

A simple global carbon and energy coupled cycle model for global warming simulation: sensitivity to the light saturation effect

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ABSTRACT

A simple Earth system model, the Four-Spheres Cycle of Energy and Mass (4-SCEM) model, has been developed to simulate global warming due to anthropogenic CO₂ emission. The model consists of the Atmosphere–Earth Heat Cycle (AEHC) model, the Four Spheres Carbon Cycle (4-SCC) model, and their feedback processes. The AEHC model is a one-dimensional radiative convective model, which includes the greenhouse effect of CO₂ and H₂O, and one cloud layer. The 4-SCC model is a box-type carbon cycle model, which includes biospheric CO₂ fertilization, vegetation area variation, the vegetation light saturation effect and the HILDA oceanic carbon cycle model. The feedback processes between carbon cycle and climate considered in the model are temperature dependencies of water vapor content, soil decomposition and ocean surface chemistry. The future status of the global carbon cycle and climate was simulated up to the year 2100 based on the “business as usual” (IS92a) emission scenario, followed by a linear decline in emissions to zero in the year 2200. The atmospheric CO₂ concentration reaches 645 ppmv in 2100 and a peak of 760 ppmv approximately in the year 2170, and becomes a steady state with 600 ppmv. The projected CO₂ concentration was lower than those of the past carbon cycle studies, because we included the light saturation effect of vegetation. The sensitivity analysis showed that uncertainties derived from the light saturation effect of vegetation and land use CO₂ emissions were the primary cause of uncertainties in projecting future CO₂ concentrations. The climate feedback effects showed rather small sensitivities compared with the impacts of those two effects. Satellite-based net primary production trends analyses can somewhat decrease the uncertainty in quantifying CO₂ emissions due to land use changes. On the other hand, as the estimated parameter in vegetation light saturation was poorly constrained, we have to quantify and constrain the effect more accurately.

1. Introduction

Global warming, resulting from anthropogenic greenhouse gas emissions such as CO₂, is one of the most significant of the Earth’s environmental prob-

lems. Anthropogenic CO₂ has been emitted into the atmosphere since the 18th century, and it was estimated to be approximately 7.0 GtC yr^{–1} during the 1980s (Prentice et al., 2001). Approximately one half of the emissions remain in the atmosphere, and the other half are absorbed by the ocean and terrestrial biosphere (Prentice et al., 2001). Since global warming is caused by the coupled feedback cycles of energy and carbon (Charlson 2000; Prentice et al., 2001), it is essential to assess the impact of various feedback processes on CO₂ uptake by the land and oceans to

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enable projection of the carbon cycle and climate of the future.

Most of the previous studies on projection of atmospheric CO₂ and climate have not considered these feedback processes between the carbon cycle and climate systems. For instance, simple carbon cycle models were used to predict atmospheric CO₂ concentrations under a given scenario of anthropogenic CO₂ emissions, and then the predicted atmospheric CO₂ variations were input into the General Circulation Model (GCM) for the future climate simulation. Then, dynamic carbon cycle models were applied under simulated climate scenarios for the analysis of future biosphere and ocean carbon cycle conditions (Cao and Woodward, 1998; Sarmiento et al., 1998).

Earth system models that integrate models of climate and carbon cycle were only recently developed. Cox et al. (2000) showed that the feedback processes could accelerate climate change based on the fully coupled, three-dimensional carbon–climate model. However, changes in CO₂ emission levels due to land use change are not removed directly from the terrestrial biosphere in their model. Another problem is that their model generates higher atmospheric CO₂ levels and temperatures compared with current levels, which may likewise be amplified in projections and may result in an exaggerated climate feedback effect.

Simple Earth system models were also developed for analysis of climate and carbon cycle feedback mechanisms (Lenton, 2000; Prentice et al., 2001). Lenton (2000) developed a simple Earth system model that consists of atmospheric, terrestrial and ocean carbon cycle models and an atmospheric climate model. In the third assessment report of the Intergovernmental Panel on Climate Change (IPCC), the Bern model and the Integrated Science Assessment Model (ISAM) were used as the simplified carbon–climate coupled models. Due to a complicated process-based dynamic global vegetation model (the LPJ model; Sitch, 2000), the Bern model cannot be regarded as a simple model. The simple models did not consider variation in vegetation area in a physically based manner or the impact of light availability on vegetation growth saturation (if vegetation density becomes high, vegetation growth is limited due to light availability), both of which have an important effect on the biospheric CO₂ uptake.

In the present study, we developed a simple Earth system model, the Four Spheres Cycle of Energy and Mass (4-SCEM) model, for analysis of the coupled cycles of energy and carbon. The feedback processes

included are (1) water vapor feedback, (2) the effect of CO₂ fertilization on vegetation photosynthesis and (3) temperature effects on soil decomposition, and on the atmosphere–ocean carbon exchange. The predictions of future carbon cycle and climate conditions were analyzed based on the given scenario of anthropogenic CO₂ emission levels. Sensitivities of the model parameters were also tested in this study. Details of the 4-SCEM model are described in section 2. Section 3 describes model calibrations, and the projection of future carbon cycle and climate conditions. The results of sensitivity studies are shown in section 4. Finally, section 5 presents the study's conclusions.

2. Model

The 4-SCEM model consists of two sub-models, one that is a carbon cycle model (Four Spheres Carbon Cycle model; 4-SCC) and an energy cycle model (Atmosphere–Earth Heat Cycle model; AEHC). The outline of 4-SCEM is shown in Fig. 1. Feedback processes included in the model are (1) CO₂ fertilization effects on the biosphere, (2) biospheric productivity saturation effect due to increases in vegetation density, (3) temperature dependence of soil decomposition, (4) temperature dependence of ocean surface chemistry and (5) water vapor feedback in the atmosphere. A detailed description of each model and feedback process is provided below.

2.1. Atmosphere–Earth Heat Cycle (AEHC) model

The AEHC model is a one-dimensional radiative convective model based on Hayashi (1991), shown in Fig. 1. The modeled atmosphere consists of 13 layers from surface to tropopause (assumed to be 12 km in height). The convective adjustment scheme (e.g. Manabe and Strickler, 1964) was used to make a realistic temperature profile in the troposphere.

The time variation of temperature in an atmospheric layer at height z can be written as

$$\frac{\partial T(z)}{\partial t} = -\frac{1}{\rho(z)C_p} \left\{ \int_{SW} \frac{\partial S(z, \nu)}{\partial z} d\nu + \int_{LW} \frac{\partial L(z, \nu)}{\partial z} d\nu \right\}, \quad (1)$$

where $T(z)$, $\rho(z)$, C_p , $S(z, \nu)$ and $L(z, \nu)$ are temperature, density, specific heat and net flux of short-wave and longwave at height z and wavenumber ν ,

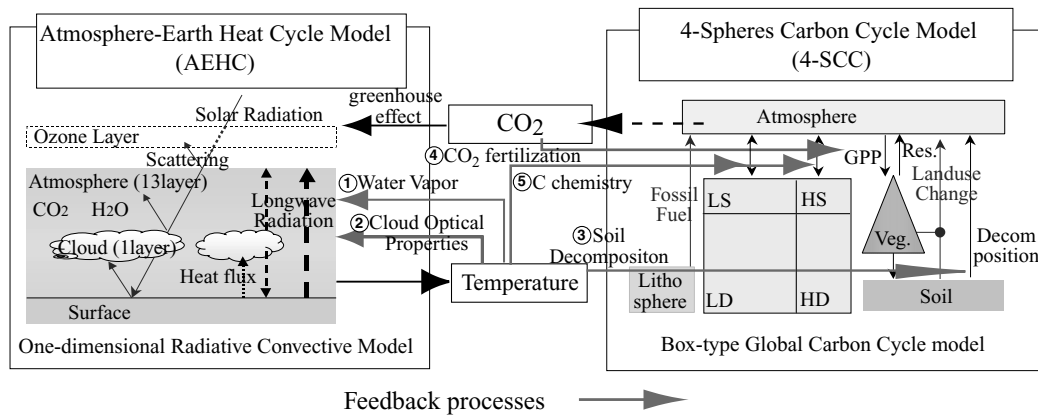


Fig. 1. Outline of four spheres cycle of energy and mass model. It consists of atmospheric one-dimensional radiative convective model and terrestrial and ocean carbon cycle model. Each model is connected with feedback processes as described in the text.

respectively. The shortwave and longwave flux were obtained from the transmittance of each layer, incident solar flux at the top of the atmosphere and longwave flux from each atmospheric layer and the Earth's surface.

