Examination of a climate stabilization pathway via zero-emissions using Earth system models

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Keywords: zero emissions, cumulative emissions, Earth system model, carbon cycle, climate stabilization

Abstract
Long-term climate experiments up to the year 2300 have been conducted using two full-scale complex Earth system models (ESMs), CESM1(BGC) and MIROC-ESM, for a CO2 emissions reduction pathway, termed Z650, where annual CO2 emissions peak at 11 PgC in 2020, decline by 50% every 30 years, and reach zero in 2160. The results have been examined by focusing on the approximate linear relationship between the temperature increase and cumulative CO2 emissions. Although the temperature increase is nearly proportional to the cumulative CO2 emissions in both models, this relationship does not necessarily provide a robust basis for the restriction of CO2 emissions because it is substantially modulated by non-CO2 forcing. CO2-induced warming, estimated from the atmospheric CO2 concentrations in the models, indicates an approximate compensation of nonlinear changes between fast-mode responses to concentration changes at less than 10 years and slow-mode response at more than 100 years due to the thermal inertia of the ocean. In this estimate, CESM1(BGC) closely approximates a linear trend of 1.7 °C per 1000 PgC, whereas MIROC-ESM shows a deviation toward higher temperatures after the emissions peak, from 1.8 °C to 2.4 °C per 1000 PgC over the range of 400–850 PgC cumulative emissions corresponding to years 2000–2050. The evolution of temperature under zero emissions, 2160–2300, shows a slight decrease of about 0.1 °C per century in CESM1(BGC), but remains almost constant in MIROC-ESM. The fast-mode response toward the equilibrium state decreases with a decrease in the airborne fraction owing to continued CO2 uptake (carbon cycle inertia), whereas the slow-mode response results in more warming owing to continued heat uptake (thermal inertia). Several specific differences are noted between the two models regarding the degree of this compensation and in some key regional aspects associated with sustained warming and long-term climate risks. Overall, elevated temperatures continue for at least a few hundred years under zero emissions.

1. Introduction
Numerous studies have suggested that the globally averaged temperature response is approximately proportional to cumulative CO2 emissions (e.g., Matthews and Caldeira 2008, Allen et al 2009, Matthews et al 2009, Solomon et al 2009). This means that the temperature increases with CO2 emissions and that the elevated temperature would remain approximately constant even if emissions reach zero in the future. This proportional relationship essentially arises from the compensation of radiative forcing change and thermal inertia of the ocean (Matthews et al 2009), which involves several nonlinear effects such as the logarithmic nature of CO2 forcing, changes in the airborne fraction of CO2 resulting from carbon cycle processes and their interaction with climate, and long-term evolution of ocean heat uptake. Recent studies have explored these effects quantitatively by using analytical formulation and have discussed the reasoning behind the compensation (Goodwin et al 2015, MacDougall and Friedlingstein 2015). In the framework of cumulative CO2 emissions, the role of non-CO2 greenhouse gases has also been considered.
regarding emission trajectories (Raupach 2013) and the persistent nature of warming (Solomon et al 2010, Frölicher and Joos 2010).

The proportional relationship, or persistent warming following the cessation of CO\(_2\) emissions, has been well established by numerical experiments with Earth system models (ESMs) of intermediate complexity (e.g., Plattner et al 2008, Eby et al 2009, Zickfeld et al 2013, Herrington and Zickfeld 2014).

Although full-scale complex ESMs essentially support these findings (Love et al 2009, Gillett et al 2011, Gillett et al 2013), exceptional behavior has been found in the long-term temperature response depending on the model and emissions pathway used in a particular experiment (Nohara et al 2013, Frölicher et al 2014). Despite the uncertainties, using complex ESMs is the only way to explore the spatial distribution of the temperature response and other relevant climate variables (e.g., Zickfeld et al 2012).

