Damage of Land Biosphere due to Intense Warming by 1000-Fold Rapid Increase in Atmospheric Methane: Estimation with a Climate–Carbon Cycle Model

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ABSTRACT
Decadal-time-scale responses of climate and the global carbon cycle to warming associated with rapid increases in atmospheric methane from a massive methane release from marine sedimentary methane hydrates are investigated with a coupled climate–carbon cycle model. A 1000-fold methane increase (from <1 to 1000 ppmv) causes surface air temperatures to increase with a global warming of >6°C within 80 yr. The amount of carbon stored in the land biosphere decreases by >25%. This is mostly due to a large decrease in tropical net primary production during the first few years (~40%), which is caused by a decrease in photosynthesis and an increase in plant maintenance respiration with the early warming of ~3°C, leading to tropical forest dieback (>20%) and the largest decrease in vegetation carbon of >50% (~80% of the decrease in global vegetation carbon). The decrease in global land carbon is also partly due to forest diebacks (mainly boreal forest dieback by heat stress) at northern middle latitudes. In contrast, vegetation increases by >50% at northern high latitudes because of the amelioration to warm and wet conditions. Sensitivity experiments show that the warming of >6°C consists mainly of >5°C by the 1000-fold atmospheric methane and an additional increase of 1°C by the atmospheric CO2 increase due to the land CO2 release and that the CO2 fertilization of land plants prevents further warming of 1°C by limiting the atmospheric CO2 increase. The large decrease in land biomass estimated in this study suggests a critical situation for the land ecosystem or agricultural production, especially in the tropics. Because global methane content of marine methane hydrates is estimated at ~10 000 Gt, more intense warming leading to greater damage to the land biomass than the authors’ experiment (~2000 Gt) is possible in the future methane release event that would be caused by the ongoing anthropogenic warming.

1. Introduction
Methane (CH4) is a greenhouse gas whose global warming potential is larger than carbon dioxide (CO2) by an order of magnitude per the same amount of carbon (Houghton et al. 2001). In the present-day atmosphere, the methane concentration is about 1.8 ppm by volume (ppmv), which is much lower than the CO2 concentration of about 390 ppmv (World Meteorological Organization 2009), and thus the greenhouse effect of CO2 is dominant. However, methane is considered to have played an important role in several thermal maxima of the paleoclimate. For example, paleoclimatic records suggest that, during the transition from the Paleocene to the Eocene around 55 megannum (Ma: million years before present), massive methane release from destabilization of marine sedimentary methane hydrates caused a rapid global warming of more than several degrees known as the Paleocene–Eocene thermal maximum (PETM) (e.g., Dickens et al. 1995, 1997; Matsumoto 1995; Zachos et al. 2003). The warming of the PETM is considered to be due to intense greenhouse effects of large amounts of methane and CO2 produced from oxidation of the methane. The amount of methane released into the atmosphere during the PETM is estimated at more than 2000 gigatons of carbon (GtC) from the large negative δ13C anomalies (Dickens et al. 1995; Matsumoto 1995). This amount of carbon is about 1000 times the present-day (preindustrial) atmospheric methane and is comparable to fossil fuel CO2 emissions projected for a few hundred years into the future.

Although CO2 is regarded as the main cause of the ongoing global warming (Forster et al. 2007), it is possible that methane-induced warming analogous to the PETM might also occur in the future. With future global warming, a large amount of marine sedimentary methane...
hydrates on continental slopes between \(~(500–1000)-m\) depths may be destabilized by water temperature increases and rapidly released as methane into the atmosphere to enhance global warming. Because global estimates of the CH$_4$ content of marine methane hydrates range from \(~500\) to \(~24000\) Gt (Kennett et al. 2003), more intense warming than the PETM is possible.

When atmospheric methane rapidly increases, the heat by the intense greenhouse effect would cause a decrease in photosynthesis of terrestrial vegetation from thermal damage of plant cells or thermally induced drought conditions and an increase in decomposition of soil organic matter, leading to a decrease in carbon (wane) of the land biosphere. On the longer time scale, however, the high concentration of atmospheric CO$_2$ produced from oxidation of methane may cause land biomass to increase from CO$_2$ fertilization (Beerling 2000). Part of the newly produced CO$_2$ will be absorbed into the ocean and acidify the seawater, which would damage the ocean ecosystem, especially calcite shell-producing organisms. These changes in the biosphere will feedback to further influence atmospheric CO$_2$ concentration and climate. The methane-induced changes in the Earth system can qualitatively be understood as described above but have not yet been quantified.

For the study of changes in the Earth system, a numerical model that simulates the coupling of atmosphere–ocean general circulation, atmospheric chemistry, and the carbon cycle is useful because of the consistent interaction between geophysical and biogeochemical processes. With regard to a massive methane release event, although atmosphere–ocean models (Schmidt and Shindell 2003; Renssen et al. 2004) and a carbon cycle model (Zeebe et al. 2009) have partly captured the features of the PETM in terms of the atmosphere–surface ocean warming and the marine sedimentary calcite dissolution, respectively, feedbacks between climate and the global carbon cycle have not yet been identified by an Earth system model. The long time scale of the event (\(>1000\) yr; Zachos et al. 2003, 2005) requires large numerical resources and makes the study by the complicated model difficult.

In this study, as a preliminary stage for a more comprehensive understanding of the Earth system response to a massive methane release, the simulated response of a climate and the global carbon cycle on the time scale of several decades is investigated by using a coupled climate–carbon cycle model and instantaneously increasing atmospheric methane concentration to an extremely high level (\(1000\) ppmv: about a 1000-fold increase from the present level). The model is based on the general circulation model (GCM) of the Meteorological Research Institute (MRI). The prescribed methane concentration refers to the released carbon amount estimated for the PETM (\(~2000\) GtC; Dickens et al. 1995; Matsumoto 1995). Although methane has been released during the respective epoch in three separate pulses, each corresponding to cumulative emission of several hundred GtC and separated in time by intervals of about 10 000 yr (Bains et al. 1999), this study simulates the methane release in a single pulse and is thus not a simulation of the PETM.

