Terrestrial Carbon Cycle: Climate Relations in Eight CMIP5 Earth System Models

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ABSTRACT

Eight Earth System Models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) are evaluated, focusing on both the net carbon dioxide flux and its components and their relation with climatic variables (temperature, precipitation, and soil moisture) in the historical (1850–2005) and representative concentration pathway 4.5 (RCP4.5; 2006–2100) simulations. While model results differ, their median globally averaged production and respiration terms from 1976 to 2005 agree reasonably with available observation-based products. Disturbances such as land use change are roughly represented but crucial in determining whether the land is a carbon source or sink over many regions in both simulations. While carbon fluxes vary with latitude and between the two simulations, the ratio of net to gross primary production, representing the ecosystem carbon use efficiency, is less dependent on latitude and does not differ significantly in the historical and RCP4.5 simulations. The linear trend of increased land carbon fluxes (except net ecosystem production) is accelerated in the twenty-first century. The cumulative net ecosystem production by 2100 is positive (i.e., carbon sink) in all models and the tropical and boreal latitudes become major carbon sinks in most models. The temporal correlations between annual-mean carbon cycle and climate variables vary substantially (including the change of sign) among the eight models in both the historical and twenty-first-century simulations. The ranges of correlations of carbon cycle variables with precipitation and soil moisture are also quite different, reflecting the important impact of the model treatment of the hydrological cycle on the carbon cycle.

1. Introduction

The global carbon cycle consists of the combined interactions among a series of carbon reservoirs in the earth system (such as CO₂ in the atmosphere, soil organic carbon and vegetation, and carbonate and phytoplankton in the ocean) and all the fluxes and feedbacks that regulate dynamics in the sizes of these reservoirs. Most of the sensitivity and uncertainty in coupled carbon-climate projections lie in the terrestrial (rather than oceanic) carbon cycle (e.g., Zeng et al. 2005; Friedlingstein et al. 2006; Denman et al. 2007; Booth and Jones 2011). The importance of the terrestrial component of the carbon cycle to future model projections is widely recognized, in large part because the terrestrial component
is so greatly influenced by anthropogenic activities such as restoration from past disturbances and changes in land use or management (Houghton 2007). The interannual and interdecadal variability in the growth rate of atmospheric CO2 concentration is dominated by the response of the land biosphere sink to climate variations (Denman et al. 2007), and recent analyses from global observations have revealed that these sinks, while remaining stable in their long-term averages, have become increasingly sensitive to subtle changes in climate (Ballantyne et al. 2012). The terrestrial carbon sink is affected by temperature, precipitation, soil moisture, water vapor, CO2 concentration, and solar visible radiation and hence will be altered by climate change (Long et al. 2004; Boer and Arora 2010; Wolkovich et al. 2012; Rutishauser et al. 2012). In the twenty-first century, the land carbon storage is expected to be enhanced by the CO2 increase but dampened by climate change (Friedlingstein et al. 2006). In turn, feedbacks from global carbon sinks to the atmospheric reservoir of CO2 can accelerate or constrain climate change (e.g., Qian et al. 2009).

It is important to critically evaluate the terrestrial carbon cycle model using observations, since its uncertainty is comparable to that attributed to all the physical parameters (Gregory et al. 2009; Booth and Jones 2011). Despite much progress in model development and observational dataset development, large uncertainties still exist in the model simulations of climate (Good et al. 2013) and climate–carbon feedbacks (Cadule et al. 2009; Gregory et al. 2009) and in carbon-related observations (Todd-Brown et al. 2012). Previous model intercomparisons, such as the Coupled Carbon Cycle Climate Model Intercomparison Project (C4MIP), focused mostly on the net surface–atmosphere carbon exchange that directly impacts the global climate (e.g., Friedlingstein et al. 2006), and much less on other carbon cycle components. On the other hand, the net carbon flux represents the small difference between two large gross fluxes (i.e., global respiration and gross primary productivity), and its value (including its sign) can be easily changed by uncertainties in either of the gross fluxes and/or other smaller components associated with natural and anthropogenic disturbances. To improve our understanding of processes related to the net carbon flux exchange and ensure that the right answers on the flux exchange are for the right reasons, it is crucial to evaluate both the net flux exchange and other components of the terrestrial carbon cycle. One of the purposes of this study is to fill this gap.

Specifically, we will analyze the terrestrial carbon cycle from the fifth phase of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012; see Table 1 for a complete list of all model expansions).
Both CMIP5 and the previous C4MIP incorporate the carbon cycle. Compared with C4MIP, the main advantage of CMIP5 is the representation of land cover change that is considered to be a significant driver for the terrestrial carbon cycle (Ballantyne et al. 2012; Houghton 2012), and the inclusion of generally more processes [such as the nitrogen (N) cycle and dynamics in regional fire frequency]. Explicit land use change (LUC) projections are provided along with future CO2 emission projections and are represented in the region-specific simulations through different land use classifications, parameter settings, allocation rules, and biogeographical patterns in CMIP5. Furthermore, CMIP5 includes non-CO2 anthropogenic emissions (e.g., sulfur and black and organic carbon) and more diverse coupled carbon cycle experiments than C4MIP [e.g., historical and future simulations with prescribed CO2 concentration or prescribed anthropogenic emission of CO2 and other idealized experiments to assess the climate–carbon cycle feedback; Taylor et al. (2012)].

Complementary to previous carbon–climate model intercomparison studies (e.g., Friedlingstein et al. 2006), in this study we will evaluate both the net and individual carbon fluxes between land and atmosphere against available observations and their responses to climate using CMIP5 historical simulations. We will also analyze how such responses of the terrestrial ecosystem differ between the historical hindcast and future projection. Furthermore, we will quantify the perturbations of disturbances to the net CO2 flux exchange between land and atmosphere. Our focus is placed on the CO2 flux between the atmosphere and land rather than carbon storage (or “net sequestration”), partly due to the availability of large-scale observations of the former and lack thereof for the latter.

2. Model and data descriptions

a. CMIP5 models and experiments

Eight earth system models (ESMs) from CMIP5 were selected based on the availability of monthly outputs of specific carbon fluxes, and a set of their output fields relating to the terrestrial carbon budget and climate is analyzed. These ESMs differ in their representations of vegetation types, soil properties, human disturbances, carbon, and nitrogen pools, as well as horizontal and vertical resolutions on the surface and in the atmosphere and ocean, respectively. Table 2 summarizes the components and characteristics of each ESM and their terrestrial carbon cycle; appendix A provides more details.