Among others, Nohara et al (2013) reported that the behavior of the Atlantic meridional overturning circulation (AMOC) after the cessation of emissions depends on the emissions pathways even for the same total amount and that weakening of AMOC contributes little to changes in the global ocean carbon cycle.

These past studies are largely based on idealized emissions pathways resulting in a large signal-to-noise response to facilitate better analysis of the model behavior. In contrast, the present study focuses on the response of complex ESMs to more plausible emissions pathways in the context of climate change mitigation policies and socioeconomic feasibility. One such emissions pathway, Z650, has been proposed and discussed by Matsuno et al (2012a, 2012b). Z650 is designed on the basis of the ‘zero-emissions stabilization’ concept, which targets the reduction of CO\(_2\) emissions to zero in the distant future, typically in the middle of the 22nd century. Under this framework, the atmospheric CO\(_2\) concentration is reduced by natural removal processes and eventually reaches stable equilibrium. The amount of CO\(_2\) emissions is 652 PgC during the 21st century, and the cumulative total from the middle of the 19th century, a common starting point for historical climate experiments, is 1101 PgC. The name ‘Z650’ implies reaching zero following approximately 650 PgC during the 21st century.

Z650 is intended to allow more emissions during the 21st century from a socioeconomic viewpoint compared with the rather stringent mitigation scenarios such as RCP2.6, the lowest forcing level among the four representative concentration pathways (RCPs) (Moss et al 2010). In fact, the CO\(_2\) emissions of 650 PgC are between RCP2.6 and the second-lowest, RCP4.5. According to the scenario categories of the Intergovernmental Panel on Climate Change Working Group III (Clarke et al 2014), Z650 belongs to the 530–580 ppm CO\(_2\)eq category in case of a simple climate modeling exercise conducted by Matsuno et al (2012b), in which non-CO\(_2\) forcing of 0.65 W m\(^{-2}\) is assumed at the end of the 21st century and beyond.

Moreover, Z650 is intended to avoid long-term serious climate risks, such as ice-sheet melting in Greenland, by a long-lasting decrease in the atmospheric CO\(_2\) concentration under zero emissions and a subsequent decrease in the temperature by about 1 °C or less on a multiple century timescale. This concept, however, may not be compatible with the persistent nature of the temperature response. The present study investigates the details of climate system behavior in the Z650 pathway by using two different ESMs and discusses the implications for long-term climate stabilization in light of the zero-emissions stabilization concept.

2. Method

2.1. Models

The ESMs used in this study are the National Center for Atmospheric Research Community Earth System Model version 1 Biogeochemistry (CESM1(BGC)) (Hurrell et al 2013) and the ESM based on Model for Interdisciplinary Research on Climate (MIROC-ESM; Watanabe et al 2011). Both models are included in the fifth phase of the Coupled Model Intercomparison Project (CMIP5) (Taylor et al 2012) and are categorized as among the most comprehensive ESMs, incorporating full biogeochemical processes with no flux adjustments to represent the climate–carbon cycle in atmosphere–ocean general circulation.

CESM1(BGC) has a uniform horizontal resolution of 1.25 × 0.9° (zonal/meridional grid) with 26 vertical levels in the atmosphere and a horizontal resolution of 1.11° × 0.27–0.54° with 60 vertical levels in the ocean. The terrestrial carbon cycle is coupled to biogeophysical and hydrological processes, simulating 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).

MIROC-ESM has a horizontal resolution of T42, approximately 2.8°, with 80 vertical levels in the atmosphere and a horizontal resolution of 1.4° × 0.5–1.7° with 44 levels in the ocean. MIROC-ESM is coupled with the spatially explicit individual-based dynamic global vegetation model (SEIB-DGVM) for the terrestrial ecosystem (Sato et al 2007), which adopts a scheme that explicitly captures light competition among trees rather than using parameterized schemes. The ocean carbon cycle component in this model is also based on the nutrient–phytoplankton–zooplankton–detritus approach (Oschlies 2001).

