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Increase of the transient climate response to cumulative carbon emissions with decreasing CO₂ concentration scenarios

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Abstract

Near-constancy of the transient climate response to cumulative carbon emissions (TCRE) facilitates the development of future emission pathways compatible with temperature targets. However, most studies have explored TCRE under scenarios of temperature increase. We used an Earth system model (MIROC-ESM) to examine TCRE in scenarios with increasing and stable CO₂ concentrations, as well as overshoot pathways in which global mean temperatures peak and decline. Results showed that TCRE is stable under scenarios of increasing or stable CO₂ concentration at an atmospheric CO₂ concentration (pCO₂) double the pre-industrial level. However, in the case of overshoot pathways and a stable pCO₂ scenario at a quadrupled pCO₂ level, the TCRE increases by 10%–50%, with large increases over a short period just after pCO₂ starts to decrease. During the period of pCO₂ increase, annual ocean heat uptake (OHU) and ocean carbon storage (Cᵪ) (or cumulative ocean carbon uptake from the start of the experiment) exhibit similar changes, resulting in a stable TCRE. During the pCO₂ decrease period, after a sudden TCRE increase when pCO₂ starts to decrease, the OHU decreases and Cᵪ increases (relative to the pCO₂ increase period) balance each other out, resulting in a stable TCRE. In overshoot pathways, the temperature distribution when the global mean temperature anomaly cools to 1.5 °C reveals small warming over land and large warming over the oceans relative to the 1% per annum pCO₂ increasing scenario, particularly in some high-latitude areas of both hemispheres. The increase in TCRE with overshoot pathways decreases the carbon budget for the temperature anomaly targets in such scenarios. Our analysis showed a 16%–35% decrease in the remaining carbon budget for the 1.5 °C global warming target, in comparison with the reference scenario with a 1% per year pCO₂ increase, for pathways peaking at the doubled pCO₂ level followed by decline to the pre-industrial level.

1. Introduction

The transient climate response to cumulative carbon emissions (TCRE) is defined as the global mean surface temperature change per 1000 GtC (PgC) of anthropogenic CO₂ emission (Allen et al 2009, Matthews et al 2009, Collins et al 2013). A constant TCRE that is independent of time and pathway implies that global mean temperature rise is approximately proportional to cumulative carbon emissions (Cₑ). Such proportionality implies that temperature increases are dependent only on the integrated amount of carbon emissions and not on the emission pathway, thereby facilitating the development of future emission pathways compatible with temperature targets. Future warming can be estimated by multiplying the TCRE by the Cₑ (Allen and Stocker 2014). Likewise, a carbon budget for a given temperature target can be calculated by dividing the target temperature anomaly by the TCRE (e.g. Rogelj et al 2018).

Near-constancy of the TCRE results from certain combinations of two parameters (Allen et al 2009, Matthews et al 2009): the cumulative airborne fraction —ratio of increase of atmospheric carbon (Cₐ) to total anthropogenic emissions (Keeling et al 1995, Forster et al 2007)—and the change in temperature per unit increase in Cₐ, both of which are nonlinear functions of CO₂ emissions. Initially, it was assumed that TCRE
stays constant only while temperature is increasing (Collins et al 2013) and that it is related to transient climate response. Estimates of TCRE include: Estimates of TCRE include 1.4 °C–2.5 °C/1000 PgC (Allen et al 2009; 5–95 percentiles, the same range is quoted hereafter, unless otherwise noted), 1.0 °C–2.1 °C/1000 PgC (Matthews et al 2009), 0.7 °C–2.0 °C/1000 PgC (Gillett et al 2013), 1.1 °C±0.5 °C/1000 PgC (Goodwin et al 2015; ±1 S.D. range), 1.1 °C–1.7 °C/1000 PgC (Tachiiri et al 2015), 1.3 °C–2.2 °C/1000 PgC (Steinacher and Joos 2016; 68% range), 1.1 °C–2.0 °C/1000 PgC (Frölicher and Paynter 2015; ±1 S.D. range for CMIP5-ESMs), among others.

