https://doi.org/10.5194/esd-2019-82
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Constraints on long term warming in a climate mitigation scenario

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Abstract. Cumulative emissions budgets and net-zero emission target dates are often used to frame climate negotiations (Frame et al., 2014; Millar et al., 2016; Van Vuuren et al., 2016; Rogelj et al., 2015b; Matthews et al., 2012). However, their utility for near-term policy decisions is confounded by an uncertainties in future negative emissions capacity (Fuss et al., 2014; Smith et al., 2016; Larkin et al., 2018; Anderson and Peters, 2016) and in long term Earth System response to climate forcers (Rugenstein et al., 2019; Knutti et al., 2017; Armour, 2017) which may impact the utility of an indefinite carbon budget if peak temperatures occur significantly after net zero emissions are achieved, the likelihood of which in a simple model is conditional on prior assumptions about the long term dynamics of the Earth System. Here we illustrate that the risks associated with near term positive emissions can be framed using a definite cumulative emissions budget with a 2040 time horizon, allowing the necessity and scope for negative emissions deployment later in the century to be better informed by observed warming over coming decades.

Introduction

The climate policy discussion has adopted some convenient frameworks which act as proxies for the drivers and consequences of climate change. For example, it is broadly assumed that climate risks scale with global mean temperature (O'Neill et al., 2017). International climate agreements have thus been framed in this context (United Nations, 2015), necessitating Earth system parameters which relate future emissions trajectories to temperatures. This relationship is often framed through the Transient Climate Response to cumulative carbon Emissions (TCRE - the ratio of the globally averaged transient surface temperature change per unit carbon dioxide emitted, Rogelj et al. (2019a); Allen et al. (2009); Millar et al. (2016); Matthews et al. (2009); Gillett et al. (2013)). This convenient relationship allows the direct translation of temperature targets into available carbon budgets and derives from the apparent near linear relationship between cumulative emissions and surface temperatures seen in many climate simulations (England et al., 2009). Under the TCRE framework, the emissions budget constraint is dependent on a single parameter and the range of TCRE values observed in Earth System Models (ESMs) can be used to infer model-based carbon budgets which are compatible with the 1.5 and 2 degree Celsius targets of the Paris agreement (Gillett et al., 2013).

Meanwhile, understanding of how the Earth System reaches equilibrium in response to climate forcing has advanced; recent studies have highlighted that existing 150 year simulations are insufficiently short to assess the Equilibrium Climate Sensitivity (ECS, the equilibrium response of surface temperatures to a doubling of carbon dioxide concentrations) of General Circulation

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Models, and assuming a single feedback parameter associated with Effective Climate Sensitivity (Gregory et al., 2004) can lead to a significant underestimation of long term response (Gregory and Andrews, 2016; Geoffroy et al., 2013; Senior and Mitchell, 2000; Winton et al., 2010; Armour et al., 2013; Li et al., 2013; Rose et al., 2014; Andrews et al., 2018).

What is less clear at present is whether these findings have any relevance for the use of TCRE in emissions policy decisions.. Though TCRE is relatively robust in transient scenarios in which emissions remain mostly positive (Zickfeld et al., 2012; Krasting et al., 2014; Herrington and Zickfeld, 2014), its robustness in complex models under large negative emissions is relatively unexplored (Boucher et al., 2012; Vichi et al., 2013; Cao and Caldeira, 2010). Although an experimental design to test the long term robustness of TCRE under zero or negative emissions (Jones et al., 2019) or the dynamics of equilibrium response to forcing (Rugenstein et al., 2019) have been proposed and would be highly valuable, Earth System Models have not generally performed this type of experiment to date. Though such experiments have been performed in simple climate (Ricke and Caldeira, 2014; Millar et al., 2017c) and some intermediate complexity models (Zickfeld and Herrington, 2015) where cumulative emissions-temperature proportionality has been observed, it rests to thoroughly test whether these findings arise due to oversimplified model structure or prior assumptions on model parameters.

A number of modifications to the cumulative emissions/ carbon budgeting framework have been proposed (Rogelj et al., 2019a) to allow for additional corrections for non-CO_2 forcings (Rogelj et al., 2015a), continued post-zero emissions temperature evolution (Jones et al., 2019) and unforeseen earth-system feedbacks or 'tipping-points' which change biosphere or climate feedbacks (Brook et al., 2013). An complementary framework proposes a policy framework focused on net zero emissions and associated peak warming (Rogelj et al., 2019b). However, these frameworks are most useful if the zero emissions commitment is a small and finite correction to the net carbon budget, which is only true if peak warming occurs within a small number of decades of net-zero emissions.