Transmittance of an atmospheric layer at height z and wavenumber ν is based on the Random model, and band equivalent widths of H_2O and CO_2 are derived from the absorption line data of Houghton (1977). H_2O was expressed with the temperature feedback effect based on the Clausius–Clapeyron equation:

$$\rho_w = 0.217 \frac{H_R P_0 \exp(-l_e/R_w T)}{T}, \quad (2)$$

where ρ_w , H_R , P_0 , l_e and R_w are the atmospheric water density, relative humidity, a constant for the water

vapor saturation curve (1.4×10^{11} Pa; Nakajima et al., 1992), latent heat of water (43655 J mol^{-1}) and gas constant ($8.314 \text{ J mol}^{-1} \text{ K}^{-1}$), respectively. A value of relative humidity H_R was determined iteratively, as discussed later. We have also included the Rayleigh scattering (Manabe and Strickler, 1964) and ozone absorption (Lacis and Hansen, 1974) to estimate the incoming shortwave flux at the top of troposphere.

We have set one homogeneous cloud layer in the AEHC model. The cloud data, cloud cover ratio, height and optical thickness are shown in Table 1 derived from remote sensing data sets (Rossow et al., 1996). We have obtained the cloud radiative properties based on Stephens's parameterization (Stephens, 1978), which estimates the cloud optical properties from the cloud liquid water content. Temperature

Table 1. Parameters used in AEHC model

Parameter	Description	Value	Notice
H_R	Relative humidity	0.68	
C_p	Specific heat	$1004.0 \text{ J kg}^{-1} \text{ K}^{-1}$	
P_0	Water vapor saturation constant	$1.4 \times 10^{11} \text{ Pa}$	Nakajima et al., 1992
F	Cloud feedback parameter	0	
	Cloud cover ratio	0.623	Rossow et al., 1996
	Cloud height	4 km	Rossow et al., 1996
A	Surface albedo	0.132	Rossow et al., 1996
LWC(0)	Pre-industrial cloud liquid water content	23.5 g m^{-2}	Derived from ISCCP data and Stephens (1978)
C_h	Heat capacity of Earth's surface	$4.69 \times 10^{23} \text{ J K}^{-1}$	
A_e	Surface area	$5.101 \times 10^{14} \text{ m}^2$	
H_f	Sensible and latent heat flux from surface	38 W m^{-2}	

dependency of cloud liquid water content was derived from Somerville and Remer (1984):

$$\text{LWC}(t) = \text{LWC}(0)\{1 + f[T(t) - T(0)]\}, \quad (3)$$

where $\text{LWC}(t)$ and $T(t)$ are cloud liquid water content and temperature at time t ($t = 0$ indicates its pre-industrial value), and f is the cloud liquid water feedback parameter. In this study, we set $f = 0$ (no cloud–temperature feedback), and its sensitivity will be analyzed in a future study.

Upwelling shortwave and longwave fluxes from the Earth's surface were calculated from surface albedo and temperature. Surface albedo was derived using the 8 yr averaged value from the ISCCP C2 data set (Rossow et al., 1996), and set at 0.12. Time variation of temperature on the Earth's surface (T_s) can be expressed as

$$C_h \frac{\partial T_s}{\partial t} = (F_d - \sigma T_s^4) a_e \quad (4)$$

where C_h , F_d , σ and a_e are the heat capacity of the Earth's surface ($4.69 \times 10^{23} \text{ J K}^{-1}$; Huntingford and Cox, 2000), the net incoming surface flux (shortwave, longwave and sensible and latent heat flux), the Stefan–Boltzmann constant ($5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$) and surface area ($5.101 \times 10^{14} \text{ m}^2$), respectively. Ocean heat uptake may have a significant effect on the climate, and we can include the effect in the parameter of heat capacity of earth surface by tuning its value.

Finally, we determined the value of relative humidity and latent and sensible heat transport. These values were obtained by iterative procedure to satisfy the following conditions: (1) the Earth's surface temperature of 288 K (average surface temperature) and (2) continuous temperature between the Earth's surface and the lowest atmospheric layer. We set the relative humidity parameter to reproduce atmospheric temperature. A relative humidity value of 0.67 was assumed, which is less than the average observed surface value of ~ 0.77 . Then, the latent and sensible heat transport value was determined to make the temperature profile continuous in the boundary between the Earth's surface and atmosphere. A smaller value of latent and sensible heat transport (this sum is 38 W m^{-2} , which is smaller than the IPCC report of 102 W m^{-2} , Kiehl and Trenberth, 1997) was used to make the surface temperature reasonable.

2.2. Four Spheres Carbon Cycle (4-SCC) model

The 4-SCC model is a box-type simple carbon cycle model based on Ichii et al. (2001), utilized for past and future carbon cycle simulation. It deals with carbon flows among the four spheres (atmosphere, biosphere, lithosphere and hydrosphere). The biosphere consists of two boxes, the terrestrial living biosphere and soil organic matter. The ocean model is a high-latitude exchange/interior diffusion–advection (HILDA) model (Siegenthaler and Joos, 1992; Shaffer and Sarmiento, 1995). Anthropogenic effects included in the model are industrial emissions (fossil fuel and cement production) and land-use changes. Details of this model are described below.

2.2.1. Terrestrial biospheric component. The terrestrial biospheric model consists of two carbon reservoirs (vegetation and soil) and five carbon flows (gross primary production, respiration, litterfall, soil decomposition and land use change). Mass balances in the carbon reservoirs are as follows:

$$\frac{dC_{\text{veg}}(t)}{dt} = \text{GPP}(t) - \text{RES}(t) - \text{LIT}(t) - L_p \text{LU}(t) \quad (5)$$

$$\frac{dC_{\text{soil}}(t)}{dt} = \text{LIT}(t) - \text{DEC}(t) - (1 - L_p) \text{LU}(t), \quad (6)$$

where $C_{\text{veg}}(t)$ and $C_{\text{soil}}(t)$ are carbon masses of the vegetation and soil reservoirs, and $\text{GPP}(t)$, $\text{RES}(t)$, $\text{LIT}(t)$, $\text{LU}(t)$ and $\text{DEC}(t)$ are carbon fluxes of gross primary production (GPP), dark respiration, litterfall, CO_2 emission due to land use change and soil decomposition, respectively. The anthropogenic carbon emission due to land-use changes is directly removed from the vegetation and soil organic matter box using a fractionation factor of removing carbon from biosphere and soil organic matter (L_p). In an approach that is different from other similar simple models (e.g. Bacastow and Keeling, 1973; Wigley, 1993; Kwon and Schnoor, 1994; Craig and Holmén, 1995; Jain et al., 1995; Kheshgi et al., 1996; Lenton, 2000), we included (1) vegetation area variation and (2) photosynthesis limitation due to light saturation in the process of GPP.