Table 1 summarizes the primary features of climate change and the carbon cycle in CESM1(BGC) and MIROC-ESM. The values for equilibrium climate sensitivity (\(T_{\text{ES}}\)) is defined as the local
The carbon sensitivity is the atmospheric CO2 in response to CO2 emissions, as evaluated by Arora et al. For carbon feedback, positive weak carbon sensitivity was estimated by Forster and moderate carbon sensitivity was determined by the strength of natural carbon sinks ([Levis 2010](#)). The climate drift of the temperature anomalies has been removed. The carbon flux and temperature data are smoothed using a 9-year running mean.

Table 1. Primary features of climate change and the carbon cycle.

| Equilibrium climate sensitivity (°C) | CESM1(BGC) | MIROC-ESM |
|-------------------------------------|------------|------------|
| Adjusted radiative forcing (2×CO2) (Wm$^{-2}$) | 2.89 | 4.67 |
| Carbon sensitivity (terrestrial) | High | Moderate |
| Carbon sensitivity (ocean) | Moderate | Moderate |
| Carbon-climate feedback | Positive weak | Positive strong |
| Reference | Hurrell et al (2013) | Watanabe et al (2011) |

* Estimated by Forster et al (2013).
* Based on a multi-model intercomparison by Arora et al (2013).

Figure 1. Time series of biogeochemical and physical climate variables: (a) CO2 emission pathways, (b) cumulative CO2 emissions, (c) atmospheric CO2 concentration, (d) cumulative carbon uptake by land, (e) cumulative carbon uptake by ocean, and (f) globally averaged surface air temperature anomalies relative to the period 1850–1869. Panel (a) contains historical emissions from fossil fuels combustion and land use changes and two of representative concentration pathways. Panels (a), (c)–(f) contain reference data: (a), (c)–(f) contain reference data: (a), (c) Meinshausen et al (2011), (d) and (e) cumulative carbon uptake at present by the best estimate (open circle) and uncertainty (bar) (Ciais et al 2013), (f) HadCRUT4 (Morce et al 2012). The climate drift of the temperature anomalies has been removed. The carbon flux and temperature data are smoothed using a 9-year running mean.

The equilibrium surface–air–temperature change in response to doubling of the atmospheric CO2 concentration, are 2.89 °C and 4.67 °C in CESM1(BGC) and MIROC-ESM, respectively (Andrews et al 2012, Forster et al 2013). The carbon sensitivity in CESM1(BGC) (MIROC-ESM) is moderate (highest) in the CMIP5 model range, from 2.08 °C to 4.67 °C (Andrews et al 2012, Forster et al 2013). The carbon sensitivity and climate–carbon feedback were evaluated by Arora et al (2013). The carbon sensitivity is defined as the increase in atmospheric CO2 in response to CO2 emissions, as determined by the strength of natural carbon sinks (Matthews et al 2009). The climate–carbon feedback is defined as the reduction of terrestrial and oceanic CO2 uptake caused by the increase in temperature. Among CMIP5 ESMs, the terrestrial model processes in CESM1(BGC) result in higher carbon sensitivity, which is explained mainly by a suppressed CO2 fertilization effect due to nitrogen limitation (Bonan and Levis 2010). In contrast, MIROC-ESM has a moderate carbon sensitivity on land; that in the ocean is moderate for the both models. The behavior of the carbon–climate feedback differs considerably between CESM1 (BGC) and MIROC-ESM. In positive carbon–climate feedback, higher temperatures increase the flux of carbon from the land and ocean into the atmosphere, and the resulting higher atmospheric CO2 concentration enhances warming. Although both models have positive feedback, its magnitude is weak in CESM1(BGC) but strong in MIROC-ESM. On the whole, CESM1 (BGC) and MIROC-ESM have a tendency to project a higher airborne fraction of cumulative emissions among CMIP5 ESMs (Arora et al 2013).