We focus on two suites of CMIP5 experiments—concentration-driven historical and representative concentration pathway 4.5 (RCP4.5). The former is also referred to as the twentieth-century simulations from the mid-nineteenth century to near present and is suitable for comparison with observations. The RCP4.5 experiment provides a future projection of climate from 2006 to 2100 based on a mitigation or stabilization scenario in which the total radiative forcing is stabilized at 4.5 W m$^{-2}$ roughly before 2100 with CO2 concentration being approximately 525 ppm; Thomson et al. (2011)]. The historical simulations are initialized with a quasi-equilibrium state from the preindustrial control integrations, and the RCP4.5 simulations are a continuous extension from the end of the historical simulations. For the RCP4.5 scenario, the areas used for crops, pasture, and wood harvesting increase in the first half of the twenty-first century for food and bioenergy use, assuming an increase in population. After 2050, areas of crops and pasture decline, and forested areas expand as a result of the slowed growth rate in food demand, which is assumed to accompany a global population approaching equilibrium. More information about the RCPs and their development process can be found in Moss et al. (2010). The scenario drivers and technology options are detailed in Clarke et al. (2007) and Thomson et al. (2011). Additional specifications of land use and terrestrial carbon emissions are given by Wise et al. (2009).

The analyzed model outputs that we use are from the first realization in the ensemble of simulations with different initial conditions. These outputs and the observational data are regridded to a common resolution (T62 Gaussian grid of 1.9° × 1.9° with fractional land in grid cells considered). Note that because CO2 concentration is prescribed for the historical and RCP4.5 simulations, the computed net carbon uptake in all simulations did not affect CO2 concentration as a feedback.

There are many more models in CMIP5 now than the eight models that were available at the beginning of this project. Most of these models also include several ensemble members. Furthermore, besides RCP4.5, there are three other scenarios (RCP2.6, RCP6, and RCP8.5) (Moss et al. 2010). With our focus on all the flux components of the terrestrial carbon cycle, we consider that the eight models (with one ensemble member from each) under RCP4.5 used here represent enough diversity (Table 2 and appendix A) in our model evaluations, and it is beyond the scope of this paper to include all models and their ensemble members under all scenarios.

b. Carbon fluxes and model output

While the eight ESMs contain different land models, there are some similarities in their treatment of the terrestrial carbon cycle. For example, plants are categorized into several different plant function types (PFTs). For each PFT, the parameterizations for leaf photosynthesis (A), autotrophic respiration (Ra, as the sum of maintenance and growth respiration, Rm and Rg, respectively),
Table 2. Comparison of the selected attributes and processes related to the terrestrial carbon cycle in the eight ESMs. “Vegetation model” refers to the land component on vegetation phenology and carbon cycle, while “dynamic vegetation” refers to the land model component on plant competition and dynamic change of plant coverage.

<table>
<thead>
<tr>
<th>ESMs</th>
<th>CanESM2</th>
<th>CCSM4</th>
<th>GFDL-ESM2M</th>
<th>HadGEM2-ES</th>
<th>INM-CM4.0</th>
<th>MIROC-ESM</th>
<th>MPI-ESM-LR</th>
<th>NorESM1-M</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land model</td>
<td>CLASS</td>
<td>CLM4</td>
<td>LM3</td>
<td>JULES</td>
<td>No name</td>
<td>MATSIRO</td>
<td>JSBACH</td>
<td>CLM4</td>
</tr>
<tr>
<td>Vegetation model</td>
<td>CTEM</td>
<td>CLM4CN</td>
<td>LM3V</td>
<td>TRIFFID</td>
<td>No name</td>
<td>SEIB-DGVM</td>
<td>BETHY</td>
<td>CLM4CN</td>
</tr>
<tr>
<td>No. of PFTs</td>
<td>9</td>
<td>15</td>
<td>5</td>
<td>5</td>
<td>11</td>
<td>13</td>
<td>13</td>
<td>15</td>
</tr>
<tr>
<td>Dynamic vegetation</td>
<td>Yes</td>
<td>No</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>N cycle</td>
<td>No</td>
<td>Yes</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>No. of soil layers</td>
<td>3</td>
<td>15</td>
<td>20</td>
<td>4</td>
<td>23</td>
<td>6</td>
<td>5</td>
<td>15</td>
</tr>
<tr>
<td>Soil depth (m)</td>
<td>4.1</td>
<td>43.7</td>
<td>10</td>
<td>3</td>
<td>15</td>
<td>14</td>
<td>9.58</td>
<td>42.1</td>
</tr>
<tr>
<td>Human activity</td>
<td>Crop</td>
<td>Crop, pasture, wood harvest</td>
<td>Crop, pasture, wood harvest, deforestation</td>
<td>Crop, pasture</td>
<td>Deforestation</td>
<td>Crop, pasture</td>
<td>Crop, pasture, wood harvest, deforestation</td>
<td>Crop, pasture, wood harvest</td>
</tr>
<tr>
<td>Horizontal resolution (lat × lon)</td>
<td>2.8° × 2.8°</td>
<td>0.9° × 1.3°</td>
<td>2.0° × 2.5°</td>
<td>1.3° × 1.9°</td>
<td>1.5° × 2.0°</td>
<td>2.8° × 2.8°</td>
<td>1.9° × 1.9°</td>
<td>1.9° × 2.5°</td>
</tr>
</tbody>
</table>
carbon allocation, and phenology are similar across the models, though their specific parameters and limiting conditions are different. The parameterization of $A$ mostly follows Farquhar et al. (1980) for C3 plants and Collatz et al. (1992) for C4 plants. Its interplay with stomatal conductance ($SC$), which is linked to water vapor and $CO_2$ fluxes, is generally taken from the Ball–Berry conductance formulation (Collatz et al. 1991; Sellers et al. 1996). Appendix B describes the general formulations for these processes and the heterotrophic respiration ($Rh$) and their connections to climate [specifically, temperature ($T$), precipitation ($Pr$), and soil moisture ($SM$)]. More details specific to each ESM are given in appendix A and references therein. Here, we briefly summarize the definitions of several carbon fluxes that are analyzed in the subsequent sections.

Gross primary production (GPP; $kgC m^{-2} yr^{-1}$) is the aggregate amount of $A$ ($gC m^{-2} s^{-1}$) at every model time step ($\Delta t$; s), and net primary production (NPP; $kgC m^{-2} yr^{-1}$) is the difference between GPP and a similarly aggregated amount of $Ra$ ($kgC m^{-2} yr^{-1}$):

$$GPP = \sum A \Delta t \quad \text{and}$$

$$NPP = GPP - Ra = GPP - Rm = Rg.$$  \hspace{1cm} (2)

NPP is allocated to various plant pools such as leaves, stems, and roots. Then the carbon in these pools is partitioned to the soil as litter at a specified rate for each PFT and plant tissue according to natural turnover rates, as well as mortality and disturbance events. Among these pools, leaf carbon pool $C_{leaf}$ ($kgC m^{-2}$) can be converted to the leaf area index (LAI) as

$$LAI \approx C_{leaf} \times SLA,$$  \hspace{1cm} (3)

where SLA is the specific leaf area ($m^2 kgC^{-1}$) that is either a PFT-specific constant or a value that varies along a vertical gradient in the canopy (e.g., Thornton and Zimmermann 2007).