Based on the Production Efficiency model ($\text{NPP} \text{ or } \text{GPP} = \text{PAR} \cdot \text{FPAR} \cdot \varepsilon$) PAR is photosynthetically active radiation, FPAR is the fraction of PAR absorbed by the canopy, and ε is the light utilization coefficient (e.g. Kumar and Monteith, 1981), $\text{GPP}(t)$ becomes

$$\text{GPP}(t) = \text{SOL} \, dt \, S(t) \left[1 - \exp\left(-k_{\text{eff}} k_{\text{lai}} \frac{C_{\text{veg}}(t)}{S(t)}\right) \right] \varepsilon \times \left[1 + \beta \ln\left(\frac{C_{\text{atm}}(t)}{C_{\text{atm}}(0)}\right) \right], \quad (7)$$

where SOL is incoming photosynthetic active radiation (PAR), dt is integrated time, $S(t)$ is the vegetation area at time t , k_{eff} is light extinction coefficient, k_{lai} is the vegetation density conversion factor from density to leaf area index (LAI) ($\text{LAI} \, \text{kg}^{-1} \, \text{m}^2$), β is the CO_2 fertilization factor and $C_{\text{atm}}(t)$ is atmospheric carbon mass at time t . The first term, $\text{SOL} \, dt \, S(t)$, indicates the total available incoming PAR over the vegetation area, the second term, $1 - \exp[-k_{\text{eff}} k_{\text{lai}} C_{\text{veg}}(t)/S(t)]$, is FPAR, which follows the Lambert–Beer law. In this study, we assumed that the leaf area index (LAI) is proportional to the vegetation density, $C_{\text{veg}}(t)/S(t)$. The last term, $\varepsilon \{1 + \beta [C_{\text{atm}}(t)/C_{\text{atm}}(0)]\}$, shows light utilization efficiency, and includes the CO_2 fertilization factor in logarithmic form (e.g. Bacastow and Keeling, 1973).

The carbon flux due to respiration and litterfall is proportional to the mass of living biosphere, and we included the temperature dependence of soil decom-

position based on the Q_{10} equation (e.g. Raich and Schlesinger, 1992). These equations follow:

$$\text{RES}(t) = K_{\text{res}} C_{\text{veg}}(t) \quad (8)$$

$$\text{LIT}(t) = K_{\text{lit}} C_{\text{veg}}(t) \quad (9)$$

$$\text{DEC}(t) = K_{\text{dec}} C_{\text{soil}}(t) Q_{10}^{\frac{T(t)-T(0)}{10}}, \quad (10)$$

where K_{res} , K_{lit} and K_{dec} are constant and Q_{10} is the Q_{10} factor.

Although there are many feedback effects included in more detailed terrestrial models such as geographic distribution in vegetation and vegetation structural variation and precipitation, it is very difficult to include in the simplified model and is not well validated even in the detailed terrestrial models. The simplified model must include these features as far as possible, and these effects in the terrestrial biosphere will be included in a future study.

The reference model parameters were defined as follows and are shown in Table 2. The pre-industrial carbon contents of vegetation and soil organic matter were derived from Siegenthaler and Sarmiento (1993). The initial terrestrial net primary production (NPP) was assumed to be $55 \, \text{GtC} \, \text{yr}^{-1}$ based on the average of

Table 2. Parameters used in the 4-SCC model

Parameter	Description	Value	Notice
$C_{\text{atm}}(0)$	Atmospheric in pre-industrial era	588 GtC	Etheridge et al., 1999
$C_{\text{veg}}(0)$	Vegetation in pre-industrial era	610 GtC	Siegenthaler and Sarmiento, 1993
$C_{\text{soil}}(0)$	Soil C in pre-industrial era	1600 GtC	Siegenthaler and Sarmiento, 1993
$\text{GPP}(0)$	GPP in pre-industrial era	120 GtC yr^{-1}	Prentice et al., 2001
$\text{RES}(0)$	Vegetation respiration in pre-industrial era	60 GtC yr^{-1}	Prentice et al., 2001
$\text{LIT}(0)$	Litterfall flux in pre-industrial era	60 GtC yr^{-1}	Prentice et al., 2001
$\text{DEC}(0)$	Soil decomposition in pre-industrial era	60 GtC yr^{-1}	Prentice et al., 2001
L_p	Fractionation factor of CO_2 emission due to land use change	0.7	Ichii et al., 2001
K_{res}	Plant respiration rate constant	60/610 yr^{-1}	$\text{RES}(0)/C_{\text{veg}}(0)$
K_{lit}	Plant litterfall rate constant	60/610 yr^{-1}	$\text{LIT}(0)/C_{\text{veg}}(0)$
K_{dec}	Soil decomposition rate constant	60/1600 yr^{-1}	$\text{DEC}(0)/C_{\text{soil}}(0)$
Q_{10}	Q_{10} factor	2.0	Potter et al., 1993; Cox et al., 2000
K	Vertical diffusion constant	4700.0 $\text{m}^2 \, \text{yr}^{-1}$	Siegenthaler and Joos, 1992
W	Upwelling velocity in the interior	0.73 $\text{m} \, \text{yr}^{-1}$	Siegenthaler and Joos, 1992
Q	Lateral exchange	0.002 381 $\text{m} \, \text{yr}^{-1}$	Siegenthaler and Joos, 1992
U	Exchange between surface and deep polar ocean	53.0 $\text{m} \, \text{yr}^{-1}$	Siegenthaler and Joos, 1992
Delta	The part of polar ocean	0.16	Shaffer and Sarmiento, 1995
Deltas	The part of surface layer free from ice	0.10	Shaffer and Sarmiento, 1995
D	Depth of interior	3800 m	Shaffer and Sarmiento, 1995
D_s	Depth of surface layer	50 m	Shaffer and Sarmiento, 1995

values estimated using terrestrial carbon cycle models (Cramer et al., 1999). The carbon flux in terrestrial GPP was assumed to be double that of NPP, 110 GtC yr⁻¹, and respiration is 55 GtC yr⁻¹. The fluxes of litter flow and soil decomposition are also 55 GtC yr⁻¹ for each. The L_p value was the same as that in our previous study (Ichii et al., 2001) and set to be 0.7. The value of $Q_{10} = 2.0$ was derived from a commonly used value (e.g. Cox et al., 2000; Potter et al., 1993).

The product of SOL dt $S(0)\varepsilon$, which we denote as $F_{\text{gpp}} = \text{SOL } dt S(0)\varepsilon$, was estimated as follows: SOL = 198 (W m⁻²) (incoming solar radiation at the Earth's surface; Kiehl and Trenberth, 1997) $\times 0.5$ (half of solar radiation is PAR; McCree, 1981) $\times dt = 86\,400 \times 365$ (s yr⁻¹), $S(0) = 8.7722 \times 10^{13}$ m² (the pre-industrial forest area from Houghton and Hackler, 2001), and $\varepsilon = 1.25 \times 10^{-6}$ gC J⁻¹ (Heimann and Keeling, 1989). The result was roughly 350 GtC yr⁻¹. Although there is a large uncertainty in each parameter, the values were used in the reference simulation. The product of k_{ext} and k_{lai} can be estimated based on the pre-industrial GPP and other parameters. The remaining parameter, β , was determined iteratively to fit the observed time variation of atmospheric CO₂.

2.2.2. Ocean carbon cycle model. We used the HILDA model (Siegenthaler and Joos, 1992; Shaffer and Sarmiento, 1995) as an ocean carbon cycle model. The ocean is divided into low and high latitudinal zones. The low latitude zone consists of a well mixed surface layer and a one-dimensional advective-diffusive deep ocean layer. The high latitude zone consists of a well mixed polar surface layer and a deep ocean layer. We show the HILDA model with tracer time variation only briefly, because Shaffer and Sarmiento (1995) did not describe the model equation in the time-variable version.

