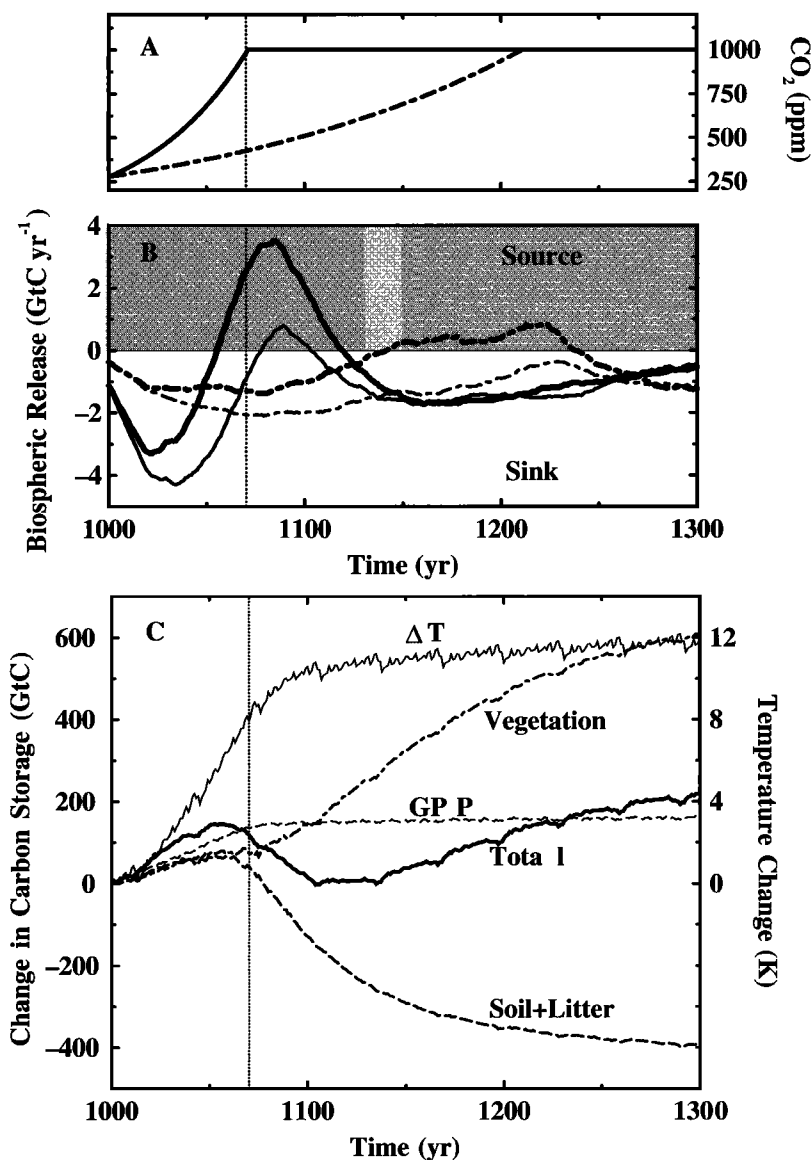


**Plate 2.** Simulated changes up to 2100 in (a) terrestrial carbon storage in  $\text{kg C m}^{-2}$  (negative numbers denote uptake) and (b) the area covered by grass and tropical, temperate, and boreal trees per  $2.5^\circ$  latitudinal zone for scenario A1B, standard experiment (T25).

ecosystem warming experiments with duration between 2 and 9 years in different biomes shows that soil respiration rates increase in general with warming, but the change in respiration rate is found to be small (or even negative) at some sites. In experiment R, soil and litter respiration rates are assumed to be independent

of global warming [Giardina and Ryan, 2000]. Global carbon storage is then increased by 188 Gt C, relative to experiment T25, by 2100 (Figure 4a).

