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 CO₂ and the dependence of radiative forcing from atmospheric CO₂ 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-CO₂ 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 CO₂ 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-CO₂ radiative forcing, as represented by decadal changes in the dependences of surface warming on radiative forcing and the fractional radiative forcing contribution from atmospheric CO₂. 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
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 atmospheric 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-CO₂ 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 \( T(t_\text{pre}) \). The cumulative amount of carbon (PgC) emitted to the atmosphere from all anthropogenic sources at time \( t \) is similarly defined, \( I(t) = I(t_\text{pre}) + \Delta I(t) \), with \( I(t_\text{pre}) \) 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 \( \Delta T / \Delta I \),

\[
\Delta T(t) = \left( \frac{\Delta T}{\Delta I} \right) \Delta I(t),
\]

where the \( \Delta \) 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 CO₂, which is defined by \( R_{\text{CO}_2}(t) = R_{\text{CO}_2}(t_\text{pre}) + \Delta R_{\text{CO}_2}(t) \). 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 CO₂, \( \Delta T / \Delta R_{\text{CO}_2} \), and the dependence of radiative forcing on carbon emissions, \( \Delta R_{\text{CO}_2} / \Delta I \):

\[
\text{TCRE} = \left( \frac{\Delta T}{\Delta R_{\text{CO}_2}} \right) \left( \frac{\Delta R_{\text{CO}_2}}{\Delta I} \right).
\]  

More generally, the radiative forcing driving climate change includes contributions from atmospheric CO₂ and other non-CO₂ contributions, so that the change in radiative forcing since the pre-industrial era is given by \( \Delta R(t) = \Delta R_{\text{CO}_2}(t) + \Delta R_{\text{nonCO}_2}(t) \). The change in non-CO₂ radiative forcing, \( \Delta R_{\text{nonCO}_2}(t) \), includes contributions from aerosols emitted via anthropogenic and volcanic activity (Ottera et al 2010, Booth et al 2012) and other non-CO₂ greenhouse gases, such as methane with a shorter lifetime than CO₂ (Pierrehumbert 2014).

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

\[
\text{TCRE}_{\text{allR}} = \left( \frac{\Delta T}{\Delta I} \right) = \left( \frac{\Delta T}{\Delta R} \right) \left( \frac{\Delta R}{\Delta I} \right),
\]

while this definition loses the clarity of the original TCRE definition in (2) in terms of isolating the climate response to cumulative carbon emissions, this generalisation for TCRE\(_{\text{allR}}\) in (3) is needed if the climate response is to be diagnosed from a climate model including non-CO₂ radiative forcing. To extend the TCRE definition in (3), the surface temperature change in (1) is written as

\[
\Delta T(t) = \Delta T / \Delta I \Delta I(t) = \left( \frac{\Delta T}{\Delta R} \right) \left( \frac{\Delta R_{\text{CO}_2}}{\Delta I} \right) \Delta I(t) + \left( \frac{\Delta T}{\Delta R_{\text{nonCO}_2}} \right) \Delta R_{\text{nonCO}_2}(t),
\]

which makes explicit the surface warming from the separate effects of the radiative forcing from CO₂ and non-CO₂ contributions. Exploiting the definition of \( \Delta R(t) = \Delta R_{\text{CO}_2}(t) + \Delta R_{\text{nonCO}_2}(t) \), the surface warming is then equivalently written as

\[
\Delta T(t) = \Delta T / \Delta I \Delta I(t) = \left( \frac{\Delta T}{\Delta R} \right) \left( \frac{\Delta R}{\Delta R_{\text{CO}_2}} \right) \Delta R_{\text{CO}_2}(t) \Delta I(t).
\]

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

\[
\text{TCRE}_{\text{allR}} = \left( \frac{\Delta T}{\Delta I} \right) = \left( \frac{\Delta T}{\Delta R} \right) \left( \frac{\Delta R}{\Delta R_{\text{CO}_2}} \right) \left( \frac{\Delta R_{\text{CO}_2}}{\Delta I} \right).
\]

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 CO₂, \( \Delta R / \Delta R_{\text{CO}_2} \); and the dependence of radiative forcing from atmospheric CO₂ on carbon emissions, \( \Delta R_{\text{CO}_2} / \Delta I \) (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 CO₂ on cumulative carbon emissions, \( \Delta R_{\text{CO}_2} / \Delta I \), as appearing in the original TCRE definition in (2), but is now...
Our framework de
non-CO2 greenhouse gases.

ecause of ocean theory and diagnos
tics of two Earth System Models, GENIE only including radiative forcing from atmospheric CO2 and IPSL-CM5A-LR including CO2 and non-CO2 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 CO2 on carbon emissions, \( \Delta R_{\text{CO2}} / \Delta I \).

