Climate Change, Allowable Emission, and Earth System Response to Representative Concentration Pathway Scenarios

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Abstract

To understand the mechanisms of long-term climate and carbon cycle feedback with anthropogenic impact, past simulations (1850–2005) and projection experiments (2006 to 2100) were conducted using a new Earth system model named “MIROC-ESM”, forced by four representative concentration pathway (RCP) scenarios that describe how greenhouse gases (GHGs), aerosols, and land-use will develop in the future. From these projections, temperature rise from 1850 to 2100 ranged from 2.4 K in the RCP2.6 scenario to 6.2 K in the RCP8.5 scenario. We found that there are discrepancies between the RCPs and the estimates of our model in both allowable fossil fuel and land-use change emissions. The former showed systematic discrepancies likely due to strong positive feedbacks in the model, but the latter did not. The likely reason for the difference in land-use emissions is the modeling of land-use change processes or definitions for the emission. Climate response to the increase of atmospheric carbon showed large variation among scenarios, strongly affected by ocean heat uptake efficiency that could depend on the rate of atmospheric CO₂ increase in each scenario. Large variation between scenarios was also found in carbon cycle sensitivity measured by cumulative airborne fraction. The variation in carbon cycle sensitivity may be attributable to the dependence of concentration-carbon feedback on the rate of atmospheric CO₂ increase. The earth system would show a similar response to emitted carbon during the 21st century if the difference of ocean heat uptake efficiency between scenarios were small. The earth system responds to RCP6.0 with less sensitivity to emitted carbon when compared with other scenarios because of high-efficiency carbon uptakes by land and ocean ecosystems. In contrast, RCP2.6 showed high sensitivity of the earth system to carbon emission, and apparently showed different behavior from other scenarios due to early reduction of GHGs.

1. Introduction

Earth system models (ESMs) that provide full representation of climate and carbon cycle processes have been developed to understand mechanisms of the whole earth system (Cox et al. 2000; Friedlingstein et al. 2006). These models are advances over coupled atmosphere-ocean general circulation models, including biogeochemical processes in both land and ocean ecosystems to represent the carbon cycle feedback on climate. Especially for long-term future projections under anthropogenic impact, carbon cycle feedback significantly affects future climate change, with a magnitude comparable to the sum of other climate feedbacks (Gregory et al. 2009).

To conduct climate change projections with these ESMs, greenhouse gas (GHG) emission scenarios from the Special Report on Emissions Scenarios (SRES) were used for the simulations in the 4th Assessment Report of the Intergovernmental Panel on Climate Change (Denman et al. 2007). Instead of SRES scenarios, the IPCC 5th Assessment Report will present the results of future projection experiments based on scenarios called “representative concentration pathways” (RCPs). These RCPs are defined by a given pathway of radiative forcing toward a stabilized future level, and completed by integrated assessment models (IAMs) that simulate an internally-consistent set of socioeconomic variables (including GHG and aerosol emissions, and land-use) (Moss et al. 2010).

Using these scenarios, numerous experiments were conducted with climate models, showing a large variety of future climate (Denman et al. 2007). To assess uncertainties in global climate models, several model intercomparison projects (MIPs) have been conducted (e.g., CMIP3: Coupled Model Intercomparison Project phase-3; C4MIP: Coupled Climate–Carbon Cycle Model Intercomparison Project by Friedlingstein et al. 2006). These projects analyzed model outputs from leading modeling centers around the world. The MIPs for ESMs have adopted the experimental setting that the ESM is driven by emissions. This setting is important for our understanding because it captures actual carbon cycle processes. Carbon (CO_2) is emitted and a

portion of the atmospheric carbon is absorbed by terrestrial and marine ecosystems, the remaining then determines the atmospheric CO_2 concentration and results in radiative forcing. Instead of emission-driven experiments, the Coupled Model Intercomparison Project phase-5 (CMIP5; Taylor et al. 2010) recommends experiments with prescribed CO_2 concentration because of the simplicity in evaluating climate and carbon cycle feedback independently. As CO_2 concentrations are prescribed in concentration-driven experiments, atmosphere-ocean and atmosphere-land CO_2 fluxes are simply diagnosed by marine and terrestrial ecosystem models. From such experiments, allowable fossil fuel emissions can be estimated by inverse analysis from the diagnosed carbon fluxes and prescribed CO_2 concentrations.

To investigate the mechanism of the earth system and evaluate uncertainty in climate models and their projections, several indices have been devised that can evaluate the earth system composed of climate and biogeochemistry. Gregory et al. (2009) proposed a uniform treatment based on global energy and carbon balance equations for evaluating all climate feedback, including that between climate and carbon cycle. Matthews et al. (2009) proposed a simple and robust metric for ESMs by measuring the net response of the earth system to anthropogenic CO_2 emissions. The measurement for evaluating carbon cycle feedback can be summarized into two terms: concentration-carbon and climate-carbon feedback. The former feedback depends on how terrestrial and marine ecosystems absorb atmospheric CO_2 in the absence of climate change. Because the increase of atmospheric CO_2 could stimulate the “fertilization” effect of ecosystems, this feedback is now considered to be negative (i.e., it acts on the earth system to reduce global warming). The latter feedback focuses on the effect of climate change (usually measured by global mean temperature change) on carbon cycles, which usually leads to carbon release from carbon reservoirs in terrestrial and marine ecosystems by enhancing ecosystem respiration. This feedback is now considered to be positive (Friedlingstein et al. 2006; Stich et al. 2008), and its magnitude indicates how the carbon cycle accelerates global warming. Both feedbacks compose the entire carbon cycle feedback on climate, which should be measured by the airborne fraction. Since this fraction indicates the ratio of emitted carbon remaining in the atmosphere, which is equivalent to the residue of emitted carbon not absorbed by land and ocean, it shows the capacity of carbon uptake by terrestrial and marine ecosystems under the influence of climate change (Gregory et al. 2009; Matthews et al. 2009).

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Although these methods for evaluating climate and carbon cycle feedback have been proposed, the analysis for the earth system response to RCP scenarios is still uncertain. This is because the response of the earth system and its feedback could be affected by non-CO₂ GHGs, aerosols, and land-use change in addition to anthropogenic CO₂ emission. Furthermore, the level of CO₂ concentration in the end of the 21st century and its trajectory during that century vary substantially between RCPs. RCP6.0 and RCP8.5 show monotonically increasing values, while RCP2.6 and RCP4.5 show a peak concentration or stabilized concentration, respectively. Thus, this variability in anthropogenic forcing may cause complex behavior of the earth system.

In this study, we investigated the responses of climate, carbon cycle, and earth system to the new-scenario RCP. To this end, we conducted a past simulation (1850–2005) and future projections (2006–2100) by applying RCPs to a newly-developed ESM. Result characteristics with RCP scenarios were analyzed, focusing mainly on projected surface air temperature and allowable emissions. Climate change was simulated based on concentrations of GHGs described in RCPs, and allowable CO₂ emissions were estimated inversely using biogeochemical responses of the land and ocean. To investigate earth system response to anthropogenic impact, we applied the metrics of Matthews et al. (2009) to the model results, considering the difference of anthropogenic forcing among RCPs.