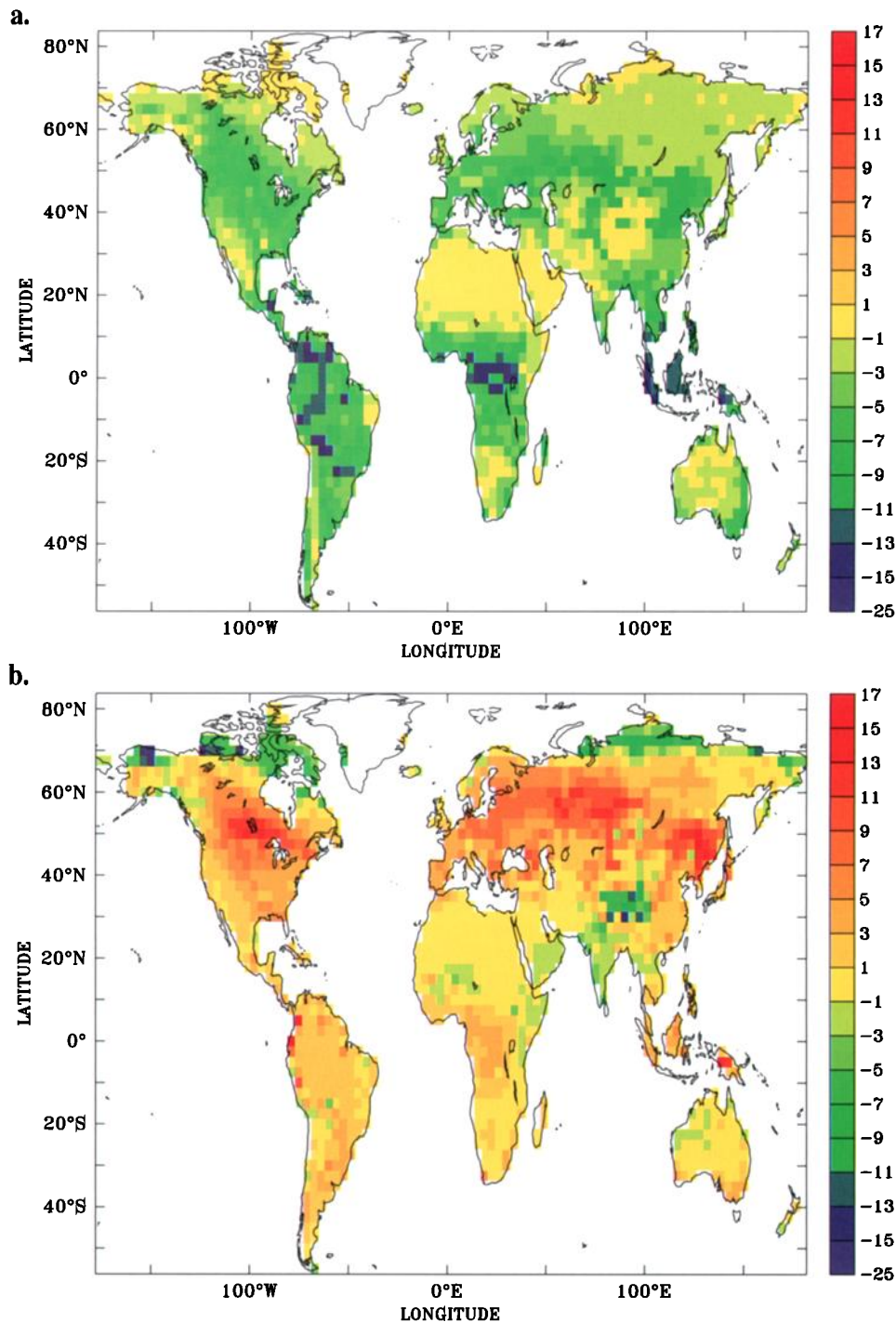
Finally, we explore the impact of uncertainties in the oceanic and terrestrial response on projected  $\text{CO}_2$  and climate. Experi-



**Figure 5.** (a) Atmospheric CO<sub>2</sub> is prescribed in transient simulations to increase exponentially within 70 and 210 years to a level of 1000 ppm and kept constant afterward. (b) Terrestrial carbon release simulated when prescribing CO<sub>2</sub> profiles shown in Figure 5a. Non-CO<sub>2</sub> radiative forcing is set to zero (thin line) or to 25% of the CO<sub>2</sub> radiative forcing (thick line). The model's climate sensitivity  $\Delta T_{2\times}$  is set to 4.5°C. Interannual variability in terrestrial release has been removed by smoothing. (c) Evolution of globally averaged surface temperature (thin solid line), carbon storage in vegetation (thick dot-dashed line), carbon storage in litter and soil (thick dashed line), and total biospheric storage (thick solid line). The increase in gross primary productivity is given in Gt C yr<sup>-1</sup> (thin dashed line; left axis). The transient simulation started at the end of the spin-up at model year 1000, and time is given in years since the start of the spin-up.

ments R and C represent upper and lower estimates for terrestrial carbon storage. The difference in the cumulative carbon uptake during this century amounts to 600 Gt C. This would correspond to a difference in projected CO<sub>2</sub> of 283 ppm (1 ppm = 2.123 Gt C). However, the ocean absorbs 131 Gt C more between year 2000 and 2100 under experiment C than under experiment R (Figure 4b) because of the higher growth in atmospheric CO<sub>2</sub> in experiment C. Thus 20% of the difference in terrestrial storage is counteracted by the ocean response, and the difference in projected CO<sub>2</sub> is accordingly reduced. Data-based estimates yield an average oceanic uptake of  $2 \pm 0.6$  Gt C yr<sup>-1</sup> for the 1980s. We have scaled all transport parameters of the ocean model (including gas exchange) by a factor of 0.5 and 1.5, thereby mimicking a faster and slower overturning ocean.

Average ocean uptake for the period from year 1980 to 1989 is 2.0 Gt C yr<sup>-1</sup> for the standard model setup, 1.46 Gt C yr<sup>-1</sup> for the "slow ocean," and 2.54 Gt C yr<sup>-1</sup> for the "fast ocean" in accordance with the range of data-based estimates. Cumulative ocean uptake during the century is  $\pm 90$  Gt C, roughly 20% different compared to the standard ocean model setup for scenario A1B. Combining the fast overturning ocean and the terrestrial model setup of experiment R (experiment R-fast) yields a lower limit in projected CO<sub>2</sub>, while combining the slow overturning ocean with experiment C (experiment C-slow) yields an upper limit. For scenario A1B, projected CO<sub>2</sub> is 620 and 920 ppm at year 2100 for the two experiments. Projected CO<sub>2</sub> varies between 820 and 1250 ppm for scenario A1FI and between 490 and 680 ppm for scenario B1 (Figure 4c). This



**Plate 3.** Simulated changes up to 2100 in terrestrial carbon storage in kg C m<sup>-2</sup> (negative numbers denote uptake) for scenario A1B and experiment (a) T0, where the models climate sensitivity is set to zero (no climate change), and (b) for experiment C0, where CO<sub>2</sub> fertilization is suppressed.

translates into a deviation in global average surface temperature at year 2100 of roughly  $-0.3^{\circ}$  to  $+0.6^{\circ}\text{C}$  compared to that of the standard A1B, A1FI, and B1 simulations. The range in projected CO<sub>2</sub> is smaller than the sum of the uncertainty ranges in terrestrial and oceanic storage. This is because uncertainties in oceanic uptake are moderated through changes in terrestrial uptake and vice versa. In conclusion, the above analyses suggest that for a given emission scenario, projected CO<sub>2</sub> at 2100 is

uncertain by  $-10$  to  $+30\%$  because of uncertainties in modeled land and ocean processes.