The diagnostics of the TCRE and TCRE\(_{\text{all R}}\) 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\(_{\text{all R}}\) become similar to each other. However, on shorter timescales, the TCRE and TCRE\(_{\text{all R}}\) differ from each other due to the
effects of other forcing agents, such as aerosols and non-CO2 greenhouse gases.

### 2.2. Applying global heat and buffered carbon integral balances

Our framework defining the TCRE and TCRE\(_{\text{all R}}\) 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, \( \Delta R(t) \) \( (\text{W} \text{m}^{-2}) \), positive downward, driving a climate response, involving additional outgoing longwave radiation, \( \lambda \Delta T(t) \) \( (\text{W} \text{m}^{-2}) \), from the increase in global-mean surface air temperature plus the net downward heat flux into the climate system, \( N(t) \) \( (\text{W} \text{m}^{-2}) \) (Gregory et al 2004, Gregory and Forster 2008)

\[
\Delta R(t) = \lambda \Delta T(t) + \varepsilon N(t),
\]

where \( \lambda \) is the equilibrium climate parameter \( (\text{W} \text{m}^{-2} \text{K}^{-1}) \) 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, \( \varepsilon \), 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

\[
\frac{\Delta T}{\Delta R} = \frac{1}{\lambda} \left(1 - \frac{\varepsilon N(t)}{\Delta R(t)} \right),
\]

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 loga
rithm of atmospheric CO2 to the sum of the cumulative carbon emission, \( \Delta I(t) \), plus the carbon undersaturation of the ocean, \( I_{\text{sat}}(t) \) \( (\text{PgC}) \), minus the increase in the terrestrial carbon uptake, \( \Delta I_{\text{ter}}(t) \) \( (\text{PgC}) \), all divided by the buffered carbon inventory, \( I_B \) \( (\text{PgC}) \) (Goodwin et al 2015):

\[
\Delta \ln CO_2(t) = \frac{1}{I_B} (\Delta I(t) + I_{\text{sat}}(t) - \Delta I_{\text{ter}}(t)),
\]

where \( CO_2 \) is the mixing ratio of atmospheric carbon dioxide \( (\text{ppm}) \), \( I_{\text{sat}}(t) \) is how much carbon the ocean needs to take up to reach an equilibrium with the atmosphere and \( \Delta I_{\text{ter}}(t) = I_{\text{ter}}(t) - I_{\text{ter}}(t_0) \) 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 CO2 enhances ocean acidity and inhibits the ability of the ocean to take up more atmospheric CO2 leading to an increasing fraction of the emitted carbon remaining in the atmosphere; thus, (8) expresses how atmospheric CO2 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 CO2, \( \Delta R_{\text{CO2}}(t) \), defined by the logarithm of atmospheric CO2 (Myhre et al 1998)

\[
\Delta R_{\text{CO2}}(t) = a \Delta \ln CO_2(t),
\]

where the logarithm conveys how there is a saturating effect of increasing atmospheric CO2 on radiative forcing.

Combining the buffered carbon balance (8) with the definition for radiative forcing from atmospheric CO2 (9) then provides an expression for the dependence of radiative forcing from atmospheric CO2 on carbon emissions (Goodwin et al 2015)

\[
\frac{\partial R_{\text{CO2}}}{\partial I} = \frac{a}{I_B} \left(1 + \frac{I_{\text{sat}}(t) - \Delta I_{\text{ter}}(t)}{\Delta I(t)} \right),
\]

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

\[
\text{TCRE}_{\text{all R}} = \frac{a}{\lambda I_B} \left(1 - \frac{\varepsilon N(t)}{\Delta R(t)} \right) \times \left(1 + \frac{I_{\text{sat}}(t) - \Delta I_{\text{ter}}(t)}{\Delta I(t)} \right) \left(\frac{\Delta R(t)}{\Delta R_{\text{CO2}}(t)} \right),
\]

The \text{TCRE}_{\text{all R}} \text{ } \text{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
Atmospheric carbon inventory at 2013

Implied TCRE, Effective global ocean buffer factor

Temperature, salinity, alkalinity

atmosphere and ocean

forcing is only from atmospheric CO₂, the TCRE

2.3. Equilibrium response for the TCRE

For a long-term equilibrium when the radiative forcing is only from atmospheric CO₂, the TCRE

asymptotes to a response given by the time-independent factors \( a/\lambda I_h \) (Williams et al 2012)

\[
\frac{\Delta T}{\Delta I_{t \to \text{equil}}} = \frac{a}{\lambda I_h}.
\]

To evaluate the equilibrium response, \( a \ (\text{W m}^{-2}) \) is the coefficient for the radiative forcing dependence on atmospheric CO₂ (Myhre et al 1998, Forster et al 2013); \( \lambda \ (\text{W m}^{-2} \text{K}^{-1}) \) is the climate parameter (Gregory et al 2004) and equivalently \( \lambda^2 \) is the equilibrium climate sensitivity (Knutti and Hegerl 2008); and \( I_h \ (\text{PgC}) \) is the buffered carbon inventory for the atmosphere and ocean (Goodwin et al 2007).

