Dependency of climate change and carbon cycle on CO₂ emission pathways

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Abstract

Previous research has indicated that the response of globally average temperature is approximately proportional to cumulative CO₂ emissions, yet evidence of the robustness of this relationship over a range of CO₂ emission pathways is lacking. To address this, we evaluate the dependency of climate and carbon cycle change on CO₂ emission pathways using a fully coupled climate–carbon cycle model. We design five idealized pathways (including an overshoot scenario for cumulative emissions), each of which levels off to final cumulative emissions of 2000 GtC. The cumulative emissions of the overshoot scenario reach 4000 GtC temporarily, subsequently reducing to 2000 GtC as a result of continuous negative emissions. Although we find that responses of climatic variables and the carbon cycle are largely independent of emission pathways, a much weakened Atlantic meridional overturning circulation (AMOC) is projected in the overshoot scenario despite cessation of emissions. This weakened AMOC is enhanced by rapid warming in the Arctic region due to considerable temporary elevation of atmospheric CO₂ concentration and induces the decline of surface air temperature and decrease of precipitation over the northern Atlantic and Europe region. Moreover, the weakened AMOC reduces CO₂ uptake by the Atlantic and Arctic oceans. However, the weakened AMOC contributes little to the global carbon cycle. In conclusion, although climate variations have been found to be dependent on emission pathways, the global carbon cycle is relatively independent of these emission pathways, at least superficially.

Keywords: CO₂ emission pathways, climate change, overshoot, carbon cycle, Earth system model, meridional overturning circulation

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1. Introduction

Assessments of a wide variety of CO₂ emission pathways have been conducted previously to assist in the design of feasible and effective emission strategies in order to avoid threats posed by climate change and expand global economic activity (e.g., IPCC 2007a). Such assessments have indicated that the globally averaged temperature response is approximately proportional to cumulative CO₂ emissions (e.g., Matthews and Caldeira 2008; Allen et al 2009; Lowe et al 2009; Matthews et al 2009; Solomon et al 2009; Gillett et al 2011). In such cases, temperatures increase with increasing CO₂ but remain approximately constant afterward, despite zero further emissions. This irreversible temperature change is induced by cancellation of a decreasing airborne fraction of the cumulative emissions and an increasing temperature change per unit atmospheric CO₂ with time. Accordingly, the irreversibility relates in part to the uptake of heat and carbon by the ocean being driven by the same
deep ocean mixing processes over long timescales (Matthews and Caldeira 2008, Matthews et al 2009, Solomon et al 2009). If we are to utilize the proportional relationship described above in the design of feasible and effective CO2 emission pathways, evidence must be produced to indicate that the relationship holds over a wide range of CO2 emission pathways. To achieve this, the independence of climate change with respect to different CO2 emission pathways with equivalent cumulative emission is necessary.

Many studies based on an Earth system model of intermediate complexity (EMIC) have shown that century-scale temperature responses are independent of emission pathways (Eby et al 2009, Zickfeld et al 2009). In contrast, nonlinearities of millennium-scale temperature responses for very high cumulative emissions have been confirmed by EMIC to be due to flushing events in the Southern Ocean (Eby et al 2009). Recently, Zickfeld et al (2012) used a state-of-the-art complex Earth system model to conclude that the response of most surface climate variables such as temperature, precipitation, and sea-ice area is largely independent of emissions pathways once emissions have ceased; their model was fully coupled with biogeochemical processes in the atmosphere–ocean general circulation model.