The time variations of tracer concentrations at the low latitude surface (Φ_{LS}), low latitude interior [$\Phi_{\text{I}}(z)$], high latitude surface [$\Phi_{\text{HS}}(z)$] and high latitude deep [Φ_{HD}] regions are as follows:

$$D \frac{d\Phi_{\text{LS}}}{dt} = g_{\text{LS}}(\Phi_{\text{A}} - \Phi_{\text{LS}}) - k \frac{d\Phi_{\text{I}}(z)}{dz(z=0)} + w[\Phi_{\text{I}}(0) - \Phi_{\text{LS}}] + S_{\text{LS}} \quad (11)$$

$$\frac{d\Phi_{\text{I}}(z)}{dt} = k \frac{d\Phi_{\text{I}}^2(z)}{dz^2} - w \frac{d\Phi_{\text{I}}(z)}{dz} - q[\Phi_{\text{I}}(z) - \Phi_{\text{HD}}] + S_{\text{I}}(z) \quad (12)$$

$$D_s \delta \frac{d\Phi_{\text{HS}}}{dt} = \delta_s g_{\text{HS}}(\Phi_{\text{A}} - \Phi_{\text{HS}}) + (1 - \delta)w(\Phi_{\text{LS}} - \Phi_{\text{HS}}) - \delta u(\Phi_{\text{HS}} - \Phi_{\text{HD}}) + \delta S_{\text{HS}} \quad (13)$$

$$D\delta \frac{d\Phi_{\text{HD}}}{dt} = (1 - \delta)w(\Phi_{\text{HS}} - \Phi_{\text{HD}}) + \delta u(\Phi_{\text{HS}} - \Phi_{\text{HD}}) + (1 - \delta)qD \times [\langle \Phi_{\text{I}}(z) \rangle - \Phi_{\text{HD}}] + \delta S_{\text{HD}}. \quad (14)$$

Table 2 shows detailed explanations of each model parameter, and detailed explanations of the model and its parameters are given in Shaffer and Sarmiento (1995). The boundary conditions are as follows:

$$\Phi_{\text{I}}(0) = \Phi_{\text{HD}} \quad (Z = 0) \quad (15)$$

$$k \frac{d\Phi_{\text{I}}(z)}{dz} - w[\Phi_{\text{I}}(z) - \Phi_{\text{HD}}] + S_{\text{B}} = 0 \quad (Z = -D). \quad (16)$$

In the case of ΣCO_2 , we set the source/sink terms as $S_{\text{I}}(z) = 0$, $S_{\text{HD}} = 0$ and $S_{\text{B}} = 0$. The surface CO₂ exchange was modeled based on Fujii et al. (2000), and was added to the source/sink term (S_{LS} and S_{HS}) for the ocean surface at low and high latitudes, instead of removing the g_{LS} and g_{HS} terms. CO₂ piston velocity and alkalinity of low and high latitude ocean were set to 3.04×10^{-5} m s⁻¹, 2360 $\mu\text{eq kg}^{-1}$ and 2290 $\mu\text{eq kg}^{-1}$ (Fujii, M. personal communication), respectively. Solubility (Weiss, 1974) and equilibrium constants of CO₂ dissolution (Goyet and Poisson, 1989) were calculated as a function of temperature, assuming that salinity is 35‰. We assumed that surface temperatures in low and high latitude ocean are $T_{\text{S}} + 7$ K and $T_{\text{S}} - 13$ K, respectively, where T_{S} is the Earth's surface temperature calculated by the AEHC model.

2.2.3. Atmosphere carbon balance. The carbon balance equation in the atmosphere was established by considering the anthropogenic industrial carbon emissions (e.g. fossil fuel combustion and cement production), and it becomes

$$\frac{dC_{\text{atm}}(t)}{dt} = -\text{GPP}(t) + \text{RES}(t) + \text{DEC}(t) + \text{LU}(t) + \text{FOS}(t) - \text{ASL}(t) - \text{ASH}(t), \quad (17)$$

where $C_{\text{atm}}(t)$, FOS(t), ASL(t) and ASH(t) are carbon mass in atmosphere, industrial carbon emissions, air to

ocean surface flux in low latitude areas and air to ocean surface flux in high latitude areas. $GPP(t)$, $RES(t)$ and $DEC(t)$ are determined from the terrestrial carbon cycle model, and $ASL(t)$ and $ASH(t)$ are from the ocean carbon cycle model.

3. Historical anthropogenic forcing and model calibration

3.1. Anthropogenic effects

We assumed that the global carbon cycle was at steady state in the pre-industrial era, before 1750. The model was forced with anthropogenic CO_2 emissions due to industrial activities such as fossil fuel combustion and cement production (Marland et al., 2000), land use change (Houghton and Hackler, 2001) and vegetation area variation. Emissions due to land use change between 1750 and 1850 were assumed to increase linearly from zero in 1750, since Houghton and Hackler did not include data before 1850. Historical anthropogenic CO_2 emissions are shown in Fig. 2(a).

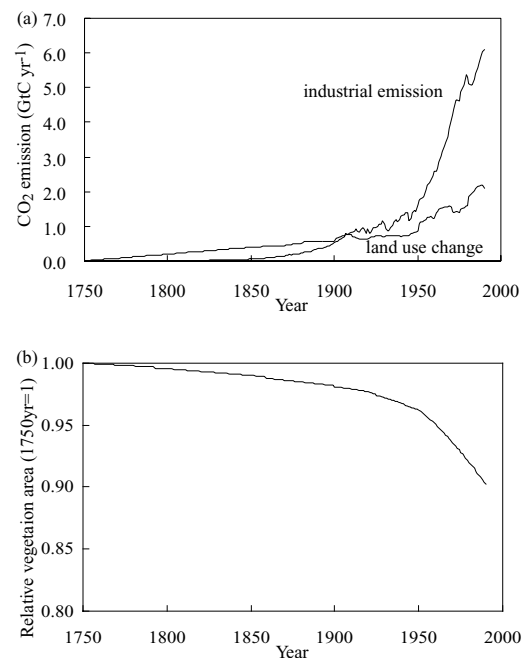


Fig. 2. Historical changes in (a) anthropogenic CO_2 emission from 1750 to 1990 owing to industrial activities such as fossil fuel combustion (thick line) (Marland et al., 1999) and land use change (thin line) (Houghton and Hackler, 2001), and (b) vegetation area relative to pre-industrial era (1750).

Time variation in vegetation area was estimated based on the world population variation (United Nations, 1999) by assuming that a decrease in vegetation area is linearly related to the increase in world population. Based on the world total forest area of $8.7772 \times 10^{13} \text{ m}^2$ in 1750 and $6.2482 \times 10^{13} \text{ m}^2$ in 1990 (Houghton and Hackler, 2001) and a world population of 0.79 billion in 1750 and 5.27 billion in 1990 (United Nations, 1999), vegetation area variation relative to 1750 [$S_r(t)$] becomes

$$S_r(t) = 1 - V_{\text{dec}} \frac{0.24}{4.48} [\text{Pop}(t) - \text{Pop}(1750)], \quad (18)$$

where $\text{Pop}(t)$ is world population at year t . The coefficient 0.0536 was obtained from total forest area and world population in 1750 and 1990. In eq. (18), we introduced a new parameter of V_{dec} (actual vegetation decrease parameter), because Houghton and Hackler's estimation did not consider all vegetation areas, only forested areas. Many forested areas were not only converted to non-vegetation areas, but also to vegetation areas such as agricultural land and pastures. We used $V_{\text{dec}} = 0.4$ based on the assumption of an actual vegetation area decrease of 10% in the reference simulation. Consequently, time variation in vegetation area is shown in Fig. 2(b).

Lastly, we assumed that the CO_2 fertilization factor is the most unknown parameter, and it was adjusted to minimize the root mean square error between observed and simulated atmospheric CO_2 from 1750 to 1990 by fixing other parameters.

3.2. Model calibration and validation

The CO_2 fertilization factor (β) was determined through iterative analysis to reproduce past atmospheric CO_2 concentrations (1750–1964, Etheridge et al., 1998; and 1965–1990, Keeling and Whorf, 1999) and was used to calibrate model parameters, because it was impossible to determine β from existing data. The obtained β value is 0.16, which is smaller than the experimental value of 0.35 (Harrison et al., 1993). However, we consider that this value is acceptable, because the real β value should be smaller than the experimental value because nutrient and water limitation reduces the fertilization effect.

