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A framework to understand the transient climate response to emissions

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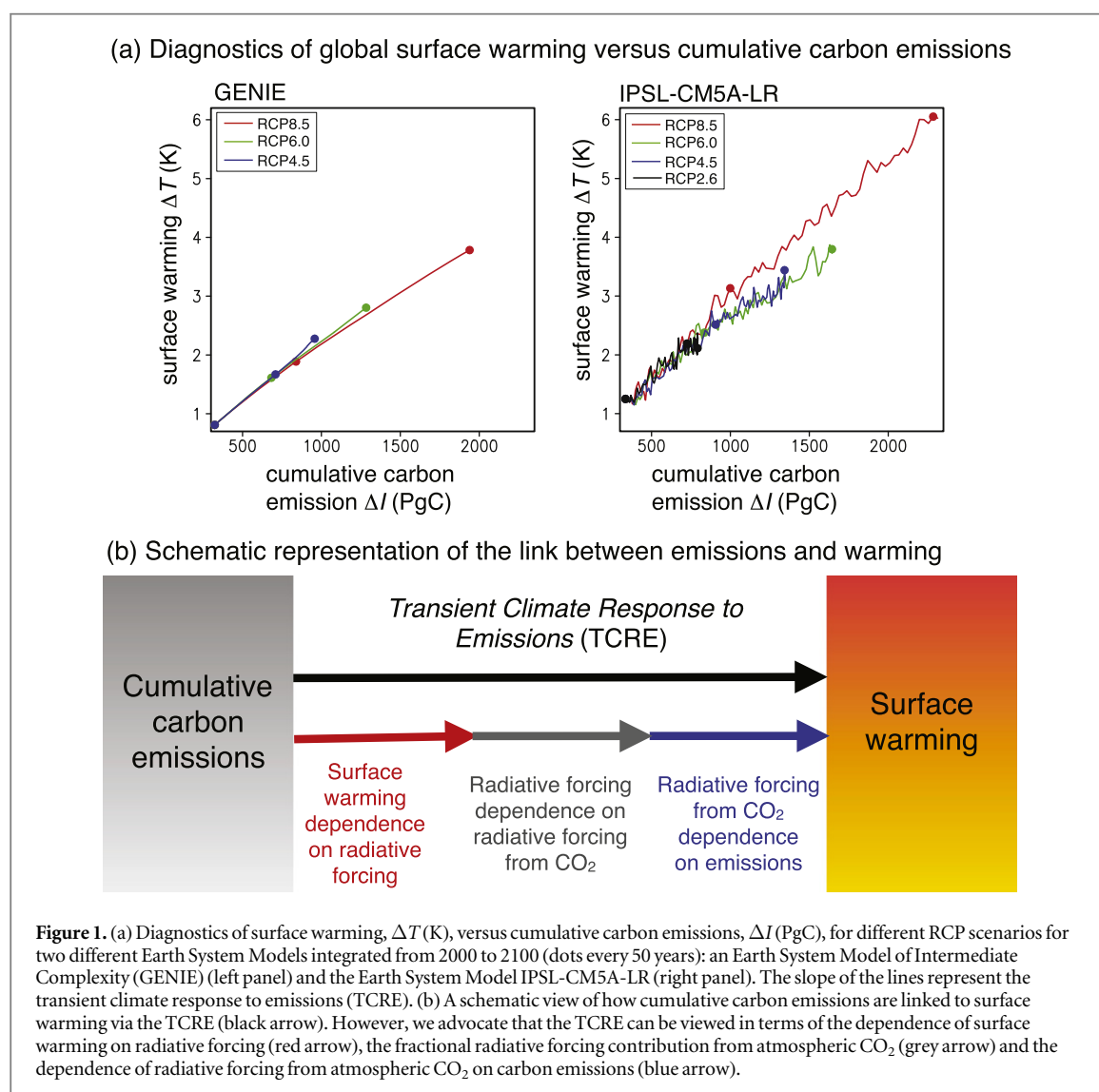# A framework to understand the transient climate response to emissions

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Abstract

Global surface warming projections have been empirically connected to carbon emissions via a climate index defined as the transient climate response to emissions (TCRE), revealing that surface warming is nearly proportional to carbon emissions. Here, we provide a theoretical framework to understand the TCRE including the effects of all radiative forcing in terms of the product of three terms: the dependence of surface warming on radiative forcing, the fractional radiative forcing contribution from atmospheric CO_2 and the dependence of radiative forcing from atmospheric CO_2 on cumulative carbon emissions. This framework is used to interpret the climate response over the next century for two Earth System Models of differing complexity, both containing a representation of the carbon cycle: an Earth System Model of Intermediate Complexity, configured as an idealised coupled atmosphere and ocean, and an Earth System Model, based on an atmosphere–ocean general circulation model and including non-CO_2 radiative forcing and a land carbon cycle. Both Earth System Models simulate only a slight decrease in the TCRE over 2005–2100. This limited change in the TCRE is due to the ocean and terrestrial system acting to sequester both heat and carbon: carbon uptake acts to decrease the dependence of radiative forcing from CO_2 on carbon emissions, which is partly compensated by changes in ocean heat uptake acting to increase the dependence of surface warming on radiative forcing. On decadal timescales, there are larger changes in the TCRE due to changes in ocean heat uptake and changes in non-CO_2 radiative forcing, as represented by decadal changes in the dependences of surface warming on radiative forcing and the fractional radiative forcing contribution from atmospheric CO_2. Our framework may be used to interpret the response of different climate models and used to provide traceability between climate models of differing complexity.

1. Introduction

Surface global warming is empirically found in climate models to be nearly linearly dependent on the cumulative amount of carbon emitted to the climate system. This climate relationship was first set out in terms of how the peak warming of a range of climate models depends on the cumulative amount of carbon emitted (Allen *et al* 2009) and how the global warming in climate model projections is proportional to cumulative carbon emissions (Matthews *et al* 2009,

Zickfeld *et al* 2009). Both these responses are largely independent of emission pathway, but the amplitude of the climate response varies with individual climate models. The proportionality of global surface warming to cumulative carbon emissions is used to define the transient climate response to emissions (TCRE, in K per 1000 PgC) (Collins *et al* 2013, Gillet *et al* 2013). The TCRE is found to be approximately independent of time and emission scenario; for example, the proportionality of the surface warming to cumulative carbon emissions is depicted here in figure 1(a) by the

Figure 1. (a) Diagnostics of surface warming, ΔT (K), versus cumulative carbon emissions, ΔI (PgC), for different RCP scenarios for two different Earth System Models integrated from 2000 to 2100 (dots every 50 years): an Earth System Model of Intermediate Complexity (GENIE) (left panel) and the Earth System Model IPSL-CM5A-LR (right panel). The slope of the lines represent the transient climate response to emissions (TCRE). (b) A schematic view of how cumulative carbon emissions are linked to surface warming via the TCRE (black arrow). However, we advocate that the TCRE can be viewed in terms of the dependence of surface warming on radiative forcing (red arrow), the fractional radiative forcing contribution from atmospheric CO_2 (grey arrow) and the dependence of radiative forcing from atmospheric CO_2 on carbon emissions (blue arrow).

slope of the warming response versus cumulative emissions for two Earth System Models of differing complexity for a range of emission scenarios.

Recently Goodwin *et al* (2015) provided a single equation to connect surface warming to carbon emissions, drawing upon theoretical relations for the global heat balance and carbon inventories for the climate system. Based on this equation, the near constancy in the TCRE is explained in terms of partly compensating effects of the ocean heat and carbon uptake, extending heuristic arguments of Solomon *et al* (2009), where the cooling effect of ocean uptake of atmospheric CO_2 is accompanied by a partly-compensating surface warming effect by declining ocean heat uptake. This time-dependent relationship for the TCRE asymptotes to a long-term equilibrium response, defined by the climate and carbon parameters for the climate system (Williams *et al* 2012).

In this study, we develop a framework to interpret how the TCRE is controlled, the dependence of surface warming on cumulative carbon emissions, which is separated into the product of the dependence of

surface warming on the radiative forcing from atmospheric CO_2 and the dependence of the radiative forcing from atmospheric CO_2 on carbon emissions. This TCRE definition is also extended to include the effect of other non-CO_2 radiative forcing via a non-dimensional term, the fractional radiative forcing contribution from CO_2. The TCRE framework is then combined with our theoretical relations for the global heat balance and buffered carbon inventories (Goodwin *et al* 2015). Our TCRE framework is illustrated using diagnostics of an Earth System Model of Intermediate Complexity (GENIE) and an Earth System Model (IPSL-CM5A-LR). GENIE is configured as an idealised atmosphere–ocean model and only includes radiative forcing from atmospheric CO_2, while IPSL-CM5A-LR additionally includes non-CO_2 radiative forcing and a representation of the land carbon cycle. Both models are integrated from a pre-industrial state and their climate response diagnosed in terms of our framework for their projections from 2005 to 2100.