However, the Earth system incorporates many forms of irreversibility and hysteretic behavior into its climate and carbon cycle responses owing to complex interactions between many internal processes, such as carbon and heat uptakes, general circulation in the atmosphere and ocean, and climate feedbacks (e.g., Frölicher and Joos 2010; Gillett et al 2011; Boucher et al 2012). For example, Gillett et al (2011) reported that temperature change in the Southern Ocean exhibits hysteresis over timescales longer than those in the Northern Hemisphere, owing to deep mixing of heat. As a result, cooling occurs over the terrestrial Northern Hemisphere and the Arctic, with pronounced ongoing warming over the Southern Ocean and Antarctic after cessation of emissions, although globally averaged temperature remains almost constant. Similarly, it has been identified that the heterogeneous response of climate change assists nonuniform spatial distribution of carbon uptake by oceans or land (e.g., Lowe et al 2009; Roy et al 2011; Vichi et al 2011). In addition, uncertainty relating to the relationship between climate change and the carbon cycle is known to contribute to the uncertainty of projections of future climate change based on multimodel intercomparison experiments (Friedlingstein et al 2006, Huntingford et al 2009). Accordingly, there are many uncertainties inherent in the irreversibility and hysteresis of interactions between CO2 emission pathways and the carbon cycle, in addition to those uncertainties that relate to climate change itself in the Earth system. Therefore, this study aims to use a state-of-the-art Earth system model to evaluate the dependency of climate change and the carbon cycle across a range of CO2 emission pathways with the same cumulative emissions.

2. Methods

2.1. Model

The Earth system model used in this study is the National Center for Atmospheric Research Community Earth System Model (CESM) version 1, which incorporates biogeochemical processes fully in the assessment of the climate–carbon cycle of the atmosphere–ocean general circulation model (Hurrell et al 2013). The terrestrial carbon cycle is coupled to the biogeoophysical and hydrological processes in this model, allowing simulation of photosynthesis, respiration, litter, and soil carbon and leaf phenology (Thornton et al 2007). The ocean carbon cycle component is based on the nutrient–phytoplankton–zooplankton–detritus approach represented by the Biogeochemical Element Cycling model (Moore et al 2004). CESM has a uniform horizontal resolution of 1.25º × 0.9º (longitude/latitude grid) with 26 vertical levels in the atmosphere and a horizontal resolution of 1.11º × 0.27º–0.54º with 60 vertical levels in the ocean. No flux adjustments are used in CESM.

In this study, all experiments used prescribed CO2 emission scenarios. Then, a preindustrial control is integrated from the initial state of the CESM (Lindsay et al 2013) for 500 yrs with no anthropogenic carbon input. Although the control performed using CESM is basically stable in terms of climate, a long-term cooling trend for the ocean (∼0.00025 °C yr−1) and an increasing trend in atmospheric CO2 concentration (0.003 ppm yr−1) remain in the control. In the context of the present study, these trends are negligible.

2.2. Emission pathways

Figure 1(a) illustrates the five types of idealized emission pathways applied to diagnose dependency. All other forcings (land cover change, non-CO2 greenhouse gases, and aerosols) are maintained at their preindustrial values. All experiments are calculated for 490 years branched from the control at year 10, with the same cumulative emissions (2000 GtC) from year 10 to 210.

Here, we impose several idealized emission pathways, which are ‘rectangular’ in the sense that they employ an abrupt change, with emissions maintained at a constant level for a given period before being stopped (see figure 1), in order to test carbon and climate responses. The pathways are classified, according to annual emissions, as low, middle, and high. L-RECT is run with annual emissions of 10 GtC yr−1 for 200 years until the cumulative emissions reach 2000 GtC. Subsequently, the annual emissions are reduced suddenly to zero and fixed at zero for 290 years. Annual emissions of 10 GtC yr−1 are comparable to the levels of anthropogenic CO2 emission in the year 2010 (Peters et al 2012). Similarly, M-RECT (H-RECT) is run with 20(40) GtC yr−1 for 100 (50) years before being run with fixed zero emissions for 390 (440) years. OVSHT represents an overshoot scenario for the emission pathways, in which cumulative emissions exceed 2000 GtC temporarily. OVSHT is run with 40 GtC yr−1 annual emissions for 100 years. Subsequently, cumulative
emissions reach 4000 GtC. Thereafter, annual emissions are reduced by negative emissions of $-20$ GtC yr$^{-1}$ and the model is run with these negative emissions for 100 years. Consequently, the cumulative emissions are reduced to 2000 GtC. In PULSE, 2000 GtC emissions are injected in the space of one year at year 10.