#### 4. Discussion and Conclusion

Limitations of our modeling studies are that land use changes are not explicitly considered and that modeled land surface changes do

not affect the circulation and the hydrological cycle in the atmosphere. Regional model results need to be interpreted with caution also because different climate models yield different regional changes in temperature and precipitation, which could imply important differences in the response of the terrestrial biosphere to global climate change.

For a range of model assumptions, we find that the terrestrial biosphere could act as a strong carbon source or sink during this century. Factors that influence terrestrial carbon storage are the assumed sensitivity of climate to radiative forcing [Houghton *et al.*, 1996], the sensitivity of soil respiration to temperature [Giardina and Ryan, 2000; Lloyd and Taylor, 1994], and the rate of increase in radiative forcing by both CO<sub>2</sub> and other GHGs. For a given emission scenario, projected atmospheric CO<sub>2</sub> is estimated to be uncertain by -10 to +30%. From these results and from earlier studies investigating feedbacks on oceanic carbon uptake [Maier-Reimer *et al.*, 1996; Sarmiento and Le Quéré, 1996; Matear and Hirst, 1999; Joos *et al.*, 1999b; Plattner *et al.*, 2001], we conclude that uncertainties in our understanding of basic mechanisms have a significant impact on projected CO<sub>2</sub> and climate. Nevertheless, the difference among the SRES scenarios are large, and it is clear that anthropogenic carbon emissions themselves will represent the dominant control over atmospheric CO<sub>2</sub> concentration during the next 100 years.

Cox *et al.* [2000] applied version HadCM3 of the Hadley Centre climate model coupled to a carbon cycle model for scenario IS92a. They projected atmospheric CO<sub>2</sub> to rise to 980 ppm by 2100. For the same scenario and year the Bern CC model yields 700 ppm (range of 630–900 ppm). The difference in simulated CO<sub>2</sub> concentration is due in part to a much slower surface to deep water transport in the Hadley Centre model than in the Bern CC model and in part to the representation of soil and litter carbon by one single reservoir in combination with a high simulated warming rate, which causes a substantial fraction of this carbon to be rapidly released. In comparison with observations, the atmospheric CO<sub>2</sub> increase up to 2000 is overestimated by 15%, and the increase in global mean surface temperature is overestimated by more than a factor of 2 in the simulation by Cox *et al.* [2000]. The high simulated warming rate is due to the neglect of cooling by sulphate aerosols, and the relatively high climate sensitivity of HadCM3. When combined with a conventional formulation of the dependence of soil respiration rates on temperature, the high warming rate results in a massive loss of carbon from soil and litter. Current ocean carbon uptake in HadCM3 is at the lower end of both data-based estimates and the range of current three-dimensional ocean models, suggesting that the simulated surface to deep water mixing rate may be too sluggish. In simulations where CO<sub>2</sub> reaches around 700 ppm at 2100, ocean uptake for the period from 2000 to 2100 is 410 Gt C in the Bern CC model but only 250 Gt C in the Hadley Centre model.

Shifts in ecosystem structure are found for all six illustrative nonintervention scenarios in the Bern CC model. Boreal trees in middle northern latitudes are replaced to a greater or lesser extent by grasses and temperate trees, whereas boreal trees invade high northern latitudes. Paleodata confirm that analogous vegetation changes have taken place (albeit in response to much slower climate changes) as those simulated for the next hundred years, and comparably rapid changes in vegetation structure are known to have occurred earlier during rapid climate change episodes, such as beginning and end of the Younger Dryas cold event [e.g., *Cooperative Holocene Mapping Project Members*, 1988; Birks and Ammann, 2000]. Such changes taking place in the modern world would inevitably impact regional forest industries and economies. The extent of the regions affected depends on the magnitude of the climate change, which in turn depends on GHG emissions.

High GHG emissions and a high contribution of non-CO<sub>2</sub> agents to radiative forcing tend to reduce the fraction of the

anthropogenic carbon emissions that is taken up by the terrestrial biosphere and the ocean. For example, 57% of the anthropogenic emissions are still airborne at 2100 for scenario A1FI, whereas only 41% remain airborne for scenario B1. This suggests that low GHG emissions yield a double dividend, allowing a lower rate of warming and a lower fraction of CO<sub>2</sub> emissions remaining in the atmosphere.

## Appendix A

### A1. Abundances and Burdens of Non-CO<sub>2</sub> Greenhouse Gases

Calculating the abundance of chemically reactive gases from emissions requires a model that can predict how the lifetimes of these gases are changed by an evolving atmospheric chemistry. Here we rely on results of a modeling workshop called OxComp where 14 state-of-the-art chemistry transport models were run under a set of emission scenarios. From the results of these simulations, simplified expressions were developed to estimate the evolution of OH and O<sub>3</sub> as a function of pollutant emissions and the impact of changing N<sub>2</sub>O emissions on the lifetime of N<sub>2</sub>O as summarized below [Prather *et al.*, 2001; M. Prather, personal communication, September 2000].