The paper is set out as follows: the theoretical framework is first presented to understand the TCRE and

our extension to include the non-CO_2 radiative forcing contribution (section 2); the Earth System Model of Intermediate Complexity and Earth System Model are diagnosed in terms of changes in the TCRE and its dependences (section 3); the mechanistic implications from our framework are then discussed (section 4) and finally conclusions are provided (section 5).

2. Theory

A new framework is set out to understand the dependence of the TCRE. This framework draws upon our prior studies for the time-dependent climate response (Goodwin *et al* 2015) and the long-term equilibrium climate response (Williams *et al* 2012) to cumulative carbon emissions.

2.1. Theoretical framework
In order to understand our framework, assume that the global-mean surface air temperature (K) at time t, $T(t)$, is defined by a time-dependent anomaly, $\Delta T(t)$, relative to the temperature at the pre-industrial at time t_o, $T(t) = T(t_o) + \Delta T(t)$. The cumulative amount of carbon (PgC) emitted to the atmosphere from all anthropogenic sources at time t is similarly defined, $I(t) = I(t_o) + \Delta I(t)$, with $I(t_o)$ taken to be zero. The surface temperature anomaly is then related to the cumulative amount of carbon emitted since the pre-industrial via the transient climate response to cumulative carbon emissions, the TCRE, represented mathematically by $\Delta T / \Delta I$,

$$\Delta T(t) = \left(\frac{\Delta T}{\Delta I} \right) \Delta I(t), \qquad (1)$$

where the Δ notation denotes differences relative to the pre-industrial (rather than a small incremental interval). Our aim is now to connect this surface warming definition (1) to the controlling factors for the TCRE, which are assumed to be the dependence of surface warming on radiative forcing and the dependence of radiative forcing on cumulative carbon emissions.

Start by assuming that the radiative forcing (W m^{-2}) driving climate change is only from atmospheric CO_2, which is defined by $R_{CO_2}(t) = R_{CO_2}(t_o) + \Delta R_{CO_2}(t)$. Here we choose to express the TCRE in (1) in terms of the product of the dependence of surface warming on radiative forcing from atmospheric CO_2, $\Delta T / \Delta R_{CO_2}$, and the dependence of radiative forcing on carbon emissions, $\Delta R_{CO_2} / \Delta I$:

$$\text{TCRE} = \frac{\Delta T}{\Delta I} = \left(\frac{\Delta T}{\Delta R_{CO_2}} \right) \left(\frac{\Delta R_{CO_2}}{\Delta I} \right). \qquad (2)$$

More generally, the radiative forcing driving climate change includes contributions from atmospheric CO_2 and other non-CO_2 contributions, so that the change in radiative forcing since the pre-industrial

era is given by $\Delta R(t) = \Delta R_{CO_2}(t) + \Delta R_{nonCO_2}(t)$. The change in non-$CO_2$ radiative forcing, $\Delta R_{nonCO_2}(t)$, includes contributions from aerosols emitted via anthropogenic and volcanic activity (Ottera *et al* 2010, Booth *et al* 2012) and other non-CO_2 greenhouse gases, such as methane with a shorter lifetime than CO_2 (Pierrehumbert 2014).

The definition of the TCRE may be extended to include the effect of all contributions to the radiative forcing, such that

$$\text{TCRE}_{allR} = \frac{\Delta T}{\Delta I} = \left(\frac{\Delta T}{\Delta R} \right) \left(\frac{\Delta R}{\Delta I} \right). \qquad (3)$$

While this definition looses the clarity of the original TCRE definition in (2) in terms of isolating the climate response to cumulative carbon emissions, this generalisation for $\text{TCRE}_{all\,R}$ in (3) is needed if the climate response is to be diagnosed from a climate model including non-CO_2 radiative forcing. To extend the TCRE definition in (3), the surface temperature change in (1) is written as

$$\Delta T(t) = \frac{\Delta T}{\Delta I} \Delta I(t) = \left(\frac{\Delta T}{\Delta R} \right) \left(\frac{\Delta R_{CO_2}}{\Delta I} \right) \Delta I(t)$$
$$+ \left(\frac{\Delta T}{\Delta R} \right) \Delta R_{nonCO_2}(t), \qquad (4a)$$

which makes explicit the surface warming from the separate effects of the radiative forcing from CO_2 and non-CO_2 contributions. Exploiting the definition of $\Delta R(t) = \Delta R_{CO_2}(t) + \Delta R_{nonCO_2}(t)$, the surface warming is then equivalently written as

$$\Delta T(t) = \frac{\Delta T}{\Delta I} \Delta I(t) = \left(\frac{\Delta T}{\Delta R} \right) \left(\frac{\Delta R}{\Delta R_{CO_2}} \right)$$
$$\times \left(\frac{\Delta R_{CO_2}}{\Delta I} \right) \Delta I(t). \qquad (4b)$$

Accordingly, the definition for the TCRE for all radiative forcing contributions in (3) is then expressed as

$$\text{TCRE}_{allR} = \frac{\Delta T}{\Delta I} = \left(\frac{\Delta T}{\Delta R} \right) \left(\frac{\Delta R}{\Delta R_{CO_2}} \right) \left(\frac{\Delta R_{CO_2}}{\Delta I} \right). \qquad (5)$$

This definition of the TCRE for all radiative forcing contributions in (5) is made up of the product of three differential terms: the dependence of surface warming on radiative forcing, $\Delta T / \Delta R$; the fractional radiative contribution from atmospheric CO_2, $\Delta R / \Delta R_{CO_2}$; and the dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2} / \Delta I$ (depicted in figure 1(b) by red, grey and blue arrows respectively). This definition of the TCRE for all radiative forcings shares the same dependence of the radiative forcing from CO_2 on cumulative carbon emissions, $\Delta R_{CO_2} / \Delta I$, as appearing in the original TCRE definition in (2), but is now

generalised to include the non-CO_2 radiative forcing contribution.

Our aim is now to assess the behaviour of the TCRE on centennial timescales using a combination of ocean theory and diagnostics of two Earth System Models, GENIE only including radiative forcing from atmospheric CO_2 and IPSL-CM5A-LR including CO_2 and non-CO_2 radiative forcing.

Our ocean theory is only appropriate for the dependence of surface warming on radiative forcing, $\Delta T / \Delta R$ and the dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2} / \Delta I$. The diagnostics of the TCRE and TCRE$_{all\ R}$ using (2) and (5) respectively are automatically the same for GENIE, but differ for IPSL-CM5A-LR. On centennial and millennial timescales, we expect that the climate response is dominated by the effect of carbon emissions, so that the TCRE and TCRE$_{all\ R}$ become similar to each other. However, on shorter timescales, the TCRE and TCRE$_{all\ R}$ differ from each other due to the effects of other forcing agents, such as aerosols and non-CO_2 greenhouse gases.

2.2. Applying global heat and buffered carbon integral balances

Our framework defining the TCRE and TCRE$_{all\ R}$ in (2) and (5) respectively is now combined with theoretical balances for a global heat balance and buffered carbon inventory following Goodwin *et al* (2015).

The global heat balance for climate change involves an increase in radiative forcing since the pre-industrial, $\Delta R(t)$ (W m^{-2}, positive downward) driving a climate response, involving additional outgoing longwave radiation, $\lambda \Delta T(t)$ (W m^{-2}), from the increase in global-mean surface air temperature plus the net downward heat flux into the climate system, $N(t)$ (W m^{-2}) (Gregory *et al* 2004, Gregory and Forster 2008)

$$\Delta R(t) = \lambda \Delta T(t) + \varepsilon N(t), \qquad (6)$$

where λ is the equilibrium climate parameter (W m^{-2} K^{-1}) and $\varepsilon N(t)$ is the scaled net heat flux into the climate system. The net heat flux $N(t)$ is dominated by the ocean heat uptake, accounting for over 90% of the anthropogenic warming of the climate system (Church *et al* 2011). The net heat flux $N(t)$ is scaled in (6) by a non-dimensional parameter, ε, to take into account the enhanced effect of ocean heat flux in increasing global-mean surface temperature relative to global-mean radiative forcing (Hansen *et al* 1984, Winton *et al* 2010). This heat balance is rearranged to provide an expression for the dependence of surface warming on radiative forcing

$$\frac{\partial T}{\partial R} = \frac{1}{\lambda}\left(1 - \frac{\varepsilon N(t)}{\Delta R(t)}\right), \qquad (7)$$

where the fractional form, $\Delta T / \Delta R$, is replaced by a differential form as there is a functional relationship and the effective ocean heat flux is normalised by the radiative forcing, $\varepsilon N(t)/\Delta R(t)$.