Some studies have argued that TCRE is scenario dependent. Hajima et al (2012), using an Earth system model (ESM), reported that TCRE was slightly larger in scenarios with low radiative forcing, such as in the Representative Concentration Pathway (RCP) 2.6 (Meinshausen et al 2011), although there might have been some effect of model drift (Sueyoshi et al 2013). Krasting et al (2014) found that TCRE has a complex relationship with emission rates and is largest for both low (2 GtC per annum (p.a.)) and high (25 GtC p.a.) emissions. Tachiiri et al (2015) reported that the uncertainty increases when atmospheric CO₂ concentration (pCO₂) stabilizes. Using a model of intermediate complexity, Zickfeld et al (2016) found that TCRE in periods of pCO₂ decrease (temperature decrease for decreasing cumulative carbon emissions) is lower than TCRE in scenarios of pCO₂ increase, and indicated some increase in TCRE in overshoot scenarios (see supplementary information (SI) is available online at stacks.iop.org/ELRL/14/124067/mmedia for additional description). Collins et al (2013) concluded that we have only limited evidence for nearly constant TCRE for CΔ > 2000 PgC. For such large cumulative carbon emissions, recent studies using an ESM of intermediate complexity (Herrington and Zickfeld 2014, Leduc et al 2015, Steinacher and Joos 2016) confirmed some decrease in the TCRE, while research using a full ESM (Tokarska et al 2016) showed no decrease.

In recent years, efforts have been made to formulate theories on TCRE constancy (Goodwin et al 2015, MacDougall and Friedlingstein 2015). MacDougall and Friedlingstein (2015) focused on certain scenario types, while Goodwin et al (2015) treated cases approaching equilibrium. Williams et al (2015) applied the approach of Goodwin et al (2015) to transient scenarios to discuss the relative contributions of different terms to TCRE. MacDougall (2017) concluded that TCRE is nearly constant (i.e. <5% change), only when changes in pCO₂ are between 0.3% and 1.2% p.a. at 400 ppm (or between 0.5% and 2.5% p.a. at >1000 ppm).

In this study, we assessed the scenario-dependence of TCRE using a full ESM (MIROC-ESM, see below) by exploring pathways with stable pCO₂ as well as those in which pCO₂ peaks and declines (i.e. an overshoot scenario). We determined contributions of the atmosphere, ocean (heat and carbon uptake), and land to TCRE change by examining the change in each term through decomposition, focusing on how TCRE can be kept relatively stable even when each term is changing. Finally, we discuss the influence of the scenario-dependence of TCRE on carbon budget estimation, as well as the scenario-dependence of the spatial temperature distribution, when the global mean temperature anomaly is 1.5 °C.

2. Method

2.1. Model

We performed experiments using the MIROC-ESM (Watanabe et al 2011). Compared with other models of the Coupled Model Intercomparison Project Phase 5 (CMIP5) (Taylor et al 2012), the equilibrium climate sensitivity of the MIROC-ESM is high (4.7 °C, Andrews et al 2012). The model also has a relatively high transient climate response (2.2 °C, Flato et al 2013), high TCRE (i.e. as high as 2.2 °C/1000 PgC) (Gillett et al 2013), and low compatible future emissions for RCPs (Jones et al 2013), implying a need for stringent emission cuts to achieve mitigation targets.

The model has nearly average (1.56 PgC ppm⁻¹) carbon-concentration feedback and strong (~100.7 PgC K⁻¹) carbon-climate feedback (these values are for quadruple the pre-industrial CO₂ level, 4× CO₂; Arora et al 2013), resulting in high TCRE. This is because the linearized formulation for TCRE can be written as α/(1 + β + αγ) (Gregory et al 2009), where α(K PgC⁻¹), β(PgC PgC⁻¹ in this case), and γ(PgC K⁻¹) represent linear transient climate sensitivity to CΔ, carbon sensitivity to CΔ, and carbon sensitivity to climate change, respectively (i.e. parameters presented by Friedlingstein et al 2006). At the doubled pre-industrial CO₂ level (2 × CO₂), β and γ of our model are 0.91(PgC PgC⁻¹) and –68(PgC K⁻¹); combined with an α value of 0.003 676 K PgC⁻¹ calculated from a transient climate response of 2.2 °C, we derived a TCRE of 2.2 °C/1000 PgC.