Global mean atmospheric abundances or burdens are projected using anthropogenic emissions from Nakićenović *et al.* [2000]. Changes in CH<sub>4</sub>, N<sub>2</sub>O, the fully fluorinated species SF<sub>6</sub>, CF<sub>4</sub>, C<sub>2</sub>F<sub>6</sub>, C<sub>4</sub>F<sub>10</sub>, and the halocarbons HFC-23, HFC-32, HFC-125, HFC-134a, HFC-143a, HFC-152a, HFC-227ea, HFC-245ca, and HFC-43-10mee are estimated from the budget equation:

$$\frac{dC}{dt} = E - \frac{1}{\tau} C, \quad (\text{A1})$$

where  $C$  is the concentration (in ppb),  $E$  is total (natural and anthropogenic) emission (in ppb yr<sup>-1</sup>), and  $\tau$  is the lifetime (in years).

Natural emissions for CH<sub>4</sub> and N<sub>2</sub>O are estimated from (A1) by specifying all other terms for today. For CH<sub>4</sub> the present lifetime is taken to be 8.4 years, the concentration is taken to be 760 ppb, corresponding to a burden of 4885 Tg(CH<sub>4</sub>), the atmospheric increase is taken to be 22 Tg(CH<sub>4</sub>) yr<sup>-1</sup>, and anthropogenic emissions are 323 Tg(CH<sub>4</sub>) [Prather *et al.*, 2001; Nakićenović *et al.*, 2000]. This yields a natural flux to the atmosphere of 301 Tg(CH<sub>4</sub>). For N<sub>2</sub>O the present lifetime is taken to be 120 years, the concentration is taken to be 316 ppb, corresponding to a burden of 1520 Tg N, the atmospheric increase is taken to be 3.8 Tg N yr<sup>-1</sup>, and anthropogenic emissions are 7.0 Tg N [Nakićenović *et al.*, 2000]. This yields a natural flux to the atmosphere of 9.5 Tg N [Prather *et al.*, 2001].

The lifetime of N<sub>2</sub>O is estimated taking into account the effect of a changing N<sub>2</sub>O burden on its own life time by

$$\tau(\text{N}_2\text{O}) = 120 \text{ years} \left( \frac{e\text{N}_2\text{O}(t)}{e\text{N}_2\text{O}(2000)} \right)^{-0.055}, \quad (\text{A2})$$

where  $e\text{N}_2\text{O}$  are total N<sub>2</sub>O emissions (in Tg N yr<sup>-1</sup>).

The relative change in tropospheric OH with respect to year 2000,  $r\text{OH}$ , is used to scale lifetimes of greenhouse gases reacting with OH. It is estimated by

$$\ln(r\text{OH}) = \ln \frac{\text{OH}(t)}{\text{OH}(2000)} = -0.32 \ln(\text{CH}_4(t)/\text{CH}_4(2000)) + 0.0042(e\text{NO}_x(t) - e\text{NO}_x(2000)) - 0.000105(e\text{CO}(t) - e\text{CO}(2000)) - 0.000315(e\text{VOC}(t) - e\text{VOC}(2000)), \quad (\text{A3})$$

where  $\text{CH}_4$  is atmospheric methane (in ppb),  $e\text{NO}_x$  denotes anthropogenic  $\text{NO}_x$  emissions in  $\text{Tg N yr}^{-1}$ , and  $e\text{CO}$  and  $e\text{VOC}$  are anthropogenic emissions of CO and volatile organic carbon compounds, respectively (in  $\text{Tg yr}^{-1}$ ).

The lifetime  $\tau$  of  $\text{CH}_4$  is then estimated from the present lifetimes with respect to reaction with OH (9.6 years), destruction in the stratosphere (120 years) and in soils (160 years), and the changes in OH:

$$\frac{1}{\tau(\text{CH}_4)} = \frac{r \text{ OH}}{9.58 \text{ years}} + \frac{1}{68.2 \text{ years}} \quad (\text{A4})$$

The lifetimes of the fully fluorinated species ( $\text{SF}_6$ ,  $\text{CF}_4$ ,  $\text{C}_2\text{F}_6$ , and  $\text{C}_4\text{F}_{10}$ ) are taken as time invariant, whereas the lifetimes of other halocarbons are scaled with OH:

$$\tau(t) = \frac{\tau(2000)}{r \text{ OH}} \quad (\text{A5})$$

Tropospheric  $\text{O}_3$  (in Dobson units (DU)) is estimated from

$$\begin{aligned} \text{O}_3(t) = & \text{O}_3(2000) + 5.0 \ln [\text{CH}_4(t)/\text{CH}_4(2000)] \\ & + 0.125 [e\text{NO}_x(t) - e\text{NO}_x(2000)] \\ & + 0.0011 [e\text{CO}(t) - e\text{CO}(2000)] \\ & + 0.0033 [e\text{VOC}(t) - e\text{VOC}(2000)]. \end{aligned} \quad (\text{A6})$$

$\text{CH}_4$  denotes atmospheric methane,  $e\text{NO}_x$  is anthropogenic emissions of  $\text{NO}_x$  (in  $\text{Tg N yr}^{-1}$ ),  $e\text{CO}$  and  $e\text{VOC}$  are anthropogenic emissions of CO and VOC (in  $\text{Tg yr}^{-1}$ ). The coefficients in the above equation represent recently updated values (M. Prather, personal communication, April 2001), whereas ~25% larger coefficients (6.7, 0.17, 0.0014, and 0.0042) were used in this study.