The buffered carbon inventory connects the logarithm of atmospheric CO_2 to the sum of the cumulative carbon emission, $\Delta I(t)$, plus the carbon undersaturation of the ocean, $I_{Usat}(t)$ (PgC), minus the increase in the terrestrial carbon uptake, $\Delta I_{ter}(t)$ (PgC), all divided by the buffered carbon inventory, I_B (PgC) (Goodwin *et al* 2015):

$$\Delta \ln CO_2(t) = \frac{1}{I_B}(\Delta I(t) + I_{Usat}(t) - \Delta I_{ter}(t)), \qquad (8)$$

where CO_2 is the mixing ratio of atmospheric carbon dioxide (ppm), $I_{Usat}(t)$ is how much carbon the ocean needs to take up to reach an equilibrium with the atmosphere and $\Delta I_{ter}(t) = I_{ter}(t) - I_{ter}(t_o)$ is the change in the residual terrestrial carbon sink since the pre-industrial era and I_B is the buffered carbon inventory for the atmosphere and ocean (Goodwin *et al* 2007). This buffered carbon inventory takes into account a positive feedback; increasing atmospheric CO_2 enhances ocean acidity and inhibits the ability of the ocean to take up more atmospheric CO_2 leading to an increasing fraction of the emitted carbon remaining in the atmosphere; thus, (8) expresses how atmospheric CO_2 increases exponentially with the sum of the cumulative carbon emission plus the ocean undersaturation and minus the terrestrial uptake.

The global heat balance (6) and buffered carbon inventory (8) are connected together via the radiative forcing from atmospheric CO_2, $\Delta R_{CO_2}(t)$, defined by the logarithm of atmospheric CO_2 (Myhre *et al* 1998)

$$\Delta R_{CO_2}(t) = a\Delta \ln CO_2(t), \qquad (9)$$

where the logarithm conveys how there is a saturating effect of increasing atmospheric CO_2 on radiative forcing.

Combing the buffered carbon balance (8) with the definition for radiative forcing from atmospheric CO_2 (9) then provides an expression for the dependence of radiative forcing from atmospheric CO_2 on carbon emissions (Goodwin *et al* 2015)

$$\frac{\partial R_{CO_2}}{\partial I} = \frac{a}{I_B}\left(1 + \frac{I_{Usat}(t)}{\Delta I(t)} - \frac{\Delta I_{ter}(t)}{\Delta I(t)}\right), \qquad (10)$$

where the effects of the ocean carbon undersaturation and increase in terrestrial carbon store are normalised by the cumulative carbon emission, $(I_{Usat}(t) - \Delta I_{ter}(t))/\Delta I(t)$. By combining the differential relations (7) and (10), the TCRE definition for all radiative forcing contributions in (5) becomes

$$TCRE_{allR} = \frac{a}{\lambda I_B}\left(1 - \frac{\varepsilon N(t)}{\Delta R(t)}\right)$$
$$\times \left(1 + \frac{I_{Usat}(t)}{\Delta I(t)} - \frac{\Delta I_{ter}(t)}{\Delta I(t)}\right)\left(\frac{\Delta R(t)}{\Delta R_{CO_2}(t)}\right). \qquad (11)$$

The TCRE$_{all\ R}$ depends on the time-independent factors, $a/(\lambda I_B)$, multiplied by the non-dimensional time-dependent terms contained within the three pairs of parentheses: the dependence of surface

Table 1. Climate variables in the two Earth System Models of different complexity.

		Earth System Model of Intermediate Complexity, GENIE	Earth System Model IPSL-CM5A-LR
Climate equilibrium parameter	λ	1.28 W m^{-2} K^{-1} (Goodwin *et al* 2015)	0.77 W m^{-2} K^{-1} based on a least-square regression (figure 3(a)) (assuming $\varepsilon = 1.14$)
Radiative forcing coefficient from CO_2	a	5.77 W m^{-2}	4.47 W m^{-2} (Andrews *et al* 2012)
Efficacy	ε	1.07 (Goodwin *et al* 2015)	1.14 (Geoffroy *et al* 2013)
Atmospheric carbon inventory at the preindustrial for 1850 at 280 ppmv (using a molar mass 1.77×10^{20} moles)	I_a	594 PgC	594 PgC
Temperature, salinity, alkalinity and Phosphate of global ocean at the pre-industrial	T, S, Alk, PO_4	3.61 °C, 34.90, 2.43 mol m^{-3}, 2.21×10^{-3} mol m^{-3}	3.14 °C, 34.73, 2.43 mol m^{-3}, 2.16×10^{-3} mol m^{-3}
DIC of global ocean at the pre-industrial	DIC	(12 g mol^{-1}) $\times$ 2.29 mol m^{-3}	(12 g mol^{-1}) $\times$ 2.347 mol m^{-3}
Saturated DIC at the pre-industrial	DIC$_{sat}$	(12 g mol^{-1}) $\times$ 2.218 mol m^{-3}	(12 g mol^{-1}) $\times$ 2.221 mol m^{-3}
Ocean saturated carbon inventory at the preindustrial (assuming ocean volume V of 1.341×10^{18} m^3)	V DIC$_{sat}$	34 707 PgC	35 730 PgC
Effective global ocean buffer factor at the preindustrial	B	12.66	12.55
Buffered carbon inventory of the atmosphere and ocean	I_B	3340 PgC	3440 PgC
Implied TCRE, $\Delta T / \Delta I$, at a long-term equilibrium	$a/(\lambda\, I_B)$	1.35 K(1000 PgC)$^{-1}$	1.69 K(1000 PgC)$^{-1}$

warming varying with ocean heat uptake; the dependence of radiative forcing from atmospheric CO_2 varying with ocean carbon undersaturation; and the fractional dependence of radiative forcing on atmospheric CO_2. As derived by Goodwin *et al* (2015), if there is only radiative forcing from CO_2, the TCRE simplifies to

$$\text{TCRE} = \frac{\Delta T}{\Delta I} = \frac{a}{\lambda I_B}\left(1 - \frac{\varepsilon N(t)}{\Delta R(t)}\right)$$
$$\times \left(1 + \frac{I_{Usat}(t)}{\Delta I(t)} - \frac{\Delta I_{ter}(t)}{\Delta I(t)}\right). \quad (12)$$

2.3. Equilibrium response for the TCRE

For a long-term equilibrium when the radiative forcing is only from atmospheric CO_2, the TCRE asymptotes to a response given by the time-independent factors $a/(\lambda I_B)$ (Williams *et al* 2012)

$$\frac{\Delta T}{\Delta I}\bigg|_{t\to\text{equilib}} = \frac{a}{\lambda I_B}. \quad (13)$$

To evaluate the equilibrium response, a (W m^{-2}) is the coefficient for the radiative forcing dependence on atmospheric CO_2 (Myhre *et al* 1998, Forster *et al* 2013); λ (W m^{-2} K^{-1}) is the climate parameter (Gregory *et al* 2004) and equivalently λ^{-1} is the equilibrium climate sensitivity (Knutti and Hegerl 2008); and I_B (PgC) is the buffered carbon inventory for the atmosphere and ocean (Goodwin *et al* 2007).

The buffered carbon inventory of the atmosphere and ocean, I_B, represents the accessible amount of

carbon in the atmosphere and ocean, and takes into account how the buffering of the ocean carbonate system inhibits the release of carbon from the ocean to the atmosphere: I_B is defined for the pre-industrial era (Goodwin *et al* 2007) as

$$I_B = I_A + V\, \text{DIC}_{sat}/B, \quad (14)$$

where I_A is the atmospheric inventory of carbon (PgC), DIC$_{sat}$ is the saturated ocean dissolved inorganic carbon (gC m^{-3}) and B is the non-dimensional buffer factor, and V is the volume of the ocean (m^3) (table 1). The buffer factor, B, represents the enhanced fractional changes in atmospheric CO_2, relative to the fractional changes in the saturated dissolved inorganic carbon DIC$_{sat}$, where $B = (\delta CO_2/CO_2)/(\delta \text{DIC}_{sat}/\text{DIC}_{sat})$; B and DIC$_{sat}$ are evaluated using CO_2 and global-mean temperature, salinity, alkalinity and phosphate values for the pre-industrial ocean (Follows *et al* 2006, Williams and Follows 2011).

3. Model diagnostics of the TCRE and its dependence on surface warming and radiative forcing

The transient climate responses for the two Earth System Models of differing complexity are now investigated, using our framework for how the TCRE relates to dependences of the surface warming and radiative forcing.