Because the model used here does not include atmospheric chemistry or marine sedimentary calcite, changes in atmospheric methane and CO$_2$ through oxidation of methane or long-term carbon cycle responses cannot be investigated. However the model includes the climate–carbon cycle response mechanism on the short time scale (\(<100\) yr) such as an enhanced warming by land biomass CO$_2$ release or ocean CO$_2$ solubility decrease due to warming (Friedlingstein et al. 2006). Besides, in case of the massive methane release, the high atmospheric methane concentration can be maintained longer through the exhaustion of hydroxyl (OH) radical in the methane oxidation (e.g., Sze 1977; Schmidt and Shindell 2003). Thus, the model is expected to be able to investigate the early period of the massive methane release event. Although it is an extreme case that atmospheric methane is instantaneously increased to \(1000\) ppmv and that oxidation of the methane and the resultant CO$_2$ production in the atmosphere are not considered, the purpose of this study is, as a first study of this kind, to explore the temporally, quantitatively extreme region of the climate–carbon cycle variability in the methane release event and to investigate climate–carbon cycle changes caused by greenhouse effect of the methane increase alone. Such an extreme case is possible because of the ongoing anthropogenic global warming and the massive storage of marine methane hydrates, and thus must be investigated for the further understanding of the future Earth system variability.

In addition to the experiment of \(1000\)-ppmv methane, sensitivity experiments are conducted to examine the relative importance of atmospheric methane concentration, CO$_2$ radiative forcing, and CO$_2$ fertilization of land plants on the climate–carbon cycle system.

2. Methods

a. Model

The numerical model employed here is a coupled climate–carbon cycle model in which land and ocean carbon cycle models are coupled to a global atmosphere–ocean general circulation model (AOGCM). They were
developed at the MRI. The model is improved from the previous version (Obata 2007), especially for the land carbon cycle, and is described briefly below.

The global AOGCM is the MRI Coupled General Circulation Model, version 2 (CGCM2; Yukimoto and Kitamura 2003). The atmospheric component (AGCM) is a spectral model with a horizontal resolution of T42, approximately 2.8° × 2.8° transformed grid, and 30 vertical levels with the top at 0.4 hPa. The longwave radiation is calculated by a multiparameter random model formulated by Shibata and Aoki (1989), which explicitly addresses absorption by methane (CH4) and nitrous oxide (N2O) in addition to water vapor (H2O), carbon dioxide (CO2), and ozone (O3). The oceanic component is a Bryan–Cox type of ocean general circulation model (OGCM; Bryan 1969). The horizontal resolution is 2.0° latitude (ranging from 0.5° around the equator between 4°S and 4°N to 2.0° at 12° latitude) and 2.5° longitude. The vertical resolution is 23 levels, and 13 of these levels are placed between the surface and the 300-m depth. The diapycnal diffusion coefficient increases from 0.3 × 10−4 m2 s−1 at the surface to 1.0 × 10−4 m2 s−1 at the depth of 5000 m (the maximum depth in the OGCM), which is a modified point from the version of Yukimoto and Kitamura (2003) (0.1 × 10−4 m2 s−1 for all the depths) to reproduce more realistic vertical stratifications and thermohaline circulations as suggested by Tsujino et al. (2000).

The land carbon cycle model follows, on the organism–leaf level, models of the biochemical photosynthesis processes dependent on the enzyme and stomatal conductance (Woodward et al. 1995; Sellers et al. 1996). On the ecosystem–biogeochemical level, the model follows the Lund–Potsdam–Jena dynamic global vegetation model (LPJ-DGVM; Sitch et al. 2003) in which the ecosystem is divided into several components [leaves, woody parts, and roots of plant functional types (PFTs); litter; and humus] is driven by net primary production (NPP; CO2 absorption from the atmosphere; photosynthesis from the organism-leaf model minus plant respiration) partitioned among vegetation components, carbon flow between components, and soil respiration (CO2 emission into the atmosphere) in each terrestrial grid area of the AOGCM. The biochemical photosynthesis processes are fundamental improvements on the previous model (Obata 2007) in which NPP is empirically calculated by temperature, precipitation, and atmospheric CO2 concentration. The distributions of PFTs are allowed to change across the earth’s surface as environmental conditions change in the simulations. Regarding the exchange of heat, water, and momentum between the land and atmosphere, the simple biosphere model of Sato et al. (1989) is used with the parameters of the time-varying dominant PFT in each grid area. The present perennial ice distributions of Greenland and Antarctica are prescribed and remain the same during the time integration. Thus, these areas are not included in simulations of the global carbon cycle.

The ocean carbon cycle model represents ocean biogeochemical processes related to CO2, such as sea–air CO2 exchange, seawater carbonate chemistry, surface biological export production driven by light and nutrient (phosphate), and dissolution of the produced matters in deep waters, in addition to advection and diffusion of the dissolved biogeochemical elements by the OGCM. The details and the reproducibility of climatological fields [e.g., the deep waters become oxygen poorer and older in age (lower δ13C and Δ14C), containing more dissolved inorganic carbon (DIC), alkalinity, and phosphate from the North Atlantic to the North Pacific] and time variations are described by Obata and Kitamura (2003) and Obata (2007). Riverine flux, sediments, and dissolved organic matter cycle are not included in the model.

b. Experiments of rapid increases in atmospheric methane

In this study, the initial condition of the experiment of atmospheric methane increase is the preindustrial state of the climate–carbon cycle model, which is calculated as follows: The ocean component of the model is integrated for 10 000 yr by the surface forcings of climatological temperature, salinity (Levitus 1982), and wind stress (Hellerman and Rosenstein 1983) from a homogeneous state including globally averaged observational concentrations of the biogeochemical tracers. The ocean component is coupled to the atmospheric component (AGCM), and they are integrated for 200 yr to reach a steady state. The land carbon cycle component is integrated for 5000 yr by the forcing of this AOGCM steady state. All components are finally coupled together and integrated for 200 yr to reach a realistic preindustrial steady state in which the global-mean surface air temperature is about 14°C, the atmospheric CO2 concentration is about 290 ppmv (=610 GtC), and the intensity of the North Atlantic thermohaline circulation is 17 Sv (1 Sv = 106 m3 s−1). Carbon of the land vegetation, soil, and ocean amounts to 460, 1390, and 37 900 GtC, respectively. Land NPP is 56 GtC yr−1, and export production of particulate organic carbon in the ocean is 8.5 GtC yr−1. These amounts are consistent with the previous estimates (e.g., Prentice et al. 2001), though land NPP in our model is slightly smaller because of the weaker CO2 fertilization effect at the preindustrial level of 290 ppmv. Land NPP and leaf area index (LAI) in the steady state are shown in Figs. 1a,b, which are consistent with the observations on the global
scale in terms of the feature of the tropical high values and boreal low values.