Lifetimes and concentrations at year 2000 and at preindustrial time  $t_0$  are given in Table A1. Concentrations of species controlled under the Montreal Protocol are not calculated but are prescribed following United Nations Environment Programme (UNEP)/World Meteorological Organization (WMO) Scientific Assessment of Ozone Depletion [UNEP/WMO, 1998].

## A2. Radiative Forcing by Greenhouse Gases

Radiative forcing provides a convenient first-order measure of the climatic importance of perturbations to the planetary radiation balance [Ramaswamy *et al.*, 2001; Shine *et al.*, 1990; Shine and Forster, 1999]. Global radiative forcing (RF) by well-mixed greenhouse gases is well defined ( $\pm 10\%$ ), whereas present and future radiative forcing by changes in tropospheric and stratospheric ozone are much more uncertain (approximately factor 2) [Shine and Forster, 1999]. Here radiative forcing is estimated from projected abundances using published simplified expressions [Ramaswamy *et al.*, 2001; Myhre *et al.*, 1998; Harvey *et al.*, 1996].

Radiative forcing for  $\text{CO}_2$  is

$$\text{RF}(\text{CO}_2) = 5.35 \text{ W m}^{-2} \left( \ln \frac{\text{CO}_2(t)}{\text{CO}_2(t_0)} \right), \quad (\text{A7})$$

where  $t_0$  is the preindustrial reference time. Radiative forcing for  $\text{CH}_4$  (in ppb) is

$$\begin{aligned} \text{RF}(\text{CH}_4) = & 0.036 \text{ W m}^{-2} \left[ \sqrt{\text{CH}_4(t)} - \sqrt{\text{CH}_4(t_0)} \right] \\ & - f[\text{CH}_4(t), \text{N}_2\text{O}(t_0)] - f[\text{CH}_4(t_0), \text{N}_2\text{O}(t_0)]. \end{aligned} \quad (\text{A8})$$

The function  $f$  that accounts for the overlap in  $\text{CH}_4$  and  $\text{N}_2\text{O}$  (in ppb) bands is

$$\begin{aligned} f(M, N) = & 0.47 \\ & \cdot \ln \left( 1 + 2.01 \times 10^{-5} (MN)^{0.75} + 5.31 \times 10^{-15} M (MN)^{1.52} \right). \end{aligned} \quad (\text{A9})$$

A similar expression holds for  $\text{N}_2\text{O}$ :

$$\begin{aligned} \text{RF}(\text{N}_2\text{O}) = & 0.12 \left[ \sqrt{\text{N}_2\text{O}(t)} - \sqrt{\text{N}_2\text{O}(t_0)} \right] \\ & - f[\text{CH}_4(t_0), \text{N}_2\text{O}(t)] - f[\text{CH}_4(t_0), \text{N}_2\text{O}(t_0)]. \end{aligned} \quad (\text{A10})$$

Radiative forcing by tropospheric ozone,  $\text{SF}_6$ , the gases controlled under the Montreal Protocol, and the other halocarbons included in this model is calculated by the expression

$$\text{RF} = \alpha [C(t) - C(t_0)], \quad (\text{A11})$$

where  $C$  is the concentration (in ppb (DU for  $\text{O}_3$ )) and  $\alpha$  is the radiative efficiency (in  $\text{W m}^{-2}$  per ppb or per DU for  $\text{O}_3$ ; Table A1).

The radiative forcing due to the stratospheric  $\text{O}_3$  depletion is calculated as a function of equivalent effective stratospheric chlorine (EESC) (in ppb) as taken from WMO/UNEP, chapter 11 [UNEP/WMO, 1998]. Important bromines are taken into account in the WMO/UNEP data. Thus

$$\text{RF}[\text{O}_3(\text{stratosphere})] = \alpha [\text{EESC}(t) - \text{EESC}(1970)], \quad (\text{A12})$$

where  $\alpha$  is in  $\text{W m}^{-2} \text{ ppb}^{-1}$ .

Finally, the small radiative forcing from changing concentrations in stratospheric  $\text{H}_2\text{O}$  due to  $\text{CH}_4$  oxidation is estimated to be 5% of the pure radiative forcing by  $\text{CH}_4$ :

$$\text{RF}(\text{H}_2\text{O}) = 0.05 \left[ 0.036 \left( \sqrt{\text{CH}_4(t)} - \sqrt{\text{CH}_4(t_0)} \right) \right]. \quad (\text{A13})$$

## A3. Radiative Forcing by Aerosols

The magnitude of radiative forcing by aerosols is not well understood and uncertain [e.g., Penner *et al.*, 2001; Ramaswamy *et al.*, 2001; Houghton *et al.*, 1996; Shine and Forster, 1999]. Its quantification using complex models or by simplified expressions approximating model results is highly tentative. Our motivation to estimate aerosol forcings from emissions of aerosols and aerosol precursors is to take into account at least at first order their significant contribution to the total radiative forcing of all anthropogenic agents on the global scale. We note that aerosol forcing varies strongly between different regions. Here we rely on published expressions and best estimates for present-day forcing [Ramaswamy *et al.*, 2001; Harvey *et al.*, 1996; Shine and Forster, 1999].