The remaining carbon after subtracting both $Ra$ and $Rh$ is defined as the net ecosystem production (NEP):

$$NEP = GPP - Ra - Rh = NPP - Rh.$$  \hspace{1cm} (4)

Further, the remaining portion of NEP after considering the disturbances ($D$) such as fires and LUC is defined as the net biome production (NBP), following the CMIP5 list of requested model output (available at http://cmip-pcmdi.llnl.gov/cmip5/data_description.html):

$$NBP = NEP - D.$$  \hspace{1cm} (5)

A positive sign for NEP and NBP indicates net uptake of carbon from the atmosphere by the land, while a negative sign signifies a net release of carbon from the land back to the atmosphere.

Not all of these variables were available for each model at the time we downloaded the CMIP5 outputs, and in such cases they are derived from other variables. For instance, NPP of INM-CM4.0 was determined by subtracting Ra from GPP, and for NorESM1-M it was taken as the sum of the carbon allocated to leaf, wood, and root. Three physical variables, $T$ ($^\circ C$), $Pr$ ($mm yr^{-1}$), and $SM$ ($kg m^{-2}$), were obtained directly from the CMIP5 outputs. SM is computed on a standard depth down to 3 m. Note that outputs of MPI-ESM-LR do not include SM in each soil layer, and its reported soil moisture over the whole soil column with a depth of 9.58 m is used here. The variation of SM reflects the combining effects of climate, soil, and vegetation, particularly in water-limited ecosystems (D’Odorico et al. 2007).

c. Observational data

For our model evaluations, data from a variety of sources are used (Table 3). The actual data uncertainties are generally larger than the standard deviations provided in Table 3. The temperature data (Jones et al. 2012) are based on in situ measurements, and the precipitation data (Xie and Arkin 1997) are based on in situ measurements and satellite remote sensing. The LAI data (Lawrence and Chase 2007, 2010) are based on the satellite Moderate Resolution Imaging Spectroradiometer (MODIS) products. In general, these data are reliable

<table>
<thead>
<tr>
<th>Variables</th>
<th>Mean (SD)</th>
<th>Temporal coverage</th>
<th>Spatial coverage</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T$ ($^\circ C$)</td>
<td>13.48 ($\pm 0.21$)</td>
<td>Monthly; 1850-2012</td>
<td>Global; $5^\circ \times 5^\circ$</td>
<td>Jones et al. (2012)</td>
</tr>
<tr>
<td>$Pr$ ($mm yr^{-1}$)</td>
<td>733 ($\pm 55$)</td>
<td>Monthly; 1979-2000</td>
<td>Global; $2.5^\circ \times 2.5^\circ$</td>
<td>Xie and Arkin (1997)</td>
</tr>
<tr>
<td>LAI ($m^2 m^{-2}$)</td>
<td>1.03 ($\pm 0.12$)</td>
<td>Monthly; 2001-03</td>
<td>Global; $0.05^\circ \times 0.05^\circ$</td>
<td>Lawrence and Chase (2007, 2010)</td>
</tr>
<tr>
<td>GPP ($kgC m^{-2} yr^{-1}$)</td>
<td>118 ($\pm 1.8$)</td>
<td>Monthly; 1982-2008</td>
<td>Global; $0.5^\circ \times 0.5^\circ$</td>
<td>Jung et al. (2011)</td>
</tr>
<tr>
<td>NPP ($kgC m^{-2} yr^{-1}$)</td>
<td>56 ($\pm 1.6$)</td>
<td>Annual; 2001-03</td>
<td>Global; 1 km</td>
<td>Zhao et al. (2005)</td>
</tr>
<tr>
<td>Ra ($kgC m^{-2} yr^{-1}$)</td>
<td>60</td>
<td></td>
<td></td>
<td>Chapin et al. (2002)</td>
</tr>
<tr>
<td>Rh ($kgC m^{-2} yr^{-1}$)</td>
<td>60</td>
<td></td>
<td></td>
<td>Schlesinger (1997)</td>
</tr>
</tbody>
</table>
with well-characterized uncertainties. The GPP data (Jung et al. 2011) are derived by upscaling flux tower measurements to the global scale using a machine learning technique that is limited by the uncertainties because of the effects of disturbance and lag. Jung et al. (2011) estimated the uncertainty of the globally averaged GPP to be $66 \pm 6 \text{ kgC m}^{-2} \text{ yr}^{-1}$. The NPP data (Zhao et al. 2005) are derived from MODIS products, and the data quality is affected by the uncertainties in the descriptions of biome type and meteorological input data as well as in the algorithm that translates measured parameters into inferred process rates. Zhao et al. (2005) indicated that these uncertainties may be large in some regions or during some seasons. The global annual-mean $Ra$ (Chapin et al. 2002) and $Rh$ (Schlesinger 1997) values are empirically inferred, and their uncertainties are not exactly known. In general, the uncertainty of the estimated $Ra$ is expected to be smaller than that of $Rh$ since $Ra$ is related to the amount of plant carbon that is more readily observed than the soil carbon. Based on these discussions, we subjectively assign a relatively large uncertainty of $\pm 15\%$ for GPP, NPP, Ra, and Rh in Table 3 in our model evaluations.

3. Historical simulations of carbon cycle from 1850 to 2005

While the global sum or average values of carbon cycle processes and products are useful for easily quantifying the carbon–climate interactions, as emphasized in some model intercomparison studies (e.g., Friedlingstein et al. 2006), the comparison of zonal distributions and regional averages may provide greater insight. Therefore, here we first evaluate the global and zonal averages from the eight ESMs and then assess the averages in three regions over the tropics, midlatitudes, and high latitudes, respectively.

a. Global and zonal-mean carbon budget

The latitudinal distribution of the zonal-mean $T$ over land agrees well among models (Fig. 1a). Compared with observations, $T$ is overestimated by most models.
between 10° and 30°N and underestimated in two narrow belts centered near 10°S and 35°N. The difference between the observed and modeled global average (over land between 60°S and 90°N) T is less than 1°C except INM-CM4.0 (Fig. 2a). Pr shows more variability among the models than T, particularly over the tropics, and it is overestimated in most latitude bands by most models (Fig. 1b). Most models also overestimate the global Pr (Fig. 2b). While global SM data do not exist, there are large SM variations among the models (Fig. 2c) resulting from the differences with regard to model Pr, T, and other factors. These biases in the physical climate may affect the performance of carbon cycle submodels and vice versa.

Figures 2e–h show the various components of the terrestrial carbon cycle. The simulated GPPs averaged over the last three decades range from 122 to 168 PgC yr\(^{-1}\) (Fig. 2e). Even considering the uncertainty in observation-based estimates of GPP (Beer et al. 2010; Jung et al. 2011), GPPs above 160 PgC yr\(^{-1}\) from GFDL-ESM2M and MPI-ESM-LR are overestimated. The modeled global NPPs
vary from 45 to 87 PgC yr\(^{-1}\) with the multimodel average close to the remote sensing estimate that also contains a relatively large uncertainty (Zhao et al. 2005; Fig. 2f). Systematic biases in NPP would affect the accuracy of the simulated LAI, all of which are projected as greater than observations (Lawrence and Chase 2007, 2010; Fig. 2d). The underestimate of NPP and overestimate of LAI in CCSM4 and NorESM1-M suggest that the relationship between LAI and biomass production is unrealistic and needs to be improved in these two models.