2.2. Experimental design

Figure 1(a) shows a time series of Z650 carbon emissions from the year 2005 following historical carbon emissions by fossil fuels and land use change; this series is compared with RCP 2.6 and RCP 4.5 from the year 2005 and their extension to 2300 (Meinshausen et al 2011). The Z650 pathway, located between RCP 2.6 and RCP 4.5, peaks in 2020 with 11 PgC and reaches zero in 2160. The amount of annual emissions declines during this period roughly at a rate of 50% every 30 years. Figure 1(b) shows a time series of the cumulative total of the historical and Z650 CO2 emissions.
emissions. The cumulative emissions reach 400 PgC in the year 2000 and 860 PgC in the year 2050.

We conducted historical and Z650 emission-driven experiments by using CESM1(BGC) and MIROC-ESM. These models prognostically compute global atmospheric CO2 mole fractions, which represent an integration of physical, chemical, and biological processes with land use change, and their interactions and feedbacks with the climate system. The historical experiments, corresponding to experiment 5.2 or esmHistorical of CMIP5 (Taylor et al. 2012), are forced by spatially distributed CO2 emissions reconstructed from fossil fuel consumption estimates (Andres et al. 2011) since 1850. The historical forcing also includes land use change and concentrations of non-CO2 greenhouse gases such as methane, nitrous oxide, and halocarbons in addition to aerosols. At the end of the historical period, CO2 forcing is altered to match Z650, which is implemented as the reconstructed fossil emissions in 2005 scaled by the global total emissions along the Z650 pathway. Land use is not changed after 2005, although some delayed emissions from land use changes in the historical period are accounted for. Other non-CO2 greenhouse gases and aerosols follow RCP 2.6 (van Vuuren et al. 2011) for the period 2005–2100 and are kept constant after 2100.

### 3. Results and discussion

#### 3.1. Biogeochemical and climate changes

The atmospheric CO2 concentrations calculated by the two ESMs both show a peak-and-decline evolution in response to the Z650 CO2 emissions, as illustrated in figure 1(c). The peak level and time in CESM1 (BGC) and MIROC-ESM are 520 ppm in the year 2080 and 540 ppm in the year 2120, respectively. This difference simply reflects the different amounts of carbon uptake by natural processes because land use is unchanged by the experiment design. At the end of the experiment period, at year 2300, the concentrations and their decreasing trends since the emissions level has reached zero, at year 2160, are about 455 ppm and 510 ppm, and $-0.32$ ppm yr$^{-1}$ and $-0.16$ ppm yr$^{-1}$, respectively, in CESM1(BGC) and MIROC-ESM (table 2). It should be noted that both models have a positive bias; the concentration at the end of the historical period is 20 ppm and 12 ppm, respectively, above the observed level compiled by Meinshausen et al. (2011).

| CO2 concentration (ppm yr$^{-1}$) | Temperature ($^\circ$C per century) |
|-----------------------------------|-------------------------------------|
| CESM1(BGC)                        | MIROC-ESM                          |
| $-0.32$                            | $-0.16$                             |
| $-0.083$                           | $0.004$                             |

The amounts of accumulated land and ocean carbon are shown in figures 1(d) and (e) with observed estimates in the year 2010 (Clais et al. 2013) for reference. At this reference time for both models, land carbon is underestimated whereas ocean carbon is comparable to the observed estimates. In response to the Z650 emissions under fixed land use, land carbon increases toward an apparent equilibrium level in CESM1(BGC), whereas it peaks by the end of the 21st century, then declines to negative values in MIROC-ESM owing to its strong carbon–climate feedback. The cumulative carbon uptake by the year 2300 is 332 PgC in CESM1(BGC) and $-35$ PgC in MIROC-ESM.

Carbon accumulation in the ocean continues throughout the experiment period in both ESMs. However, a decline in the accumulation growth rate noted in both models is more discernible in CESM1 (BGC). This difference is understood to originate from an interaction of the land and ocean uptake processes such that a larger land uptake in CESM1(BGC) makes the atmospheric concentration lower, which suppresses the ocean uptake owing to a smaller difference of CO2 pressure between the atmosphere and ocean. The cumulative carbon uptake by the year 2300 is 475 PgC in CESM1(BGC) and 520 PgC in MIROC-ESM.