### 3. Results and discussion

Figure 1(b) illustrates the globally averaged atmospheric CO$_2$ concentration simulated by CESM. The concentration in the control is remarkably stable for several centuries. The highest peak in atmospheric CO$_2$ concentration (1744 ppm) is reached in OVSHT and is caused by temporal overshoot of 2000 GtC cumulative emissions. Of the other experiments, those with higher annual emissions (such as PULSE) are likely to exhibit higher peak concentrations. The peak concentrations of PULSE, H-RECT, M-RECT, and L-RECT are 1163 ppm, 999 ppm, 932 ppm, and 857 ppm, respectively. A rectangular emission pathway yields a triangular CO$_2$ concentration, reflecting the long lifetime of CO$_2$ and its corresponding accumulation.

Concentrations continue to reduce in all experiments after reaching their corresponding peak concentration. The difference in concentration between the five emission pathways at the end of the simulation is relatively smaller than the difference in peak concentration, with the highest concentrations exhibited by OVSHT and L-RECT (705 ppm) and the lowest by PULSE (686 ppm). With the exception of OVSHT, there is a tendency toward reduction of concentration in pathways with higher annual emissions, which is comparable to that found in a previous study using EMIC (Eby et al. 2009).

Figure 1(c) illustrates annual CO$_2$ uptake by land. Such uptake increases rapidly in all experiments after the initiation of emissions. Higher annual emissions, such as in PULSE, are likely to result in higher peak CO$_2$ uptake. In all experiments except PULSE, CO$_2$ uptakes are maintained or slightly reduced throughout periods of positive annual emissions. In OVSHT, land becomes a source of CO$_2$ during periods of negative emission. After cessation of emissions, CO$_2$ uptakes converge to zero in all experiments. At the end of simulation, about 200 GtC (10% of the cumulative emissions) is sequestered on land in all experiments.

Figure 1(d) illustrates annual CO$_2$ uptake by the ocean. When atmospheric CO$_2$ concentration is increased, CO$_2$ in the atmosphere is absorbed partly by the oceans, at a rate that depends on differences in the partial pressure of CO$_2$ between the atmosphere and the surface of the ocean. Therefore, peak CO$_2$ uptake in all experiments is clearly listed in order of annual emissions. The highest peak in CO$_2$ uptake (66 GtC yr$^{-1}$) is exhibited by PULSE. After peaking, annual CO$_2$ uptakes in higher emission pathways such as OVSHT or H-RECT reduce gradually during the period of positive annual emissions. Since CO$_2$ transport by deep ocean mixing processes is limited, the difference in
Figure 2. Time series of physical climate variables: (a) globally averaged surface air temperature, (b) sea water temperature averaged over the entire ocean, (c) globally averaged net radiation at top of atmosphere, and (d) North Atlantic meridional overturning circulation in Sverdrup (1 Sv = $10^6$ m$^3$ s$^{-1}$). Data are smoothed using a 9 year running mean.

In contrast, the annual CO$_2$ uptake in L-RECT is maintained at 3 GtC yr$^{-1}$ because CO$_2$ uptake can be balanced with deep ocean mixing processes. After the cessation of CO$_2$ emission, the oceans in all experiments continue to absorb CO$_2$ (about 0.5 GtC yr$^{-1}$ in year 500). Until the end of the simulations, the oceans accumulate an additional 900 GtC CO$_2$ in all experiments.

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Figure 2(a) illustrates the temporal evolution of the global surface air temperature anomaly from preindustrial climatology, averaged for 30 years from year 10 to year 39 in the control. The temperature in the control remains approximately constant for several centuries. After cessation of CO$_2$ emissions, the temperature converges to similar values (3.8$^\circ$C peak temperature and 3.6$^\circ$C at the end of simulation) in all experiments except OVSHT. However, the temperature in OVSHT reaches a peak of 5.6$^\circ$C temporarily owing to the temporarily higher CO$_2$ concentration. During the period of negative emission, the temperature reduces gradually to 3.0$^\circ$C; this temperature is lower than that in other experiments for the same period and persists for 150 years. Conversely, after year 400, the temperature is higher than that in the other experiments. In year 420, the temperature in OVSHT exceeds 4.3$^\circ$C. Thereafter, the temperature declines gradually, but remains warmer than that of the other experiments at the end of the simulation.