2. Model and experiments

2.1 Earth system model

We used the Model for Interdisciplinary Research on Climate ESM (MIROC-ESM), which is based on the model previously referred to in C4MIP as the Frontier Research Center for Global Change model (Friedlingstein et al. 2006; Yoshikawa et al. 2008). MIROC-ESM consists of atmospheric, ocean, and terrestrial components, and includes both marine and terrestrial biogeochemical processes. The atmospheric model (MIROC-AGCM; Watanabe et al. 2008) has T42 spatial resolution (about 2.8×2.8 degrees) and 80 vertical layers, including the stratosphere and mesosphere. An online aerosol module (Spectral Radiation-Transport Model for Aerosol Species or SPRINTARS; Takemura et al. 2000, 2005) predicts major tropospheric aerosols. Direct, semi-direct, and indirect effects of aerosol are self-consistently calculated in the model. Ocean physical processes, including sea ice dynamics, were simulated by the Center for Climate System Research ocean component model (Hasumi and Emori 2004), which consists of grids of 1.4 degrees in longitude and 0.5–1.7

degrees in latitude and has 44 vertical layers. Ocean biogeochemical processes are described via a nutrient-phytoplankton-zooplankton-detritus model (Kawamiya et al. 2000). Physical and biophysical processes on the land surface are calculated using the “minimal advanced treatment of surface interaction and runoff” (MATSIRO) model (Takata et al. 2003), which consists of a single canopy layer, three snow layers, and six soil layers. MATSIRO mainly treats energy transfer and hydrological processes. Terrestrial biogeochemical processes and the effects of anthropogenic land-use change on carbon cycles are calculated using the “spatially explicit individual-based dynamic global vegetation model” (Sato et al. 2007; Ise et al. 2009). Leaf area index and net land-atmosphere carbon fluxes are obtained dynamically through the process of competition among 13 plant functional types. A detailed description of MIROC-ESM and its ability to reproduce 20th century climate and carbon cycle have been reported by Watanabe et al. (2011).

2.2 Experiments

a. Experiments to quantify carbon cycle feedback with 1% p.a. CO₂ increase

To quantify the carbon cycle feedbacks in our model, three experiments were conducted with 1% p.a. CO₂ increase, as described by Gregory et al. (2009). The first is the “fully coupled” (FULL) experiment, in which both climate and carbon cycle processes respond to the prescribed atmospheric CO₂ concentration. In this experiment, the carbon balance is described as

$$C_E = C_A + \beta C_A + \gamma T, \quad (1)$$

where C_E is the cumulative anthropogenic carbon emission, which is equal to the allowable anthropogenic emission inversely estimated by the prescribed CO₂ concentration and diagnosed carbon flux by terrestrial and marine ecosystems. C_A is carbon increase in the atmosphere from the preindustrial state, and T is the change in global mean surface air temperature from the unperturbed state. β and γ represent the sensitivity of land and ocean carbon storage to C_A and T , respectively. In this equation, βC_A represents the magnitude of concentration-carbon feedback, and γT represents climate-carbon feedback.

The second is the “biogeochemically coupled” (BGC) experiment to quantify the concentration-carbon feedback. In this experiment, terrestrial and marine ecosystems respond to the CO₂ increase, but the radiation process continues to refer to the fixed preindustrial CO₂ level; i.e., no climate change occurs from atmo-

Fig. 1. (a) Past (1850–2005; gray) and future (2006–2100; RCP2.6 in black, RCP4.5 in red, RCP6.0 in green, and RCP8.5 in blue) CO₂ concentrations used in the experiments. (b) Reference radiative forcing of CO₂ (thick lines) and net radiative forcing of other anthropogenic agents (non-CO₂ GHGs CH₄, N₂O, aerosols; thin lines) approximated by Integrated Assessment Models (Meinshausen et al. 2011, available at <http://www.iiasa.ac.at/>).

spheric CO₂ increase. Thus, the carbon balance equation in this experiment should be

$$C_E = C_A + \beta C_A. \quad (2)$$

The third is the “radiatively coupled” (RAD) experiment to quantify the climate-carbon feedback. In this case, only climate processes are affected by atmospheric CO₂ increase; terrestrial and marine ecosystems are unaffected by CO₂ concentration change. Since the carbon cycle responds only to climate change, the carbon balance should be

$$C_E = C_A + \gamma T. \quad (3)$$

From these three experiments, we obtain the sensitivities of our ESM in response to the increase of atmospheric CO₂ concentration.

b. Spinup, historical simulation, and future projections with RCP scenarios

To simulate 20th Century Climate in Coupled Models (20C3M) and make projections with RCP scenarios, spinup runs were done by recursively executing the spinups with the entire ESM and off-line models, establishing initial conditions of the earth system in 1850. The total simulation years for spinups with the ESM, off-line terrestrial ecosystem and marine ecosystem models embedded in the ocean general circulation model were 480, 6350, and 1245 years, respectively. Then, historical simulations for 1850–2005 were made. Finally, four experiments were executed for 95 years (2006 to 2100) under the prescribed CO₂ concentration path-

ways of RCPs 2.6, 4.5, 6.0, and 8.5 (Fig. 1a). The dynamics of other GHGs were given as the model forcing (Fig. 1b), following the CMIP5 protocols (Taylor et al. 2010; datasets for model forcing are available at <http://www.pik-potsdam.de/~mmalte/rcps/>) as well as the treatments of historical changes in solar radiation and volcanic effects. Scenarios for areal changes in land-use types (i.e., primary vegetation, secondary vegetation, pasture, cropland, and urban areas) were also used to reproduce past and future land-use and their emissions. A detailed description of the settings of 20C3M and its initialization in MIROC-ESM are found in Watanabe et al. (2011).

c. Evaluating net land-use change emission using offline terrestrial ecosystem model

To estimate past and future net land-use change emission and evaluate its impacts on the global carbon cycle, supplementary experiments were conducted using the offline terrestrial ecosystem model within MIROC-ESM (see Appendix A for the incorporation of land-use change in our model). Net land-use change CO₂ emission was defined as the difference in carbon emissions resulting from land-use change, such as forest harvesting and carbon uptake by plant regrowth after land-use; for example, that caused by agricultural abandonment. This anthropogenic impact directly alters global carbon partitioning because land-use change induces land carbon decrease (increase) and consequent increase (decrease) of CO₂ emission. From Gregory et al. (2009), the carbon balance should be

$$C_{FF} = C_A + C_L + C_O,$$

where C_{FF} is a cumulative fossil fuel emission, and $C_{L/O}$ is change in land/ocean carbon. An explicit expression for the effect of land-use change on carbon balance is

$$C_{FF} = C_A + (C_L^{noLUC} - C_{LUC}) + C_O. \quad (4)$$

C_{LUC} is a cumulative land-use change emission, and C_L^{noLUC} is change in land carbon without the land-use change effect. C_{LUC} can be calculated by C_L obtained from two experiments:

$$C_{LUC} = -(C_L - C_L^{noLUC}), \quad (5)$$

where C_L is the cumulative net carbon uptake estimated from the experiment in which terrestrial carbon cycles are affected by land-use change as well as climate and atmospheric CO_2 level; C_L^{noLUC} is estimated from the experiment in which land-use status is fixed to the pre-industrial condition. To obtain C_{LUC} , offline simulations using the terrestrial ecosystem model were performed, both with and without land-use change for the historical and each RCP run. Each simulation was forced by MIROC-ESM climate output. For the without-land-use-change runs, a fixed 1850s land-use map was used throughout the simulations.