2.2. Experimental design

The scenarios (sections) used in this study are summarized in table 1 (and figure 1(a)). They represent idealized experiments designed to supply fundamental information, in which rates of increase and decrease are constant, and only CO₂ is taken into account. For each pathway, the pre-industrial control run output was taken at 10 year intervals to initialize three ensemble members, and the ensemble means were analyzed. The D1.0%, 4x experiment has been used to investigate the reversibility of the climate system (Boucher et al 2012, Zickfeld et al 2016); it was included in the Carbon Dioxide Removal Model Intercomparison Project (CDR-MIP) as a tier 1 experiment (Keller et al 2018).
The I1.0% experiment, included in the standard CMIP5 experimental protocol, was also explored in this study. It was used as the reference scenario and the pathway to reach $2 \times \text{CO}_2$, which is the initial condition for $S_{2x}$, $D0.5\%$, and $D1.0\%_{2x}$, and $4 \times \text{CO}_2$, which is the initial condition for $S_{4x}$ and $D1.0\%_{4x}$.

Table 1. Description of each scenario used in the experiment. No.: number of ensemble members. In addition, existing data with the I1.0% (1% p.a. increase from $1 \times \text{CO}_2$) scenario are analyzed.

| Scenario | pCO$_2$ change rate | Start CO$_2$ level | End CO$_2$ level | Duration (years) | No. |
|----------|---------------------|---------------------|------------------|-----------------|-----|
| I0.5%    | 0.5% p.a. increase  | $1 \times \text{CO}_2$ ($\text{Initial}$) | $4 \times \text{CO}_2$ | 280 | 3   |
| I2.0%    | 2% p.a. increase    | $1 \times \text{CO}_2$ ($\text{Initial}$) | $4 \times \text{CO}_2$ | 70 | 3   |
| S$_{2x}$ | Fixed concentration | $2 \times \text{CO}_2$ (Branched from I1.0%) | $2 \times \text{CO}_2$ | 70 | 3   |
| S$_{4x}$ | Fixed concentration | $4 \times \text{CO}_2$ (Branched from I1.0%) | $4 \times \text{CO}_2$ | 140 | 3  |
| D0.5%    | 0.5% p.a. decrease  | $2 \times \text{CO}_2$ (Branched from I1.0%) | $1 \times \text{CO}_2$ | 140 | 3   |
| D1.0%$_{2x}$ | 1.0% p.a. decrease | $2 \times \text{CO}_2$ (Branched from I1.0%) | $1 \times \text{CO}_2$ | 140 | 3   |
| D1.0%$_{4x}$ | 1.0% p.a. decrease | $4 \times \text{CO}_2$ (End of I1.0%) | $1 \times \text{CO}_2$ | 140 | 3   |
Compared with other models, the MIROC-ESM has a relatively large warming drift (Sueyoshi et al 2013) of about 0.07 K/century in global mean surface air temperature. To minimize its effect, the linear regression of the control run (for a 300 year period) was removed from all variables prior to analysis (however, we still need to be careful in discussing TCRE when anomalies in temperature and cumulative carbon emission are very small).

From model results, carbon emissions each year were calculated as $\Delta C_A + C_{A-1} + C_{A-2}$, where $\Delta C_A$ is the difference between the airborne carbon of the year concerned and that of the previous year, and $C_{A-1}$ and $C_{A-2}$ are the annual total carbon flux from atmosphere to land and to ocean, respectively.

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3. Results and discussion

3.1. Experimental results

For all experiments listed in table 1, trajectories of changes in global mean surface air temperature ($\Delta T$, figure 1(b)) and $C_t$ (figure 1(c)) follow the shapes of CO$_2$ concentration pathways (figure 1(a)). Consequently, TCREs are similar in magnitude, with values between 2 and 3°C/TtC (or °C/1000 PgC) (figure 1(d); for each member see figure S1).