With regard to a massive methane release from methane hydrates, atmospheric methane concentration in the model is instantaneously increased from the pre-industrial level of 0.75 ppmv to an exceedingly high level, and the model is integrated from the preindustrial state with the atmospheric methane concentration fixed at the high level. Oxidation of the methane and the resultant CO₂ production in the atmosphere are not included because the purpose of this study is to investigate the effect of the methane increase alone. In addition, the lifetime of such a high methane concentration can be much longer (at least several decades but <100 yr) than the present-day lifetime of about 10 yr because of the exhaustion of the atmospheric OH radical used for the methane oxidation (e.g., Sze 1977; Schmidt and Shindell 2003). The model is therefore integrated for about 80 yr in the sensitivity experiments described below, though the main experiment (M-CH4) is carried out for longer than 200 yr to examine how the system approaches another steady state.

Concerning the prescribed atmospheric methane concentration, two experiments are carried out: 1000 ppmv [1000(mille)-fold methane (M-CH4)] and 100 ppmv [100(cent)-fold methane (C-CH4)]. The M-CH4 is concerned with the methane release of 2000 GtC estimated for the PETM (e.g., Dickens et al. 1995, 1997; Matsumoto 1995), though the M-CH4 is carried out by instantaneously prescribing the release. The C-CH4 is concerned with a weaker methane release compared to the M-CH4 and is a sensitivity experiment to the atmospheric methane concentration. The feedbacks of the carbon cycle on climate change in the M-CH4 are examined by conducting two additional simulations. In the first simulation, atmospheric CO₂ concentration for the radiation code of the model is fixed at the initial level of 290 ppmv [radiation by 290 ppmv CO₂ (R290CO₂)] and the results are compared to the M-CH4 simulation results to examine the influence of CO₂ radiative forcing.
on the climate–carbon cycle system. In the second simulation, the atmospheric CO\textsubscript{2} concentration for the biochemical photosynthesis processes is fixed at 290 ppmv [fertilization by 290 ppmv CO\textsubscript{2} (F290CO\textsubscript{2})] and the results are compared to the M-CH\textsubscript{4} simulation results to examine the influence of CO\textsubscript{2} fertilization of land plants on the climate–carbon cycle system. The other procedures of the R290CO\textsubscript{2} and F290CO\textsubscript{2} are the same as the M-CH\textsubscript{4}. The climate–carbon cycle responses for 80 yr of these experiments are summarized in Table 1.

### 3. Results

#### a. Climate–carbon cycle response to 1000-fold atmospheric methane

Global climate–carbon cycle response to the rapid 1000-fold increase in atmospheric methane (from <1 to 1000 ppmv) in the experiment M-CH\textsubscript{4} (Fig. 2) is characterized by rapid increases in temperature and soil respiration and a rapid decrease in land NPP that lead to a large decrease in land carbon (\sim 30\%) and a large increase in atmospheric CO\textsubscript{2} (>50\%) during the first several decades. The large changes are concentrated within the first several years. The instantaneous radiative forcing at the top of the atmosphere reaches about 10 W m\textsuperscript{-2} during the first year and exponentially decreases to 1.5 W m\textsuperscript{-2} through the increase in the outward longwave radiation by the surface warming. During the first year, the rapid increase in air temperature (+1.5°C; Fig. 3a) by the intense radiative forcing increases the saturation specific humidity faster than the water vapor supply from the ocean especially in the subsidence of the subtropics and middle latitudes. This change decreases relative humidity, cloudiness, and thus precipitation over the subtropical and midlatitude ocean, resulting in a 3% decrease in global precipitation (Fig. 3b). After that, the global precipitation increases by about 10% during the first 40 yr (Fig. 2b) because of the intensified hydrologic cycle by the atmospheric water vapor increase with the air temperature rise.

Global-mean surface air temperature (Fig. 2a) rises by about 6°C during the first 40 yr. Warming over the land is stronger than over the ocean (Fig. 4a). The magnitude of the warming is larger in northern high latitudes (Fig. 4a) because of the ice–albedo feedback due to decreases in snow cover and sea ice extent. Precipitation increases in the tropics and high latitudes and decreases in the subtropics and middle latitudes (Fig. 4b) in association with the intensified hydrologic cycle and a poleward expansion of the Hadley circulation. Decreases in precipitation are also evident in Central America (Fig. 4b) in association with an El Niño–like climate change in the tropics, a weakening of the Hadley and Walker circulations, due to the atmospheric vertical stabilization caused by the warming-induced water vapor increase. These features of climate change in the M-CH\textsubscript{4} are qualitatively similar to those of global warming projections for the twenty-first century by other AOGCMs (Meehl et al. 2007) but are quantitatively more significant: anomalies of temperature and precipitation in the M-CH\textsubscript{4} are more than 2 times those of the multimodel mean for a medium scenario of greenhouse gas concentrations (derived from scenario A1B of the Intergovernmental Panel on Climate Change Special Report on Emissions Scenarios; Nakićenović and Swart 2000); these features become stronger in the later period of the M-CH\textsubscript{4} (Figs. 4c,d). These climate changes are occurring much more rapidly (within 10 yr) so that the terrestrial biota is not able to adapt to these changes (e.g., large decreases in biomass and vegetation redistribution) as described below.

The rapid warming causes large damages to the land biosphere. The maximum damage to the global vegetation occurs during the first 3 yr, in which global NPP decreases from 56 to 40 GtC yr\textsuperscript{-1} (29\%), with more than 3/5 of the NPP decrease occurring in the first year (Fig. 3c). The NPP decrease is mostly due to lowland tropical forests (from Fig. 1a to Fig. 1c; Fig. 5a shows the difference), and LAI markedly decreases especially in tropical America and Africa (from Fig. 1b to Fig. 1d). In addition to the NPP