While the Ra values from five models may be comparable to the estimate of 60 PgC yr\(^{-1}\) (Chapin et al. 2002), Ra is overestimated by 30% in CCSM4, MPI-ESM-LR, and NorESM1-M (Fig. 2g). In contrast, compared with the estimate of 60 PgC yr\(^{-1}\) (Schlesinger 1997), Rh is underestimated by 25% in CCSM4 and NorESM1-M (Fig. 2h). While the amount of aboveground biomass is empirically inferred through LAI, the values of belowground biomass (e.g., inferred by using allometric relationships) and soil organic matter are more uncertain. In general, Rh remains the least constrained component of the terrestrial carbon cycle because of the less quantitative knowledge of belowground carbon and associated microorganisms (Bond-Lamberty and Thomson 2010), and a more detailed Rh evaluation of these eight ESMs using in situ observations is reported in a separate study (Shao et al. 2013, manuscript submitted to Environ. Res. Lett.).

Note that the performances of CCSM4 and NorESM1-M are similar in predicting global GPP, NPP, Ra, and Rh because they share the same land model (CLM4CN). For instance, their Rg is higher than others, which is calculated as a higher fraction of the fixed carbon \([A - R_m; Eq. (B5)]\) than other models (e.g., 30% in CLM4CN versus 10% for SEIB-DGVM and 25% for TRIFFID). For some ESMs such as MIROC-ESM, the obvious biases in productivity and respiration are mutually compensatory, leading to a reasonable apparent NPP. Thus, there is a potential problem with equifinality in the model intercomparison: several of the models may be arriving at the same estimates for net carbon exchange but for different reasons in the differences between the gross fluxes.

Globally averaged \(D\) is much smaller than the production and respiration terms in Figs. 2e–h. For example, in NorESM1-M, the annual carbon fluxes from the harvest, fire, and LUC are just 1.0, 2.0, and 0.36 PgC, respectively, while its GPP is as high as 126.3 PgC. At regional scales, however, \(D\) is an important determinant of NBP. For instance, Fig. 3 shows that most land regions are carbon sinks based on NEP in Eq. (4). When \(D\) is considered, however, some regions (e.g., East Asia) would become carbon sources based on the sign of NBP in Eq. (5). Even at global scale, \(D\) is important for the accumulated NBP (Table 4). For example, the accumulated \(D\) in GFDL-ESM2M from harvest, grazing, LUC, and fire is 280, 124, 471, and 230 PgC by the end of
Table 4. The cumulated global terrestrial NEP, NBP, and $D$ by the end of the historical (1850–2005) and RCP4.5 (2006–2100) periods, respectively. A positive (or negative) sign of NEP and NBP indicates carbon uptake from the atmosphere (or release to the atmosphere).

<table>
<thead>
<tr>
<th></th>
<th>NEP (PgC)</th>
<th>NBP (PgC)</th>
<th>$D$ (PgC)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CanESM2</td>
<td>24/100</td>
<td>18/98</td>
<td>6.5/2.2</td>
</tr>
<tr>
<td>CCSM4</td>
<td>371/444</td>
<td>−59/60</td>
<td>430/384</td>
</tr>
<tr>
<td>GFDL-ESM2M</td>
<td>978/4418</td>
<td>−120/255</td>
<td>1099/4163</td>
</tr>
<tr>
<td>HadGEM2-ES</td>
<td>50/306</td>
<td>6.4/290</td>
<td>43/16</td>
</tr>
<tr>
<td>MIROC-ESM</td>
<td>730/795</td>
<td>−57/69</td>
<td>787/726</td>
</tr>
<tr>
<td>MPI-ESM-LR</td>
<td>1730/2274</td>
<td>0.8/456</td>
<td>1729/1819</td>
</tr>
<tr>
<td>NorESM1-M</td>
<td>403/403</td>
<td>−58/22</td>
<td>461/381</td>
</tr>
<tr>
<td>INM-CM4.0</td>
<td>214/243</td>
<td>211/239</td>
<td>3.3/3.8</td>
</tr>
</tbody>
</table>

Historical period, respectively, most of which are much higher than past empirical estimates [e.g., 108–188 PgC from LUC; Houghton (2010)]. This leads to the most negative and unrealistic estimate of the cumulative total NBP from 1850 to 2005 (−120 PgC) by GFDL-ESM2M (Table 4). Note that while the accumulative NBP varies between −120 and 211 PgC for the two extreme models (GFDL-ESM2M and INM-CM4.0), the majority of the models lay between −59 and 18 PgC (Table 4).

For the zonal mean, most models overestimate GPP between 30°S and 15°N and LAI in all latitudes (Figs. 4a,b). For the band between 15° and 30°N, all models underestimate GPP but overestimate LAI, implying that the subtropical ecosystems are not well modeled in regard to the relationship between the capture of solar photons by leaves and carbon assimilation. The zonal patterns of NPP, $R_a$, and $R_h$ are similar to GPP with maxima occurring in the tropics and secondary maxima in the mid–high latitudes corresponding to the tropical rain forest and temperate/boreal forest, respectively (not shown). In contrast, there is considerable variation in the zonal-mean NBP among the ESMs, particularly in the tropics (Fig. 4c). The model-averaged NBP magnitude (black dashed line in Fig. 4c) is relatively high in the tropics, implying that these models have either a large carbon sink or source there (depending on the sign of NBP).

The NPP to GPP ratio reflects the ecosystem carbon use efficiency, and it is also directly used in some studies to infer NPP from GPP. Zhang et al. (2009) showed that this ratio increases with latitude and its global average of 0.52 does not change as much as NPP or GPP with interannual climate variability. Malhi et al. (2009) found that disturbances appear to drive a shift toward increased values in this ratio in the field. In the CMIP5 simulations, this ratio is generally lower in the tropics and higher at high latitudes, which is consistent with Zhang et al. (2009), showing that the plants tend to store carbon more efficiently with increasing latitude (Fig. 4d). This is probably because as mean temperature decreases, respiration decreases more than gross photosynthesis in all models, reflecting good solid physiological foundations.

Fig. 4. Zonal distributions of (a) GPP (kgC m$^{-2}$ yr$^{-1}$), (b) LAI (m$^2$ m$^{-2}$), (c) NBP (kgC m$^{-2}$ yr$^{-1}$), and (d) the ratio of NPP to GPP, averaged over the last 30 years in the historical experiment (color lines) compared to the observation (black solid line) in (a) and (b). In (c), the black dashed line represents the average of the absolute values of NBP across the models to highlight the latitudes with large NBP deviations from zero (i.e., large carbon sources or sinks).
in these models. NorESM1-M and CCSM4 systematically underestimate this ratio compared with the results in Fig. 2 of Zhang et al. (2009) because of their low estimation of NPP (Fig. 2f).