The past trend of atmospheric CO_2 was generally well reproduced by our model, as shown in Fig. 3(a). However, there is a tendency to underestimate CO_2 during the period from 1860 to 1960, with the maximum deviation of -13 ppmv in 1940. The root mean

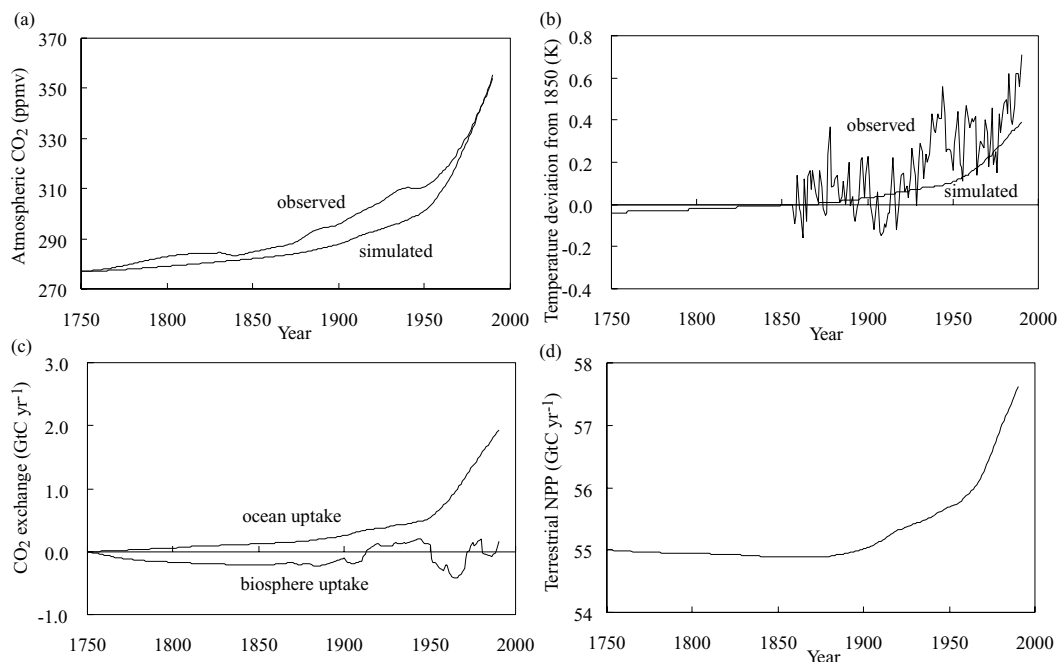


Fig. 3. Simulated and observed time variation of carbon cycle and climate. (a) Simulated (thick line) and observed (thin line) variation of atmospheric CO_2 (ppmv). (b) Simulated (thick line) and observed (thin line) variation of land temperature. (c) Simulated net biospheric (thick line) and ocean carbon uptake (thin line). (d) Simulated terrestrial net primary production.

square error of CO_2 concentration from 1750 to 1990 was approximately 6 ppmv. The tendency of underestimation in these years was also found in the carbon cycle model of Lenton (2000). Time variation of simulated temperature showed a slight small increase trend as shown in Fig. 3(b), and its general tendency is coincident with the observed one (Jones et al., 2000). Again, the simulated temperature is lower than the observed one, and the deviation is as much as 0.2 K.

Simulated carbon budgets during the 1980s, shown in Fig. 3(c), were compared with the recent IPCC estimation based on atmospheric O_2 and CO_2 observation. The simulated oceanic carbon uptake of 1.7 GtC yr^{-1} and biospheric carbon uptake of 0.0 GtC yr^{-1} are near the central value of the IPCC estimation (Prentice et al., 2001). In addition, we can compare these values with the trend of terrestrial NPP [Fig. 3(d)] obtained from remotely sensed observations. The simulated NPP trend shows a 1.2% increase in the 1980s, close to the satellite-based estimations of 1.8–4.4% 10 yr^{-1} increase during the 1980s (Ichii et al., 2001).

Lastly, CO_2 uptake projected by the terrestrial and ocean carbon cycle model in the 4-SCC model was compared with the output of the other process-based

models. The 4-SCC model was forced with the projected atmospheric CO_2 concentrations and temperature values derived from the IS92a emissions scenario, as it was in Prentice et al. (2001). We have tested two cases, one under the projected CO_2 and constant climate, and the other under the projected CO_2 and variable climate. The ocean CO_2 uptake ability is 5.3 GtC yr^{-1} (constant climate) and 4.1 GtC yr^{-1} (variable climate), near the central values of the other ocean carbon cycle model studies (Orr et al., 2000). The terrestrial model response shows that the CO_2 uptakes of 7.9 GtC yr^{-1} (constant climate) and 2.4 GtC yr^{-1} (variable climate) in 2100 are within the possible range provided by the model intercomparison studies (Cramer et al., 2001).

4. Future carbon cycle and climate simulation

4.1. Future projection in the reference simulation

The IPCC IS92-a “business as usual” scenario (Leggett et al., 1992) was used as an anthropogenic

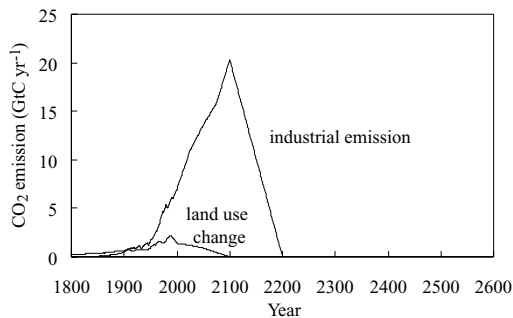


Fig. 4. Future carbon emission scenario used in the simulation. IPCC IS92-a "business as usual" scenarios were applied for 1990–2100 with the linear decline to zero at 2200.

forcing for the years 1990–2100. Beyond 2100, industrial CO₂ emissions were assumed to decline linearly and reach 0 in 2200, following Houghton et al. (1996) and Lenton (2000), as shown in Fig. 4.

Figure 5 and Table 3 show the projected carbon cycle and climate based on the reference simulation.

The projected atmospheric CO₂ becomes 645 ppmv in 2100 and reaches a peak of 760 ppmv in approximately 2170. Then, it declines and stabilizes at 600 ppmv. The temperature becomes 1.5 K higher in 2100 than at the pre-industrial level, and the peak level becomes 1.8 K higher than the pre-industrial level. The net biospheric carbon uptake (NPP – soil decomposition – land use change) becomes positive at the end of the 20th century, and reaches a peak of 6.6 GtC yr⁻¹ in 2100; it then decreases to become negative in 2200. The net oceanic uptake reaches the maximum 4.7 GtC yr⁻¹ in 2110, and then decreases.

The climate feedback effect was evaluated in the reference simulation. Figure 5 and Table 3 show a comparison of the two simulation results, with and without the climate feedback effect. The climate feedback effect increased the atmospheric CO₂ throughout the simulation, and decreased the ocean and biosphere CO₂ uptake. The general mechanism of the climate feedback effect on the carbon cycle can be characterized by decreases in (1) net terrestrial uptake due to an