3.1. Earth System Model configurations

The Earth System Model of Intermediate Complexity (GENIE) is configured as a coarse-resolution atmosphere–ocean system, containing coupled circulation and biogeochemistry with 16 ocean layers and 36×36 equal-area grid elements over the globe (Ridgwell *et al* 2007). This version of GENIE includes climate feedbacks, in which increased CO_2 increases the radiative forcing and warms the climate; in this application of GENIE, sediment interactions are disabled. The model configurations are forced to reproduce historical CO_2 concentrations to 2005 and then to follow Representative Concentration Pathways (RCPs, Moss *et al* 2010) until 2100 for those RCPs with significant warming (RCP 4.5, 6.0 and 8.5).

IPSL-CM5A-LR is an ocean–atmosphere general circulation model including an explicit representation of the carbon cycle on land and in the ocean (Dufresne *et al* 2013). In this version, the atmospheric component (LMDZ) is integrated with 39 vertical levels and a horizontal resolution of $3.75° \times 1.9°$, and the oceanic component (NEMOv3.2) is integrated with 31 vertical levels and a horizontal resolution ranging from $0.5°$ to $2°$. For the simulations diagnosed here, the model is forced with prescribed atmospheric CO_2 concentrations, following historical CO_2 values until 2005, and the different RCP from 2005 to 2100; RCP 2.6, 4.5, 6.0 and 8.5. Other forcings, such as land-use changes, other greenhouse gases and anthropogenic aerosols are also included.

The carbon cycle components, ORCHIDEE for the land (Krinner *et al* 2005) and PISCES for the ocean (Aumont and Bopp 2006), provide net carbon fluxes from the atmosphere to the terrestrial biosphere and the ocean, respectively. These fluxes include the net effect of changing atmospheric CO_2 and those of climate feedbacks. Compatible anthropogenic emissions are diagnosed a posteriori, using modelled air-to-sea and air-to-land carbon fluxes as well as imposed atmospheric CO_2 trajectories, as applied in Jones *et al* (2013).

3.2. Assessment using an Earth System Model of Intermediate Complexity

The transient climate response is now assessed using GENIE, configured as an idealised atmosphere–ocean model; see Goodwin *et al* (2015) for further model details. There is only radiative forcing from atmospheric CO_2 with no changes to other radiative forcing agents, such as aerosols, and no changes in the terrestrial sink of carbon, so that the TCRE dependence in (12) simplifies to

$$\text{TCRE} = \frac{\Delta T}{\Delta I} = \frac{a}{\lambda I_B}\left(1 - \frac{\varepsilon N(t)}{\Delta R(t)}\right)\left(1 + \frac{I_{\text{Usat}}(t)}{\Delta I(t)}\right).$$

$$(15)$$

Within GENIE, the equilibrium TCRE from $a/(\lambda I_B)$ is given by $1.35 \, \text{K}(1000 \, \text{PgC})^{-1}$, where

$a = 5.35 \, \text{W m}^{-2}$, $\lambda = 1.28 \, \text{W m}^{-2} \, \text{K}^{-1}$ and $I_B = 3340 \, \text{PgC}$ (table 1). The efficacy factor weighting the effect of the ocean heat flux on surface temperature is 1.07 (table 1; Goodwin *et al* 2015).

The temporal evolution of the TCRE is determined by (i) how the dependence of surface warming on radiative forcing, $\Delta T/\Delta R$, varies with the normalised ocean heat uptake, $\varepsilon N(t)/\Delta R(t)$, versus (ii) how the dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2}/\Delta I$, varies with the normalised ocean undersaturation in carbon, $I_{\text{Usat}}(t)/\Delta I(t)$.

Within the GENIE model, the dependence of surface warming on radiative forcing, $\Delta T/\Delta R$, increases in time (figure 2(a)). This increase in $\Delta T/\Delta R$ is due to changes in ocean heat uptake as the ocean interior warms and its temperature increases from its pre-industrial state; the ocean surface temperature then increases more rapidly for a specified increase in radiative forcing, since a smaller fraction of the radiative forcing is taken up by the ocean heat flux to warm the ocean interior. This response is equivalent to surface warming leading to a thinning of the surface mixed layer and increasing the stratification in the underlying ocean interior.

The dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2}/\Delta I$, decreases in time (figure 2(b)). This decrease in $\Delta R_{CO_2}/\Delta I$ is due to the ocean taking up carbon, so that the ocean carbon undersaturation decreases. The increase in radiative forcing from atmospheric CO_2 then becomes smaller for additional increases in emitted carbon with a smaller change in the logarithm of CO_2 induced per unit carbon emitted through the decrease in ocean undersaturation.
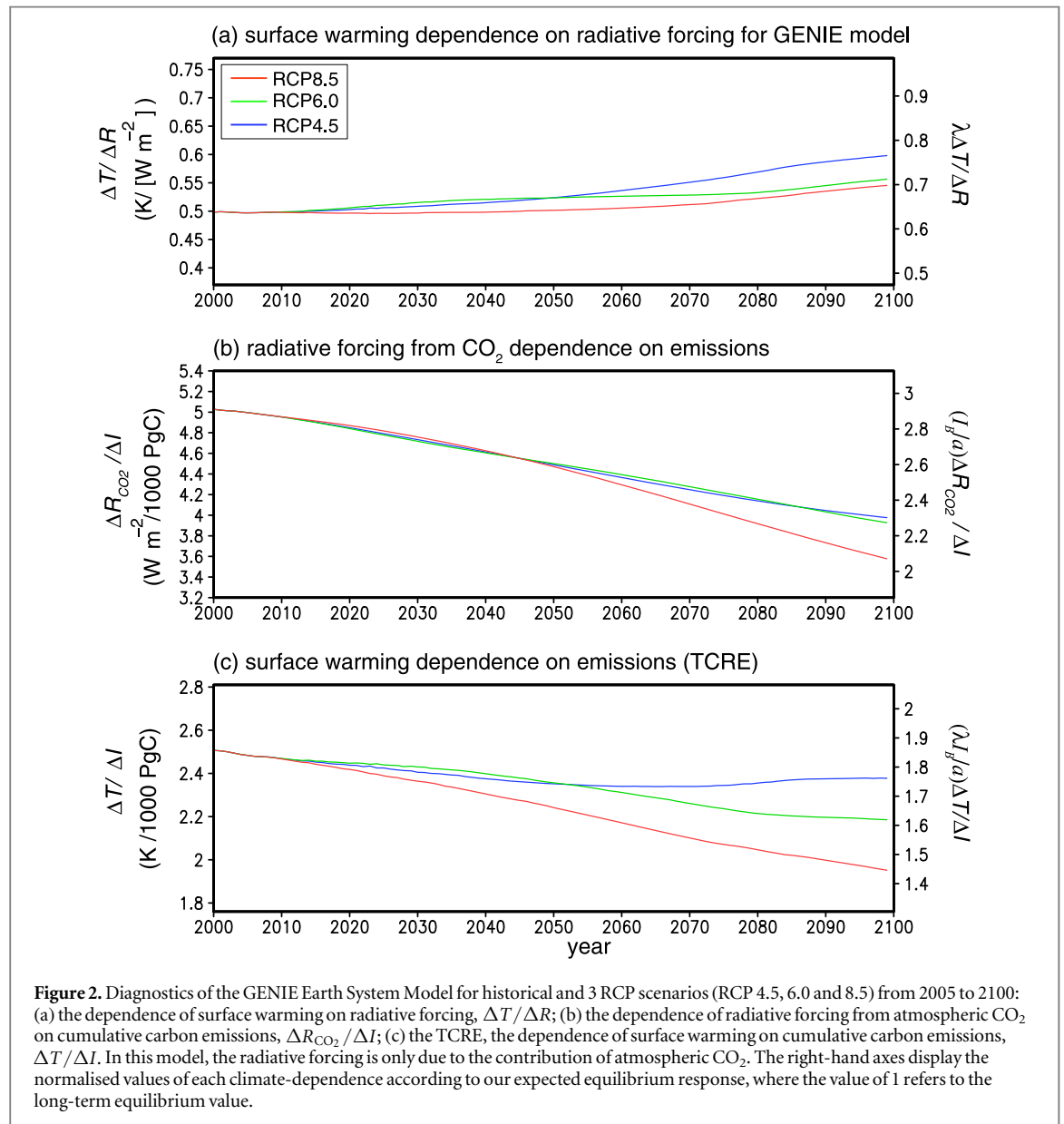
In consequence, there is a relatively small decrease in the TCRE (figure 2(c)) through a slightly larger decrease in $\Delta R_{CO_2}/\Delta I$ being partly compensated by a smaller increase in $\Delta T/\Delta R$. There is only a relatively weak sensitivity to emission pathway on this centennial timescale (different coloured lines in figure 1(a), left panel and figure 2).