The buffered carbon inventory of the atmosphere and ocean, \( I_h \), 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; \( I_h \) is defined for the pre-industrial era (Goodwin et al 2007) as

\[
I_h = I_A + V \frac{\text{DIC}_{\text{sat}}}{B},
\]

where \( I_A \) is the atmospheric inventory of carbon (PgC), \( \text{DIC}_{\text{sat}} \) is the saturated ocean dissolved inorganic carbon (g C m\(^{-2}\)) 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 CO₂, relative to the fractional changes in the saturated dissolved inorganic carbon \( \text{DIC}_{\text{sat}} \), where \( B = (\delta \text{CO}_2 / \text{CO}_2) / (\delta \text{DIC}_{\text{sat}} / \text{DIC}_{\text{sat}}) \); and \( \text{DIC}_{\text{sat}} \) are evaluated using CO₂ 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 CO2 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 CO2 concentrations, following historical CO2 values until 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° × 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 CO2 concentrations, following historical CO2 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 CO2 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 CO2 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 CO2 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

\[
\text{TCRE} = \frac{\Delta T}{\Delta T} = \frac{a}{\lambda_\text{fb}} \left( 1 - \frac{\varepsilon N(t)}{\Delta R(t)} \right) \left( 1 + \frac{I_{\text{sat}}(t)}{\Delta I(t)} \right).
\]

Within GENIE, the equilibrium TCRE from \(a/(\lambda_\text{fb})\) is given by 1.35 K/(1000 PgC)−1, where \(a = 5.35 \text{ W m}^{-2}, \lambda = 1.28 \text{ W m}^{-2} \text{ K}^{-1}\) and \(I_{\text{fb}} = 3340 \text{ PgC}\) (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 CO2 on carbon emissions, \(\Delta R_{\text{CO2}}/\Delta I\), varies with the normalised ocean undersaturation in carbon, \(I_{\text{sat}}(t)/\Delta I(t)\).

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 CO2 on carbon emissions, \(\Delta R_{\text{CO2}}/\Delta I\), decreases in time (figure 2(b)). This decrease in \(\Delta R_{\text{CO2}}/\Delta I\) is due to the ocean taking up carbon, so that the ocean carbon undersaturation decreases. The increase in radiative forcing from atmospheric CO2 then becomes smaller for additional increases in emitted carbon with a smaller change in the logarithm of CO2 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 \(\Delta R_{\text{CO2}}/\Delta I\) 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-CO2 radiative forcing, including changes from anthropogenic aerosols, non-CO2 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 \(a/(\lambda_\text{fb})\), each of these input parameters differ slightly from those for GENIE. The buffered carbon
I is diagnosed as 3440 PgC, similar to the estimate of 3340 PgC for GENIE (table 1). The climate parameter is diagnosed as $0.77 \text{ W m}^{-2} \text{ K}^{-1}$ 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, $\Delta T$ (K) (figure 3(a), table 2). Our estimate of the climate parameter, $\lambda$, 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 $\times$ CO$_2$ increase, respectively, from Andrews et al (2012), Dufresne et al (2013) and Geoffroy et al (2013). The climate parameter, $\lambda$, 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 $\lambda$ 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 $\times$ 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 $a$, $\lambda$ 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).
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, $I_{\text{Usat}}(t)$, is diagnosed from the difference in the saturated DIC anomaly and the actual DIC anomaly over the global ocean, $I_{\text{Usat}}(t) = V (\Delta \text{DIC}_{\text{sat}}(t) - \Delta \text{DIC}(t))$.

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 $\lambda$ 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, $\Delta R = \varepsilon N (\text{W} \cdot \text{m}^{-2})$ and surface temperature change $\Delta T$ (K) for an efficacy $\varepsilon$ of 1.14 (Geoffroy et al. 2013). The average value of $\lambda$ for all RCPs is 0.77; all correlation coefficients are significant ($r > 0.26$ for a 99% confidence limit).

| Emission pathway | $\lambda$ (W m$^{-2}$ K$^{-1}$) | $r$ |
|------------------|-------------------------------|----|
| RCP 2.6          | 1.02                          | 0.78 |
| RCP 4.5          | 0.76                          | 0.92 |
| RCP 6            | 0.68                          | 0.96 |
| RCP 8.5          | 0.63                          | 0.97 |

atmospheric CO$_2$, $\Delta R_{\text{CO}_2}(t)$ (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 smaller. The ocean carbon undersaturation, $I_{\text{USat}}(t)$ (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, $I_{\text{USat}}(t)/\Delta I(t)$, reduces in time and there is also a relatively small increase in the terrestrial store of carbon, $\Delta I_{\text{ter}}(t)$ (figure 3(c), green line). Thus, $(I_{\text{USat}}(t) - I_{\text{USat}}(0))/\Delta I(t)$ 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 dependences making up the TCREE$_{\text{all} \ R}$ than for the TCREE$_{\text{all} \ R}$; 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$_2$, $\Delta R/\Delta R_{\text{CO}_2}$ (figure 4(b)), and a larger decrease in the dependence of radiative forcing on carbon emissions, $\Delta R_{\text{CO}_2}/\Delta I$ (figure 4(c)). The resulting TCREE$_{\text{all} \ R}$ then only slightly decreases over the centennial timescale (figure 4(d)).