2.3 Parameters for response of earth system to anthropogenic impact

The metric of Matthews et al. (2009) was applied to the model output to investigate earth system response to anthropogenic impact. They proposed a simple index called carbon-climate response (CCR) for evaluating the response of the earth system to carbon emissions:

$$CCR = \frac{T}{C_E}. \quad (6)$$

The CCR aggregates the response of both climate and carbon cycle systems to emitted carbon.

To separate the earth system response into climate and carbon cycle responses, CCR can be rewritten as

$$CCR = \frac{T}{C_A} \frac{C_A}{C_E} = \alpha A, \quad (7)$$

$$\alpha = \frac{T}{C_A}, \text{ and} \quad (7)$$

$$A = \frac{C_A}{C_E}. \quad (8)$$

α is the temperature change per unit atmospheric carbon increase, and A is the airborne fraction of cumulative

carbon emissions. Because A is the result of carbon cycle response to atmospheric CO_2 and climate changes, it indicates the net effect on the carbon cycle of all climate and carbon feedbacks (Gregory et al. 2009 and Eq. A6).

These equations in Eqs. (6)–(8) were originally formulated to analyze results from experiments with idealized scenarios, such as sensitivity analysis in which only CO_2 induces climate change. Since our 20C3M simulation and projections with RCPs include influences of radiative forcing of non- CO_2 GHGs and aerosols, we devised formulas that enable approximation of earth system response parameters in the absence of these influences and then applied model outcomes to them. These formulas are analytically obtained from physically-based energy and carbon balance equations (see Appendix B for derivation for these formulas).

$$\alpha^C = \frac{\alpha^{ALL}}{1 + \frac{F_{NC}}{\phi C_A}} \quad (9)$$

$$A^C = \frac{A^{ALL}}{1 - \alpha^C \gamma \frac{F_{NC}}{\phi C_E}} \quad (10)$$

$$CCR^C = \frac{CCR^{ALL}}{1 + (1 + \beta) \frac{F_{NC}}{\phi C_E}}. \quad (11)$$

α^{ALL} , A^{ALL} , and CCR^{ALL} are the climate, carbon cycle, and earth system response parameters obtained by simply applying the model outcomes to Eqs. (6)–(8). Thus, these values with “ALL” superscripts reflect the influence of non- CO_2 radiative forcing, in addition to that from CO_2 . F_{NC} is an additional anthropogenic radiative forcing on the atmosphere, arising from agents other than atmospheric CO_2 . α^C , A^C , and CCR^C are respective approximations for α , A , and CCR , in which the effect of additional radiative forcing F_{NC} is omitted. ϕ is a radiative forcing per unit increase of carbon in the atmosphere. Here, we approximate $\phi = 5.35 \ln(1 + C_A/C_{A0})/C_A$, where 5.35 is the conversion factor from atmospheric CO_2 to its radiative forcing (Ramaswamy et al. 2001), C_{A0} is the carbon amount in the atmosphere at the preindustrial CO_2 concentration level. β and γ were obtained from the sensitivity analyses described in Subsection 2.2a, taking values of the 140th year shown in Table 2. For F_{NC} , we used the reference values from the IAM used for producing CMIP5 socio-economic variables (Fig. 1b; Meinshausen et al. 2011, available at <http://cmip-pcmdi.llnl.gov/>).

Table 1. Notations for symbols with superscripts “ALL”, “C”, “no ϵ ”, and “no γ ”.

Symbols	Notations
$\alpha^{ALL}, A^{ALL}, CCR^{ALL}$	Variables obtained from Eqs. (6)–(8), directly applying ESM outputs to these equations. These variables reflect all effects from CO ₂ , non-CO ₂ GHGs and aerosols, and other forcing and feedbacks internally computed in MIROC-ESM.
α^C, A^C, CCR^C	Variables approximated by Eqs. (9)–(11) with α^{ALL} , A^{ALL} , and CCR^{ALL} . These variables represent the climate, carbon cycle, and earth system response in the absence of anthropogenic radiative forcing from non-CO ₂ GHGs and aerosols.
$\alpha^{C,no\epsilon}, A^{C,no\epsilon}, CCR^{C,no\epsilon}$	Variables approximated by Eqs. (13)–(15) with α^{ALL} , A^{ALL} , and CCR^{ALL} . These variables exclude additional energy flux due to the difference of ocean heat uptake efficiency among scenarios, in addition to anthropogenic non-CO ₂ agents.
$A^{ALL,no\gamma}$	Variable approximated by Eq. (A9) with A^{ALL} . This variable excludes the effects of climate-carbon feedback from A^{ALL} .

We also consider the additional heat flux stemming from the decrease of ocean heat uptake efficiency. Ocean heat uptake affects the degree of atmospheric warming because it regulates the heat balance between ocean and lowest atmospheric layer. Raper et al. (2002) suggested that ocean heat uptake efficiency may be altered by the rate of global warming. For example, in monotonically increasing scenarios such as RCP8.5, ocean heat uptake Q could be linearized in association with T using ocean heat uptake efficiency κ : $Q = \kappa T$. However, particularly in RCP2.6 or RCP4.5 where the rate of anthropogenic emission is slow and early declining or stabilization of radiative forcing is achieved in the 21st century, ocean heat uptake may not be proportional to temperature increase. Therefore, the reduction of ocean heat uptake in these “slow” scenarios may act as additional radiative forcing on the atmosphere. Here, we define the additional radiative forcing from the decrease of ocean heat uptake efficiency:

$$F_E = \kappa_{RCP85}(T_{RCP85} - T_{RCPX}), \quad (12)$$

where subscript $RCPX$ represents variables obtained from the result of each RCP simulation. In our definition, F_E in RCP8.5 equals zero.

Analogous to Eqs. (9)–(11), the effects of extra heat flux caused by the reduction of ocean heat uptake efficiency F_E should be eliminated from α^C , A^C , and CCR^C :

$$\alpha^{C,no\epsilon} = \frac{\alpha^{ALL}}{1 + \frac{F_{NC} + F_E}{\phi C_A}}, \quad (13)$$

$$A^{C,no\epsilon} = \frac{A^{ALL}}{1 - \alpha_c \gamma \frac{F_{NC} + F_E}{\phi C_E}}, \quad \text{and} \quad (14)$$

$$CCR^{C,no\epsilon} = \frac{CCR^{ALL}}{1 + (1 + \beta) \frac{F_{NC} + F_E}{\phi C_E}}. \quad (15)$$

$\alpha^{C,no\epsilon}$, $A^{C,no\epsilon}$, and $CCR^{C,no\epsilon}$ are respective approximations for α , A , and CCR , in which the influence of additional radiative forcing F_{NC} and F_E are omitted.