The 1.0%, 2.0%, 4x and 1.0% 4x experiments are defined for $C_t$ up to 2500 PgC. We could confirm that the variation in TCRE was as large as 30%–40% in comparison with the reference TCRE of the model (2.2 °C/1000 PgC) obtained from the 1.0% scenario (Gillet et al 2013). Although the TCRE variation is somewhat enhanced by a small accumulated carbon emission period, TCRE remains quasi-linear in the temperature-cumulative carbon emission plot (figure S2), consistent with results of Zickfeld et al (2016) using an ESM of intermediate complexity. Relative variation in TCRE is <50% in the scenarios of slow CO$_2$ increase, stable CO$_2$ concentration, and overshoot. Considering the averages of each scenario, we found a 10% increase during CO$_2$ decline in overshoot pathways as well as in the stabilization scenario for 4 × CO$_2$. This change could be significant when calculating a carbon budget using the TCRE. In scenarios of CO$_2$ decrease (D0.5%, D1.0% 4x, and pCO$_2$ decrease in part of D1.0% 4x) and stabilization at a 4 × CO$_2$ level (S 4x), TCREs averaged approximately 2.5 °C/TtC. Meanwhile, in the scenarios of CO$_2$ increase, TCREs averaged approximately 2.2 °C/TtC (results from the first 30 years were removed because of instability). Generally, when pCO$_2$ declined (or stabilized at a high level), TCREs increased slightly. In particular, TCREs were larger in the D0.5%, D1.0% 2x, and D1.0% 4x scenarios, when pCO$_2$ returned to approximately pre-industrial values (285 ppm). Large TCRE variations in D1.0% 2x (figure 1(d)) were caused by $C_t$ approaching zero.

3.2. Contributions of the atmosphere, ocean, and land to changes in transient climate response TCREs

3.2.1. Overview

TCRE is a result of the contributions of atmosphere, ocean, and land. To discuss their relative contributions to TCRE change, we can rewrite TCRE as equation (1):

$$\text{TCRE} = \frac{\Delta T}{\text{CE}} = \frac{\text{RF} - \text{OHU}}{\lambda \times \left( C_A + C_O + C_L \right)}, \tag{1}$$

where RF(W m$^{-2}$) is CO$_2$-induced radiative forcing; $C_A$(PgC), $C_O$(PgC), and $C_L$(PgC) are carbon storage in the atmosphere, ocean, and land, respectively; $\lambda$(W m$^{-2}$ K$^{-1}$) is the climate feedback parameter; and OHU(W m$^{-2}$) is OHU (which is equal to the top of atmosphere imbalance of the ESM). Herein, we consider the contributions of RF, OHU, $C_A$, $C_O$, $C_L$, and $\lambda$ to TCRE. Changes in OHU, $C_A$, and $C_L$ or ocean carbon uptake (OCU) are presented in figure 2 and discussed below.

If we did not have land carbon uptake or ocean heat and carbon uptake in the Earth system, and if $\lambda = 1$, then equation (1) could be reduced to RF/$C_A$. Here, $C_A$ is prescribed, and RF is assumed to be a log function of $C_A$. When we consider peak and decline scenarios with 1% p.a. of pCO$_2$ increase and decrease, in which both RF and $C_A$ are symmetrical for the periods of pCO$_2$ increase and decrease, then through a change in $\lambda$ that is calculated as $\Delta T/(\text{RF} - \text{OHU})$ for the year concerned (black line in figure 3(a) for the result of II.0% and D1.0% 4x scenarios), we see that RF/$C_A$, $\lambda$, i.e. the red line in figure 3(a), stabilizes around peak pCO$_2$, but remains nearly symmetrical for the periods of pCO$_2$ increase and decrease. When OHU and $C_O$, and then $C_L$ are added to the numerator and denominator in equation (1), the stability of TCRE increases, and an increase within the pCO$_2$-decline period becomes evident (figure 3(a)). In figure 3(b), the contribution of $C_L$ to $C_{E}$ is small and stable, indicating that $C_L$ does not play an important role in TCRE change. Thus, we focused on the contributions of OHU and $C_O$ in pCO$_2$ increase, decrease, and stabilization periods. We did not focus on $C_A$ because it is given (as pCO$_2$).

3.2.2. pCO$_2$ increase

In scenarios with pCO$_2$ increase, TCRE stabilizes at approximately the 30th year (blue, green and dotted black lines in figure 1(d)), although the spread among each ensemble member becomes larger when the rate of increase of pCO$_2$ is small (figure S1). In the numerator of equation (1), OHU varies almost linearly with time (figure 2(a)). Approximated by a log function of pCO$_2$ that increases exponentially in these scenarios, RF also varies linearly with time (not shown). This causes RF − OHU, which is the energy input that causes global atmospheric temperature to increase, also to vary linearly with time (figure S3). Owing to the quasi-linear decrease during the pCO$_2$-increase period in $\lambda$ (figure 3(a); a decrease in $\lambda$...
after around the 40th year is common for 10.5%, 12.0% (not shown), the relationship between $\Delta T$ and time has weak nonlinearity (figure 1(b)). In the denominator of equation (1), the air-to-sea carbon flux increases during the first several decades and then becomes nearly constant (figure 2(b)). Ocean carbon storage ($C_O$ in equation (1)), which is more directly associated with TCRE, varies almost linearly with time because of the stabilization of the air-to-sea carbon flux (figure 2(c)). The exponential increase of $C_A$ is moderated when $C_O$, which varies quasi-linearly with time (figure 2(c)), is added. The nonlinearity is further reduced by slightly decreasing the relative contribution of $C_A$ over time (figure 2(d)). Consequently, the trajectory of $C_E$ is similar to that of $\Delta T$ (figure 1(c)), resulting in a stable TCRE in scenarios with $pCO_2$ increase.