Direct radiative forcing by sulphate aerosols is taken to be proportional to anthropogenic sulphur emissions,  $e\text{SO}_x$  (in  $\text{Tg S yr}^{-1}$ ) with an RF of  $-0.4 \text{ W m}^{-2}$  for today:

$$\text{RF}(\text{S} - \text{direct}) = -0.4 \text{ W m}^{-2} \frac{e\text{SO}_x(t)}{e\text{SO}_x(2000)}. \quad (\text{A14})$$

Indirect aerosol effects via changing cloud properties are approximated as a function of sulphur emissions [Harvey *et al.*, 1996]

$$\begin{aligned} \text{RF}(\text{S} - \text{indirect}) = & -0.8 \text{ W m}^{-2} \ln \frac{\text{Enat} + e\text{SO}_x(t)}{\text{Enat}} \\ & \cdot \left( \ln \frac{\text{Enat} + e\text{SO}_x(2000)}{\text{Enat}} \right)^{-1}. \end{aligned} \quad (\text{A15})$$

$\text{Enat}$  denotes natural sulphur emissions taken to be 42 Tg S.

Finally, the combined direct effect of black carbon (BC) and organic carbon (OC) from biomass burning and fossil fuel use is

**Table A1.** Concentrations, Lifetimes, and Radiative Efficiencies<sup>a</sup>

Gas	$C(t_0)$ , ppb or ppt	$C(2000)$ , ppb or ppt	$\tau(2000)$ , years	$\alpha$ , $\text{W m}^{-2} \text{ ppb}^{-1}$
CO <sub>2</sub>	278,600	368,000	...	see text
CH <sub>4</sub>	742.2	1760	8.4	see text
N <sub>2</sub> O	272.0	316	120	see text
tropo-O <sub>3</sub> <sup>b</sup>	25	34	...	0.042
<i>Gases Controlled Under the Montreal Protocol</i>				
CFC-11	0	267	45	0.25
CFC-12	0	535	100	0.32
CFC-113	0	85	85	0.30
CFC-114	0	16	300	0.31
CFC-115	0	9	1700	0.18
CCl <sub>4</sub>	0	92	35	0.13
CH <sub>3</sub> CCl <sub>3</sub>	0	44	4.8	0.06
HCFC-22	0	145	11.9	0.20
HCFC-141b	0	13	9.3	0.14
HCFC-142b	0	15	19	0.20
HCFC-123	0	0	1.4	0.20
CF <sub>2</sub> BrCl	0	4	11	0.30
CF <sub>3</sub> Br	0	3	65	0.32
EESC <sup>c</sup>	1.25	3.28	...	-0.07317
<i>Other Halocarbons and SF<sub>6</sub></i>				
CF <sub>4</sub>	44	82.0	50,000	0.08
C <sub>2</sub> F <sub>6</sub>	0	3.2	10,000	0.26
C <sub>4</sub> F <sub>10</sub>	0	0.0	2,600	0.33
SF <sub>6</sub>	0	4.7	3,200	0.52
HFC-23	0	15.0	260	0.16
HFC-32	0	0.0	5.0	0.09
HFC-125	0	0.0	29	0.23
HFC-134a	0	12.0	13.7	0.15
HFC-143a	0	0.0	52	0.13
HFC-152a	0	0.7	1.4	0.09
HFC-227ea	0	0.0	33	0.30
HFC-245ca	0	0.0	5.9	0.23
HFC-43-10mee	0	0.0	15.2	0.40

<sup>a</sup> Concentrations  $C$  at preindustrial time ( $t_0$ ) taken to be nominal year 1765 and for today in parts per billion for CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O and in parts per trillion for halocarbons and SF<sub>6</sub>. Here  $\tau$  denotes lifetime (for CO<sub>2</sub>, no single lifetime exists) and  $\alpha$  radiative efficiency. The lifetimes of the fully fluorinated species (SF<sub>6</sub>, CF<sub>4</sub>, C<sub>2</sub>F<sub>6</sub>, and C<sub>4</sub>F<sub>10</sub>) are taken as time invariant, whereas the lifetimes of other halocarbons are scaled with tropospheric OH.

<sup>b</sup> Tropospheric ozone burden is given in Dobson units, and the radiative efficiency (in  $\text{W m}^{-2} \text{ DU}^{-1}$ ) was calculated assuming a present-day forcing of  $0.378 \text{ W m}^{-2}$ .

<sup>c</sup> Equivalent effective stratospheric chlorine (EESC; including bromine components) is given in parts per billion for 1970 ( $t_0$ ) and today. It is used to calculate radiative forcing from stratospheric O<sub>3</sub> depletion since 1970; the coefficient  $\alpha$  relating radiative forcing and EESC is obtained assuming a present-day forcing of  $-0.15 \text{ W m}^{-2}$ .

taken to be proportional to anthropogenic emissions of CO,  $e\text{CO}$ , with a present-day forcing of  $-0.1 \text{ W m}^{-2}$ , consistent with recent estimates [Ramaswamy *et al.*, 2001]

$$\text{RF}(\text{OC} + \text{BC} - \text{direct}) = -0.1 \text{ W m}^{-2} \frac{e\text{CO}(t)}{e\text{CO}(2000)}. \quad (\text{A16})$$

Here the CO emissions tabulated in SRES [Nakićenović *et al.*, 2000] are offset by a constant value to yield a year 2000 emission of 1036 Tg(CO).

cover, and ocean thermal expansion [Hooss *et al.*, 1999, Hooss *et al.*, 2001].