2.4 Summary of methods

We did a 20C3M simulation and projections by applying RCP scenarios to MIROC-ESM. For evaluating the anthropogenic impact on climate, we chose global temperature change (T) and cumulative allowable emission (C_E) as fundamental indices of climate change and anthropogenic impact, respectively. CCR describes the sensitivity of the earth system to C_E . α and A are the measures for the climate and carbon cycle response to anthropogenic impacts. The net land-use change emission composing C_E was estimated in the experiments with an offline terrestrial ecosystem model. The climate, carbon cycle, and earth system response to anthropogenic impacts were further analyzed by applying Eqs. (6)–(15) to model outputs. Definitions of α , A , and CCR are listed in Table 1. In addition to the experiments for 20C3M and RCP scenarios, we conducted sensitivity analyses with 1% CO₂ p.a. increase to examine the strength of carbon cycle feedback in our model. Results are used as the sensitivity parameters in Eqs. (9)–(15).

Table 2. Changes in carbon storage and their sensitivities in MIROC-ESM in the 140th simulation years of three 1% CO₂ increment experiments (CO₂ concentration in the 140th year corresponds to about four times the preindustrial CO₂ level). Land and ocean ecosystems are under the influence of climate change and atmospheric CO₂ increase in FULL, only affected by CO₂ increase in BGC, and only affected by climate change in RAD. Subscripts *A*, *L*, *O*, and *L + O* denote the variables of atmosphere, land, ocean, and sum of land and ocean, respectively. *T* is temperature change, *C* is the increase of carbon storage, β is the sensitivity of carbon storage to *C_A*, and γ is the sensitivity of carbon storage to *T*. Variables in FULL-BGC correspond to those following experimental settings of Friedlingstein et al. (2006), in which there was no RAD experiments and influences of climate change on the carbon cycle were evaluated as the difference between the FULL and BGC experiments. Note that γ in FULL-BGC is calculated with *T* of the FULL experiment.

	<i>T</i>	<i>C_A</i>	<i>C_L</i>	<i>C_O</i>	<i>C_{L+O}</i>	β_L	β_O	β_{L+O}	γ_L	γ_O	γ_{L+O}
	[K]	[PgC]	[PgC]	[PgC]	[PgC]	[PgC PgC ⁻¹]	[PgC PgC ⁻¹]	[PgC PgC ⁻¹]	[PgC K ⁻¹]	[PgC K ⁻¹]	[PgC K ⁻¹]
FULL	5.66	1807	57	576	632	—	—	—	—	—	—
BGC	0.24	1807	661	653	1314	0.37	0.36	0.73	—	—	—
RAD	5.54	1807	−489	−59	−548	—	—	—	−88.3	−10.6	−98.9
FULL-BGC	5.43	1807	−604	−78	−682	—	—	—	−106.6	−13.7	−120.4

3. Results and discussion

3.1 Strength of carbon cycle feedback in MIROC-ESM

Results of the sensitivity analyses of 1% p.a. CO₂ increase experiments are summarized in Table 2. CO₂ concentration in the 140th year reaches about 1130 ppmv, which corresponds to an atmospheric carbon increase of about 1800 PgC. Since CO₂ increase is not allowed to generate radiative forcing in the BGC experiment, the degree of global warming was not great. However, slight warming is confirmed, which is likely caused by the alternation of terrestrial biophysical processes (e.g., albedo, water and heat balance on the land surface). Carbon accumulation in the BGC, which is stimulated by the CO₂ fertilization effect, was about 1310 PgC in total. The contributions of land and ocean to the carbon uptake were almost identical, and the total carbon cycle sensitivity to CO₂ concentration (β) was about 0.73.

In the RAD experiment, in which CO₂ increase causes the climate change, global temperature rose about 5.5 K, comparable to the 5.7 K in the FULL experiment (transient climate sensitivity, which corresponds to the temperature increase in the 70th year, was 2.3 K in that experiment). Climate-carbon feedback reduced carbon storage in both land and ocean by about 550 PgC in total. The sensitivity of land carbon storage to temperature change (γ) was far larger (more negative) than that of oceans, accounting for 89% of total carbon release.

The change in total carbon storage in the FULL experiment, in which the carbon cycle is affected by both concentration-carbon and climate-carbon feed-

back, was 632 PgC, comparable to the 630 PgC change in the Hadley Centre coupled climate-carbon cycle model (HadCM3LC; Gregory et al. 2009). This shows that the total carbon cycle feedback in MIROC-ESM and HadCM3LC are similar. But the strength from concentration-carbon and climate-carbon feedbacks was different between these two ESMs. The increase of carbon storage to *C_A* in the BGC experiment was 1400 PgC in HadCM3LC; total carbon decrease in the RAD experiment was 480 PgC in HadCM3LC. Given that HadCM3LC is at the high end of climate and carbon cycle feedback in C4MIP (Friedlingstein et al. 2006), it may be said that MIROC-ESM is one of the “pessimistic” models for simulating climate projections because it has such large positive feedback.

It is helpful to compare the carbon cycle feedback of our model with past MIP results, but differences in settings, scenarios, and protocols adopted in such projects prevent us from simply comparing the strength of carbon cycle feedback; for example, C4MIP adopted emission-driven experiments, whereas our settings are concentration-driven. Additionally, Friedlingstein et al. (2006) evaluated the strength of climate-carbon feedback as the difference between the FULL and BGC experiments. However, as suggested by Gregory et al. (2009) and Tachiiri et al. (2012), the sum of carbon storage change in the RAD and BGC experiments is not necessarily equal to that of the FULL experiments, and similar results were obtained from our analyses (Table 2). Large nonlinearity was confirmed in our results, especially in land carbon. *C_L* in FULL is much smaller than in RAD+BGC, about 115 PgC. As a result, the climate-carbon feedback in our model evaluated by

Fig. 2. (a) Sum of cumulative diagnosed CO₂ emission from fossil fuel and land-use change (thick lines) and from land-use change only (thin lines), in historical (gray), RCP2.6 (black), RCP4.5 (red), RCP6.0 (green), and RCP8.5 (blue) experiments. (b) Changes in global averages of simulated surface air temperature from 1850 from the prescribed CO₂ concentration experiments.

Table 3. Summary of projected climate, carbon emissions in each experiment: 20th Century Climate in Coupled Models (20C3M) from 1850 to 2005, and RCP2.6/RCP4.5/RCP6.0/RCP8.5 projections from 2006 to 2100. Notations are: atmospheric CO₂ concentration as a driving force ($p\text{CO}_2$) at the end of each experiment, changes in surface air temperature (T), cumulative total emission (C_E), cumulative allowable fossil fuel emission (C_{FF}), and cumulative net land-use emission (C_{LUC}). Subscript “ref” represents reference values approximated during RCP scenario-making, using integrated assessment models that incorporate the earth system model with simplified representation for climate system and ecosystems (Meinshausen et al. 2011).

Experiment	$p\text{CO}_2$ at the end of experiment [ppmv]	T [K]	C_E [PgC]	$C_{E,ref}$ [PgC]	C_{FF} [PgC]	$C_{FF,ref}$ [PgC]	C_{LUC} [PgC]	$C_{LUC,ref}$ [PgC]
20C3M	379	0.9	395	469	270	314	125	155
RCP2.6	421	1.5	212	382	195	321	17	61
20C3M+RCP2.6		2.4	607	850	465	635	142	215
RCP4.5	538	2.5	572	807	653	784	-81	23
20C3M+RCP4.5		3.3	967	1276	923	1098	44	177
RCP6.0	670	3.1	969	1219	1016	1220	-48	-1
20C3M+RCP6.0		4.0	1363	1687	1286	1534	77	153
RCP8.5	936	5.4	1479	1971	1467	1918	12	53
20C3M+RCP8.5		6.2	1874	2439	1737	2231	137	208

FULL-BGC includes this nonlinearity of carbon cycle feedback, and thus sensitivity parameters γ should differ between RAD and FULL-BGC.