In contrast, the temporal evolution of seawater temperature averaged over the entire ocean differs in each of the emission pathways (figure 2(b)). The temperature in the control declines continuously as a consequence of climate drift in CESM. All experiments expect OVSHT indicate a general rising trend until the end of simulation. Higher annual emissions, such as in PULSE, are likely to increase the temperature. The temperature in OVSHT almost reaches equilibrium state at year 380, superficially.

To evaluate the differential variation of the surface air temperature in OVSHT, we focus firstly on the global energy balance in the Earth system. Figure 2(c) illustrates the temporal evolution of the global average of annual net radiation at the top of atmosphere. The net radiation increases with a rise in atmospheric CO$_2$ concentration but tends to decrease with increasing temperature owing to increased outgoing longwave radiation following the Planck response. In the long term, the net radiation approaches zero as a result of climate feedbacks caused by changes in distribution of cloud cover, water vapor, and surface albedo in addition to the Planck response. After the cessation of CO$_2$ emission, the net radiation in all experiments reduces gradually to zero. Nevertheless, at the end of simulation, the net radiation is maintained at 0.3 W m$^{-2}$ in all experiments except OVSHT. The principal effect of the positive net radiation is the warming of seawater temperature. In addition, the slightly negative net radiation in the control contributes to
the decreasing trend of seawater temperature. Conversely, in OVSHT, the net radiation increases slightly from year 220 to 300 during the period of cooler temperature, while net radiation is almost zero from year 400 to 500. Such a temporal evolution does not explain the causes of temporal variations in cooling or warming of the surface air temperature in OVSHT.

The variation of the surface air temperature in OVSHT can be interpreted as an impact of the internal processes of climate variability. Figure 2(d) illustrates the temporal evolution of the North Atlantic meridional overturning circulation (AMOC). The AMOC is defined as the maximum stream function of the meridional overturning transport for the Atlantic basin between 30°N and 50°N below 500 m. The AMOC in the control is maintained at about 27 Sv (1 Sv = 10^6 m^3 s^-1) for 500 years. The AMOC is weakened temporarily in all experiments as a result of rapid warming over the Arctic region. Therefore, the maximum weakening of the AMOC is roughly proportional to the peak CO_2 concentration in the following order: OVSHT, PULSE, H-RECT, M-RECT, and L-RECT. The proportionality is similar to that described by Jahn and Holland (2013), who explored the long-term climate impacts of the different emission pathways (referred to as representative concentration pathways or RCP; Meinshausen et al. (2011)). The AMOC recovers to control level gradually after the maximum weakening, except in the case of OVSHT. Although the AMOC in all other experiments recovers after the cessation of CO_2 emissions, that in OVSHT continues to weaken until year 300. The maximum weakening of AMOC in OVSHT is 8 Sv, which corresponds to one-third of the AMOC in the control. The weakening of the AMOC is driven by low-density water formed in the subpolar North Atlantic, induced by increasing freshwater supply due to rapid warming over high-latitude parts of the Northern Hemisphere (supplement 1 available at stacks.iop.org/ERL/8/014047/mmedia). The weakened AMOC and the low-density water formed in the subpolar North Atlantic were also represented in several water-hosing experiments that were designed to study the sensitivity of the AMOC to an external source of freshwater (e.g., Rahmstorf 1995; Stouffer et al. 2006; Hofmann and Rahmstorf 2009). However, it is known that the sensitivity differs among climate models (Plattner et al. 2008; Hofmann and Rahmstorf 2009). During periods of the weakened AMOC, poleward heat transport is decreased remarkably owing to cessation of the North Atlantic Current. Thereafter, the AMOC strengthens to 36 Sv in around year 400, after which the North Atlantic Current activates rather than the control. Finally, the AMOC recovers to the control level gradually. These results suggest that the particular variation of the surface air temperature in OVSHT is caused by internal processes such as the AMOC, rather than by external CO_2 forcing.