3.3. Assessment using an Earth System Model

In the more complex Earth System Model IPSL-CM5A-LR, the warming response is explored for four RCP emissions scenarios (figure 1(a), right panel). The climate projections are affected also by non-CO_2 radiative forcing, including changes from anthropogenic aerosols, non-CO_2 greenhouse gases, land-use changes and terrestrial changes in carbon storage. Accordingly, the TCRE is now diagnosed using the full expression in (11).

3.3.1. Equilibrium response

Our theory suggests that the TCRE asymptotes to an equilibrium response given by the time-independent factors $a/(\lambda I_B)$, each of these input parameters differ slightly from those for GENIE. The buffered carbon

Figure 2. Diagnostics of the GENIE Earth System Model for historical and 3 RCP scenarios (RCP 4.5, 6.0 and 8.5) from 2005 to 2100: (a) the dependence of surface warming on radiative forcing, $\Delta T / \Delta R$; (b) the dependence of radiative forcing from atmospheric CO_2 on cumulative carbon emissions, $\Delta R_{CO_2} / \Delta I$; (c) the TCRE, the dependence of surface warming on cumulative carbon emissions, $\Delta T / \Delta I$. In this model, the radiative forcing is only due to the contribution of atmospheric CO_2. The right-hand axes display the normalised values of each climate-dependence according to our expected equilibrium response, where the value of 1 refers to the long-term equilibrium value.

inventory I_B is diagnosed as 3440 PgC, similar to the estimate of 3340 PgC for GENIE (table 1). The climate parameter is diagnosed as $\lambda = 0.77$ W m^{-2} K^{-1} by solving a least-squares regression of the surface heat balance (Gregory *et al* 2004); comparing the time series of the radiative forcing minus ocean heat uptake, $\Delta R - \varepsilon N$ (W m^{-2}), versus surface temperature change, ΔT (K) (figure 3(a), table 2). Our estimate of the climate parameter, λ, is very close to estimates of 0.75, 0.76 and 0.79 W m^{-2} K^{-1} based on climate-model projections for an abrupt 4 × CO_2 increase, respectively, from Andrews *et al* (2012), Dufresne *et al* (2013) and Geoffroy *et al* (2013). The climate parameter, λ, is assumed to remain constant over time and is evaluated over the climate model integrations to 2100, rather than from an integration to an equilibrium state: the resulting values of λ vary with the RCP scenario (table 2), which might possibly reflect a time-dependence in its response (Senior and

Mitchell 2000) or a sensitivity to regional changes in warming and ocean circulation that differ with each RCP (Armour *et al* 2013, Winton *et al* 2013).

The coefficient for the radiative forcing dependence on atmospheric CO_2, *a*, is taken as 4.47 W m^{-2} from the effective radiative forcing diagnosed for a climate model projection with an abrupt 4 × CO_2 increase (Andrews *et al* 2012). This effective radiative forcing estimate is less than the expected theoretical value of 5.35 W m^{-2} (Myhre *et al* 1998) and takes into account rapid adjustments in clouds and other tropospheric and land-surface changes (Forster *et al* 2013).

The diagnosed values of *a*, λ and I_B, then suggest an equilibrium dependence of surface warming to cumulative carbon emissions for the Earth System Model of 1.69 K(1000 PgC)$^{-1}$, slightly larger than the equilibrium dependence of 1.35 K(1000 PgC)$^{-1}$ for GENIE (table 1).
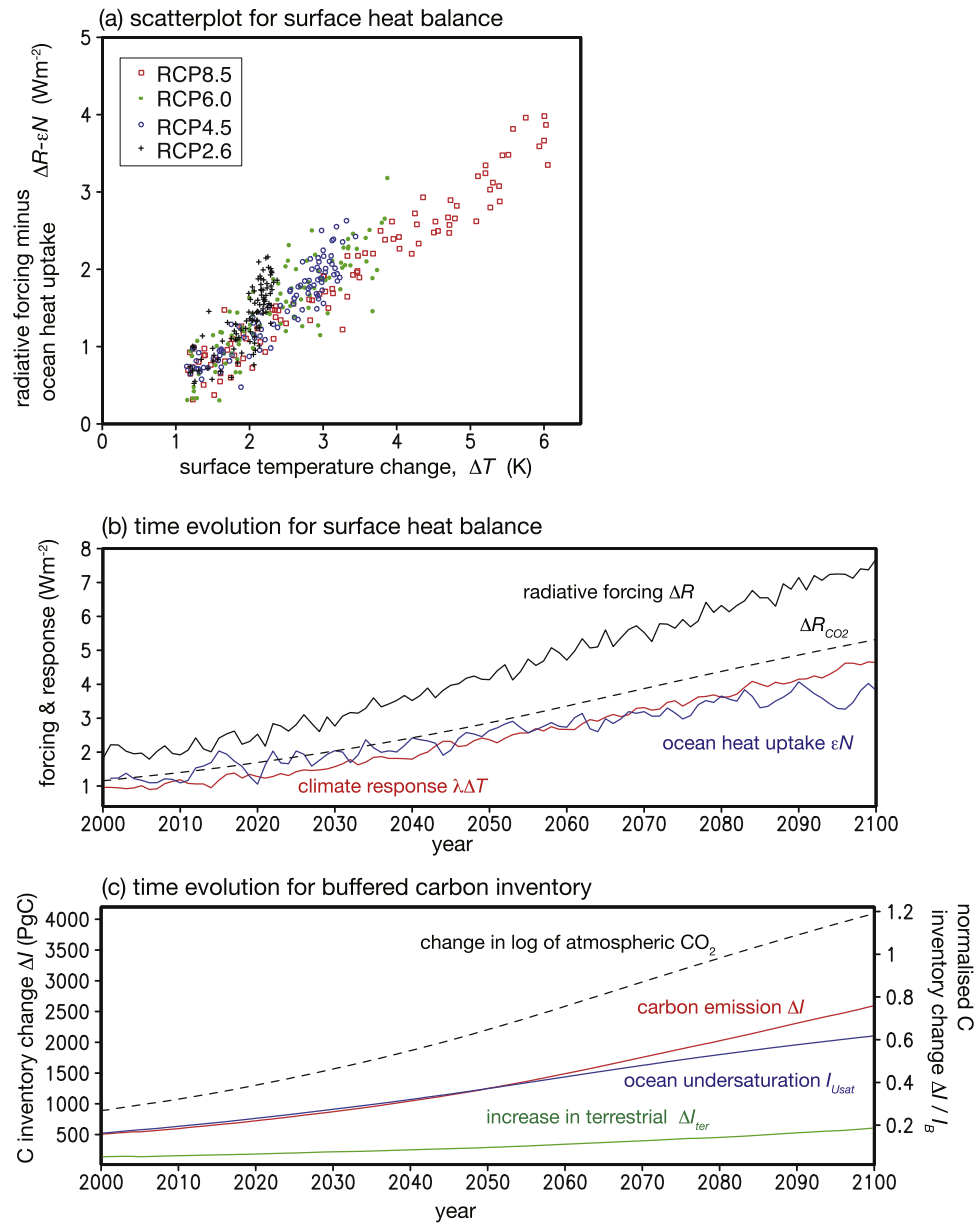
Figure 3. Diagnostics of IPSL-CM5A-LR simulations for the surface heat balance and buffered carbon inventory from 2005 to 2100: (a) scatterplot showing the relationship between the radiative forcing minus ocean heat uptake, $\Delta R - \varepsilon N$ (W m^{-2}) versus surface temperature change, ΔT (K), for RCP2.6, RCP4.5, RCP6.0 and RCP8.5. A least squares fit to the slope provides an estimate of the climate parameter λ with a mean value of 0.77 W m^{-2} K^{-1} for the four RCPs (table 2); this scatterplot is for an efficacy ε of 1.14 (Geoffrey *et al* 2013); (b) temporal variation of the surface heat balance (W m^{-2}) for RCP8.5 (assuming $\lambda = 0.77$ W m^{-2} K^{-1} and $\varepsilon = 1.14$). The radiative forcing $\Delta R(t)$ (black line) is mainly from the effect of atmospheric CO_2 (black dashed line). The radiative forcing $\Delta R(t)$ is partly offset by an ocean heat uptake $N(t)$ (blue line), and the resulting net heating drives the climate response of a surface warming and increase in outgoing longwave radiation, $\lambda \Delta T(t)$ (red line). At the start of the integration, the climate response is slightly less than the ocean heat uptake, while by the end of the integration, the climate response slightly exceeds the ocean heat uptake. Thus, the dependence of surface warming on radiative forcing increases in time. In (c), accompanying temporal changes in the carbon inventory (PgC) for carbon emission (red line), ocean undersaturation (blue line) and change in terrestrial store (green line) together with the change in the log of atmospheric CO_2 (black dashed line) that balances the sum of the inventory changes normalised by $I_B = 3440$ PgC (right axis). The climate emission is initially comparable to the ocean undersaturation, but then proceeds to exceed the ocean undersaturation as the ocean sequesters carbon from the atmosphere. Thus, the dependence of radiative forcing from atmospheric CO_2 on carbon emissions decreases in time.