Most models simulate positive trends in global-averaged terrestrial productivity, respiration, and T with near-zero trends of Pr and SM over the historical period (figure not shown). Model-averaged trends of GPP, NPP, Ra, and Rh are 8.6, 5.3, 2.9, and 3.5 PgC century\(^{-1}\), respectively, with the trend of GPP comparable to the observation-based product from 1982 to 2008 (Jung et al. 2011). Most models have a positive trend of D and hence a negative trend of NBP.

To better understand the carbon cycle–climate interactions, it is useful to evaluate the temporal correlations between global annual-mean carbon and climatic variables over land in the historical simulation. Light, T, and Pr are generally regarded as the most important meteorological drivers of plant growth, and hence carbon fluxes (e.g., GPP and NPP) are expected to be positively correlated with T and Pr. Based on the formulations for carbon fluxes at individual time steps (of \(\sim 15\) min) in appendix B, GPP, NPP, Ra, and Rh would all increase with T and SM. Pr affects vegetation through SM. Therefore, it is usually assumed that these fluxes would be positively correlated with T, Pr, and SM, but few studies have actually evaluated such correlations. Indeed, Fig. 5a shows that the median of the correlations of GPP, NPP, Ra, and Rh with T and Pr from the eight models are positive and statistically significant at the \(p = 0.01\) level. However, the median correlations of these quantities with SM are close to zero or even negative. Furthermore, while the correlations for individual models are usually positive, correlations (e.g., between Rh and SM and between GPP and T) are negative and significant (at the \(p = 0.01\) level) for some models, which is contrary to our expectations. The median correlation between LAI and Pr is positive and significant at the \(p = 0.01\) level, but it is close to zero between LAI and T (or SM). Most correlation coefficients between NBP and climatic variables, especially those exhibiting the highest values, have opposite signs from those between D and climatic variables because of the constraint of Eq. (5).

**b. Regional carbon budget**

To further evaluate climate–carbon cycle interactions, we looked at three regions over the tropics, midlatitude, and high latitude in more detail: Amazonia (0°–10° S, 50°–70° W; tropical rain forest), the western United States (30°–50° N, 130°–105° W; semiarid), and eastern Siberia (50°–66.5° N, 90°–140° E; northern temperate/boreal forest).

Amazonia has the highest productivity and respiration among the three regions, indicating the most favorable environmental conditions for biological activities (highest mean T, Pr, and SM) (Fig. 6; Table 5). The median T and GPP among the eight models agree well with observations, while the median Pr (or LAI) is lower (or higher) than observations (Table 5). The term D differs substantially among the models (Fig. 6j) and plays an important role in determining whether Amazonia is a carbon sink or source (positive or negative NBP; Fig. 6i) in the simulations. The results from CanESM2 are particularly poor: its overestimate of T and substantial underestimate of Pr lead to underestimates of productivity and respiration (Figs. 6d–h). More analysis is also needed to understand why SM in MPI-ESM-LR is much lower than in other models (Fig. 6c), while other quantities from this model are overall consistent with other models.

In eastern Siberia, the median T among the eight models agrees well with the observational value, while the median Pr, GPP, and LAI are higher than observations (Table 5). In particular, GFDL-ESM2M shows a high positive bias in GPP and LAI and has a large NPP and Rh relative to other models, probably a result of its representation of nearly 100% tree cover in this region. The spread of the predicted vegetation cover among the ESMs (between ~10% and ~100%) is the largest here in the three regions (not shown). This broad range in represented vegetation cover may be caused by the
differences in simulated $T$ and Pr. The most striking difference among the models caused by the physical climate appears in SM that ranges from 188 to 1224 kg m$^{-2}$ (Table 5). It is partly related to the different treatment of soil ice versus soil water and the inclusion or exclusion of soil ice in the soil moisture output among the models.

In the western United States, the median $T$ and GPP among the models agree well with the observation-based product, while the median Pr and LAI are higher than observations (Table 5). The range of $D$ between 0 and 181 gC m$^{-2}$ yr$^{-1}$ is greatest among the three regions. Overall, the regional differences in the climatic and terrestrial carbon cycle variables among the eight models and between models and observations are greater than global differences, and the models performing well in globally averaged variables from the historical simulation may have poor performances reproducing regional variables and vice versa.

**FIG. 6.** Climatological annual averages over Amazonia (0°–10°S, 50°–70°W). (a) $T$ (°C), (b) Pr (mm), (c) SM (kg m$^{-2}$), (d) LAI (m$^2$ m$^{-2}$), (e) GPP (kgC m$^{-2}$ yr$^{-1}$), (f) NPP (kgC m$^{-2}$ yr$^{-1}$), (g) Ra (kgC m$^{-2}$ yr$^{-1}$), (h) Rh (kgC m$^{-2}$ yr$^{-1}$), (i) NBP (gC m$^{-2}$ yr$^{-1}$), and (j) $D$ (gC m$^{-2}$ yr$^{-1}$). The model results are averaged over the last 30 years in the historical and RCP4.5 experiments. The horizontal lines represent the observations in Table 3. Note that the unit of NBP and $D$ differs from that of GPP, NPP, Ra, and Rh by a factor of 1000.
The trends in each region are also analyzed but figures are not shown here for brevity. Temperature has a positive trend over all regions in each model, while Pr show negative trends over the western United States and Amazonia in some models. Accordingly, most models show positive trends in GPP, NPP, Ra, Rh, and LAI that are generally the greatest over Amazonia. The NBP trend is positive in half of the models over Amazonia, while it is positive in most ESMs in other regions. The Pr trend is positive in most models in all regions.

4. Twenty-first-century projections of carbon cycle

Under RCP4.5, the terrestrial annual T and Pr are projected to increase by 1.7°C-3.8°C and 10.4-57.1 mm, and the modeled SM changes are projected to increase from −32.7 to 2.8 kg m⁻² by the end of twenty-first century (Figs. 2a–c). The model-averaged trend of T is 2.5°C ± 0.84°C century⁻¹ compared to 0.6°C ± 0.19°C century⁻¹ in the historical experiment. It is also higher than the ensemble average using different CMIP5 models (2.0°C) (Good et al. 2013). The model-averaged Pr trend changes from negative (~2.7 ± 12.1 mm century⁻¹) in the historical simulation to positive (34.4 ± 17.6 mm century⁻¹) in the twenty-first-century simulation. The increased T and Pr along with the prescribed CO₂ concentration evolution in the twenty-first century under RCP4.5 lead to significant jumps in productivity and respiration in all models (Figs. 2d–h). The fertilization effect of enhanced CO₂ concentration is inherent in the photosynthesis calculation [Eq. (B1)], and the moderate increase in global-averaged LAI (Fig. 2d) also agrees with field studies for the elevation of CO₂ concentration under open-air conditions (Long et al. 2004).