3.2 Allowable emissions, global temperature, and carbon cycle

a. Allowable emissions and global temperature

Figure 2a shows the cumulative allowable emissions in 20C3M and projections with RCPs, as diagnosed by MIROC-ESM for C_{FF} and estimated by the offline ter-

restrial ecosystem model for C_{LUC} . In 2100, RCP8.5 showed the largest total allowable emission, 1874 PgC. This was about seven times larger than that of RCP2.6 (Table 3), where carbon capture and storage techniques were assumed to achieve an early reduction in atmospheric CO₂ concentrations. As a result, the cumulative allowable emission in RCP2.6 showed a decrease after 2070 (i.e., an annual net negative anthropogenic emission). Cumulative allowable emissions during 2006–2100 obtained from MIROC-ESM (C_E in Table 3) were

Fig. 3. Comparisons of estimates for cumulative (a) fossil fuel and (b) land-use change emissions from 2006 to 2100: Estimates by MIROC-ESM (white bars), IAMs used for RCP scenario (grey bars), and their difference (calculated as “IAMs” - “MIROC-ESM”; black bars).

lower than those assumed in the RCP scenarios ($C_{E,ref}$ in Table 3), ranging from 56% in RCP2.6 to 80% in RCP6.0. This means that the terrestrial and marine ecosystems in MIROC-ESM have less carbon uptake capability in the model-produced climate than assumed in scenario-making processes. Especially in fossil fuel emissions, the difference of MIROC-ESM (C_{FF}) and RCP ($C_{FF,ref}$) in each scenario shows systematic increase as CO_2 level assumed in the scenario becomes higher (Table 3 and Fig. 3a). This suggests that the discrepancies in fossil fuel emissions may be caused by the difference in strength of climate and carbon cycle feedback between IAMs and MIROC-ESM. On the other hand, such systematic discrepancies cannot be confirmed in land-use emissions (Table 3 and Fig. 3b). Especially for RCP4.5 and 6.0 scenarios, land-use emissions are negative in MIROC-ESM, whereas RCPs show positive or almost zero. This suggests that the definitions for land-use emissions (for example, Eq. 5 and related experimental settings in our analyses) or modeling of emissions from harvested biomass or excessive carbon uptake by secondary forest regrowth might be very different in these two models. In MIROC-ESM, C_{LUC} in each RCP scenario diverged from the beginning of the 21st century, forming two groups (Fig. 2b): increasing (RCP2.6 and 8.5) and decreasing (RCP4.5 and

6.0). This divergence was largely because of land-use change scenarios in the RCPs, i.e., whether the scenario assumed an increase in agricultural area during the 21st century (Kato et al. 2011).

The global mean surface air temperature increased until 2100 in all scenarios, including RCP2.6, in which there is a reduction of atmospheric CO_2 . This suggests that the climate warming will occur even after stabilizing or declining of atmospheric CO_2 concentrations. For RCP2.6, 4.5, 6.0, and 8.5, global-mean surface air temperature increased from 1850 to 2100 by 2.4, 3.3, 4.0, and 6.2 K, respectively. These values are comparable to those of Arora et al. (2011) with Canadian Earth System Model 2. Their ESM showed temperature increases from 1850 to 2100 of 2.3, 3.2, and 5.8 K for RCP2.6, RCP4.5, and RCP8.5, respectively. The results of MIROC-ESM in the RCP8.5 scenario showed slightly higher temperatures than those of Arora et al. (2011), by about 0.4 K. However, we note that their temperature changes were ensemble-averaged, whereas our results were obtained from a single experiment in historical and each RCP projection.

b. Changes in land and ocean carbon storage

Changes in land and ocean carbon storages reproduced by MIROC-ESM are shown in Fig. 4. For land carbon, C_L^{noLUC} in Eq. (5) is also shown. Total carbon storage in the terrestrial ecosystem shows a different trend in each scenario (Fig. 4a). After a decrease in land carbon, especially around the mid-20th century, RCP4.5 and RCP6.0 showed carbon increases, whereas RCP2.6 and RCP8.5 forced a reduction of terrestrial carbon storage. Since these trends correspond to that in C_{LUC} with different sign during the 21st century, land-use change, as well as climate change, may perturb terrestrial carbon storage. As a result, carbon in terrestrial ecosystems increased in RCP4.5 and RCP6.0, and declined in RCP2.6 and RCP8.5.

Figure 4c shows land carbon trends that represent terrestrial carbon change, excluding the effect of land-use change on the carbon cycle (i.e., C_L^{noLUC} in Eq. 5). After removal of these effects, total carbon in terrestrial ecosystems is always larger than that in the pre-industrial state. RCP6.0 showed the largest accumulation at the end of the 21st century, about 160 PgC, while RCP8.5 showed the smallest, less than 20 PgC. The greatest accumulation in RCP6.0 is likely from the lesser contribution of non- CO_2 GHGs to global warming (Fig. 1b). At the end of the 21st century, the contribution of non- CO_2 GHGs in RCP6.0 to total anthropogenic radiative forcing is about 14%, while this contribution in RCP8.5 is about 22%. This contribution of

Fig. 4. (a) Changes in terrestrial carbon C_L , (b) changes in ocean carbon C_O , and (c) changes in terrestrial carbon without impacts of land-use change C_L^{noLUC} . C_L^{noLUC} in each scenario was calculated by adding C_L to cumulative net land-use change emission C_{LUC} based on results from off-line terrestrial ecosystem model.

non- CO_2 GHGs to global warming could alter the earth system response, and will be discussed in Subsection 3.5. The largest carbon reduction in the last two decades of RCP8.5 may be explained by the increase of ecosystem respiration with global warming. As described in Friedlingstein et al. (2006) and Yoshikawa et al. (2008), the increase of ecosystem respiration should intensify climate-carbon feedback.

As opposed to terrestrial carbon storage, the ocean continued as a carbon sink during the entire simulation period in all scenarios, including RCP2.6, in which atmospheric CO_2 decreases after the mid-21st century (Figs. 1a, 3b). Because air-sea exchange of CO_2 is determined by the difference in CO_2 partial pressure ($p\text{CO}_2$) between air and sea, a naive assumption would be that the ocean turns into a source when atmospheric CO_2 decreases, but this is not the case. The most likely reason for continued ocean carbon uptake in RCP2.6 is the transport of inorganic carbon into the deeper ocean.

Miyama and Kawamiya (2009) demonstrated that even after atmospheric CO_2 stabilizes, inorganic carbon continues to be transported from the ocean surface into intermediate and deep ocean layers. Because of such transport, $p\text{CO}_2$ in the surface ocean can never surpass that of the atmosphere (even under RCP2.6), so that the ocean can absorb CO_2 even in a stabilized or peak-decline CO_2 concentration scenario (such as RCP2.6).