<table>
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<th>Expt</th>
<th>Warming (°C)</th>
<th>Carbon amount change (GtC)</th>
<th>Land NPP min (GtC yr\textsuperscript{-1})</th>
<th>Soil respiration max (GtC yr\textsuperscript{-1})</th>
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<td>+125</td>
<td>-200</td>
<td>48</td>
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Table 1. Climate–carbon cycle responses for 80 yr of the experiments with rapid increase in atmospheric methane. The initial condition is the preindustrial state with 0.75-ppmv methane. M-CH\textsubscript{4} is driven by 1000-ppmv methane. R290CO\textsubscript{2} is the same as M-CH\textsubscript{4}, but the atmospheric CO\textsubscript{2} concentration for the radiation code is fixed at the initial level of 290 ppmv. F290CO\textsubscript{2} is the same as M-CH\textsubscript{4}, but the land CO\textsubscript{2} fertilization effect is fixed at the initial level of 290 ppmv. C-CH\textsubscript{4} is driven by 100-ppmv methane. Warming and carbon amount change are shown for about year 80. The initial value of land NPP and soil respiration are 56 GtC yr\textsuperscript{-1}.
decrease, global soil respiration increases from 56 to 73 GtC yr\(^{-1}\) (+30\%) (Figs. 3c, 5b) resulting in a minimum of global net ecosystem production (NEP; NPP minus soil respiration) (−33 GtC yr\(^{-1}\); Figs. 3c, 5c) at year 3.

Although the global response of the land biosphere is characterized by the damage, the regional responses are different as shown by Fig. 6: for example, decreases in NPP and soil respiration (wane of ecosystem) in the tropics (Fig. 6a) and increases in NPP and soil respiration (wax of ecosystem) at high latitudes (Fig. 6c). The tropical responses are further analyzed by dividing into two regions with different features: large damages in America and Africa and small damages in Southeast Asia and India (Fig. 7). As described in detail below, Figs. 6 and 7 indicate that the initial damages to the land biosphere consist mainly of 1) rapid decrease in NPP by a decrease in photosynthesis and an increase in plant maintenance respiration with warming, leading to negative NPP-induced dieback at low and middle latitudes, and 2) boreal forest dieback by heat stress (based on the annual accumulated degree-day sum above a specific temperature base), leading to more decrease in NPP (mortality scheme in the model follows LPJ-DGVM; Sitch et al. 2003).

After the methane release event, tropical NPP decreases from 800 to 400 gC m\(^{-2}\) yr\(^{-1}\) by year 11 (−50\%; Fig. 6a). In the first year, the NPP decrease is the largest (−23\%), with the strongest warming of 2°C (Fig. 6g).
During the first few years, surface air temperatures in tropical lands increase by about 4°C (Fig. 7a). In tropical South America and Africa (red lines in Fig. 7), this rapid warming decreases relative humidity (Fig. 7b), which increases the evaporative demand of the atmosphere, leading to a decrease in soil water (Fig. 7c). Owing to these warming and drying, C₃ plants much decrease stomatal conductance (Woodward et al. 1995) to limit evapotranspiration and retain soil water. Although the evapotranspiration is limited (Fig. 7d), the reduced stomatal conductance decreases CO₂ flux through the stomata at the same time, leading to a large decrease in photosynthesis (Fig. 7e). Soil water continues to decrease through the direct evaporation from the soil surface, despite the decrease in stomatal conductance.

Such initial drying and associated decrease in photosynthesis do not occur in Southeast Asia and India (blue lines in Fig. 7) because of the sufficient water supply from the ocean. However, plant maintenance respiration increases with the warming (Fig. 7f), leading to a decrease in NPP during the first 3 yr (−40%; Fig. 7g). In tropical South America and Africa, both the decrease in photosynthesis and the increase in plant maintenance respiration cause a large decrease in NPP (−50%; Fig. 7g), leading to a large decrease in vegetation (Fig. 7h for LAI). Thus, the total tropical NPP rapidly decreases during the first 3 yr (Fig. 6a).

After the initial few years, air temperatures of land ecosystems in tropical South America and Africa continue to increase, at a much slower rate, by about 5°C by year 30 (Fig. 7a). Relative humidity continues to decrease with the warming (Fig. 7b), causing soil water to decrease (Fig. 7c). Evapotranspiration decreases (Fig. 7d) because of the LAI decrease (Fig. 7h), which contributes to the stabilization of soil water toward a lower state after year 15 (Fig. 7c). Despite this decrease in water availability, C₃ photosynthesis of the canopy top leaf begins to partly recover (Fig. 7c) because of the fertilization effect of the rapid increase in atmospheric CO₂ (e.g., +50 ppmv during the first 4 yr) by the land carbon release (Fig. 2d). Plant maintenance respiration per unit biomass increases with the warming, but the total plant maintenance respiration decreases (Fig. 7f) with the decrease in vegetation biomass as represented by woody LAI (Fig. 7h). Thus, the NPP decrease in tropical South America and Africa is much slower after year 3 (Fig. 7g).
In contrast, in Southeast Asia and India, C3 photosynthesis of the canopy top leaf becomes larger than initially (Fig. 7e) because of the weaker warming (~+2°C after year 3; Fig. 7a) by the oceanic large thermal inertia effect, the wet condition (Figs. 7b,c) by the sufficient water supply from the ocean, and the CO₂ fertilization, leading to a recovery of NPP and LAI from the initial decrease (Figs. 7g,h). Thus, as the whole of the tropics, the decreasing NPP becomes much more stable toward a lower state after year 3 (Fig. 6a).

NPP becomes negative in some areas of the tropics, particularly in tropical America and Africa, leading to dieback of tropical forests. This dieback is significant during the first 10 yr and amounts to a decrease in forest carbon of ~20% (~1.3 kgC m⁻² versus the initial forest of 6.6 kgC m⁻²; Fig. 6d), which is more than half of the total decrease of ~36% for the first 10 yr (~2.4 kgC m⁻²). The remaining ~16% is due to plant tissue turnover larger than the decreased NPP in the living forest. The negative NPP-induced dieback increases the litter pool by 24% during the first 3 yr (from 0.46 to 0.57 kgC m⁻²; Fig. 6d). Soil respiration is enhanced by the warming especially in the first year and by the increased litter pool in the following 2 yr (Fig. 6a). After year 3, the litter pool rapidly decreases because of decreases in carbon flux from vegetation by the NPP decrease, and thus the soil respiration decreases.