Six ESMs project a negative NBP trend (reduction of land carbon uptake) in contrast to the mixed trends in the historical period (positive in three ESMs, negative in two ESMs, and nearly zero in three ESMs). While NBP varies across different regions, a pronounced disagreement on NBP among the ESMs occurs in tropical regions (e.g., Amazonia) where the model results show a decrease in both Pr and SM and an increase in T (Fig. 6). The global cumulative NBP is positive in all ESMs with substantial variation across models from 22 to 456 PgC by 2100, in contrast to half of the ESMs showing negative NBP accumulations by 2005 (Fig. 7b and Table 4). Note that while all the models show a terrestrial carbon sink in the twenty-first century, there is no feedback to CO₂ concentration in the atmosphere in these concentration-driven simulations. It will be a future task to compare these results versus those from emission-driven simulations in which atmospheric CO₂ concentration is affected by land and ocean processes.

The correlations between the carbon and climatic variables change in the RCP4.5 simulation from those in the historical simulation (Fig. 5). The median correlations across the eight models between most carbon cycle variables (GPP, NPP, Ra, Rh, and LAI) and T (or Pr) become stronger. The range of these correlations also become much smaller, and all these correlations are significant at p = 0.01 level. The median correlations of NBP with T and Pr remain nearly the same in the historical and RCP4.5 simulations. The median correlation between D and T is significantly decreased (at p = 0.01 level) in the twenty-first century, but it remains nearly the same between D and Pr. Compared with these changes, the change of the median correlations between the carbon cycle variables and SM from the historical to the RCP4.5 simulations is overall smaller, while the range of correlations among models becomes much larger for all variables except NBP. These results indicate that, with the dominant effect of global warming
(relative to the small change in Pr and SM) in the twenty-first century, the correlations between carbon flux variables and $T$ (or Pr) become consistent with our expectations. However, the large discrepancies in the correlations between carbon flux variables and SM in these models are surprising, as similar model formulations for carbon fluxes in appendix B are used. This may reflect the important impact of the model treatment of the hydrological cycle on the carbon cycle.

The zonal means of GPP, NPP, Ra, Rh, and LAI in the RCP4.5 simulation increase more in the tropics than in boreal latitudes compared with the historical simulation (figure not shown), consistent with Raddatz et al. (2007). There is a negligible change in the NPP to GPP ratio in all models (Fig. 8d), which is consistent with the conclusion from field studies that this ratio is nearly constant across a range of CO$_2$ concentration and $T$ for herbaceous and woody plants (e.g., Tjoelker et al. 1999; Chen et al. 2000). The effect of increased $T$ (Fig. 8a) in the twenty-first century on this ratio is partly counteracted by the increase of Pr (Fig. 8b) because this ratio generally decreases with Pr (Zhang et al. 2009).

The changes are more complicated and inconsistent among the models in zonal NBP (Figs. 4c and 7a), particularly in the tropics where the impact of Pr appears to be more important than that of $T$ (Zeng et al. 2005). Again, the differences in $D$ (not shown) contribute to the different signs of NBP among the models in the tropics (Fig. 7a) as we noted in the historical simulation (Fig. 3). Both the tropical and boreal latitudes represent major sinks based on accumulated NBP in the twenty-first century (Fig. 7b).

Regional analyses indicate that over Amazonia $T$ increases in all models, but Pr and SM decrease in most models in the twenty-first century. Accordingly, $D$ increases and NBP decreases in most models (Fig. 6). In contrast, NBP increases in the twenty-first century over eastern Siberia. This could be attributed to the increase of $T$ and Pr since conifers are physiologically less sensitive to increases in CO$_2$ concentration (Long et al. 2004).
Analyses of the temporal correlations of annual average carbon cycle variables with climatic variables over Amazonia show that the median correlations of GPP, NPP, Ra, Rh, and LAI with Pr and SM are stronger than those with $T$ in most models in the historical simulation, indicating the stronger control of Pr and SM on these carbon cycle variables in these models. In contrast, the median correlation between NBP and $T$ is negative, smallest (among the median correlations between carbon cycle and climatic variables), and significant at the $p = 0.01$ level, consistent with the results in Figs. 6a and 6i. These correlations do not change much in the twenty-first century. Over eastern Siberia, $T$ has a stronger control of carbon cycle variables, and the median correlations of GPP, NPP, Ra, Rh, and LAI with $T$ are stronger than those with Pr or SM in both the historical and twenty-first-century simulations.

What is surprising in these correlation analyses is the substantial range of correlations. For instance, the correlations between NPP and $T$ over Amazonia vary from $-0.89$ to $0.63$ (or from $-0.89$ to $0.83$) among the eight models in the historical (or twenty-first century) simulation. Over eastern Siberia, the correlations between NPP and SM vary from $-0.40$ to $0.61$ (or from $-0.83$ to $0.85$) among the eight models in the historical simulation (or twenty-first century). More detailed analyses are needed for individual models to understand such large disparities in correlations.

Besides the above analyses of global, zonal, and regional mean variables, we have analyzed variables in individual grid cells. Figure 9 shows that the eight ESMs are quite diverse in predicting the gridded NBP trends. Compared with the historical runs, the histograms of most models under RCP4.5 slightly shift to the negative trends, increasing the number of grid cells in an elevated CO$_2$ atmosphere and with warmer wetter conditions to become carbon sources. In contrast, CanESM2 predicts more grid cells with increased trends in the capacity of storing carbon. The tails in the RCP4.5 histograms are more pronounced than during the historical period, indicating the growing importance of extreme cases (i.e., grid cells as large carbon sinks or sources).

As mentioned earlier, besides RCP4.5, there are three other scenarios (RCP2.6, RCP6, and RCP8.5; Moss et al. 2010). To preliminarily explore the sensitivity of our conclusions to the scenarios, we have analyzed the CCSM4 results under RCP8.5 (with the total radiative forcing stabilized at 8.5 W m$^{-2}$ roughly before 2100) and found that the results from the historical to RCP4.5 simulations are overall consistent with those from the RCP4.5 to RCP8.5 simulations. For instance, the ratio of NPP to GPP changes little ($<0.03$) over all latitudes, the tails in the RCP8.5 histograms become more pronounced than in the RCP4.5, and correlations between the carbon cycle and climate variables increase from the historical to RCP4.5 and from the RCP4.5 to RCP8.5 simulations, except those between $D$ and $T$ and between NBP and $T$, SM, or Pr.

5. Conclusions and further discussion

Eight CMIP5 ESMs have been systematically evaluated here with a focus on the terrestrial carbon cycle in the historical period (1850–2005) and twenty-first century
In contrast to previous model intercomparison studies primarily focusing on the net carbon exchange (e.g., Friedlingstein et al. 2006), we evaluated both the net CO$_2$ flux and its components, and their relations with climate, and how these relations will change in the future at both the global and regional scales.