There are therefore decadal changes in the TCREE$_{\text{all} \ R}$ (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$_2$, $\Delta R/\Delta R_{\text{CO}_2}$ (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, $\Delta R_{\text{CO}_2}/\Delta I$, 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 TCREE$_{\text{all} \ R}$ highlights how there are broadly similar responses in the Earth System Models, GENIE and IPSL-CM5A-LR. The centennial trends of the TCREE$_{\text{all} \ R}$ 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 TCREE. 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 TCREE$_{\text{all} \ R}$ 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 TCREE$_{\text{all} \ R}$ are due to (i) a larger dependence of the surface warming on radiative forcing in IPSL-CM5A-LR than in GENIE, and (ii) 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 TCREE draws upon the fundamental link between the logarithm of atmospheric CO$_2$ and the definitions of the radiative forcing from atmospheric CO$_2$ (6) and a buffered carbon inventory (6) accounting for ocean carbonate chemistry.
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 CO₂, \( \Delta T / \Delta F \) (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 CO₂ and changes in the air-borne fraction of emitted carbon.

Secondly, the sensitivities of climate models have been understood in terms of empirical parameters, \( \alpha \), \( \beta \) and \( \gamma \), representing the sensitivity of temperature to atmospheric CO₂, and the sensitivity of carbon inventories to atmospheric CO₂ 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.

| 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 |

(a) Earth System model of Intermediate Complexity, GENIE

| Year | 2005 | 2100 | 2005 | 2100 | 2005 | 2100 |
|------|------|------|------|------|------|------|
| RCP 4.5 | 0.68 | 0.94 | 3.46 | 2.21 | 3.55 | 2.67 |
| RCP 6 | 0.67 | 0.91 | 3.46 | 2.11 | 3.55 | 2.56 |
| RCP 8.5 | 0.71 | 0.81 | 3.46 | 2.33 | 3.55 | 2.31 |

(b) Earth System Model IPSL-CM5A-LR

$\Delta T / \Delta R$ denotes the dependence of surface warming on radiative forcing, $R_{CO2}$ denotes the dependence of radiative forcing from atmospheric CO$_2$ on cumulative carbon emissions, and TCRE denotes the dependence of surface warming on cumulative carbon emissions.
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 CO2 drawing upon the buffered carbon inventory \((8)\) and the definition of radiative forcing from atmospheric CO2 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 \(\text{Friedlingstein et al. 2006}\).

### 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, \(\Delta T / \Delta R\), increases as ocean heat uptake declines; (b) the dependence of radiative forcing from atmospheric CO2 on carbon emissions, \(\Delta R_{CO2} / \Delta I\), declines as the ocean takes up more excess CO2 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) \(\Delta R_{CO2} / \Delta I\) decreasing in time through the contrasting effects of ocean heat and carbon uptake, respectively.

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 CO2 on carbon emissions, \(\Delta R_{CO2} / \Delta I\), declines as the ocean takes up more excess CO2 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) \(\Delta R_{CO2} / \Delta I\) decreasing in time through the contrasting effects of ocean heat and carbon uptake, respectively.

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 CO2 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\textsubscript{all} 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, $\Delta R_{CO_2} / \Delta I = \frac{a}{\lambda b} \left( 1 + \frac{I_{uat} (t)}{\Delta I (t)} - \frac{\Delta I_{ter} (t)}{\Delta I (t)} \right)$. The carbon undersaturation of the ocean, $I_{uat} (t)$, 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, $\Delta R_{CO_2} / \Delta I$:

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

This definition of the TCRE\textsubscript{all} 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$_2$:

$$ \text{TCRE}_{\text{all}} = \frac{a}{\lambda b} \left( 1 - \frac{\varepsilon N (t)}{\Delta R (t)} \right) \times \left( 1 + \frac{I_{uat} (t)}{\Delta I (t)} - \frac{\Delta I_{ter} (t)}{\Delta I (t)} \right) \left( \frac{\Delta R (t)}{\Delta R_{CO_2} (t)} \right). $$

Applying this framework, the near constancy of the TCRE on multi-decadal to centennial 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 primarily 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).

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