Hooss *et al.* [1999] have applied an empirical orthogonal function analysis to extract the climate change signal from an 850 year simulation of the ECHAM3/LSG where atmospheric CO<sub>2</sub> was prescribed to rise exponentially to reach a value four-fold of the preindustrial value at model year 120; afterward, CO<sub>2</sub> was kept constant [Voss and Mikolajewicz, 2001]. The perturbation of a climate variable,  $\Delta v$ , is represented as the superposition of a set of mutually orthogonal spatial patterns,  $\text{EOF}_i^{\Delta v}(\mathbf{x})$  and the time-dependent scalar coefficients termed principal components,  $\text{PC}_i^{\Delta v}(t)$ :

$$\Delta v(\mathbf{x}, t) = \sum_i \text{PC}_i^{\Delta v}(t) \text{EOF}_i^{\Delta v}(\mathbf{x}). \quad (\text{A17})$$

#### A4. Changes in the Fields of Surface Temperature, Precipitation, Cloud Cover, and Ocean Thermal Expansion

The spatiotemporal response of an AOGCM to an increase in radiative forcing can be captured by a combination of IRFs that take into account the inertia in the climate system and EOFs that describe the spatial patterns of climate change. Here we apply an IRF-EOF substitute of the ECHAM3/LSG AOGCM to project the evolution of surface temperature, precipitation, cloud

Each of the PC-EOF pairs  $i$  is computed from the AOGCM output in successive order to explain the maximum possible variance in a climate variable.

In the substitute the global mean perturbation since preindustrial time ( $t_0$ ) of each climate variable  $\Delta v_{av}(t)$  is calculated from the

**Table A2.** Parameters of the IRF-EOF substitute of ECHAM3/LSG<sup>a</sup>

Variable	$a_1^v$	$a_2^v$	$\tau_1^v$ , years	$\tau_2^v$ , years	$S^v$	$S_{2x}^v$
Temperature	0.290	0.710	448	14.4	0.674	2.5°C
Precipitation	0.345	0.655	1098	35.0	17.3	64 mm yr <sup>-1</sup>
Cloud cover	0.218	0.782	391	12.0	-0.238	-0.88%
Thermal expansion	0.960	0.040	836	31.0	34.5	128 cm

<sup>a</sup> Here  $a_1^v$  and  $\tau_1^v$  are the coefficients used to calculate the impulse response function. The climate sensitivity  $S^v$  is given in °C (W m<sup>-2</sup>)<sup>-1</sup> for surface temperature, in mm yr<sup>-1</sup> (W m<sup>-2</sup>)<sup>-1</sup> for precipitation, in % (W m<sup>-2</sup>)<sup>-1</sup> for cloud cover, and in cm (W m<sup>-2</sup>)<sup>-1</sup> for thermal expansion.  $S_{2x}^v$  is the climate sensitivity for a doubling of CO<sub>2</sub>.

convolution integral of the appropriate IRF  $R^v$  and the change in radiative forcing (RF):

$$\Delta v_{av}(t) = S^v \int_{t_0}^t dt' [R^v(t-t')] \left( \frac{d}{dt'} \text{RF}(t') \right). \quad (\text{A18})$$

$S^v$  is the climate sensitivity expressed as equilibrium change in  $\Delta v_{av}$  for a change in RF of 1 W m<sup>-2</sup> (Table A2). The change in  $v$  at location  $\mathbf{x}$  and time  $t$  is

$$\Delta v(\mathbf{x}, t) = \left[ \Delta v_{av}(t) \right] \text{EOF}_1^{\Delta v}(\mathbf{x}), \quad (\text{A19})$$

when  $\text{EOF}_1^{\Delta v}$  is normalized to yield unity when averaged over the globe. The (normalized) IRFs,  $R^v$ , to a step increase in radiative forcing are expressed analytically by

$$R^v(t) = \sum_i a_i^v \left[ 1 - \exp\left(-\frac{t}{\tau_i^v}\right) \right]. \quad (\text{A20})$$

The coefficients of  $R^v$  (equation (A20) and Table A2) were determined such as to minimize the sum of the squared deviations between the values of  $\Delta v_{av}(t)$  (from (A18)) and  $PC_1^{\Delta v}$  from the 850 year transient AOGCM simulation.

The substitute is built from the first EOF-PC pair only. Only the first EOF-PC pair of each of the four climate variables showed a long-term trend identifying them as climate change signal. The first EOF and PC capture 97% of the variability in near-surface temperature, 43% of the variability in cloud cover, and 31% of the variability in precipitation. Changes in the frequency of extreme events such as droughts and other changes in the variability of precipitation and cloud cover not described by the first EOF-PC pair are neglected in the substitute. Radiative forcing in the substitute is taken to be the sum of global mean radiative forcing from GHGs and aerosols, whereas the IRF-EOF pairs were determined from a CO<sub>2</sub> only simulation. This approximation seems justified because the correlation between the fields obtained from GHG only simulations and from simulations with GHGs and aerosols for a distinct AOGCM is, in general, higher than the correlation between output fields obtained with different AOGCMs.