3.3.2. Transient response

To understand the transient response, the different terms in the global heat balance (6) and buffered carbon inventory (8) are diagnosed. The heat flux into the climate system, $N(t)$, is equated with the ocean heat uptake, and diagnosed from the tendency in the global ocean heat content. The ocean carbon undersaturation, $I_{Usat}(t)$, is diagnosed from the difference in the saturated DIC anomaly and the actual DIC anomaly over the global ocean, $I_{Usat}(t) = V(\Delta DIC_{sat}(t) - \Delta DIC(t))$.

There are temporal variations in the surface heat balance (6) over the 100 year integration. The radiative forcing, $\Delta R(t)$, is dominated by the contribution from

Table 2. Estimates of the climate parameter λ from the Earth System Model IPSL-CM5a-LR for emission pathways: RCP2.6, RCP4.5, RCP6.0 and RCP8.5. The climate parameter is estimated from a least squares fit between the radiative forcing minus ocean heat uptake, $\Delta R - \varepsilon N$ (W m^{-2}) and surface temperature change ΔT (K) for an efficacy ε of 1.14 (Geoffrey *et al* 2013). The average value of λ for all RCPs is 0.77; all correlation coefficients are significant ($r > 0.26$ for a 99% confidence limit).

Emission pathway	Climate parameter	Correlation coefficient
	λ (W m^{-2} K^{-1})	r
RCP 2.6	1.02	0.78
RCP 4.5	0.76	0.92
RCP 6	0.68	0.86
RCP 8.5	0.63	0.97

atmospheric CO_2, $\Delta R_{CO_2}(t)$ (figure 3(b), black full and dashed lines for RCP8.5). The radiative forcing, $\Delta R(t)$, is partly offset by an ocean heat uptake, $N(t)$ (blue line), and the remaining net forcing drives the surface climate response, $\lambda \Delta T(t)$ (red line) in figure 3(b); assuming $\varepsilon = 1.14$ from Geoffroy *et al* (2013). At the start of the integration, the climate response is slightly less than the ocean heat uptake, while by the end of the integration, the climate response is slightly larger than the ocean heat uptake. Thus, the normalised ocean heat uptake, $\varepsilon N(t)/\Delta R(t)$, decreases in time in (11), leading to the dependence of surface warming on radiative forcing increasing in time.

There are likewise temporal changes in the buffered carbon balance (8), although the carbon inventory changes are noticeably smoother. The ocean carbon undersaturation, $I_{Usat}(t)$ (blue line), is initially comparable to the cumulative carbon emission, $\Delta I(t)$ (red line), but becomes slightly smaller over the 100 year integration in figure 3(c). Thus, the normalised ocean undersaturation, $I_{Usat}(t)/\Delta I(t)$, reduces in time and there is also a relatively small increase in the terrestrial store of carbon, $\Delta I_{ter}(t)$ (figure 3(c), green line). Thus, $(I_{Usat}(t) - \Delta I_{ter}(t))/\Delta I(t)$ reduces in time in (11), leading to the dependence of radiative forcing on carbon emissions decreasing in time.

Accordingly, there are larger centennial trends for the dependencies making up the TCRE$_{all\ R}$, than for the TCRE$_{all\ R}$: there is a long-term increase in the dependence of surface warming on radiative forcing, $\Delta T/\Delta R$ (figure 4(a)), a smaller decrease in the fractional dependence of radiative forcing from atmospheric CO_2, $\Delta R/\Delta R_{CO_2}$ (figure 4(b)), and a larger decrease in the dependence of radiative forcing on carbon emissions, $\Delta R_{CO_2}/\Delta I$ (figure 4(c)). The resulting TCRE$_{all\ R}$ then only slightly decreases over the centennial timescale (figure 4(d)).

There are though decadal changes in the TCRE$_{all\ R}$ (figure 4(d)). The decadal variability in $\Delta T/\Delta I$ originates from decadal changes in the dependence of surface warming on radiative forcing, $\Delta T/\Delta R$ (figure 4(a)) from changes in ocean heat uptake (figure 3(b), blue line), and the fractional dependence

of radiative forcing on atmospheric CO_2, $\Delta R/\Delta R_{CO_2}$ (figure 4(b)) from the effects of aerosols and non-CO_2 greenhouse gases. In contrast, the dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2}/\Delta I$, varies relatively smoothly in time (figure 4(c)) due to the smooth changes in carbon emissions and ocean carbon undersaturation (figure 3(c), red and blue lines).

3.4. Comparison of Earth System Models of different complexity

Our framework for the TCRE$_{all\ R}$ highlights how there are broadly similar responses in the Earth System Models, GENIE and IPSL-CM5A-LR. The centennial trends of the TCRE$_{all\ R}$ and its dependences have similar signs to each other in both Earth System Models (figures 2 and 4): both models show a positive trend in the dependence of surface warming on radiative forcing, a negative trend in the dependence of radiative forcing on carbon emissions, and a resulting slight negative trend in the TCRE. Thus, there is traceability between these Earth System Models of different complexity, as there are similar underlying trends in their response.

The exact values of the dependences differ though in each Earth System Model (table 3), as the models contain different representations of the physical circulation and biogeochemistry. The TCRE$_{all\ R}$ is higher in the IPSL-CM5A-LR model, than in GENIE, and subsequently reduces by a greater amount over the next century (table 3). These differences in the TCRE$_{all\ R}$ are due to (i) a larger dependence of the surface warming on radiative forcing in IPSL-CM5A-LR than in GENIE, and (ii) conversely a smaller dependence of the radiative forcing from atmospheric CO_2 on cumulative emissions in IPSL-CM5A-LR than in GENIE. There are also interannual and decadal differences in each model due to the IPSL-CM5A-LR containing radiative forcing contributions from aerosols and non-CO_2 greenhouse gases, as well as more complex ocean dynamics altering the ocean uptake of heat and carbon.

There are only slight differences in each of the climate dependences for the different choices of the RCP emission pathways over a century timescale (figures 1(a), 2 and 4; table 3).