Deviations in spatial patterns of surface air temperature and precipitation in OVSHT from those of the control are presented in figure 3. During the period from year 110 to 129, land areas warm faster than oceans, the Northern Hemisphere high-latitude regions experience the greatest warming, and the North Atlantic cools (figure 3(a)). In addition, precipitation increases (decreases) occur in many tropical and high-latitude (subtropical) regions (figure 3(b)). These spatial patterns correspond approximately to the results of other experiments (not shown) and multimodel projections reported in the IPCC fourth assessment report (IPCC 2007b). Incidentally, after cessation of emissions, cooling occurs over land in the Arctic region, and pronounced ongoing warming occurs over the Southern Ocean and Antarctica after cessation of emissions in all experiments except OVSHT (supplement 2 available at stacks.iop.org/ERL/8/014047/mmedia), although the surface air temperature remains almost constant. The spatial patterns are similar to those described by Gillett et al. (2011), who also discussed climate change following cessation of emissions.

The weakened AMOC during year 210–229 induces a decrease in surface air temperature over the northern North Atlantic and northern Europe, as shown in figure 3(c). A maximum cooling of 10°C is indicated over the coastal region of Greenland. Such a decrease in temperature leads to expansion of sea-ice area, which reaches a latitude of 50°N during winter (supplement 3 available at stacks.iop.org/ERL/8/014047/mmedia). In addition, the cooled temperature decreases annual precipitation over Europe and central Eurasia because of a reduction in the supply of water vapor from the North Atlantic. However, the weakened AMOC has little impact on climate change over other regions. Conversely, the temperature over the North Atlantic shifts in the period of strengthened AMOC from year 410 to 429, resulting in a warming climate. Moreover, precipitation over the northern North Atlantic and Europe increases relative to the control. Therefore, variability in the AMOC has an impact on regional climate over northern North Atlantic, Europe, and central Eurasia.

Next, we focus on the impact of regional climate change on the regional carbon cycle. Figure 4 shows cumulative carbon uptake by the oceans over the initial period (year 10–210), during the period of weakened AMOC in OVSHT (year 210–359), and over the entire study period (year 0–499). During the initial period (figure 4(a)), the highest cumulative CO_2 uptake by global oceans is 780 GtC in OVSHT. In contrast, the lowest CO_2 uptake is 600 GtC and occurs in L-RECT. The amounts of global cumulative CO_2 uptake are proportional to the intensity of annual CO_2 emissions, except in the case of OVSHT, because CO_2 uptake depends on the difference in partial pressure of CO_2 between the atmosphere and the sea surface (figure 1(d)). During the period of weakened AMOC (figure 4(b)), the global cumulative CO_2 uptake is reduced in all experiments owing to a reduction in the difference of partial CO_2 pressure. In OVSHT, which is the experiment with the lowest uptake, the global ocean absorbs about 80 GtC over a period of 150 years. However, global cumulative CO_2 uptake over the entire period indicates little dependency on emission pathways (figure 4(c)).

For the Atlantic and the Arctic oceans, it is assumed that regional carbon uptake is affected by regional climate change due to the weakened AMOC in OVSHT. The CO_2 uptake in these regions is climatologically positive: it is controlled by deep water formation processes such as winter mixing and the biological pump. During the initial period (figure 4(d)), the
Atlantic and Arctic oceans in all experiments absorb CO$_2$ of 140–150 GtC. CO$_2$ uptake by the Atlantic and Arctic oceans represents only 20% of global CO$_2$ uptake. CO$_2$ uptake in OVSHT becomes negative during the period of weakened AMOC (figure 4(e)). Finally, for the entire period shown in figure 4(f), deviations in the regional cumulative CO$_2$ uptake of each experiment expand relative to those in global cumulative CO$_2$ uptake. Therefore, the weakened AMOC reduces CO$_2$ uptake by the Atlantic and Arctic oceans through regional climate change, but the reduction in regional CO$_2$ uptake has a relatively small effect on the global carbon cycle.