3.3 Climate response: α

Figure 5a shows the time evolution of α in each scenario from 1950 to 2100. This includes the effect of GHGs and aerosols, as well as CO_2 (α_{ALL}). α_{ALL} in RCP2.6 continued to increase during the 21st century, while RCP6.0 and RCP8.5 showed declines, especially after 2050. One reason for these α reductions, especially under scenarios with higher atmospheric CO_2 concentration, is because of the saturation of radiative forcing from atmospheric CO_2 (as described in

Fig. 5. Time evolution of temperature change per unit carbon “ α ”. (a) α with all radiative forcing using MIROC-ESM results ($= \alpha^{ALL}$). (b) α in the absence of additional anthropogenic radiative forcing from non- CO_2 GHGs and aerosols ($= \alpha^C$ approximated by Eq. 9). (c) α from which additional energy flux to the atmosphere caused by decrease in ocean heat uptake efficiency was omitted, in addition to influence from non- CO_2 GHGs and aerosols ($= \alpha^{C,noE}$ approximated by Eq. 13).

Matthews et al. 2009 and Gregory et al. 2009). As a result, the range of α_{ALL} from RCP2.6 to RCP8.5 was about 3.7 K/TtC, comparable to 70% of the absolute value of α_{ALL} in 2000.

Figure 5b shows α^C , where the effect of radiative forcing by non- CO_2 agents was removed using Eq. (9) to account for other reasons of α^{ALL} variation among RCPs. Since the net anthropogenic radiative forcings other than CO_2 was positive during the 21st century (Fig. 1b), α^C was lower than α^{ALL} after 2000. At the end of the 21st century, α^C in RCP2.6 and RCP4.5 showed about a 17% reduction compared with α^{ALL} , and RCP8.5 had the greatest reduction, about 22%. The removal of radiative forcing from non- CO_2 agents made for less α variation among scenarios, but the degree of reduction was so small that it only accounts for 11%

variation in α^{ALL} .

Figure 5c shows $\alpha^{C,noE}$, for which the additional energy flux to the atmosphere from the reduction of ocean heat uptake efficiency was removed, using Eq. (13). After removing the variation of ocean heat uptake efficiency, the variation of $\alpha^{C,noE}$ among RCPs was 1.7 K/TtC. This result suggests that the difference in ocean heat uptake efficiency accounts for about 40% of the variation in α^{ALL} . Possible explanations for the residual α^{ALL} variations among RCPs is the difference in radiative forcing from CO_2 (the saturation effect of CO_2 radiative forcing, mentioned above), and the difference of climate feedbacks within climate systems.

Fig. 6. Time evolution of cumulative airborne fraction “A”. (a) A with all radiative forcing using MIROC-ESM results ($= A^{ALL}$). (b) A in the absence of additional radiative forcing from non- CO_2 GHGs and aerosols ($= A^C$ approximated by Eq. 10). (c) A from which additional radiative energy flux caused by decrease in ocean heat uptake efficiency was omitted using the same efficiency, in addition to the effect from non- CO_2 GHGs and aerosols ($= A^{C, no\epsilon}$ approximated by Eq. 14). (d) A excluding climate-carbon cycle feedback ($= A^{ALL, no\gamma}$ approximated by Eq. A9).

3.4 Carbon cycle response: A

The cumulative airborne fraction estimated from MIROC-ESM is shown in Fig. 6a. At the end of the 21st century, RCP8.5 showed the highest value of about 0.74, followed by 0.60 in RCP6.0, 0.56 in RCP4.5, and 0.48 in RCP2.6. The order of RCP4.5 and RCP6.0 changed around 2060, corresponding to a reversal of the CO_2 concentration order (Fig. 1a).

After excluding anthropogenic radiative forcing by non- CO_2 agents and the difference in ocean heat uptake efficiency among RCPs using Eqs. (10) and (14), the airborne fraction became lower than that with all forcing, A^{ALL} (Figs. 6b, c). This is because these additional radiative forcings caused additional warming accompanied by larger ecosystem respiration. However, even

after these treatments, variations of airborne fractions among RCPs were about 0.24 in A^C and 0.27 in $A^{C, no\epsilon}$, nearly the same as the 0.26 in A^{ALL} . There are three possibilities that may account for these variations: levels of CO_2 concentration, climate-carbon feedback, or concentration-carbon feedback. As noted above, there may also be variation in CO_2 radiative forcing among RCPs because it will saturate due to the increase of atmospheric CO_2 . However, this difference among RCPs cannot explain the variation of A^{ALL} because the saturation of radiative forcing efficiency in higher CO_2 concentration scenarios induces less global warming and smaller ecosystem respiration.

Figure 6d shows the airborne fraction approximation in which the effect of climate-carbon feedback is

Fig. 7. Time evolutions of carbon-climate response “CCR”. (a) CCR with all radiative forcing using MIROC-ESM results ($= CCR^{ALL}$). (b) CCR without additional radiative forcing from non- CO_2 GHGs and aerosols ($= CCR^C$ approximated by Eq. 11). (c) CCR from which additional energy flux caused by a decrease in ocean heat uptake efficiency was omitted by using the same efficiency, in addition to the effect from non- CO_2 GHGs and aerosols ($= CCR^{C,no\epsilon}$ approximated by Eq. 15). Dashed black line in each figure indicates CCR of RCP2.6 that used peak C_E (corresponding to C_E in year 2075) for its calculation.

excluded ($A^{ALL,no\gamma}$) from A^{ALL} (see Eq. A9 and Appendix C for the formulation of this approximation) assuming that the parameters for climate-carbon feedback are the same among four scenarios. $A^{ALL,no\gamma}$ still showed a large variation of 0.21, although this is less than that in A^{ALL} . This result indicates that the variation of concentration-carbon feedback may explain the large fraction of variations in A among RCPs, with the assumption that parameter γ is independent of scenario (i.e., the rate of change). Gregory et al. (2009) showed that the sensitivity of land and ocean carbon cycles to atmospheric CO_2 increase (β) depends on the rate of atmospheric CO_2 change, whereas the sensitivity to climate change (γ) had much less dependence on the rate of the scenario. Following their results, the variations

in A^{ALL} may be attributed to the rate of CO_2 increase. However, since these carbon cycle characteristics were not examined in detail, there is a need for further analyses focusing on carbon cycle dependence on the rate of the scenario and warming.

3.5 Earth system response: CCR

The variation in anthropogenic forcing also affected CCR within each RCP scenario. Figure 7a shows CCR^{ALL} calculated through direct application of ESM outputs to Eq. (6). Except for the case of RCP2.6, CCR^{ALL} in each scenario was relatively constant after 2050, which is consistent with the results of Matthews et al. (2009). At the end of the 21st century, the difference in CCR maximum (RCP2.6) and minimum (RCP6.0)

was about 1.0 K/TtC. After removing the additional radiative forcing from non-CO₂ GHGs and aerosols, the *CCR* average among all RCPs was reduced from 3.4 K/TtC in *CCR*^{ALL} to 2.8 K/TtC in *CCR*^C, although the *CCR*^C variation among RCPs was still around 1.0 K/TtC (Fig. 7b). Furthermore, after omitting additional heat flux arising from the decrease of ocean heat uptake efficiency, *CCR*^{C, noE} in each scenario showed similar trends during the 21st century and converged to 2.3–2.4 K/TtC (Fig. 7c) in 2100, except for RCP2.6 in which the characteristic increase remains after 2085.