In the northern middle latitudes (Figs. 6b,e,h), NPP is most reduced in the first year (~70 gC m⁻² yr⁻¹, ~23%) similar to the tropical latitudes. The NPP decrease is significant in Europe and eastern to central regions of North America (~300 gC m⁻² yr⁻¹; Fig. 5a). Although photosynthesis is increased by the warming (+2.5°C) within the temperature range of the middle latitudes, plant maintenance respiration much increases at the same time, leading to the NPP decrease. The NPP decrease in Europe is also caused by precipitation decrease due to northward extension of the Azores high in the global warming (Fig. 4b), though precipitation in the middle latitudes slightly increases on the average. In the following period until year 5, NPP is reduced mostly by heat stress–induced dieback of boreal forests, resulting in the total decrease of ~50% (from 300 to 150 gC m⁻² yr⁻¹).

In addition to the heat stress–induced dieback of boreal forests, NPP of temperate forests becomes negative in some areas of the precipitation decrease in Europe, leading to dieback of the temperate forests especially in the first 5 yr (Fig. 6e). The dieback of boreal and temperate forests (~61% and ~46%, respectively) is much larger than that estimated for tropical forests (~20%), accounting for 90% or more of the decrease in boreal and temperate forest carbon. The forest dieback causes the litter pool to increase from 0.5 to 0.9 kgC m⁻² (Fig. 6e) or 80% during the first 3 yr. The litter pool increase per unit area for the first 3 yr is 4 times larger than that of the tropics (0.4 and 0.1 kgC m⁻²; Figs. 6e,d). Similar to the tropics, soil respiration is enhanced by the warming, especially in the first year, and by the
increased litter pool in the following 2 yr (Fig. 6b). The enhanced soil respiration by the increased litter pool is evident north of the Great Lakes, in Europe, and in Manchuria (Fig. 5b). After year 3, the litter pool rapidly decreases because of decreases in carbon flux from vegetation by the NPP decrease, and thus the soil respiration decreases. After year 5, NPP slightly increases because of $C_3$ grass expansion in the decreased forest.
shading and because of CO2 fertilization effect on C3 plants. In the northern high latitudes (Figs. 6c,f,i), the warming enhances soil respiration (Fig. 5b), leading to negative NEP during the first 4 yr (Fig. 5c). However, the warming and precipitation increase (Fig. 6i) cause photosynthesis to increase more than plant maintenance respiration, leading to a large increase in NPP (+100% for the first 15 yr; Fig. 6c). Thus, NEP becomes positive after year 4 (Fig. 6c). Evaporation of snowpack by the warming also helps to increase photosynthesis by increasing relative humidity: a decrease in snowpack of 20% and an increase in relative humidity of 5% by year 10 (not shown). The NPP increase leads to increases in boreal forest carbon, C3 grass carbon (>30% and +200%, respectively; Fig. 6f), and LAI (>100%; from Fig. 1b to Fig. 1d). The increase in C3 grass is limited after year 15 (Fig. 6f) because of an increase in forest shading. The litter pool is slightly decreased by the warming-enhanced soil respiration during the first 4 yr and then increases by 50% (Fig. 6f) because of increases in carbon flux from vegetation by the NPP increase. In contrast to the low and middle latitudes, the response of the land ecosystem in the high latitudes is characterized by the increasing activity with the rapid warming. A similar response to the warming occurs also in mountainous regions such as the Tibetan Plateau (e.g., NPP increase of >200 gC m\(^{-2}\) yr\(^{-1}\); >50%; Fig. 5a). After year 3, the global NPP begins to recover from the minimum of 40 GtC yr\(^{-1}\) (Fig. 3c), which is partly due to the northern high-latitude NPP persistently increasing from the initial value (Fig. 6c).

As a result, tropical NPP and carbon stored in vegetation, which are initially more than 60% of the global total amounts, markedly decrease to less than 50% of those during the first 30 yr. Tropical NPP decreases from 37 to 20 GtC yr\(^{-1}\) (−46%; Fig. 8a), and vegetation carbon decreases from 300 to 130 GtC (−57%, Fig. 8b). Thus, carbon stored in global vegetation decreases from 460 to 260 GtC (−43%), with the tropics accounting for ~80% of the decrease and the northern middle latitudes accounting for ~20% of the decrease. After reaching a minimum at year 23, the global vegetation begins to increase (Fig. 8b) because of an increase in vegetation carbon of >50% by an increase in NPP of ~100% in the warming-ameliorated northern high latitudes (Figs. 8b,a) and the CO2 fertilization from enhanced land-released CO2; this effect is shown by the F290CO2 sensitivity experiment described below.

Soil carbon is basically decreased (Fig. 8c) by the warming-enhanced respiration. The decrease in soil carbon is larger in the lower latitudes (e.g., a decrease in tropical soil carbon of >50%; Fig. 8c) because of the warming-induced NPP decreases. Overall, the land biosphere releases 480 GtC into the atmosphere during the first 40 yr. About year 40, 25% (120 GtC) of the land-released carbon is absorbed by the ocean because of the increase in air–sea CO2 partial pressure difference and the remaining 75% (360 GtC; Fig. 2d) increases atmospheric CO2 concentration by 170 ppmv, from the initial 290 ppmv to the maximum 460 ppmv. Cumulative changes in land carbon by about year 80 (Fig. 9)

**FIG. 7.** Time series of tropical land ecosystems for the first 40 yr of the experiment M-CH4: (a) surface air temperature (°C), (b) relative humidity (%), (c) water content (mm) of two soil layers (upper and root zone), (d) evapotranspiration (mm yr\(^{-1}\)), (e) C3 photosynthesis of canopy top leaf (µmolC m\(^{-2}\) s\(^{-1}\)), (f) plant maintenance respiration (kgC m\(^{-2}\) yr\(^{-1}\)), (g) NPP (kgC m\(^{-2}\) yr\(^{-1}\)), and (h) LAI of woody plants. Red lines denote the time series for tropical South America and Africa (10°S–7°N, 70°W–25°E) and blue lines denote the time series for Southeast Asia and India (10°S–18°N, 70°–160°E).
clearly show the largest decreases in the tropics, large decreases in the northern middle latitudes, and increases in the high latitudes (including part of Patagonia) and mountainous regions (the Tibetan Plateau and part of the Great Rift Valley and the Andes).