Figure 1(f) displays the temporal evolution of the globally averaged surface air temperature anomalies relative to the years 1850–1869. Climate drift components, obtained from a linear trend during the period 1850–2300 in the preindustrial control of $-0.008$ °C per century in CESM1(BGC) and 0.102 °C per century in MIROC-ESM, were removed. In the year 2005, CESM1(BGC) is about 0.4 °C higher than observation by HadCRUT4 (Morice et al. 2012), although MIROC-ESM is comparable to observation. After a rapid increase until about 2050, both models show decelerated warming. In CESM1(BGC), the temperature anomaly peaks at about 2.5 °C around the year 2100, then slightly decreases at a rate of about 0.1 °C per century (table 2). The temperature anomaly in MIROC-ESM peaks at about 3.5 °C around the year 2130 and subsequently remains almost constant. Overall, both ESMs show elevated temperature anomalies that continue for at least a few hundred years in response to the Z650 emissions pathway. This long-term tendency is essentially consistent with the energy budget of the model, which includes radiative forcing and ocean heat uptake (supplement 1). In addition, this trend is explained by the thermal inertia of the ocean and the logarithmic nature of CO2 forcing, which results in a small decrease in forcing relative to the concentration decrease (Solomon et al. 2010).

#### 3.2. Proportionality of the transient climate response to cumulative emissions

Figure 2(a) shows globally averaged surface air temperature anomalies as a function of cumulative carbon
emissions. Although both ESMs show a somewhat proportional relationship between temperature change and cumulative emissions, noticeable differences are apparent in the trajectories on the temperature–cumulative emissions diagram. Overall, the trajectory is curved downward (upward) in CESM1 (BGC) (MIROC-ESM). Figure 2(b) shows the components of CO₂-induced change and residual one on the same diagram as that shown in figure 2(a), which were calculated by using a simple climate model (SCM) with an assumption of linear forcing additivity (supplement 2). The component trajectories for the CO₂-induced change in both ESMs are approximately consistent with those from the 1% per year CO₂ increase experiments of CMIP5 (figure 1(a) in Gillett et al 2013). The CO₂-induced temperature anomaly in response to 1000 PgC emissions, termed transient climate response to cumulative carbon emissions (TCRE), is 1.8 °C in CESM1(BGC) and 2.2 °C in MIROC-ESM, ranked as middle and high, respectively, across the CMIP5 models.

A comparison of the component and the total trajectories reveals that the larger total warming in CESM1(BGC) despite its lower climate sensitivity from 200 PgC to 700 PgC in 1970–2030 period is due mainly to different implementation of non-CO₂ forcing. Other factors relevant to this anomalous response include different response characteristics (supplement 2) and decadal climate variability. Although the two ESMs deal with common forcing agents, the cooling effect due to aerosol–cloud interaction is not implemented in CESM1(BGC) (table 12.1 in Collins et al 2013). The aerosol–cloud interaction is quantified as −0.45 Wm⁻² with 90% uncertainty of −1.2 Wm⁻² to 0.0 Wm⁻² in terms of effective radiative forcing from 1750 to 2011 (Myhre et al 2013). Apart from the uncertainty of the forcing magnitude and its efficacy on the temperature response, the absence of this cooling effect is large enough to create a warm bias in the early 21st century in CESM1(BGC) (Hurrell et al 2013). In the later period, when non-CO₂ forcing does not much affect differences between the two models, the residual component for MIROC-ESM is comparable to or higher than that for CESM1(BGC), where decadal fluctuations imply the models’ internal variability. Trajectories for the CO₂-induced temperature indicate an approximately linear relationship in CESM1(BGC), while MIROC-ESM shows a deviation toward higher temperatures after the emissions peak.