These analyses using *CCR* depict the characteristics of the earth system response to RCP scenarios. First, the earth system in RCP6.0 showed low sensitivity to C_E in the relationship between climate and carbon cycle (Fig. 7a). However, after excluding the radiative forcing from non-GHG agents, *CCR* in RCP6.0 fell into the range between RCP8.5 and RCP4.5 (Fig. 7b). At the end of the 21st century, the non-CO₂ radiative forcing in RCP6.0 estimated by the IAM (Fig. 1b) is at the same level as RCP4.5. These anthropogenic forcings by non-CO₂ agents act only on climate-carbon feedback, whereas CO₂ increase strengthens the concentration-carbon feedback, stimulating the “fertilization effect” in ecosystems. The low sensitivity of the earth system in RCP6.0 can be confirmed by the land carbon change without land-use effect (Fig. 4c), showing that land carbon accumulation C_L in RCP6.0 is the greatest of all scenarios because of less warming caused by additional radiative forcing of non-CO₂ agents. Therefore, RCP6.0 resulted in the most effective scenario for stabilizing CO₂ and mitigating future climate change. However, to evaluate the impact of land-use change on land biogeophysical processes and resultant local climate change, uncoupled simulations for land-use change in each scenario are needed; this will be described elsewhere.

Second, *CCR* displays the unique behavior of earth system response to anthropogenic impact in RCP2.6. Especially after 2075, RCP2.6 shows an apparently different trend from other scenarios. One reason for this trend is that the annual anthropogenic carbon emission is negative during this period. Thus, the cumulative total carbon emission, which is the denominator of Eq. (6), decreased. By taking the peak value for C_E , which corresponds to C_E in 2075 of RCP2.6, the trend after 2075 was moderated (black dashed lines in Fig. 7), especially in *CCR*^{C, noE}. In cases using the peak cumulative emission, RCP2.6 is converged to 2.3–2.6 K/TtC (black dashed line in Fig. 7c). In a peak and decline scenario such as RCP2.6, peak value for cumulative emission might be more important rather than cumulative emis-

sion itself. To investigate what will happen and how it will be induced after 2100, RCP2.6 should be helpful for understanding the earth system and projecting with mitigation scenarios.

We diagnosed the variables, particularly those associated with the carbon cycle (i.e., cumulative carbon emissions or sensitivity parameters such as β and γ), and used them for the analyses. However, other variables, especially non-CO₂ radiative forcing (F_{ETC} in Eqs. 9–11 and 13–15), were obtained from the IAM estimates used in RCPs. It is more appropriate to use the model-specific variables that can be obtained with the ESM from several uncoupled simulations to maintain analysis consistency. Especially for aerosols, large uncertainty remains for estimates of radiative forcing. In Takemura et al. (2006), where instantaneous radiative forcing was evaluated by the atmosphere and aerosol transport-radiation component of MIROC-ESM, the radiative forcing from direct/indirect aerosol effects were estimated about -0.1 and -1.0 W m⁻² in 2000. Corresponding estimates of IAM used for RCPs were about -0.4 and -0.7 W m⁻². Furthermore, radiative forcing by volcanic eruption is significantly different between these two models (i.e., peak radiative forcing from the Pinatubo eruption was about -2.4 W m⁻² in Takemura et al. (2006) and -1.4 W m⁻² in the IAM), showing the difference of model sensitivity to stratospheric aerosols. In addition, we only considered the non-CO₂ radiative forcing from atmospheric composition change. However, changes in land surface condition, such as albedo change caused by anthropogenic land-use, may also function as radiative forcing. These land surface changes also impact climate change, but further analyses and evaluations for modeling these processes are needed.

4. Conclusions

We conducted past and future simulations from 1850 to 2100 using a newly developed ESM, MIROC-ESM. Four RCP scenarios that describe future socio-economic changes and anthropogenic forcing were used in the ESM. Our results showed that temperature increase from 1850 to 2100 was between 2.4 and 6.2 [K]. This is consistent with other studies. Carbon emission allowances to achieve RCP concentration scenarios predicted by MIROC-ESM should be less than the reference amount assumed in the scenario-making processes for RCP. This is because the climate and carbon cycle feedback in our model was so strong that its strength was comparable to past “pessimistic” ESM suggested by sensitivity analyses.

Climate response measured by temperature increase

per unit atmospheric carbon increase showed substantially different trends among scenarios, e.g., greater sensitivity in slower scenarios. Climate response in each scenario was greatly affected by ocean heat uptake efficiency, which determines energy flux at the bottom of the atmosphere and depends on the rate of climate change. The difference in ocean heat uptake efficiency accounted for more than 40% of the variation in climate responses. Variation in climate response among RCPs was also produced by differences in non-CO₂ anthropogenic impact on climate, such as CH₄, N₂O, other GHGs, and aerosols.

Carbon cycle responses to RCPs, evaluated by the cumulative airborne fraction, also varied among scenarios. However, this variation could not be explained by the difference in ocean heat uptake efficiency or non-CO₂ anthropogenic radiative forcing from GHGs/aerosols. With the assumption that climate-carbon feedback is independent of the rate of climate change (i.e., there is no difference of carbon cycle sensitivity to global warming among the scenarios with different rates of atmospheric CO₂ change) as supported by Gregory et al. (2009), this large variation may be attributable to the dependence of concentration-carbon feedback on the rate of atmospheric CO₂ increase. For quantifying the dependence of carbon cycle feedback on the rate of socio-economic change and revealing its mechanism, uncoupled experiments such as “BGC”/“RAD” under different scenarios should be conducted with ESMs.

The earth system response aggregates climate and carbon cycle response to anthropogenic impacts that are assessed by temperature change per unit carbon emission. These responses also varied among all scenarios. The variation of earth system response between RCP scenarios can be explained by the difference in radiative forcing arising from anthropogenic non-CO₂ agents and variations in ocean heat uptake efficiency. The earth system behavior captured by *CCR* depicts characteristics of the response to each RCP scenario. Especially in RCP6.0, the anthropogenic impacts reinforce carbon uptake by terrestrial/marine ecosystems when compared with other scenarios. This is likely caused by lesser contributions of non-CO₂ GHGs to global warming, leading to low sensitivity of the earth system to the emitted carbon. In contrast, the earth system response in RCP2.6 was highly sensitive to emitted carbon, and its evolution in the 21st century seems very different from other scenarios. This trend was modified by applying the peak emission for calculating *CCR*, in addition to the removal of radiative forcing arising from non-CO₂ anthropogenic agents and the decrease of ocean heat uptake efficiency. To reveal the detailed mechanism of

the earth system and provide practical information on mitigation of climate change, further analyses using extended scenarios are needed.