Figure 5 illustrates the spatial distribution of cumulative CO$_2$ flux from the atmosphere to the land and ocean for 150 yr over the period from year 210 to year 359. During this period, CO$_2$ emissions are maintained at zero in all experiments. CESM in the control (figure 5(a)) reproduces roughly the observation-based distribution of ocean CO$_2$ flux reported by Takahashi et al (2009). In L-RECT, areas of positive flux anomaly compared to the control are distributed in many parts of the eastern tropical Pacific, subtropical Atlantic, and Southern Ocean (figure 5(b)). Conversely, areas of negative flux anomaly are distributed in the northern Atlantic and northwestern Pacific.

Deviations in the cumulative CO$_2$ flux of each experiment with respect to L-RECT are illustrated in figures 5(c)–(f). In PULSE, H-RECT, and M-RECT, the differences are remarkably small, although cessation of emissions in these experiments is achieved more than 100 years earlier than in L-RECT (figure 1(a)). Conversely, in OVSHT, areas of negative flux anomaly are distributed over the northern North Atlantic, Arctic Ocean, and Southern Ocean (figure 5(d)), whereas areas of positive flux anomaly extend from south...
Figure 4. Cumulative CO$_2$ uptake by (upper) global oceans and (lower) Atlantic and Arctic oceans for years (left) 10–209, (middle) 210–359, and (right) 0–499.

of Iceland to west of Ireland. In OVSHT, rapid warming in the Arctic region over the period from year 10 to 210 forms vertically stabilized shallow water in the northern Atlantic (supplement 1 available at stacks.iop.org/ERL/8/014047/mmedia). The stabilized water not only enhances weakening of AMOC but also reduces vertical mixing of heat and carbon in this region. As a result, CO$_2$ uptake in the North Atlantic region is reduced. This relationship between the weakened AMOC and the carbon cycle is consistent with previous studies, such as those of Obata (2007) and Zickfeld et al. (2008).

4. Summary

Dependency of climate and carbon cycle change on CO$_2$ emission pathways is evaluated using CESM, a fully coupled climate–carbon cycle model. To assess this dependency, five types of idealized CO$_2$ emission pathways (including one with an overshoot scenario) with cumulative emissions of 2000 GtC are designed. Although the atmospheric CO$_2$ concentration increases rapidly in pathways with higher annual emissions (such as PULSE) relative to those with lower annual emissions (such as L-RECT), concentrations in all experiments are reduced gradually as a result of CO$_2$ uptake by the oceans and land after reaching peak concentration. Eventually the concentration converges to similar values in all experiments. Therefore, atmospheric CO$_2$ concentration can be regarded as independent of emission pathways. Moreover, globally averaged surface air temperature converges to the same value in all experiments, except OVSHT. Such independence from emission pathways corresponds to the results of a previous study by Zickfeld et al. (2012).

However, we found that regional climate change such as that reflected by variations in temperature and the AMOC in OVSHT is different from that suggested by our other experiments. The overshoot scenario of cumulative emissions in OVSHT produces temporarily higher atmospheric CO$_2$ concentration than that in other experiments. This higher CO$_2$ concentration induces much a weaker AMOC that is enhanced by rapid warming in the Arctic region. Since the low-density water formed in the subpolar North Atlantic is stable, the weakened AMOC persists irreversibly for 150 years after the cessation of CO$_2$ emission. The weakened AMOC causes local decline in surface air temperature and decreases in precipitation over the northern Atlantic and Europe region. These results indicate that regional climate
change is sensitive to CO$_2$ emission pathways through the long-term pathways of CO$_2$ concentration. Moreover, the weakened AMOC reduces CO$_2$ uptake by the Atlantic and Arctic oceans. However, the reduction in regional CO$_2$ uptake has a relatively small effect on the global carbon cycle. In conclusion, although dependency on CO$_2$ emission pathways is evident in climate change patterns, the global carbon cycle is relatively independent of CO$_2$ emission pathways overall.

However in the present study, the dependency of CO$_2$ emission pathways is evaluated using only a single Earth system model, yet differences between similar models can be considerable. In particular, comparison between Earth system models has indicated significant differences in the representation of the carbon cycle (Friedlingstein et al. 2006). To accurately evaluate the details of this dependency, similar analysis must be conducted over a broader range cumulative CO$_2$ emissions based on multimodel ensembles incorporating Earth system models.

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