The CMIP5 historical simulations tend to overestimate global GPP, Ra, and LAI against reference data. Large intermodel disagreements exist in NPP, Rh, disturbance ($D$), and the cumulative NBP over the historical period. Without considering $D$, the terrestrial biosphere is a net carbon sink over most regions in the twentieth century, but it becomes a net source in many regions after considering $D$, demonstrating the critical importance of $D$ (which is less understood and only roughly represented in ESMs) in the land–atmosphere carbon exchange. While GPP and LAI are strongly dependent on latitudes with the peak over the tropics, the NPP to GPP ratio, representing the ecosystem carbon use efficiency, is less latitude dependent among all the models, consistent with previous observational analysis.
While the globally averaged temperature \((T)\) has a positive trend in all historical simulations, the signs of the precipitation (Pr) and soil moisture (SM) trends are model dependent. The GPP, NPP, Ra, Rh, and \(D\) trends are positive in most models, but the LAI and NBP trends are model dependent. These results are largely related to the temporal correlations of the annual carbon cycle variables with climate variables \((T, \text{Pr}, \text{and SM})\). For instance, the median of the correlations of GPP, NPP, Ra, and Rh with \(T\) and Pr from the eight models are positive and statistically significant at the \(p = 0.01\) level. In contrast, the median correlation between LAI (or NBP) and \(T\) is relatively close to zero and not statistically significant.

Over the three regions (Amazonia, western United States, and eastern Siberia), model results differ significantly in all carbon cycle and climatic variables. The magnitude of \(D\) is usually comparable to, or greater than, NBP in the historical simulation. The NBP trend is positive over eastern Siberia and the western United States in most models, but its sign is more model dependent over Amazonia.

In the twenty-first century under RCP4.5, all the ESMs indicate that the terrestrial biosphere will benefit (with increases in GPP, NPP, Ra, Rh, and LAI) from climate change and prescribed rise in CO2 concentration, which is consistent with the C4MIP result that the total effect of climate change on GPP, Ra, and Rh is positive in the twenty-first century (Gregory et al. 2009). In contrast, the NPP to GPP ratio does not change much, even if the nitrogen limitations are taken into account. The trend of climatic variables \((T, \text{Pr}, \text{and SM})\) and most carbon cycle variables (GPP, NPP, Ra, Rh, and LAI) is accelerated in the twenty-first century. While the cumulative NBP is positive in half of the models in the historical period, it is positive in all models in the twenty-first century. Both the tropics and boreal latitudes would become major carbon sinks. For the period of 2071–2100, it becomes negative in Amazonia but remains positive in eastern Siberia, suggesting the increased importance of boreal forests as carbon sinks (relative to tropical forests) in the twenty-first century. This is related to the increase of both \(T\) and Pr over eastern Siberia in contrast to the increase of \(T\) but decrease of Pr over Amazonia.

While model results differ, their median globally averaged carbon cycle variables (GPP, NPP, Ra, and Rh) agree reasonably well with available observation-based estimates. Based on these median values, we can synthesize the globally averaged terrestrial carbon cycle in the last 30 years of this historical period as (all in PgC yr\(^{-1}\)) \(\text{GPP} = 129 (127, 136), \text{NPP} = 62 (60, 71), \text{Ra} = 67 (64, 80), \text{Rh} = 60 (54, 61), \text{D} = 3.46 (0.40, 6.75), \text{NEP} = 3.27 (2.35, 6.71), \text{and NBP} = 0.30 (−0.06, 0.98), where the two values in parentheses represent the 29th and 71st percentiles (computed from the third and sixth ranked model results among the eight models). The changes in the last 30 years between the RCP4.5 and historical simulation would be \(\Delta \text{GPP} = 27.3 (19.1, 48.6), \Delta \text{NPP} = 10.7 (4.4, 29.4), \Delta \text{Ra} = 14.5 (10.8, 18.2), \Delta \text{Rh} = 11.6 (3.2, 25.2), \Delta \text{D} = 0.41 (−3.38, 4.22), \Delta \text{NEP} = 1.11 (−0.92, 3.21), \text{and NBP} = 0.88 (−0.18, 2.04), respectively.

While CO2 concentration is prescribed in the experiments we analyzed, CMIP5 does include experiments that incorporate the direct carbon–climate interaction (i.e., CO2 concentration is modified by terrestrial and ocean carbon cycle processes with prescribed anthropogenic...
CO₂ fluxes). Thus, further work will be required to investigate climate–carbon cycle interactions in these experiments. While the spread and uncertainties in the carbon cycle among the eight ESMs are demonstrated, the relevant mechanisms are not yet provided or quantified in this study. Further investigation is needed to explore such mechanisms with specific models (e.g., at the regional scale using flux tower observations).

Note that even the most complex CMIP5 ESMs do not incorporate all the diversity of the biogeochemical processes. For example, phosphorus limitation (Zhang et al. 2011) and mycorrhizal contributions to soil respiration (Heinemeyer et al. 2007) are not yet represented or less well represented than other processes. This omission is especially relevant to projections from wet tropical ecosystems. Methane production represents another important biogeochemical omission that is loosely coupled to the carbon cycle but is not commonly represented in the ESMs. Further, while CCSM4 and NorESM1-M consider the nitrogen cycle, which is closely coupled to the carbon cycle, it is not included in the other models evaluated here.

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APPENDIX A

Brief Description of CMIP5 Models

The second generation CanESM2 combines the atmospheric and ocean components of the fourth generation Canadian Coupled General Circulation Model (CanCM4; Chylek et al. 2011) and CLASS (Verseghy 1991; Verseghy et al. 1993) with three soil layers. The terrestrial carbon cycle involves dynamic vegetation and linear land cover transition models based on CTEM (Arora and Boer 2010). A simple crop is included, but the pasture is not considered.

The CCSM4 (Gent et al. 2011) and the NorESM1-M (Tjiputra et al. 2012; Zhang et al. 2012) share the Community Atmosphere Model, version 4 (CAM4; Neale et al. 2010) and CLM4CN (Oleson et al. 2010; Lawrence et al. 2011). However, NorESM1-M adopts CAM4-Oslo (a revised version of CAM4 with different representations of aerosols, aerosol–radiation, and aerosol–cloud interactions) and couples it with a hybrid representation derived from the modified physical ocean model (MICOM; Bleck et al. 1992) and the ocean biogeochemical model [Hamburg Ocean Carbon Cycle (HAMOCC) model] adopted for isopycnic coordinates (Tjiputra et al. 2010). CLM4CN borrows its biogeochemical component from the terrestrial biogeochemistry model Biome-BGC (Thornton and Rosenbloom 2005; Lawrence et al. 2011) along with accommodation for regional wildfire cycle (Thonicke et al. 2001). Vegetation profiles such as LAI, vegetation heights, and vegetation phenology are all computed for 15 PFTs. Global transient land use and land cover are prescribed on an annual basis (Lawrence et al. 2012).

The atmospheric component of the GFDL-ESM2M (Dunne et al. 2012) is virtually identical to that in the Geophysical Fluid Dynamics Laboratory (GFDL)'s previous Climate Model, version 2.1 (CM 2.1), but the ocean component in CM2.1 is updated with new oceanic ecology and biogeochemistry components. The land component LM3V (Shevlakova et al. 2009) is specifically designed to contain the land use and management such as cropland and pasture dynamics, shifting cultivation, and secondary regrowth. Each natural vegetation type has five daily modified carbon pools. The carbon lost due to natural processes and land use conversions is deposited into two independent soil carbon pools.