### 3.2.3. $pCO_2$ decrease

There is a clear increase in TCRE at the point when $pCO_2$ switches from increase to decrease (red, black, and cyan curves in figure 1(d)). Using D1.0%−4x as an example, we see that $C_E$ starts to decrease when $pCO_2$ starts to decrease. Even though OCU continues initially (figure 2(b)), it is unable to counteract the $C_A$ reduction at a rate of 1% p.a. After $pCO_2$ reaches its peak, temperature continues to increase for approximately 10 years because RF − OHU continues to increase. Although RF peaks simultaneously with $pCO_2$, the rapid decrease in OHU (figure 2(a)) surpasses the decrease of RF. Therefore, temperature continues to rise but $C_E$ becomes smaller, resulting in a TCRE increase.

The important role of OHU is confirmed in figure 3(c) by changes in $(RF-[OHU])/(C_E/\lambda)$, which
Figure 3. Decomposition of the transient climate response to cumulative carbon emissions (TCRE) in scenarios II.0% and D1.0%_4x. Time series of (a) RF/(C_E − λ), ΔT/(C_A + C_L + [C_O]), and ΔT/(C_A + C_O) = (RF − [OHU])/[C_E − λ]; (b) fractions of carbon storage in the atmosphere (C_A), ocean (C_O), and land (C_L) in cumulative carbon emission (C_E); (c) RF/(−[OHU])/[C_E − λ]; ΔT/(C_A + C_L + [C_O]), where [X] means average of variable X over 280 years, and TCRE; and (d) RF/(RF − OHU)/[C_E − λ] and −[OHU]/[RF − OHU] (red line). The same figure for other pCO_2 decreasing scenarios and the λ, RF/(CA + λ), (RF − [OHU])/(CE − λ), ΔT/(CA + CL + CO) of each ensemble members for all pCO_2 decreasing scenarios are presented in figures S3 and S6. Here, RF (W m^{-2}) is CO_2-induced radiative forcing; C_A (PgC), C_O (PgC), and C_L (PgC) are carbon storage in the atmosphere, ocean, and land, respectively; λ (W m^{-2} K^{-1}) is the climate feedback parameter; and OHU (W m^{-2}) is ocean heat uptake.

The TCRE is the transient climate response to cumulative carbon emissions. It is defined as the increase in climate warming that occurs due to a change in CO_2 emissions. The TCRE is calculated as the difference between the climate response in a scenario with increased CO_2 emissions and the climate response in a scenario with decreased CO_2 emissions, normalized by the change in CO_2 emissions.

3.2.4. pCO_2 stabilization

The trajectory of the TCRE after pCO_2 stabilization depends on the pCO_2 pathway prior to stabilization. Both OHU and OCU start to approach zero at pCO_2 stabilization.

If stabilization is reached after a period of pCO_2 increase (like in S_2x, S_4x), then OHU and OCU approach zero from positive values (i.e. by decreasing). OHU decrease leads to increases in RF − OHU and temperature. Positive OCU results in increasing C_O (and then C_E), and both the numerator and the denominator of the TCRE increase.