Changes in the distribution of dominant PFTs are shown by comparing the foliage projective cover (FPC; Sitch et al. 2003) between Figs. 10a,b. In the tropics, the area of tropical forest decreases because of the warming-induced NPP decrease and are replaced by C4 grass, except for little change in the dominance of tropical forests in Southeast Asia and India with the small damage shown by Fig. 7. The C4 grass also expands from the tropics to the higher latitudes by more than 600 km because of warming and drying, not only in Central and South America and Africa but also in Australia and on the Mediterranean coast. In the northern middle to high latitudes, boreal forests shift northward by about 500 km or newly occur in the north because of the heat stress dieback in the south (e.g., north of the Great Lakes, Eastern Europe, and Manchuria) and the ameliorated condition by the warming and precipitation increase in the north (Fig. 4). An exception occurs in central to eastern Eurasia where the heat stress dieback is significant and C3 grass instead becomes dominant. The C3 grass also increases its FPC from 0.2 to 0.7 in tundra as far north as 75°N. The decrease in forest as a dominant PFT, especially tropical and boreal forests, leads to the result that NPP and soil respiration rates come to an equilibrium state at lower rates than initially (Fig. 2c).

In the oceans, the vertically downward flow of the Atlantic meridional overturning circulation (AMOC) in the high latitudes weakens from the initial 17 to 9 Sv during the first 30 yr because of a decrease in surface water density due to the warming and precipitation increase and then becomes almost stable at that magnitude (Fig. 11a). In contrast, in the lower latitudes, the downward flow becomes stronger by 3 Sv at 25°–45°N because of increases in surface salinity due to the precipitation decrease. The depth of the southward North Atlantic Deep Water flow shallows by 800 m from 2000–3600 m to 1200–2800 m around the equator. In association with the weakening of the AMOC, the Atlantic northward heat transport decreases by 0.2 PW (1 PW = 10^{15} W) at 25°–45°N (Fig. 11b). This heat transport decrease corresponds to a decrease in surface heat flux of 2.5 W m^{-2} for the area north of 45°N, which is capable of decreasing northern surface temperatures by about 1°C but is overcome by the intense warming due to the increased atmospheric methane.

The global export production of particulate organic carbon decreases from the initial 8.5 to 7.5 GtC yr^{-1}. This decrease is mostly due to the decrease in the equatorial eastern Pacific (Fig. 12a), which is caused by a decrease in the trade wind–driven upwelling (nutrient supply from deep waters) due to the warming-induced weakening of the Hadley and Walker circulations.
reduced upwelling also decreases carbon supply from deep waters and thus decreases sea-to-air CO$_2$ flux in this region, though the reduced export production partly offsets the decrease in the carbon supply from deep waters (Fig. 12b). In the Arctic Ocean and the Southern Ocean, the sea ice decrease by the warming increases the export production and air-to-sea CO$_2$ flux (Fig. 12). However, these regional changes in the ocean carbon cycle hardly affect the global ocean CO$_2$ uptake induced by the land CO$_2$ release. Among the world oceans, the North Atlantic accumulates the largest amount of the land-released carbon through the carbon transport into the deep ocean by the North Atlantic thermohaline circulation (Fig. 9). The global warming and precipitation increase continue through the time integration of the M-CH4, though their rates slow down after the first several decades (Figs. 2a,b). At about year 50, the amount of global land
carbon appears to approach another steady state (see red line in Fig. 2d) when the reduced soil respiration becomes almost equal to the reduced NPP (Fig. 2c). The temporal pattern of soil respiration is, except for the fluctuation related to the litter pool during the first 10 yr, due to an adjustment of the labile humus, which dominates most of the soil carbon and has a carbon turnover time of several decades. The amount of carbon in vegetation and soil, however, continues to change throughout the simulation (Fig. 2d). The reduced vegetation gradually recovers (Fig. 8b) mainly by the NPP increase in the northern high latitudes (Fig. 8a) and partly by the atmospheric CO2 fertilization effect; the global NPP increases by more than 50% during the first 40 yr as a result of land CO2 releases and gradually decreases because of CO2 uptake by the ocean so that the atmosphere contains just under 50% more CO2 at the end of the simulation than initially. The ocean carbon (blue line in Fig. 2e) increases by only 0.6% over the simulation period.

The intense warming in the M-CH4 is caused not only by the prescribed large increase in atmospheric methane of 1000 ppmv but also by the increase in atmospheric CO2 of 150 ppmv due to the feedback of warming-induced land CO2 release (Fig. 2d). To estimate the contribution of the atmospheric CO2 feedback to the total warming in the M-CH4, a complementary experiment (R290CO2) to the M-CH4 is carried out, where atmospheric CO2 concentration for the radiation code of the model is fixed at the initial level of 290 ppmv. The other procedures of the R290CO2 are the same as the M-CH4. As a result, the global-mean surface air temperature rise during the 80 yr in the R290CO2 is 5°C, about 80% of 6°C in the M-CH4 (Fig. 13, the bottom and middle lines; Table 1). According to the temperature rise, the decrease in land carbon in the R290CO2 simulation is also about 80% of that in the M-CH4 (2.410 versus 2.500 GtC). The increase of 1°C from the R290CO2 to the M-CH4 is due to the atmospheric CO2 increase of 160 ppmv and is not negligible in the total warming of the M-CH4. This result is consistent with an estimate of the radiative forcing associated with a post-industrial warming of about 1°C (2.35 W m⁻², Myhre et al. 1998).

The atmospheric CO2 increase from 290 to 450 ppmv due to the land CO2 release in the M-CH4 would simultaneously increase land NPP by the CO2 fertilization effect. If it were not for this effect, the land carbon...
decrease and the increase in atmospheric CO2 and temperature would be larger. To this effect, a complementary experiment with the CO2 fertilization effect fixed at 290-ppmv CO2 (F290CO2) was conducted. The other procedures of the F290CO2 are the same as the M-CH4. As a result, the land NPP in the F290CO2 decreases greatly from the initial 56 to 25 GtC yr\(^{-1}\) and remains at that value because of the methane-induced thermal damage, the fixed CO2 fertilization effect, and the associated vegetation shifts (e.g., forest to grasses). The land carbon decreases by 950 GtC (vs 500 GtC in M-CH4), the atmospheric CO2 reaches 620 ppmv (vs. 450 ppmv in M-CH4), and the global-mean surface air temperature increases by 7\(^\circ\)C as compared to 6\(^\circ\)C in the M-CH4 (Fig. 13, the top and middle lines; Table 1). This comparison indicates that CO2 fertilization of NPP compensates for 50% of the warming-induced decrease in the land carbon, 170 ppmv of the additional increase in atmospheric CO2, and a 1\(^\circ\)C increase in air temperature that would have occurred without CO2 fertilization.