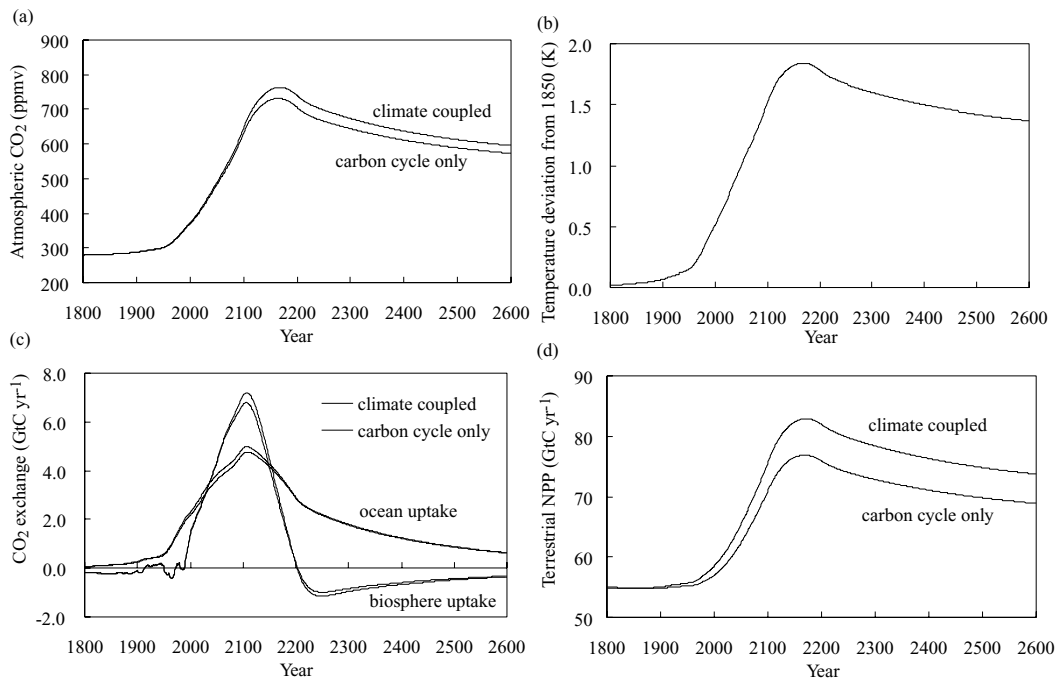


Fig. 5. Simulated future carbon cycle and climate variation forced with prescribed anthropogenic emission scenario. (a) Future atmospheric CO₂ variation based on carbon cycle and climate coupled cycle model (thick line) and carbon cycle and constant climate model (thin line). (b) Future variation of land temperature. (c) Future variation of net biosphere and ocean carbon uptake based on carbon cycle and climate coupled cycle model (thick line) and carbon cycle and constant climate model (thin line). (d) Future variation of terrestrial net primary production based on carbon cycle and climate coupled cycle model (thick line) and carbon cycle and constant climate model (thin line).

Table 3. *Test cases and results of sensitivity studies*

Test case	ΔT (K)	V_{dec}	F_{gpp} (GtC yr ⁻¹)	Q_{10}	β	L_u (GtC yr ⁻¹)	Ocean uptake (GtC yr ⁻¹)	CO ₂ in 2100 (ppmv)	Max CO ₂ (year) (ppmv)	Bio uptake (GtC yr ⁻¹)	NPP trend (% 10 yr ⁻¹)
Reference	1.3	0.4	350	2.0	0.16	2.0	1.7	645	760 (2170)	0.0	1.2
Const. climate	0.0	0.4	350	1.0	0.14	2.0	1.8	630	730 (2160)	0.0	0.7
AEHC T response	1.5	0.4	350	2.0	0.17	2.0	1.7	639	756 (2170)	0.1	1.3
	2.5	0.4	350	2.0	0.19	2.0	1.6	647	778 (2170)	0.2	0.7
	4.5	0.4	350	2.0	0.24	2.0	1.3	654	813 (2180)	0.4	2.7
Vegetation area	1.3	0.0	350	2.0	0.08	2.0	1.7	661	803 (2180)	0.0	1.1
	1.3	0.8	350	2.0	0.26	2.0	1.7	659	781 (2170)	0.0	1.0
Light saturation	1.3	0.4	110	2.0	0.58	2.0	1.7	705	879 (2180)	0.0	1.1
	1.3	0.4	1050	2.0	0.10	2.0	1.7	556	588 (2130)	0.1	1.3
Q ₁₀ factor	1.3	0.4	350	1.0	0.14	2.0	1.7	638	745 (2170)	0.1	0.8
	1.3	0.4	350	3.0	0.18	2.0	1.7	640	761 (2170)	0.1	1.5
Land use change	1.3	0.4	350	2.0	0.11	0.6	1.7	720	874 (2170)	-0.2	0.6
	1.3	0.4	350	2.0	0.18	2.5	1.7	621	728 (2170)	0.1	1.4
Ocean C uptake	1.3	0.4	350	2.0	0.18	2.0	1.3	644	769 (2170)	-0.8	1.5
	1.3	0.4	350	2.0	0.13	2.0	2.5	627	721 (2160)	0.5	0.6

increase in soil decomposition, and (2) net ocean uptake due to ocean surface warming. On the other hand, the terrestrial NPP increase was due to an increase in light availability and CO₂ fertilization as shown in Fig. 5(d). The small impact of climate feedback on atmospheric CO₂ concentrations is due to the balance of net effects described above. Atmospheric CO₂ levels become 630 ppmv in 2100, and reach a peak of 730 ppmv in 2160. As shown below, climate feedback has a smaller effect than the other, unknown parameters such as the light saturation effect and land use change-effected CO₂ emissions.

4.2. Sensitivity analysis

We tested sensitivity of the modeled projection based on the future emission scenario of an extended IS92a scenario. In each simulation, we have tuned the CO₂ fertilization factor (β) to minimize the root mean square error of observed and simulated CO₂ concentrations from 1750 to 1990. The test cases analyzed the sensitivities of (1) temperature response in the AEHC model, (2) light saturation factor (F_{gpp}), (3) vegetation area decrease factor (V_{dec}), (4) temperature dependency of soil decomposition (Q_{10}), (5) uncertainty in emissions due to land use changes (L) and (6) uncertainty in the ocean uptake ability, as shown in Table 3.

We did not test the model structure uncertainties of the biosphere and ocean. Biosphere model (subdivid-

ing soil reservoir) sensitivities were tested by Lenton (2000), who showed that there is little effect on atmospheric CO₂ concentration, but a large effect on the soil reservoir. Ocean model sensitivity, for example the ocean biospheric effect and changes in ocean circulation, must be tested. Although simplified, these models have not been well established yet, and these effects must be included in future studies.

4.2.1. Temperature responses in the AEHC model. When CO₂ concentrations were doubled from 350 ppmv, the temperature response in our AEHC model was an 1.3 K increase, which is smaller than the temperature change of 1.5–4.5 K at the doubled CO₂ concentration predicted by the general circulation model (GCM) studies (Houghton et al., 1996). We assume that the upper and lower limits of temperature response to doubled CO₂ are 1.5 and 4.5 K. We have tested three cases of temperature responses to doubled CO₂, i.e. 1.5, 2.5 and 4.5 K, by amplifying AEHC model responses. The simulated atmospheric CO₂ varies from 639 ppmv (the 1.5 K case) to 654 ppmv (the 4.5 K case) in 2100 and from 756 ppmv (the 1.5 K case) to 813 ppmv (the 4.5 K case) in the year of peak CO₂ [Fig. 6(a) and Table 3]. The sensitivity of temperature responses in the atmospheric model is weak compared with that of other parameters shown in Table 3.

4.2.2. Light saturation factor (F_{gpp}). Although we used the prescribed value in the estimation of F_{gpp} , the value of F_{gpp} has a large uncertainty because of