#### A5. Atmospheric CO<sub>2</sub> and Oceanic Carbon Uptake

Atmospheric CO<sub>2</sub>,  $p\text{CO}_{2a}$ , is projected from the following budget equation:

$$\frac{dp\text{CO}_{2a}}{dt} = F_{\text{fossil}} + F_{\text{land-use}} - F_{ab} - F_{as}. \quad (\text{A21})$$

$F_{\text{fossil}}$  and  $F_{\text{land-use}}$  represent emissions by fossil fuel burning and by land use changes.  $F_{ab}$  and  $F_{as}$  denote the uptake of excess carbon by the land biosphere as calculated with the LPJ-DGVM

and the ocean. All fluxes are expressed in units of ppm yr<sup>-1</sup> (1 ppm = 2.123 × 10<sup>15</sup> g C = 7.779 × 10<sup>15</sup> g CO<sub>2</sub> = 1.768 × 10<sup>14</sup> mol).

The oceanic uptake is either calculated by the HILDA model or its impulse response representation [Joos *et al.*, 1996]. In the mixed layer impulse response substitute, ocean uptake is

$$F_{as} = K_g(p\text{CO}_{2a} - p\text{CO}_{2s}). \quad (\text{A22})$$

$K_g$  represents the global average gas exchange coefficient that is (9.06 years)<sup>-1</sup> for HILDA, and  $p\text{CO}_{2a}$  and  $p\text{CO}_{2s}$  correspond to the global average partial pressures of CO<sub>2</sub> in the atmosphere and the surface ocean (expressed in units of ppm); total pressure at sea level is taken to be 1 atm. The perturbation in dissolved inorganic carbon in the surface ocean  $\delta\Sigma\text{CO}_2$  is obtained from the convolution integral of the mixed layer impulse response function  $r_s$  and the net air-to-sea flux  $F_{as}$ :

$$\delta\Sigma\text{CO}_2 = \frac{c}{hA_{oc}} \int_{t_0}^t F_{as}(t') r_s(t-t') dt', \quad (\text{A23})$$

where  $h$  is the mixed layer depth taken to be 75 m,  $A_{oc}$  is the ocean area taken to equal 3.62 × 10<sup>14</sup> m<sup>2</sup>, and  $c$  is a unit conversion factor ( $c = 1.722 \times 10^{17}$  μmol m<sup>3</sup> ppm<sup>-1</sup> kg<sup>-1</sup>).

The relationship between the perturbation in the sea surface partial pressure relative to the preindustrial temperature,  $T_0$ , and inorganic carbon is obtained by the following expression that holds for 0 ≤  $\delta p\text{CO}_{2s}$  ≤ 1320 ppm and 17.7°C ≤  $T_0$  ≤ 18.3:

$$\begin{aligned} \delta p\text{CO}_{2s}(T_0) = & (1.5568 - 1.3993T_0 \times 10^{-2})\delta\Sigma\text{CO}_2 \\ & + (7.4706 - 0.20207T_0) \times 10^{-3}(\delta\Sigma\text{CO}_2)^2 \\ & - (1.2748 - 0.12015T_0) \times 10^{-5}(\delta\Sigma\text{CO}_2)^3 \\ & + (2.4491 - 0.12639T_0) \times 10^{-7}(\delta\Sigma\text{CO}_2)^4 \\ & - (1.5468 - 0.15326T_0) \times 10^{-10}(\delta\Sigma\text{CO}_2)^5. \end{aligned} \quad (\text{A24})$$

The preindustrial global average surface ocean temperature  $T_0$  is taken to be 18.2°C, and  $\delta p\text{CO}_{2s}$  and  $\delta\Sigma\text{CO}_2$  are again in units of ppm and μmol kg<sup>-1</sup>. The CO<sub>2</sub> partial pressure increases exponentially with sea surface temperature [Takahashi *et al.*, 1993], and we estimate sea surface partial pressure at time  $t$  to be

$$p\text{CO}_{2s}(t) = [p\text{CO}_{2s}(t_0) + \delta p\text{CO}_{2s}(T_0)] \exp(0.0423 \text{ K}^{-1} \Delta T). \quad (\text{A25})$$

Here  $p\text{CO}_{2s}(t_0)$  is the preindustrial sea surface pressure taken to be equal the atmospheric pressure, and  $\Delta T$  is the perturbation in global mean sea surface pressure, here calculated with the IRF-EOF substitute of the ECHAM3/LSG. This approximation gives similar results for the reduction in oceanic carbon uptake by SST warming as found for spatially resolved models (see Figure 3 of Joos *et al.* [1999b] and Plattner *et al.* [2001]).



Finally, the mixed layer impulse response  $r_s$  for the HILDA model is, for  $0 \leq t \leq 2$  years,

$$r_s(t) = 0.12935 + 0.21898 \exp(-t/0.034569) + 0.17003 \exp(-t/0.26936) + 0.24017 \exp(-t/0.96083) + 0.24093 \exp(-t/4.9792) \quad (\text{A26})$$

(check value:  $r_s(t = 2 \text{ years}) = 0.32071$ ), and for  $t \geq 2$  years is

$$r_s(t) = 0.022936 + 0.24278 \exp(-t/1.2679) + 0.13963 \exp(-t/5.2528) + 0.089318 \exp(-t/18.601) + 0.037820 \exp(-t/68.736) + 0.035549 \exp(-t/232.30) \quad (\text{A27})$$

(check value:  $r_s(t = 2 \text{ years}) = 0.32068$ ).

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