b. Comparison of 1000- and 100-fold atmospheric methane experiments

At the beginning of the C-CH4, the instantaneous radiative forcing at the top of the atmosphere is 5 W m\(^{-2}\), half of the M-CH4. Time series of the global climate–carbon cycle responses of the C-CH4 are shown in Fig. 14 for the comparison with the M-CH4 in Fig. 2. The global-mean temperature rise is about 3\(^\circ\)C and the precipitation increase is about 5% during the first 40 yr in the C-CH4, which are also half of those in the M-CH4 (+6\(^\circ\)C and +10%, respectively). The time variation patterns of the carbon cycle in the C-CH4 are qualitatively similar to those in the M-CH4. However, the magnitudes of the variations in the C-CH4 are much smaller, at most 50% of the M-CH4 (Table 1). These differences in the climate–carbon cycle responses between the M-CH4 and C-CH4 characterized by a factor of \((2–3)\) are consistent with the dependence of the radiative forcing of atmospheric methane on the square root of the methane concentration (Myhre et al. 1998).

The spatial variations of the climate–carbon cycle responses in the C-CH4 (e.g., strong warming by the ice–albedo feedback in the northern high latitudes, precipitation decreases in the subtropics and middle latitudes, and land NPP decreases by the thermal damage in the tropics) are also qualitatively similar to those in the M-CH4 but are quantitatively much smaller, at most 50% of the M-CH4. For example, Fig. 15 shows surface air temperature anomaly and land NEP of the first 10 yr in the C-CH4. The land temperature increases are largely 2\(^\circ\)–4\(^\circ\)C (Fig. 15a) and the NEP in the tropical America and Africa are \(-200\) to \(-400\) gC m\(^{-2}\) yr\(^{-1}\) (Fig. 15b), while the corresponding values are 4\(^\circ\)–8\(^\circ\)C.
Changes in dominant PFT in the C-CH4 are also similar to the M-CH4 in terms of the expansion of C4 grass and the northward movement of boreal forest (Fig. 10). However, the diebacks of tropical and boreal forests in the C-CH4 are weaker because of the weaker warming than the M-CH4, which contributes to the result that NPP and soil respiration return to the same levels as before the methane increase (Fig. 14c).

4. Discussion

In the experiment M-CH4, the global soil respiration rapidly increases to the maximum (73 GtC yr\(^{-1}\)) within the first 3 yr and decreases to the initial level within the next 10 yr (Fig. 3c). Although the first increase is caused partly by the methane-induced rapid warming, the rapid increase–decrease fluctuation is caused mostly by the fast decomposition of the increased litter; the litter pool is increased by the forest dieback due to the warming-induced negative NPP and the heat stress in the tropics and northern middle latitudes, respectively (Figs. 6a, 6b, d, e). In the model, the biomass of the dieback is added to the litter pool as detritus, and both of the woody and herbaceous detritus are decomposed with the litter turnover time (time for decrease to \(1/e\)) that is about 3 yr at 10°C and decreases with surface air temperature (Lloyd and Taylor 1994; Sitch et al. 2003). In the real ecosystem, woody tissue takes a longer time to decompose than herbaceous tissue, but the buildup of woody slash also increases the probability of fire, which could result in a rapid return of carbon to the atmosphere even within the 3-yr window suggested by the M-CH4. The fire module was not included in the experiments of this study in order to interpret the results clearly. However, by use of the short turnover time also for the woody detritus in the litter pool, the above mechanism of the real ecosystem is implicitly represented, leading to the rapid change in the soil respiration during the early period of the M-CH4.

A large amount of carbon is stored at northern high latitudes and the decomposition rates are influenced by the presence of permafrost in addition to surface air temperatures. It is probable that the thermal dynamics associated with permafrost degradation delay the effects of the air temperature changes on soil respiration. In the M-CH4, however, the warming is so intense that the temperature increase of the permafrost follows the air temperature with almost no delay (the difference less than 0.5°C), resulting in the relatively rapid increase in the soil respiration (Figs. 6c, i). Moreover, the response of soil carbon in the northern high latitudes is also affected by other factors; because of the increase in carbon flux from vegetation by the NPP increase, the soil respiration increases, but the soil carbon does not decrease so much (Fig. 8c). In addition, the model in this study has a simple soil structure, three soil layers with prognostic temperatures at the depth of 1 cm (layer 1: upper, thin layer), 25 or 75 cm (layer 2: root zone), and 1 or 2.5 m (layer 3: underlying recharge layer) for grassland or forest (modified from Sato et al. 1989), and the temperature dependence of soil respiration, \(\exp\{308.56 \times [(1/56.02) - 1/(T + 46.02)]\}\) (Lloyd and Taylor 1994; Sitch et al. 2003), is determined by the temperatures of the layers 1 and 2, which may be too coarse to represent the influence of the permafrost accurately. Methane emission from thawing permafrost in the decadal-time-scale warming is estimated at 10–100 GtC (Davidson and Janssens 2006; Zhuang et al. 2006), which does not influence the result by the radiative forcing of 2000-GtC methane in this study. However, the influence of the permafrost in the rapid warming must be investigated...
strictly by the vertically high-resolution soil model with dynamics related to permafrost methane.

In the M-CH4, the amount of carbon in the land biosphere decreases by more than 25% because of the intense warming by the rapid increase in atmospheric methane. At the same time, the fertilization effect of the atmospheric CO₂ increase of 160 ppmv due to the land CO₂ release partly compensates land NPP for its warming-induced decrease, which prevents the warming-induced decrease in the land carbon by about 50% and prevents further increases in atmospheric CO₂ and surface air temperature (+170 ppmv and +1°C) as shown by the comparison of the M-CH4 and F290CO₂ (Fig. 13). This is, however, only a partial offset to the main response of the land carbon decrease. Beerling (2000) suggested that the amount of carbon in the land biosphere increases across the PETM because of the large fertilization effect of the methane oxidation-produced high CO₂ level (+1000 ppmv). This land carbon increase is reasonable but may take at least ~100 yr to occur, waiting for a slow increase in atmospheric CO₂ by the methane oxidation with the exhaustion of the OH radical (e.g., Sze 1977; Schmidt and Shindell 2003). This sort of problem must be resolved accurately by the Earth system model including realistic atmospheric chemical processes.