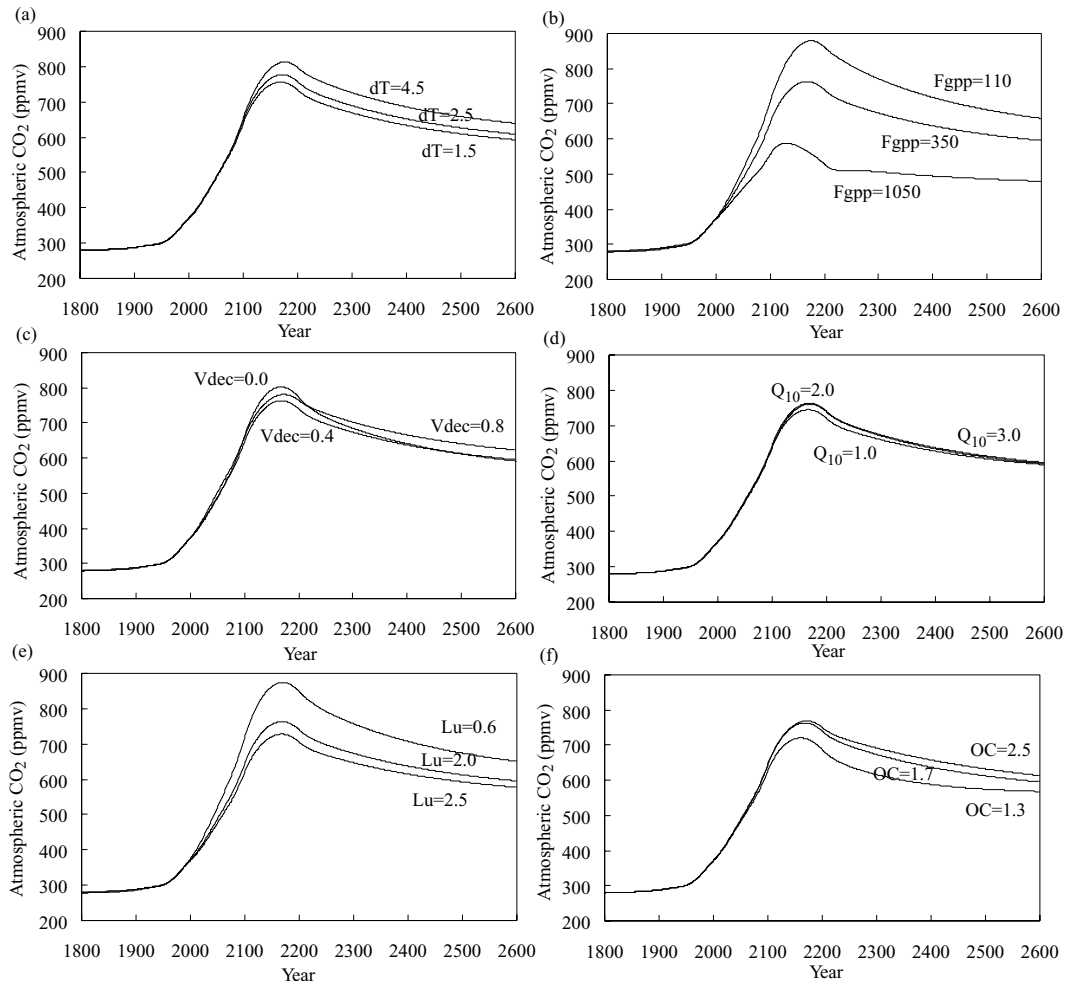


Fig. 6. Results of sensitivity analysis. (a) Climate model sensitivity to doubled CO₂ (350 to 750 ppmv), (b) vegetation growth limitation factor (F_{gpp}), (c) actual vegetation area variation (V_{dec}), (d) Q_{10} factor, (e) land use emission (L) and (f) ocean uptake capability in the 1980s.

the difficulty in estimation of globally averaged light utilization efficiencies, and total vegetation areas in the pre-industrial era. We tested two extreme cases. One is $F_{gpp} = 110 \text{ GtC yr}^{-1}$, which is a minimum value of F_{gpp} and makes the GPP equation equal to that of past simple carbon cycle models (e.g. Craig and Holmén, 1995; Wigley, 1993) as discussed later. The other is with a F_{gpp} value that is three times larger ($F_{gpp} = 1050 \text{ GtC yr}^{-1}$), because we cannot determine an optimal maximum F_{gpp} value. Although it is desirable to test the sensitivity by using a larger F_{gpp} to evaluate the model response of the Bacastow and Keeling (1973)

type (section 4.3.2), $F_{gpp} = 1050 \text{ GtC yr}^{-1}$ should be sufficient to test the sensitivity as described below.

The simulated atmospheric CO₂ varies from 556 ppmv ($F_{gpp} = 1050 \text{ GtC yr}^{-1}$) to 705 ppmv ($F_{gpp} = 110 \text{ GtC yr}^{-1}$) in 2100, and from 588 ppmv ($F_{gpp} = 1050 \text{ GtC yr}^{-1}$) to 879 ppmv ($F_{gpp} = 110 \text{ GtC yr}^{-1}$) in the year of peak CO₂ [Fig. 6(b) and Table 3]. These results show that the sensitivity of this parameter is the highest of all parameters. The uncertainties derived from F_{gpp} are almost same as the ISAM model (Prentice et al., 2001), and the parameter must be constrained primarily.

4.2.3. Vegetation area decrease factor (V_{dec}). We used the forest area at pre-industrial and present times (Houghton and Hackler, 2001) to estimate the vegetation area variations. However, the conversion of forested area to cultivated land does not always decrease the vegetation area. Therefore, the parameter, V_{dec} , was introduced to convert the forested area variation to vegetation area variation. Since a suitable data set of vegetation area variation is not available, we have set the uncertainty of V_{dec} as 0.0–0.8. The value of 0.0 indicates that there has not been vegetation area variation since the pre-industrial era, and that of 0.8 means that 80% of the deforested area was converted to a non-vegetation area.

The simulated atmospheric CO_2 varies from 659 ppmv ($V_{dec} = 0.8$) to 661 ppmv ($V_{dec} = 0.0$) in 2100, and from 781 ppmv ($V_{dec} = 0.8$) to 803 ppmv ($V_{dec} = 0.4$) in the year of peak CO_2 [Fig. 6(c) and Table 3]. These values are higher than those of the reference simulation are, but the overall sensitivity of this parameter is low. We have tested other V_{dec} values in the simulation, and found that a V_{dec} value of approximately 0.4 resulted in the lowest CO_2 concentration. A lower V_{dec} value makes GPP increase due to the vegetation area decrease, but decrease due to a tuned fertilization factor (lower β value). The GPP is determined by a balance of positive and negative effects, and $V_{dec} = 0.4$ results in the highest GPP.

4.2.4. Soil decomposition (Q_{10}). This parameter also has large uncertainty, therefore we tested the value of $Q_{10} = 1.0$ and $Q_{10} = 3.0$. The simulated atmospheric CO_2 varies from 638 ppmv ($Q_{10} = 1.0$) to 640 ppmv ($Q_{10} = 3.0$) in 2100 and from 745 ppmv ($Q_{10} = 1.0$) to 761 ppmv ($Q_{10} = 3.0$ case) in the year of peak CO_2 [Fig. 6(d) and Table 3]. Sensitivity of this parameter is also low.

4.2.5. Land use change emissions (L). Following the IPCC's estimation of land use change of 0.6–2.5 $GtC\ yr^{-1}$ in the 1980s (Prentice et al., 2001), we tested the sensitivity of the land use change emissions of 0.6 and 2.5 $GtC\ yr^{-1}$ in the 1980s. Historical land use change emissions were multiplied by appropriate constant factors, and tuned to have the above values in the 1980s. The simulated atmospheric CO_2 varies from 621 ppmv ($L = 2.5\ GtC\ yr^{-1}$) to 720 ppmv ($L = 0.6\ GtC\ yr^{-1}$) in 2100 and from 728 ppmv ($L = 2.5\ GtC\ yr^{-1}$) to 824 ppmv ($L = 0.6\ GtC\ yr^{-1}$) in the year of peak CO_2 [Fig. 6(e) and Table 3]. The sensitivity of this parameter is high.

4.2.6. Ocean uptake ability. Following the IPCC's ocean uptake estimation of $1.9 \pm 0.6\ GtC$

yr^{-1} during the 1980s (Prentice et al., 2001), we tested an ocean uptake sensitivity of 1.3 (minimum ocean uptake) and 2.5 $GtC\ yr^{-1}$ (maximum ocean uptake). The simulated atmospheric CO_2 varies from 627 ppmv (2.5 $GtC\ yr^{-1}$) to 644 ppmv (1.3 $GtC\ yr^{-1}$) in 2100 and from 721 ppmv (2.5 $GtC\ yr^{-1}$) to 769 ppmv (1.3 $GtC\ yr^{-1}$) in the year of peak CO_2 [Fig. 6(f) and Table 3]. The sensitivity of this parameter is low.