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Appendix A

Incorporation of land-use change in terrestrial ecosystem model

To represent the effects of anthropogenic land-use change on the carbon cycle and biophysical processes, the terrestrial ecosystem model incorporates the land-use datasets of Hurtt et al. (2009) that describe areal transitions of anthropogenic land-use among five categories (primary vegetation, secondary vegetation, pasture, cropland, and urban area). Following these datasets, fractional changes of land-use area are reproduced on each grid of MIROC-ESM, and then the impacts on vegetation are computed. Secondary vegetation is formed from logging or burning of primary forests or abandonment of agricultural land. The model then simulates additional carbon uptake from forest regrowth. Carbon in harvested biomass is transferred into carbon pools of linear decay (with turnover times of 1, 10, and 100 yr) according to the Grand Slam Protocol described by Houghton et al. (1983). The balance of carbon uptake from plant regrowth and emission from harvested biomass or residue determines the magnitude of net carbon emission (i.e., net carbon loss in terrestrial ecosystems) caused by land-use change.

Appendix B

Excluding an additional radiative forcing

a. Fundamental equations

Gregory et al. (2009) proposed simple but essential equations for heat and carbon balance to understand the earth system using first-order approximations. The heat

balance equation is described below.

$$F_C + F_{ETC} = (\lambda + \kappa)T = \rho T \quad (\text{A1})$$

F_C and F_{ETC} are radiative forcings from atmospheric CO_2 and other agents, respectively. λ is the sum of climate feedback except for carbon cycle feedback, and κ is ocean heat uptake efficiency. The sum of λ and κ is defined as climate resistance ρ .

Carbon balance is

$$C_E = C_A + C_L + C_O.$$

This equation represents the allocation of emitted anthropogenic carbon C_E to atmosphere (C_A), land (C_L), and ocean (C_O). This equation can be rewritten using the sensitivity parameters of carbon cycle to the increase of atmospheric carbon (β) and temperature change (γ):

$$C_E = C_A + \beta C_A + \gamma T.$$

(A2; same as described in Eq. 1)

The radiative forcing from anthropogenic carbon is a function of C_A :

$$F_C = \phi C_A. \quad (\text{A3})$$

b. Derivation of α^C

In the case of $F_{ETC} = 0$ in Eq. (A1) (i.e., climate change is induced only by CO_2 increase), the temperature change per unit carbon α and climate resistance ρ have a relationship:

$$\alpha^C = \frac{T}{C_A} = \frac{\phi}{\rho}, \quad (\text{A4})$$

using Eqs. (7), (A1), and (A3).

When the additional radiative forcing F_{ETC} is not equal to zero in Eq. (A1), we can obtain the relationship

$$\phi C_A + F_{ETC} = \rho T.$$

Thus, α_{ALL} , which corresponds to α that includes the additional radiative forcing F_{ETC} , is described as

$$\alpha_{ALL} = \frac{T}{C_A} = \frac{\phi}{\rho} + \frac{F_{ETC}}{\rho C_A} = \frac{\phi}{\rho} \left(1 + \frac{F_{ETC}}{\phi C_A} \right).$$

By using Eq. (A4), α^{ALL} must be

$$\alpha^{ALL} = \alpha^C \left(1 + \frac{F_{ETC}}{\phi C_A} \right).$$

Thus,

$$\alpha^C = \frac{\alpha^{ALL}}{1 + \frac{F_{ETC}}{\phi C_A}}. \quad (\text{A5})$$

c. Derivation of A^C

When there is no radiative forcing other than CO_2 ($F_{ETC} = 0$), by removing T from Eqs. (A1) and (A2), we can obtain

$$C_E = C_A + \beta C_A + \frac{\phi \gamma}{\rho} C_A.$$

Thus,

$$\frac{C_A}{C_E} = \frac{1}{1 + \beta + \frac{\phi \gamma}{\rho}} = A^C. \quad (\text{A6})$$

In the case where F_{ETC} is not equal to zero, solving Eqs. (A1) and (A2) by removing T gives

$$C_E = C_A \left(1 + \beta + \frac{\phi \gamma}{\rho} \right) + \frac{\gamma}{\rho} F_{ETC}.$$

Thus, we can obtain

$$\frac{C_A}{C_E} = \frac{1 - \frac{\gamma F_{ETC}}{\rho C_E}}{1 + \beta + \frac{\phi \gamma}{\rho}} = A^{ALL}. \quad (\text{A7})$$

Using Eqs. (A4), (A6), and (A7), the relationship between A_{ALL} and A^C is

$$A^{ALL} = A^C \left(1 - \frac{\gamma F_{ETC}}{\rho C_E} \right) = A^C \left(1 - \alpha^C \gamma \frac{F_{ETC}}{\phi C_E} \right).$$

Thus,

$$A^C = \frac{A^{ALL}}{1 - \alpha^C \gamma \frac{F_{ETC}}{\phi C_E}}.$$

d. Derivation of CCR^C

In the case of $F_{ETC} = 0$, by removing C_A in Eqs. (A1), (A2), and (A3), we can obtain T as a function of C_E :

$$T = \frac{\phi}{\rho + \rho \beta + \phi \gamma} C_E.$$

Thus, CCR^C is described with parameters ϕ , ρ , β , and γ :

$$CCR^C = \frac{T}{C_E} = \frac{\phi}{\rho + \rho \beta + \phi \gamma}. \quad (\text{A8})$$

In the case where F_{ETC} is not equal to zero, T as a function of C_E includes the term F_{ETC} :

$$T = \frac{\phi}{\rho + \rho \beta + \phi \gamma} C_E + \frac{1 + \beta}{\rho + \rho \beta + \phi \gamma} F_{ETC}.$$

Thus,

$$CCR^{ALL} = \frac{T}{C_E} = \frac{\phi}{\rho + \rho \beta + \phi \gamma} \left(1 + (1 + \beta) \frac{F_{ETC}}{\phi C_E} \right).$$

Using Eq. (A8), the relationship for CCR^{ALL} is

$$CCR^{ALL} = CCR^C \left(1 + (1 + \beta) \frac{F_{ETC}}{\phi C_E} \right).$$

Thus,

$$CCR^C = \frac{CCR^{ALL}}{\left(1 + (1 + \beta) \frac{F_{ETC}}{\phi C_E} \right)}.$$

Appendix C

Cumulative airborne fraction without climate-carbon feedback

When there is no climate-carbon feedback effect (the term $\gamma T = 0$ in Eq. A2), the carbon balance should be

$$C_E^{ALL, no\gamma} = C_A + \beta C_A,$$

where superscript $no\gamma$ denotes the variable in the absence of climate-carbon feedback. After subtracting Eq. (A2) by this equation, we obtain

$$C_E - C_E^{ALL, no\gamma} = \gamma T.$$

By dividing by C_A , this equation becomes

$$\frac{C_E}{C_A} - \frac{C_E^{ALL, no\gamma}}{C_A} = \frac{\gamma T}{C_A}.$$

Since $A = C_A/C_E$ and $A^{ALL, no\gamma} = C_A/C_E^{ALL, no\gamma}$, this equation becomes

$$\frac{1}{A} - \frac{1}{A^{ALL, no\gamma}} = \frac{\gamma T}{C_A}.$$

Therefore, the relationship between $A^{ALL, no\gamma}$ and A is described below.

$$A^{ALL, no\gamma} = \frac{A^{ALL} C_A}{C_A - A^{ALL} \gamma T}. \quad (A9)$$

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