4. Discussion

Our framework to understand the variation of the TCRE draws upon the fundamental link between the logarithm of atmospheric CO_2 and the definitions of the radiative forcing from atmospheric CO_2 (5) and a buffered carbon inventory (6) accounting for ocean carbonate chemistry.
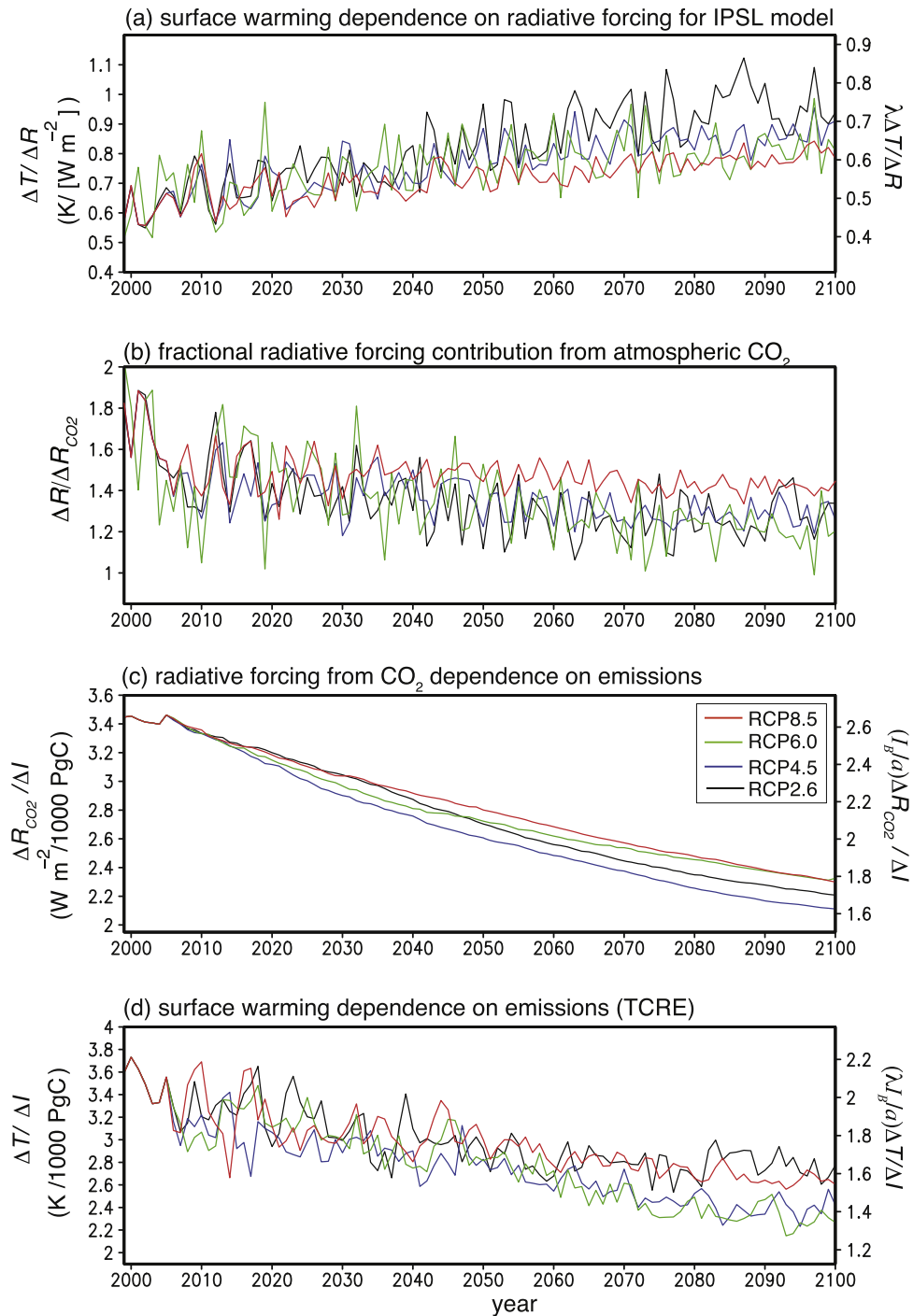
Figure 4. Diagnostics of IPSL-CM5A-LR simulations for historical and 4 RCP scenarios (RCP 2.6, 4.5, 6 and 8.5) from 2005 to 2100: (a) dependence of surface warming on effective radiative forcing, $\Delta T / \Delta R$ (K $W^{-1}m^{-3}$); (b) fractional radiative forcing contribution from atmospheric CO_2, $\Delta R / \Delta R_{CO_2}$; (c) dependence of radiative forcing from CO_2 on carbon emissions, $\Delta R_{CO_2} / \Delta I$ (W m^{-2}/ 1000 PgC) and (d) the $TCRE_{all\,R}$, the dependence of surface warming on carbon emissions, $\Delta T / \Delta I$ (K/1000 PgC) versus time. The right-hand axes display the normalised values of each climate-dependence according to our expected equilibrium response, where the value of 1 refers to the long-term equilibrium value. Note that the vertical range in (a) and (d) are a factor 2 larger than in figure 2.
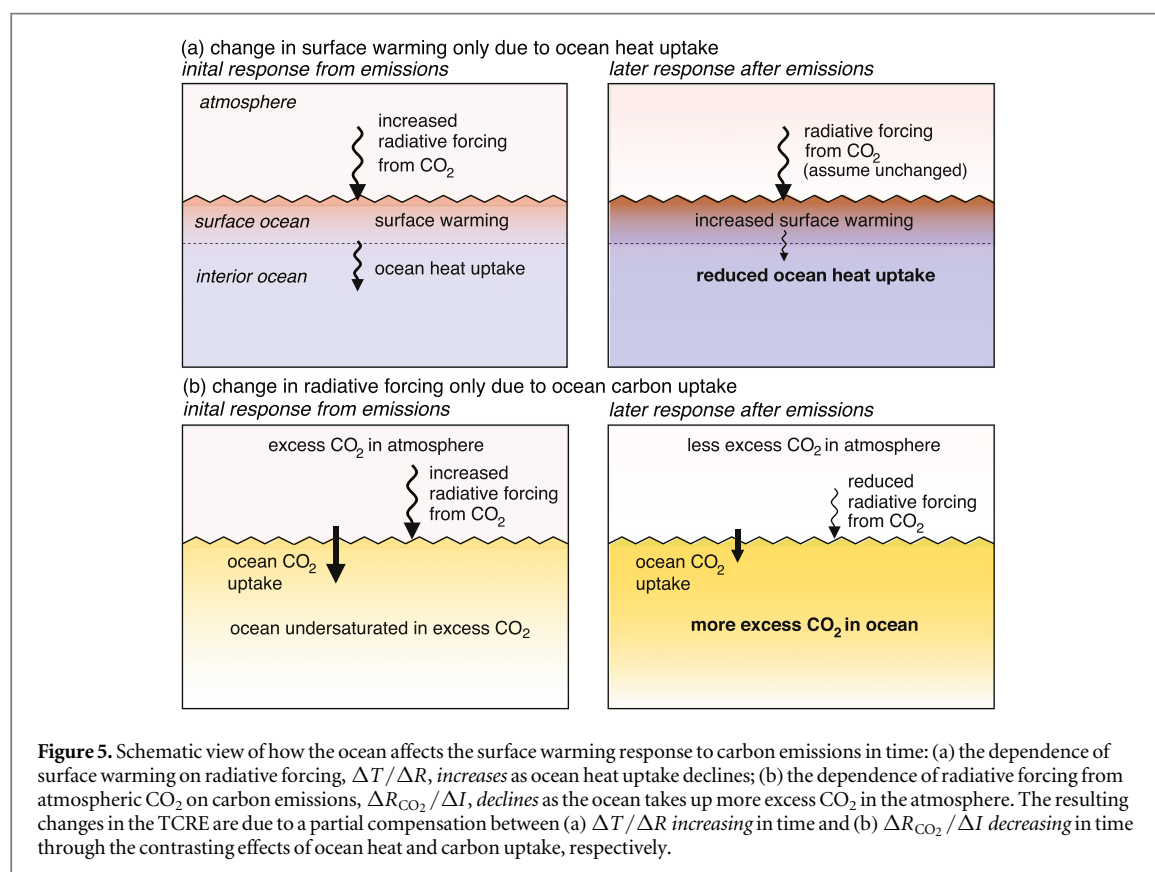
4.1. Alternative frameworks

There are two alternative approaches to our framework. Firstly, the TCRE has been previously explained in terms of the dependence of surface temperature on atmospheric CO_2, $\Delta T / \Delta CO_2$, multiplied by the dependence of atmospheric CO_2 on cumulative emissions, $\Delta CO_2 / \Delta I$ (Matthews *et al* 2009, Gillett *et al* 2013). Their approach highlights how the TCRE is nearly constant due to partial compensation between changes in the dependence of surface temperature on CO_2 and changes in the air-borne fraction of emitted carbon.

Secondly, the sensitivities of climate models have been understood in terms of empirical parameters, α, β and γ, representing the sensitivity of temperature to atmospheric CO_2, and the sensitivity of carbon inventories to atmospheric CO_2 and temperature respectively (Friedlingstein *et al* 2006, Plattiner *et al* 2008).

Table 3. Climate dependences for two Earth System Models at 2005 and 2100 for different choices of emission pathways.

	Dependence of surface warming on radiative forcing, $\Delta T/\Delta R$ K (W m^{-2})$^{-1}$		Dependence of radiative forcing from atmospheric CO$_2$ on cumulative carbon emissions, $\Delta R_{CO_2}/\Delta I$ W m^{-2} (1000 PgC)$^{-1}$		Dependence of surface warming on cumulative carbon emissions, the TCRE$_{all\,R}$, $\Delta T/\Delta I$ K (1000 PgC)$^{-1}$	
(a) Earth System model of Intermediate Complexity, GENIE						
Year	2005	2100	2005	2100	2005	2100
RCP 4.5	0.50	0.60	5.00	3.98	2.48	2.38
RCP 6	0.50	0.56	5.00	3.93	2.48	2.19
RCP 8.5	0.50	0.55	5.00	3.58	2.48	1.95
(b) Earth System Model IPSL–CM5A-LR						
Year	2005	2100	2005	2100	2005	2100
RCP 2.6	0.68	0.94	3.46	2.21	3.55	2.67
RCP 4.5	0.67	0.91	3.46	2.11	3.55	2.56
RCP 6	0.71	0.81	3.46	2.33	3.55	2.31
RCP 8.5	0.67	0.78	3.46	2.30	3.55	2.65

Figure 5. Schematic view of how the ocean affects the surface warming response to carbon emissions in time: (a) the dependence of surface warming on radiative forcing, $\Delta T / \Delta R$, *increases* as ocean heat uptake declines; (b) the dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2} / \Delta I$, *declines* as the ocean takes up more excess CO_2 in the atmosphere. The resulting changes in the TCRE are due to a partial compensation between (a) $\Delta T / \Delta R$ *increasing* in time and (b) $\Delta R_{CO_2} / \Delta I$ *decreasing* in time through the contrasting effects of ocean heat and carbon uptake, respectively.