HadGEM2-ES has its atmospheric and oceanic components derived from HadGEM1 (Johns et al. 2006). The land surface scheme is from JULES based on MOSES II and has four soil layers to 3-m depth (Blyth et al. 2006; Collins et al. 2011). The terrestrial carbon cycle consists of the TRIFFID dynamic vegetation model with the Rothamsted Carbon (RothC) soil carbon model (Cox 2001). A time-varying land use mask is applied to prevent natural plant growth in agricultural regions, thus concurrently simulating the five natural PFTs and crop dynamics.

The INM-CM4.0 (INM-RAS 2012) from Russia consists of the general atmospheric and oceanic circulation models and a land surface model with improved soil and vegetation processes from the previous version INM-CM3.0 (Volodin et al. 2010). The model is capable of simulating interactions between the contributions of
CO₂ and methane sources on atmospheric composition, including the land carbon flux due to LUC (deforestation and soil erosion; Volodin 2006).

MIROC-ESM (Watanabe et al. 2011) from Japan is based on the global climate Model for Interdisciplinary Research on Climate (MIROC) coupled GCM consisting of an atmospheric general circulation model (AGCM), an ocean GCM with an inclusive sea ice component [Center for Climate System Research (CCSR) Ocean Component Model (COCO)], and a land surface model (MATSIRO) with six layers of soil to a depth of 14 m (Takata et al. 2003). SEIB-DGVM is employed to simulate the changes of 11 tree PFTs and 2 grass PFTs. It contains two soil organic carbon pools (fast and slow decomposing) based on the RothC scheme as HadGEM2-ES. Carbon in harvested biomass is transferred into carbon pools, following HadGEM2-ES.

The MPI-ESM-LR consists of the atmospheric general circulation model (ECHAM6), ocean model [Max Planck Institute Ocean Model (MPIOM)], and modular land surface scheme JSBACH (Raddatz et al. 2007; Thum et al. 2011). JSBACH contains a carbon flow submodel, Cbalance, which considers 13 PFTs including crops and pastures. Plant carbon is divided into three pools: woody biomass, actively metabolizing plant tissues such as fine roots and leaves, and a reserve pool for starch and sugars. Its LUC can be either prescribed by sequences of land use maps or derived from the land use protocol.

**APPENDIX B**

**Formulations for Carbon Fluxes**

Leaf photosynthesis ($\dot{A}$) is calculated following Farquhar et al. (1980) for C3 plants and Collatz et al. (1992) for C4 plants as a function of $T$, solar radiation, turbulence, and nutrients (e.g., nitrogen):

$$\frac{d\dot{A}}{dt} = \begin{cases} \min \left[ \frac{V_{\text{max}}(c_i - \lambda)}{c_i + K_c(1 + o_i/K_o)} 4.6(c_i - \lambda)\text{PAR} \times \text{QE} \right] & \text{for C3} \\ \min \left[ V_{\text{max}} 4.6\text{PAR} \times \text{QE} c_i V_{\text{max}} \frac{c_i}{P} \right] & \text{for C4} \end{cases}$$

where $\text{QE}$ is the PFT-dependent quantum efficiency (mol mol$^{-1}$), $\text{PAR}$ is the absorbed photosynthetically active radiation (W m$^{-2}$), $P$ is the atmospheric pressure (Pa), and $c_i$ and $o_i$ are the internal leaf CO₂ and O₂ partial pressure (Pa), respectively. Parameters $K_c$ and $K_o$ (Michaelis–Menten constants for CO₂ and O₂, respectively) increase with $T$ according to the $Q_{10}$ function as well as $\lambda$ (the CO₂ compensation point) that is proportional to the ratio of $K_c$ over $K_o$. The coefficient $c_1$ is positive. The parameter $V_{\text{max}}$ is the maximum rate of carboxylation (μmol m$^{-2}$ s$^{-1}$), which, for example in the National Center for Atmospheric Research (NCAR) CLM4, is formulated to vary with leaf temperature ($T_v$), SM, and daylength (DL) and further scaled by nitrogen (N) limitation:

$$V_{\text{max}} \approx \frac{1.09T^{f_1}(\text{SM})f_2(\text{N})f_3(\text{DL})}{1 + c_2e^{c_3(T_v-273.15)}}$$

where $f_1$, $f_2$, and $f_3$ are functions ranging from one to near zero, and $c_2$ and $c_3$ are positive and negative coefficients, respectively.

The interplay between the assimilation rate (i.e., leaf photosynthesis) ($\dot{A}$) and SC (μmol m$^{-2}$ s$^{-1}$) that is needed for the water vapor and CO₂ fluxes is generally taken from the Ball–Berry conductance formulation (Collatz et al. 1991; Sellers et al. 1996):

$$\text{SC} \approx \frac{A}{c_4 r_e P}$$

where $c_4$ is the CO₂ partial pressure at the leaf surface (Pa), and $r_e$ is the ratio of the vapor pressure at the leaf surface to the saturation vapor pressure inside the leaf. Equation (B1) indicates that CO₂ enrichment (i.e., increasing $c_4$) would enhance $A$ and hence increase SC in Eq. (B3). The leaf-level biochemical model is commonly incorporated into the vegetation model in climate system modeling, scaling from leaf to canopies and landscapes with well-defined canopy properties.

Ra is split into maintenance and growth respiration (Rm and Rg). As an example, in the NCAR CLM4, Rm is estimated as a function of $T$ and tissue nitrogen concentration following Ryan (1991) [Eq. (B4), where $c_4$ is a positive coefficient and $C_{\text{pool}}$ is tissue carbon content], and Rg is assumed to be proportional to the total new growth on a given time step [Eq. (B5)]:

$$\text{Rm} \propto e^{-|c_4/(T-227.13)|C_{\text{pool}}f(N)}$$

and

$$\text{Rg} \propto A - \text{Rm}(A \geq \text{Rm})$$

If $A < \text{Rm}$ (e.g., at night or under conditions of low light or drought stress), then all of the current $A$ directed
toward satisfying $R_m$ and $R_g$ is taken as zero. Some models assume that $R_m$ is linearly related to the nitrogen content of living tissue and raised to the power of $T$. Thus, low $T$ generally curtails $R_a$.

$R_h$ represents the decomposition of soil carbon from dead plant tissues. In SEIB-DGVM, as an example, it varies with a pool-specific fixed turnover rate ($k_a$) that is regulated exponentially by soil temperature ($T_s$) and linearly by SM as follows:

$$Rh \propto k_a e^{-c_s(T_s+46.02)}\left(0.25 + 0.75 \frac{SM}{W \times Depth}\right), \quad (B6)$$

where $c_s$, $W$, and $Depth$ represent a positive coefficient, soil moisture at saturation point, and soil layer depth, respectively. Based on Eq. (B6), $Rh$ would increase with the increase of SM and $T$.

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