4.3. Discussion

The light saturation factor (F_{gpp}) and land use emissions produced large uncertainties in projections of future carbon cycle and climate. Therefore, the following section focuses on the uncertainty in these models and model parameters, and on an approach for decreasing the uncertainties in the model projections.

4.3.1. Model constraints determined by a satellite-based NPP trend. The sensitive parameters must be determined more precisely for the projection of carbon cycle and climate. One of the ways to constrain the range of the uncertainty, especially in the land use emission, is to use a satellite-based NPP trend. If land use emissions in the 1980s are assumed to have been 0.6 $GtC\ yr^{-1}$, the NPP increase can be estimated as 0.6% $10\ yr^{-1}$ during the 1980s (Table 3). Ichii et al. (2001) estimated the NPP increase to be 1.8–4.4% $10\ yr^{-1}$ during the 1980s based on satellite data. However, the satellite-based NPP trend seems to have a large uncertainty due to its short time coverage, large calibration uncertainty and orbital drift effects (e.g. Malmström et al., 1997). We assume that the modeled NPP trend should be more than one-third of the 3% $10\ yr^{-1}$ increase in global NPP, which is approximately the central value estimated by Ichii et al. (2001). A 1% $10\ yr^{-1}$ increase in global NPP corresponds with the land use change emissions of 1.2 $GtC\ yr^{-1}$ during the 1980s. Simulated atmospheric CO_2 concentrations in 2100 are 672 ppmv. Based on the satellite-derived NPP trend, we could show the lower uncertainty in land use CO_2 emission values and resulting future atmospheric CO_2 concentrations with higher confidence. The uncertainty in the light saturation parameter, unfortunately, could not be constrained by the satellite-based NPP trend.

4.3.2. Comparison with other carbon cycle model studies. The previous similar studies of carbon cycle models (Enting et al., 1994; Schimel et al., 1995;

Houghton et al., 1996; Prentice et al., 2001) showed higher estimations of future atmospheric CO₂ concentrations. The model intercomparison study showed that year 2100 projected CO₂ concentrations (without climate feedback) range from 668 to 734 ppmv (Schimel et al., 1996). With land use emissions ranging from 0.4 to 1.8 GtC yr⁻¹, Wigley (1997) estimated the uncertain range of 667 to 766 ppmv in 2100. The third assessment report of IPCC showed that the reference simulations of future carbon cycles reach 723 ppmv (ISAM model), based on the carbon and climate coupled model, and 682 ppmv (ISAM model), based on a carbon cycle model without climate feedback (Prentice et al., 2001). There are few simulation results beyond 2100. Lenton (2000) and Houghton et al. (1996) estimated peak CO₂ concentrations of 985 and 850 ppmv, respectively, based on the same emission scenario used in this study.

The simulated future CO₂ concentrations in 2100 and beyond 2100, presented in the current study, did not reach such high CO₂ levels as shown in Table 3. Although the highest simulation results (720 ppmv for land use change = 0.6 GtC yr⁻¹, and 705 ppmv for $F_{\text{gpp}} = 110 \text{ GtC yr}^{-1}$) are almost coincident with the referenced simulation in the ISAM models, the present study's reference simulation of CO₂ concentration is not within the range of the other studies. Moreover, the reference simulation in the current study does not fall within the range 668–734 ppmv, based on the constant climate simulation (Schimel et al., 1996), although climate feedback works as a positive feedback process.

The major difference between our model and the previous models exists in the GPP models as shown in eq. (7). We can simplify our GPP models and categorize them into the following three types:

Type 1: Constant $S(t)$, and constant $C_{\text{veg}}(t)/S(t)$ and/or large k to approximate $\{1 - \exp[-kC_{\text{veg}}(t)/S(t)]\} \simeq 1.0$, which means constant vegetation area and already saturated vegetation or constant vegetation density:

$$GPP = k_1 \{1 + \beta \ln[C_{\text{atm}}(t)/C_{\text{atm}}(0)]\}. \quad (19)$$

Type 2: $k_{\text{ext}}k_{\text{lai}} [C_{\text{veg}}(t)/S(t)] \simeq 0$ and the first-order terms in the Taylor expansion of $\{1 - \exp[-k_{\text{ext}}k_{\text{lai}} C_{\text{veg}}(t)/S(t)]\}$, which means no vegetation growth limitation due to light availability:

$$GPP = k_2 C_{\text{veg}}(t) \{1 + \beta \ln[C_{\text{atm}}(t)/C_{\text{atm}}(0)]\}. \quad (20)$$

Type 3: Constant $C_{\text{veg}}(t)/S(t)$ and/or large k to approximate $\{1 - \exp[-kC_{\text{veg}}(t)/S(t)]\} \simeq 1.0$, which

means variable vegetation area and already saturated vegetation production in terms of light availability:

$$GPP = k_3 S(t) \{1 + \beta \ln[C_{\text{atm}}(t)/C_{\text{atm}}(0)]\}. \quad (21)$$

Equation (19) is the most commonly used form (e.g. Wigley 1993; Craig and Holmén, 1995). Equation (20) was used in the pioneer carbon cycle models (e.g. Bacastow and Keeling, 1973). Equation (21) is similar to Lenton (2000), although they use different CO₂ fertilization formulas and vegetation area variations.

Although the vegetation production model in ISAM incorporates vegetation growth limitation based on the logistic function (Jain et al., 1995; Khesghi et al., 1996), the results are coincident with those of other carbon cycle models, and our highest projection. Therefore, ISAM model closely resembles our model when $F_{\text{gpp}} = 110 \text{ GtC yr}^{-1}$ (vegetation is already saturated in terms of light availability).

Our analysis showed that the choice of the GPP model (choice of F_{gpp} value) is one of the simplified carbon cycle model's most sensitive factors. Therefore, we have to quantify the light saturation effect of vegetation. However, we conclude that our model is more accurate, because the model includes the light saturation effect on vegetation growth, one of the most important model components for the projection of the future global carbon cycle and climate.

5. Conclusion

A simple climate–carbon cycle interactive model, the Four-Spheres Cycle of Energy and Mass (4-SCEM) model, was developed to project future variations in the global carbon cycle and climate. The following processes were included and compared with those of other simple carbon cycle model studies: (1) the vegetation light saturation effect and (2) vegetation area variation. The model was constrained with past variations in atmospheric CO₂, and could simulate past atmospheric CO₂ and biosphere and ocean carbon uptake during the 1980s. Moreover, the biosphere and ocean carbon cycle model showed similar responses through the year 2100 using a process-based biosphere and an ocean carbon cycle model. The satellite-based NPP trend in the 1980s was used to constrain the CO₂ emissions due to land use change, and showed the implications of decreasing the uncertainty of its estimation.

The reference simulation showed that the CO₂ concentration becomes 645 ppmv in 2100 and reaches a peak of 760 ppmv in 2170. These simulated concentrations were much lower than those of previous carbon cycle simulations. Our model included the past terrestrial production model (Bacastow and Keeling, 1973; Wigley, 1993; Jain et al., 1995; Kheshgi et al., 1996) as a special case, and the difference between CO₂ concentrations was primarily derived from the use of different GPP models. Since our estimated vegetation light saturation parameter has a large uncertainty, we have to quantify and constrain the vegetation light saturation effect more accurately.

Sensitivity analysis successfully quantified the relative importance of estimating the poorly constrained parameters and processes. The climate-carbon cycle feedback effects, which had very large positive feedback effects in Cox et al. (2000), showed a relatively low sensitivity in our model. On the other

hand, uncertainties in CO₂ emissions due to land use change and the light saturation effect resulted in high sensitivities.

As found in the current study, the future projection of carbon cycle and climate still has a large uncertainty derived from model structure and parameters. Further constraints of the aforementioned parameters and processes are indispensable to improve the confidence with which the future carbon cycle and climate are projected.

6. Acknowledgement

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