Over the range of 400–850 PgC corresponding to years 2000–2050, the warming rate per 1000 PgC is 1.7 °C in CESM1(BGC), while changes from 1.8 °C to 2.4 °C at about 600 PgC in MIROC-ESM.
Compared with idealized emissions pathways in existing studies, Z650 is a plausible mitigation pathway in which the amount of annual CO₂ emissions peaks in the year 2020 and reaches zero after 140 years. Moreover, the atmospheric CO₂ concentration experiences periods of rapid growth, decelerating growth, and decline. Here, we decompose the CO₂-induced temperature response into fast and slow components with distinct timescale separation to facilitate the analysis of the temperature and cumulative emissions relationship in these different phases. The temperature response in the SCM is formulated as the sum of three exponentials. We define the fast and slow mode responses as the sum of two exponentials with a smaller time constant of less than 10 years and that with a longer time constant of more than 100 years, respectively. Normalized amplitudes of the fast and slow mode responses are about 0.6 and 0.4, respectively, for both ESMs (table S1). Because the fast component roughly corresponds to changes in radiative forcing, its relationship to cumulative CO₂ emissions is close to a logarithmic function, depending on the extent to which the airborne fraction is stable. In addition, the fast-mode response reflects the continued CO₂ uptake, or carbon cycle inertia, predominantly by the ocean, which alters the airborne fraction. The slow component represents delayed response due to thermal inertia. Because the emissions reduction period of 140 years is comparable to the timescale of the slow component, this response essentially increases nonlinearly in terms of the cumulative emissions.

The decomposed temperature response and the airborne fraction as a function of cumulative emissions, are shown in figures 2(c) and (d). The first half of the 21st century is characterized by a peak-and-decline annual emissions change, with a peak at 11 Pg C in 2020 and a decrease to 8 Pg C in 2040. In this period, the atmospheric CO₂ concentration increases, although the rate of increase gradually becomes smaller. In addition, the carbon cycle in CESM1(BGC) behaves such that the airborne fraction remains approximately constant and then declines, whereas that in MIROC-ESM results in a temporary increase at about 600 Pg C, corresponding to the peak year of 2020. The fast temperature response should increase logarithmically for a constant airborne fraction. This logarithmic trend, however, does not deviate significantly from the linear trend for additional emissions of a few hundred petagrams of carbon during this period. In the case of MIROC-ESM, the airborne fraction increase overcomes the logarithmic forcing effect, and results in an increased gradient for the fast-mode response trajectory as well as the total one discussed above.

In the second half of the 21st century, characterized by a continued annual emissions reduction at a constant time rate of 50% every 30 years to 2 Pg C in 2100, long-term changes associated with carbon cycle inertia and thermal inertia appear more evident. The airborne fraction gradually decreases in MIROC-ESM. However, this decreasing rate is considerably smaller than that in CESM1(BGC), which causes the divergence in the fast-mode responses between the two ESMs. The slow temperature response increases with cumulative emissions at a greater rate than in the earlier period. This nonlinear increase compensates for the logarithmic fast response, resulting in the approximate linear relationship between temperature and cumulative emissions in this period for both ESMs. In the case of CESM1(BGC), this compensation is already discernible in the first half of the 21st century. The slow-mode response is greater in CESM1(BGC) because of its smaller time constant (table S1), which has a greater effect than the difference in climate sensitivity for a transient state.

Beyond the 21st century, characterized by the achievement of zero emissions in year 2160 and evolution toward a quasi-equilibrium state on a millennial timescale, the thermal inertia increases the temperature in the slow-mode response whereas the carbon cycle inertia compensates this increase through the fast-mode response. Regarding the magnitude of this compensation under zero emissions in 2160–2300, noticeable differences are apparent between the two models, as indicated by trajectories along the right edge of the diagrams in figure 2. Although the airborne fraction decreases in both ESMs, this decrease and its effect on the fast-mode response are smaller in MIROC-ESM. Moreover, the increase in the slow-mode response is greater in MIROC-ESM, which is explained by its higher climate sensitivity and longer time constant. These differences result in different developments of the CO₂-induced warming, shown as a subtle decrease (increase) in CESM1(BGC) (MIROC-ESM) of 0.07 °C (0.1 °C) per century.