In this relation, the time integration of the M-CH4 is limited to shorter than 300 yr because the extremely high atmospheric methane concentration cannot be sustained for hundreds of years because of the oxidation of methane. Besides, the ocean response governed by the vertical diffusion requires more than thousands of years to completely reach another steady state. Thus, the climate–carbon cycle response (including marine sedimentary CaCO₃) on long time scales of ~10 000 yr after the atmospheric CO₂ increase from oxidation of the released methane should now be investigated by using models of simplicity or intermediate complexity (e.g., Zeebe et al. 2009).

The land biosphere in the experiments of this study is affected mostly by the (methane induced) warming rather than by the CO₂ fertilization effect. This feature is similar to the study by Thompson et al. (2004) in which an experiment with the CO₂ fertilization saturated at the year 2000 level is compared with the fertilization case for the future projection with anthropogenic CO₂ emission. In their saturated CO₂ fertilization case, the land carbon decreases by 150 GtC during the twenty-first century because of a global warming of 3°C. The land biomass decreases at low latitudes and increases in the high latitudes. These results in their saturation case are nearly equal to the C-CH4 (+3°C and −200 GtC; Figs. 14a,d), while the results in the M-CH4 are much more intense (>+6°C and −500 GtC; Figs. 2a,d). However, the cause of the intense climate–carbon cycle change in the M-CH4 is not anthropogenic CO₂ but natural methane, whose large emission from marine sedimentary methane hydrates is possible to occur because of the intermediate-water–deep-water warming (e.g., Kennett et al. 2003) in the near future.

A comparison of the results of the 1000-fold and 100-fold methane increase experiments (M-CH4 and C-CH4) suggests that there may be a threshold in which radiative forcing begins to significantly alter the structure and function of the land biosphere. While NPP and soil respiration rates come to an equilibrium state at lower rates than before the methane release event in the M-CH4 (Fig. 2c), the rates return to the same levels as before in the C-CH4 (Fig. 14c). This difference which may be regarded as a threshold is considered to be related to the mortality scheme used in the model; when heat stress occurs on a boreal forest or forest NPP is reduced to negative by some causes such as the intense warming in the M-CH4, the forest catastrophically dies and NPP and soil respiration of the ecosystem rapidly come to a lower state. This study suggests the threshold for the land biosphere between the warming of 3°C (C-CH4) and 6°C (M-CH4). Although the model has uncertainty in the climate–carbon cycle sensitivity, more experiments using various methane concentrations must be carried out to determine the threshold more accurately.

Net primary productivity is known to be limited by nitrogen. In the model, we assume that nitrogen is sufficient in using the photosynthesis calculation of Woodward et al. (1995), and thus the effect of carbon–nitrogen cycle coupling (Thornton et al. 2007, 2009; Sokolov et al. 2008) is not considered. According to the carbon–nitrogen cycle model studies, it is probable that surface warming induces fertilization of plant growth by increased mineralization of nitrogen associated with increased decomposition of soil organic matter (Thornton et al. 2009), leading to an increase in carbon storage of the land biosphere. Even under a rapid, strong warming of 6°C such as our experiment (M-CH4), the increase in nitrogen availability may reduce the warming-induced damage to the land biosphere (CO₂ source to the atmosphere) (Sokolov et al. 2008). This point must be investigated by Earth system models with carbon–nitrogen dynamics in the future study.

5. Concluding remarks

Decadal-time-scale responses of a coupled climate–carbon cycle model to rapid increases in atmospheric methane are investigated with regard to the climatic event of massive methane release from marine sedimentary
methylene hydrates that is possible to occur with global warming. In the experiment of a 1000-fold methane increase (from <1 to 1000 ppmv; M-CH4), the global warming is >6°C. The amount of carbon in the global land biosphere decreases by >25%. This is mostly due to a large decrease in tropical net primary production during the first few years (~40%), which is caused by a decrease in photosynthesis and an increase in plant maintenance respiration with the early warming of ~3°C, leading to tropical forest dieback (>20%) and the largest decrease in vegetation carbon of >50% (~80% of the decrease in global vegetation carbon). The decrease in global land carbon is also partly due to forest diebacks (mainly boreal forest dieback by heat stress) at northern middle latitudes. In contrast, vegetation carbon increases by >50% at northern high latitudes and mountainous regions because of the amelioration to warm and wet conditions. Thus, the spatial responses of the land biosphere are characterized by wane at low and middle latitudes and wax at high latitudes (Fig. 9).

The global warming of >6°C consists mainly of >5°C by the 1000-fold atmospheric methane and an additional increase of 1°C by the increase of atmospheric CO2 (>50%) due to the warming-induced land CO2 release. The CO2 fertilization of land plants limits further warming of 1°C by limiting the atmospheric CO2 increase (Fig. 13). Although the warming in the M-CH4 (e.g., ~+5°C–6°C over the tropical ocean, >+10°C at northern high latitudes, and >+6°C for the global mean by year 80; Figs. 2a, 4c) is consistent with the Paleocene–Eocene thermal maximum (PETM) (e.g., Zachos et al. 2003), the M-CH4 is carried out by instantaneously prescribing the 1000-fold atmospheric methane, which is too rapid to explain the PETM (Bains et al. 1999).

The global land biomass decrease of more than a quarter because of the methane-induced warming estimated in this study (M-CH4) suggests a critical situation for the land ecosystem or agricultural production, especially in the low latitudes, where the land biomass is estimated to decrease by 50%. Because global estimates of the CH4 content of marine methane hydrates range from ~500 to ~24,000 Gt and an intermediate value of ~10,000 Gt is often used as a consensus value (Kennett et al. 2003), more intense warming leading to greater damage to the land biomass than our experiment (~2000 Gt) is possible in the future methane release event that would be caused by the ongoing anthropogenic global warming.

In this study, the massive release of methane from hydrates is assumed to occur under preindustrial environmental conditions. While this approach is useful for illustrating how the climate–carbon cycle system might change with such a release, we know that a massive release of methane has not occurred under preindustrial conditions or even industrial conditions. Because existing studies are insufficient, we know little about the environmental conditions that might trigger a massive release of methane from hydrates. To enhance our understanding of the environmental conditions, it is necessary to improve existing estimates of hydrate occurrences, their geographical distribution and depth profile, and dynamics of the hydrates inventory under changing environments, mainly pressure and temperature, also need to be better understood; all this requires advances in observation and modeling (Krey et al. 2009). Using these improvements, the next step would be to combine the dynamics simulated in this study with a warming from anthropogenic CO2 scenario.

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