This approach has been used to understand the contributions of the different land and ocean inventories to carbon uptake and feedback on climate.

In our view, the connection to the underlying theory is more transparent in our framework, since there are clear dependences on the logarithm of atmospheric CO_2 drawing upon the buffered carbon inventory (8) and the definition of radiative forcing from atmospheric CO_2 in (9). While our framework includes a term accounting for changes in the terrestrial carbon sink, our theory is designed to make clearer the central contribution of the ocean, rather than drawing upon empirical relationships between changes in the land sinks of carbon (Friedlingstein *et al* 2006).

4.2. Mechanistic control of the TCRE

The long-term trends for the TCRE can be simply understood in terms of the heat and carbon uptake for an idealised representation of the atmosphere–ocean-terrestrial system. The additional heat supplied to the climate system is primarily taken up by the ocean, while the additional carbon is taken up by both the ocean and terrestrial system.

In order to understand how the TCRE is controlled, now consider how the climate response to carbon emissions is likely to alter in time:

(i) the additional heat supplied from an increase in radiative forcing is used to warm both the surface and ocean interior. As the ocean interior warms in time, a smaller proportion of the radiative forcing is taken by the interior and a larger proportion of the radiative forcing then warms the surface. Thus, the dependence of surface warming on radiative forcing, $\Delta T / \Delta R$, is expected to *increase* in time as the ocean interior adjusts to a new equilibrium (figure 5(a)). Equivalently, the radiative forcing is likely to increase the stratification in time and lead to an increase in the dependence of surface warming on radiative forcing.

(ii) For a given carbon emission, radiative forcing is initially expected to be large as most of the emitted carbon is in the form of atmospheric CO_2. The radiative forcing is then expected to decline as the excess atmospheric CO_2 is taken up by the ocean and terrestrial system. Thus, the dependence of radiative forcing from atmospheric CO_2 on cumulative carbon emissions, $\Delta R_{CO_2} / \Delta I$, is expected to *decrease* in time as the ocean and terrestrial system take up the excess atmospheric CO_2 (figure 5(b)).

(iii) The resulting *surface warming response to cumulative carbon emissions* (the TCRE) then only weakly varies, depending on the partial compensation between the dependences of surface warming on radiative forcing, $\Delta T / \Delta R$, and the radiative forcing from atmospheric CO_2 on emissions, $\Delta R_{CO_2} / \Delta I$.

This inference supports the argument of Solomon *et al* (2009) as to why long-term irreversible climate change persists after the cessation of carbon emissions; the gradual reduction in radiative forcing from atmospheric CO_2 is partly compensated by a decline in ocean heat uptake, so that global warming is likely to continue to persist on a millennial timescale.

4.3. Decadal variability

The decadal changes in the $TCRE_{allR}$ can be understood via our relationship (11). The dependence of surface warming on radiative forcing, $\Delta T/\Delta R$, is controlled by the heat flux from the atmosphere to the ocean weighted by the efficacy, such that $\frac{\partial T}{\partial R} = \frac{1}{\lambda}\left(1 - \frac{\varepsilon N(t)}{\Delta R(t)}\right)$. The magnitude and sign of the ocean heat flux (figure 3(b), blue line) can easily change on interannual to decadal timescales given changes in atmosphere and ocean heat storage. For example, there is a hiatus in surface warming for 2000–2010, which is attributed to an increase in ocean heat uptake (Guemas *et al* 2013, Watanabe *et al* 2013). These changes in ocean heat uptake might be associated with coupled atmosphere–ocean phenomena, such as the El Nino Southern Oscillation, or changes in ocean ventilation, particularly associated with formation of water masses in the North Atlantic and Southern Ocean.

The dependence of radiative forcing on the contribution from atmospheric CO_2, $\Delta R/\Delta R_{CO_2}$ (figure 4(c)) can likewise easily alter through the release of atmospheric aerosols, particularly linked to volcanic and anthropogenic emissions (Ottera *et al* 2010, Booth *et al* 2012), and changes in other non-CO_2 greenhouse gases, such as methane (Pierrehumbert 2014).

The dependence of radiative forcing from CO_2 on carbon emissions, $\Delta R_{CO_2}/\Delta I$, is controlled by the carbon undersaturation of the ocean relative to the atmosphere and changes in the terrestrial sink of carbon, $\frac{\partial R_{CO2}}{\partial I} = \frac{a}{I_B}\left(1 + \frac{I_{Usat}(t)}{\Delta I(t)} - \frac{\Delta I_{ter}(t)}{\Delta I(t)}\right)$. The carbon undersaturation of the ocean, $I_{Usat}(t)$, only alters slowly in time (figure 3(c), blue line) and is less sensitive to the interannual and decadal changes in the carbon flux between the atmosphere and ocean, such as seen in climate model simulations of natural variability in air-sea CO_2 and O_2 fluxes (Resplandy *et al* 2015). The terrestrial sink probably weakly increases in time, $\Delta I_{ter}(t)$ (figure 3(c), green line), although there is a large range in the magnitude and sign in the projected terrestrial response for different climate models (Friedlingstein *et al* 2006).

5. Conclusions

The mechanisms controlling the TCRE from all radiative forcings are revealed here in a new framework, formulated in terms of the product of three climate dependences in (5): the dependence of surface warming on radiative forcing, $\Delta T/\Delta R$, the fractional dependence of radiative forcing from atmospheric CO_2, $\Delta R/\Delta R_{CO_2}$, and the dependence of radiative forcing from atmospheric CO_2 on carbon emissions, $\Delta R_{CO_2}/\Delta I$:

$$TCRE_{allR} = \frac{\Delta T}{\Delta I} = \left(\frac{\Delta T}{\Delta R}\right)\left(\frac{\Delta R}{\Delta R_{CO_2}}\right)\left(\frac{\Delta R_{CO_2}}{\Delta I}\right).$$

This definition of the $TCRE_{allR}$ can then be expressed in terms of an equilibrium response multiplied by three time-dependent terms in (11): depending on the damping of surface warming due to ocean heat flux, the enhancement of atmospheric CO_2 due to carbon undersaturation of the ocean and terrestrial system, and the fractional dependence of radiative forcing on atmospheric CO_2:

$$TCRE_{allR} = \frac{a}{\lambda I_B}\left(1 - \frac{\varepsilon N(t)}{\Delta R(t)}\right)$$
$$\times \left(1 + \frac{I_{Usat}(t)}{\Delta I(t)} - \frac{\Delta I_{ter}(t)}{\Delta I(t)}\right)\left(\frac{\Delta R(t)}{\Delta R_{CO_2}(t)}\right).$$

Applying this framework, the near constancy of the TCRE on multi-decadal to centenial timescales is explained by a partial compensation between the effects of ocean heat and carbon uptake on the dependences of surface warming on radiative forcing and radiative forcing from atmospheric CO_2 on carbon emissions. The dependence of surface warming on emissions then remains nearly constant, since the dependence of surface warming on radiative forcing is enhanced as ocean heat uptake declines, but at the same time the dependence of radiative forcing on carbon emissions is reduced as the ocean takes up excess CO_2 in the atmosphere (Goodwin *et al* 2015). Thus, this climate response primarilly depends on the physical mechanisms by which the ocean sequesters heat and carbon, involving the physical transport of surface waters into the ocean interior.

There is no need though for the ocean sequestering of heat and carbon to always mirror each other and the TCRE can respond differently to rapid or slow rates of emissions given sufficient time (Krasting *et al* 2014), as well as be affected by changes in ocean circulation (Winton *et al* 2013) and regional feedbacks (Armour *et al* 2013). There may even be continued global warming after carbon emissions cease (Frölicher *et al* 2014) due to the warming effect of decreasing ocean heat uptake exceeding the cooling effect of decreasing atmospheric CO_2.

The TCRE from all radiative forcing can easily vary on decadal timescales through changes in the sign of the annual ocean heat uptake and changes in radiative forcing from non-CO_2 contributions, such as aerosols and other greenhouse gases.

In summary, our framework for the TCRE and link to underlying theory provides an alternative way to interpret the response of climate models and understand how the climate models are mechanistically controlled. The ocean is providing a central role in controlling the TCRE through the uptake of heat and carbon in the climate system. The framework can also be used to provide traceability between simple and complex models, and understand the underlying

cause for their different responses. The robustness of the TCRE on centennial timescales also supports the view that cumulative carbon emissions provides a policy framework for reducing adverse climate change (Matthews *et al* 2012).

Acknowledgments

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