These changes still represent a transient stage from the viewpoint of climate stabilization on a millennial timescale. It is expected that long-term processes will continue beyond 2300 and that the temperature response trends may continue or move in a different direction.

3.3. Some remarks on regional changes

The Z650 concept regarding long-term climate change mitigation considers that a severe climate risk due to melting of the Greenland ice sheet is avoidable as long as sustained high temperatures are limited within a few hundred years. Here, we focus on several regional aspects associated with sustained changes and those in the Arctic region.

Figure 3 illustrates the spatial distribution of temperature change during the zero emissions period of 2160 to 2300. Both ESMs generally project a decreasing temperature trend in the Northern Hemisphere because of the reduction of radiative forcing. In contrast, the Southern Hemisphere high-latitude region surrounding the Antarctic experiences continuous
warming caused by thermal inertia of the ocean. Although these spatial distributions are generally consistent with the previous zero emission study by Gillett et al. (2011), several regional differences exist between the two models, such as that shown in the Southern Ocean. These differences might affect the global temperature response, considering that other modeling studies focusing on the role of the ocean circulation in the heat uptake (Winton et al. 2013, Frölicher et al. 2015).

Figure 4 shows the temporal evolutions of selected regional variables. Although the hemisphere-scale...
surface air temperatures show a similar long-term tendency between the two models (figure 4(a)), the temperatures over some regions, typically those in the northern Atlantic area surrounding Greenland (figure 4(b)), show opposing temperature trends (figure 3). After the year 2100, the temperature slightly decreases in MIROC-ESM but remains almost constant in CESM1(BGC). In addition, the temperature in MIROC-ESM is lower than in CESM1(BGC) despite the higher equilibrium climate sensitivity of MIROC-ESM. In general, it is expected that the decreasing temperature trend in the Northern Hemisphere reduces the risk of Greenland ice sheet melting, even though elevated temperatures continue globally for a few hundred years or more. However, such melting might be affected by several local aspects that are subject to model dependency and uncertainties.

The temperature in the northern Atlantic region is sensitive to AMOC, which is quantified by the maximum stream function of the meridional overturning transport for the Atlantic basin between 30°N and 50°N below the depth of 500 m (figure 4(c)). In CESM1(BGC) the strength of AMOC weakens temporarily and recovers to a preindustrial level. However, that in MIROC-ESM gradually weakens throughout the experiment period and reaches 4 Sv at the end, which corresponds to one-fourth of the preindustrial AMOC level. This different response may be related to the AMOC stability of models: MIROC-ESM has a characteristic bistable regime, whereas CESM1(BGC) has a monostable regime (Weaver et al. 2012). The relatively lower temperatures surrounding Greenland in MIROC-ESM are explained by decreased poleward heat transport associated with the weakened AMOC. This finding implies that internal processes, rather than external CO2 forcing, are crucial for temperature changes in this particular region.

Arctic sea ice in September disappears completely in MIROC-ESM by the year 2050. In CESM1(BGC), however, it reaches a minimum of about $3 \times 10^6$ km$^2$, or about 40% of the preindustrial level (figure 4(d)). Although the Arctic sea ice area in CESM1(BGC) slightly recovers after the year 2100, MIROC-ESM projects no sign of recovery. In either case, however, changes in the Arctic sea ice do not significantly affect the temperature trend in the Arctic region (figure 3). The Antarctic sea ice area in March in CESM1(BGC) and MIROC-ESM relatively reaches minimums of about 70% and 60% of the preindustrial level (figure 4(e)) at the end of the 21st century and later remains approximately constant in both models.