If stabilization is reached after a period of pCO_2 decrease (like in D1.0%_2x), then both OHU and OCU approach zero from negative values (i.e. by increasing). OHU increase leads to decreases in RF − OHU and temperature. Negative OCU results in decreasing C_O and then C_E, and both the numerator and the denominator of the TCRE decrease. The TCRE becomes relatively stable in both cases but
Table 2. Cumulative carbon emissions, transient climate response to cumulative carbon emissions (TCRE), and remaining carbon budgets ($B_{lim}$) calculated for different pathways for 1.5 °C warming.

| Pathways   | Cumulative emissions (PgC) | TCRE (°C/1000 PgC) | Remaining carbon budget ($B_{lim}$) using TCRE (PgC) | Ratio of $B_{lim}$ to that of 1.0% |
|------------|---------------------------|-------------------|------------------------------------------------------|---------------------------------|
| I0.5%      | 633                       | 2.37              | 77                                                   | 0.81                            |
| I1.0%      | 705                       | 2.13              | 95                                                   | 1.00                            |
| I2.0%      | 722                       | 2.08              | 99                                                   | 1.04                            |
| D0.5%      | 642                       | 2.34              | 79                                                   | 0.84                            |
| D1.0% - 2x | 570                       | 2.63              | 62                                                   | 0.65                            |
| D1.0% - 4x | 546                       | 2.75              | 56                                                   | 0.59                            |

through opposing processes for scenarios starting with 2 × CO₂ (i.e. S_2x and D1.0% - 2x).

For S_4x starting with 4 × CO₂, a TCRE increase, comparable to the pCO₂ decline, is observed. Figure 2(a) indicates some contribution of OBU change particularly in the first decades (unlike pCO₂ decline scenarios, the rapid TCRE over the first 10 years was not observed). We surmise that the difference from S_2x is mainly related to the repressed increase in C_b (figure 1(c)), reflecting cancellation of the increase in C_O by the decrease in C_b (figure 2(d)). Unlike D1.0% - 4x, in which TCRE increases because of pCO₂ decline (decrease in OHU keeps contributing throughout the section (figure 2(b))), in S_4x, high temperature reduces C_b. This reduction can play an important role in the TCRE increase, when changes in C_A and C_O are small.

3.3. Scenario-dependence of the carbon budget and spatial distribution of the atmospheric temperature anomaly in meeting the 1.5 °C global warming target

Finally, we compare the cumulative carbon emissions and temperature distributions of different pathways. The cumulative carbon emissions until each of the pathways intersects the 1.5 °C global warming target in scenarios I0.5%, I1.0%, I2.0%, D0.5%, D1.0% - 2x, and D1.0% - 4x vary by 546–722 PgC, showing a tendency for large emissions with quickly increasing pathways and small emissions with overshoot pathways (table 2; corresponding TCREs are also presented). It is not easy to directly compare these values with the total carbon budget of the actual Earth system because the pathways in the current study do not include non-CO₂ greenhouse gas (GHG) and aerosol emission scenarios. Instead, like equation (1) of Rogelj et al (2019), with an assumption that zero emission commitment and unpresented Earth system feedback are zero, we calculated the remaining carbon budget as $(T_{lim} - T_{hist} - T_{nonCO2})/TCRE$, where $T_{lim}$, $T_{hist}$, and $T_{nonCO2}$ are the temperature anomaly target, historical warming to date, and non-CO₂-induced future warming, respectively. Following Rogelj et al (2018), a $T_{hist}$ value of 0.97 °C results in a value of 0.53 °C for $T_{lim} - T_{hist}$. To estimate $T_{nonCO2}$, Rogelj et al (2019) suggested using internally consistent evolution. However, as we have no corresponding non-CO₂ GHG emission scenario, we adopted a simpler method that assumes (by considering non-CO₂-GHG-induced warming) the remaining carbon budget is reduced by 30% (estimated from Forster et al (2018)) and that 79.4 PgC (291 GtCO₂) was emitted during 2011–2017 (Rogelj et al 2018). Thus, the carbon budget after 2018, using these TCRE estimates, will be 56–99 PgC (table 2). These values can be compared with the estimates of Rogelj et al (2018) of 229, 158, and 115 PgC (840, 580, and 420 GtCO₂) for the 33rd, 50th, and 67th percentiles, respectively. As the TCRE from the MIROC-ESM (2.2 °C/1000 PgC) is larger than the value used to calculate the 67th percentile (0.55 °C/1000 GtCO₂ = 2.0 °C/1000 PgC), it is not surprising that the estimated remaining carbon budget is smaller than that of the 67th percentile value of Rogelj et al (2018). However, more importantly, TCRE shows some pathway dependence, and if we apply TCREs for overshoot and slowly increasing pathways, the remaining carbon budget is reduced by 16%–41% for pathways with slower pCO₂ increase than I1.0% (i.e. I0.5%) or overshoot patterns. Conversely, for I2.0%, we have a 4% increase in the remaining carbon budget (table 2). As such, scenario dependency of the TCRE, particularly for overshoot pathways, results in different carbon budgets for different pathways.