4. Conclusion

In this study, we conducted emission-driven experiments by using CESM1(BGC) and MIROC-ESM for the Z650 pathway, which is an illustrative long-term pathway of plausible CO2 emissions reduction. In addition, we discussed global changes in the climate and carbon cycle in light of the approximately linear relationship between the temperature increase and cumulative CO2 emissions. The Z650 annual CO2 emissions peak at 11 PgC in 2020, allowing more emissions in the early part of the 21st century compared with the stringent mitigation scenarios such as RCP2.6 and assuming long-term climate change mitigation by natural CO2 removal processes under zero emissions. Therefore, the key points are the transient temperature increase during the 21st century and the subsequent long-term temperature change under zero emissions. Because the Z650 concept for long-term climate mitigation considers the risk of Greenland’s ice sheet melting, several regional aspects were also examined.

Although the transient temperature increase is nearly proportional to cumulative CO2 emissions, this linear relationship is not necessarily a good indicator for an early part of the 21st century because it is modulated considerably by non-CO2 forcing. Model biases during the historical period also affect near-future projections. CO2-induced warming, estimated from the atmospheric CO2 concentration by using a SCM, indicates a clearer linear relationship between temperature and cumulative CO2 emissions. This linearity arises from the compensating effects of the fast (slow) mode for the temperature response with modulations by the airborne fraction, as indicated respectively by the logarithmic function (nonlinear upward-increase) shape. These components depend on characteristics of models such as climate and carbon sensitivity, climate–carbon cycle feedback, and timescales of the temperature response, as well as on emission pathways. In the case of Z650, CESM1(BGC) closely approximates a linear trend, whereas MIROC-ESM shows a deviation toward higher temperatures after the annual emissions peak due to an increase in the airborne fraction.

Regarding the long-term temperature change under zero emissions, the evolution of temperature until 2300 shows a slight decrease of about 0.1 °C per century in CESM1(BGC) compared with an almost constant value in MIROC-ESM. These changes are also affected by changes in non-CO2 forcing during the 21st century. Analysis of the simple model for CO2-induced warming indicates that the fast-mode temperature decreases in response to a decline in the airborne fraction due to carbon cycle inertia. This fast-mode decrease is largely compensated by the slow-mode increase due to thermal inertia. The magnitude of the relationship between the slow and fast modes is different between the two ESMs, resulting in different developments in the CO2-induced warming, marked by a subtle decrease (increase) in CESM1(BGC) (MIROC-ESM) of 0.07 °C (0.1 °C) per century.

Overall, elevated temperatures continue for at least a few hundred years under zero emissions. The increasing trend in MIROC-ESM may be considered a
conservative estimate in the context of climate mitigation policy because the model’s characteristics tend to produce higher airborne fractions and temperatures among the current ESMs. Nevertheless, both models project a decreasing temperature trend in the Northern Hemisphere, which reduces the risk of Greenland ice sheet melting from a long-term perspective. In the vicinity of Greenland, several differences exist between the two models associated with AMOC and the Arctic sea ice, which should be examined for both near-term and long-term climate risks. However, these aspects do not significantly affect global temperature. Because the Z650 is a plausible emissions pathway with a moderate level of cumulative emissions, these conclusions may differ from the findings of other idealized experiments forced by a much larger emissions level and its abrupt change.

Acknowledgments

We are grateful for discussions with and comments from A Oka at university of Tokyo, F O Bryan at the National Center for Atmospheric Research (NCAR), M Kawamichi at the Japan Agency for Marine-Earth Science and Technology, and M Ohba, Y Yoshida, and N Nakashiki at the Central Research Institute of Electric Power Industry (CRIEPI). CESM1(BGC) experiments were supported by joint collaborative research between CRIEPI and NCAR. This work was supported by the KAKUSHIN and SOUSEI Program of the Ministry of Education, Culture, Sports, Science, and Technology in Japan.

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