Spatial distribution of the atmospheric temperature anomaly ($\Delta T$), when the global mean anomaly is 1.5 °C, also shows some scenario-dependence (figure 4). Figure 4(a) shows clear polar intensification in the northern high latitudes in the reference scenario of I1.0%. This basic feature is repeated in other scenarios (figures 4(b)–(e)), despite differences in the value between the reference scenario and all other scenarios. The spatial distribution of temperature is maintained, when pCO₂ (and hence, temperature) increases monotonically (figures 4(b), (c)). In overshoot scenarios (figures 4(d)–(f)), temperature anomalies over land are generally smaller than those in the reference scenario, except in northeastern Siberia, where temperature anomalies are large. These features are also visible in scenario I0.5%. Temperature anomalies over the ocean are typically larger than in the reference scenario, which is considered to reflect larger cumulative OHU (i.e. ocean heat content). The strong positive
anomaly observed around 0° longitude in the high latitudes of both hemispheres likely reflects sea ice melting (see figure S7).

The difference from the reference scenario is significant in overshoot pathways. For instance, in D1.0%_2x, northeast China has less warming (<−40%) in comparison with I1.0%, while northeastern Siberia is warmer (>+40%) than when both scenarios have a 1.5 °C anomaly. Although D1.0%_4x is a scenario with an unrealistically large overshoot, results from other overshoot scenarios are more realistic and clearly show that simple pattern scaling is insufficient to describe their dynamics.

4. Conclusions

We examined the robustness of TCRE using the MIROC-ESM and various CO2 concentration pathways to confirm the quasi-constancy of TCRE during periods of pCO2 increase and stable pCO2 at a 2 × CO2 level. During pCO2 decrease and stable pCO2 at a 4 × CO2 level, we found an average TCRE increase of 10%–20% (from 2.2 °C to 2.5 °C/1000 PgC) and a maximum of 50% (20 year average). Our ESM results for overshoot scenarios are qualitatively consistent with those of Zickfeld et al. (2016). In addition, we analyzed the contributions of the atmosphere, ocean, and land to TCRE. In the case of the ocean, the contributions of OHU and OCU were distinct.

In the scenario with increasing pCO2, a stabilized OCU is considered critical for stable TCRE because it results in ocean carbon storage that increases linearly, as does OHU, with pCO2. When pCO2 starts to decrease, TCRE increases within a decade, and it then stabilizes because of a balance between OHU decrease and a slight increase in C10 (relative to the period of pCO2 increase). Typically, OCU decreases more rapidly than OHU, but cumulative OCU continues to increase for a certain period. For a stabilization at a 4 × CO2 level, negative CL and a slower increase in C10 (related to high temperature) result in a negligible increase in C10, which causes TCRE to increase, comparable to when pCO2 declines.

The increase of the TCRE during pCO2 decline will affect the carbon budget, when an overshoot pathway is planned. Indeed, analysis of different pathways to the 1.5 °C global warming target shows decreases in the remaining carbon budget in overshoot pathways. Our simulation also shows that, in overshoot pathways, temperature continues to increase for a certain period, despite strong decreases in atmospheric CO2 concentration. The results suggest that to compensate for this process, we might need extra measures to moderate climate change (e.g. geoengineering).

Changes in the spatial distributions of temperature and other variables are also of interest to climate researchers. Warming over land is smaller and warming over the ocean is larger than those in the reference (1% per annum pCO2 increase) scenario, and the warming over some parts of the ocean in high latitudes of both hemispheres is significant. These features appear even more strongly in scenarios with a rapid rate of pCO2 decrease and a high peak pCO2. The mechanism of TCRE change during pCO2 decline should be investigated to better understand the processes associated with achieving ambitious global warming targets.

We note that our study has been conducted using a single ESM, which is known to have relatively high climate sensitivity among CMIP5 models. In such models, overshooting of global temperature and the reduction of the estimated carbon budget in modeled scenarios might be overestimated; thus, TCRE...
increase might be exaggerated. To evaluate the behavior of TCRE during the decades after peak pCO₂, intercomparison studies using multiple ESMs and alternative overshoot scenarios are needed. We expect that CDR-MIP will present meaningful results for such comparisons.

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Data availability statement

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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