Aside from physical modeling uncertainties in the long term stability of the TCRE assumption, indefinite carbon budgeting in policy making requires the combination of the effects of near term emissions reductions (Knutti et al., 2016; Rogelj et al., 2016a; Eom et al., 2015) and long term carbon removal technology which is subject to large socioeconomic, technological and physical uncertainties (Fuss et al., 2014; Smith et al., 2016; Larkin et al., 2018).

Similarly, the framing of climate policy in terms of a net zero emissions target also combines decarbonization of infrastructure (of which some sectors are highly difficult (Bataille et al., 2018)) and mid-century negative emissions capacity. These two components are conceptually different - the former is at least partly a function of structural choices which are currently available, while the latter is conditional on deeply uncertain biophysical (Smith et al., 2016), technological (Lomax et al., 2015) and social (Anderson and Peters, 2016) factors.

Here, we consider long term emissions scenarios in a simple model informed by recent advances in understanding in the thermal response of the Earth system to climate forcing on a range of timescales (Armour et al., 2013; Geoffroy et al., 2013; Winton et al., 2010; Held et al., 2010; Proistosescu and Huybers, 2017; Rugenstein et al., 2016), and how prior assumptions on model parameters have an impact on the long term robustness of a cumulative carbon emissions budget and the possible commitment to long term negative emissions to maintain a stable climate. We discuss the plausibility of the Earth exhibiting

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hysteresis behavior of global mean temperature as a function of cumulative emissions and that peak warming may occur significantly after net zero emissions have been achieved.

Finally, we propose that a policy approach which relies primarily on indefinite carbon budgets is not useful in the light of large geophysical and socioeconomic uncertainties, and that more robust decisions can be made if near term mitigation priorities are decided independently from absolute commitments on long term negative emissions capacity, which can be revised later (Rogelj et al., 2019b). Furthermore, we show that global temperature evolution on the timescale of the mid 21st century would enable a better constraint on future negative emissions requirements for temperature stabilization.

1 Results

1.1 Can transient observations constrain model response?

We first consider to what degree historical observations can constrain the long term coupled carbon-climate evolution of the Earth System. To address this, we consider a two timescale thermal response model, with timescales of response representing the deep ocean and shallow ocean response (as in Proistosescu and Huybers (2017); Geoffroy et al. (2013)). This is coupled to a simple emissions driven pulse model (as in Myhre et al. (2013); Millar et al. (2017c); Smith et al. (2018), see additional material) in which each unit of emitted carbon dioxide is allocated to one of four pools with its own representative decay time. We then ask whether the physical parameters of this simple model can be constrained by historical transient information.

We consider a number of different constraint assumptions on model parameters and how they influence the range of future projections under different scenarios (Figure 1). If the model parameters are conditioned only on historical emissions and temperature (Figure 1(a,b)), transient warming under continued positive emissions is well constrained, such that temperatures follow the TCRE relationship under a high emission scenario (RCP8.5, Riahi et al. (2011)) emissions. However, the relationship is not robust under long term negative emissions in a decarbonization scenario (RCP2.6, Van Vuuren et al. (2011)) where some model variants in the posterior parameter distribution allow hysteresis in which temperatures continue to rise over the following centuries under a regime of net negative emissions.

Adding information on historical deep and shallow ocean heat content (Zanna et al., 2019) does not significantly constrain the system (Figure 1(a,c)). However, assuming addition information about long term equilibrium climate sensitivity is known from paleo-climate data (Royer et al., 2011), does provide constraint on the degree of possible hysteresis (Figure 1(d)) as does the assumption of a known Realized Warming Fraction (RWF, the fraction of present day warming relative to equilibrium warming associated with current forcing) which is a very strong constraint on TCRE-like behavior. This prior, used in Millar et al. (2017b) produces a model configuration in which a proportional relationship between cumulative emissions-temperature is robust during both positive and negative phases of the emissions scenario (Figure (Figure 1(e)).

This raises the question of the degree to which we are confident in our knowledge of the values of ECS and RWF. In Millar et al. (2017b), the RWF prior is derived from the observation that the Transient Climate Response (TCR, the warming at the time of CO_2 doubling in a transient simulation where CO_2 increases by 1 percent per year) and Effective Climate Sensitivity (EffCS) are correlated in the CMIP5 ensemble (Millar et al., 2015) (where EffCS is the estimation of equilibrium response through the

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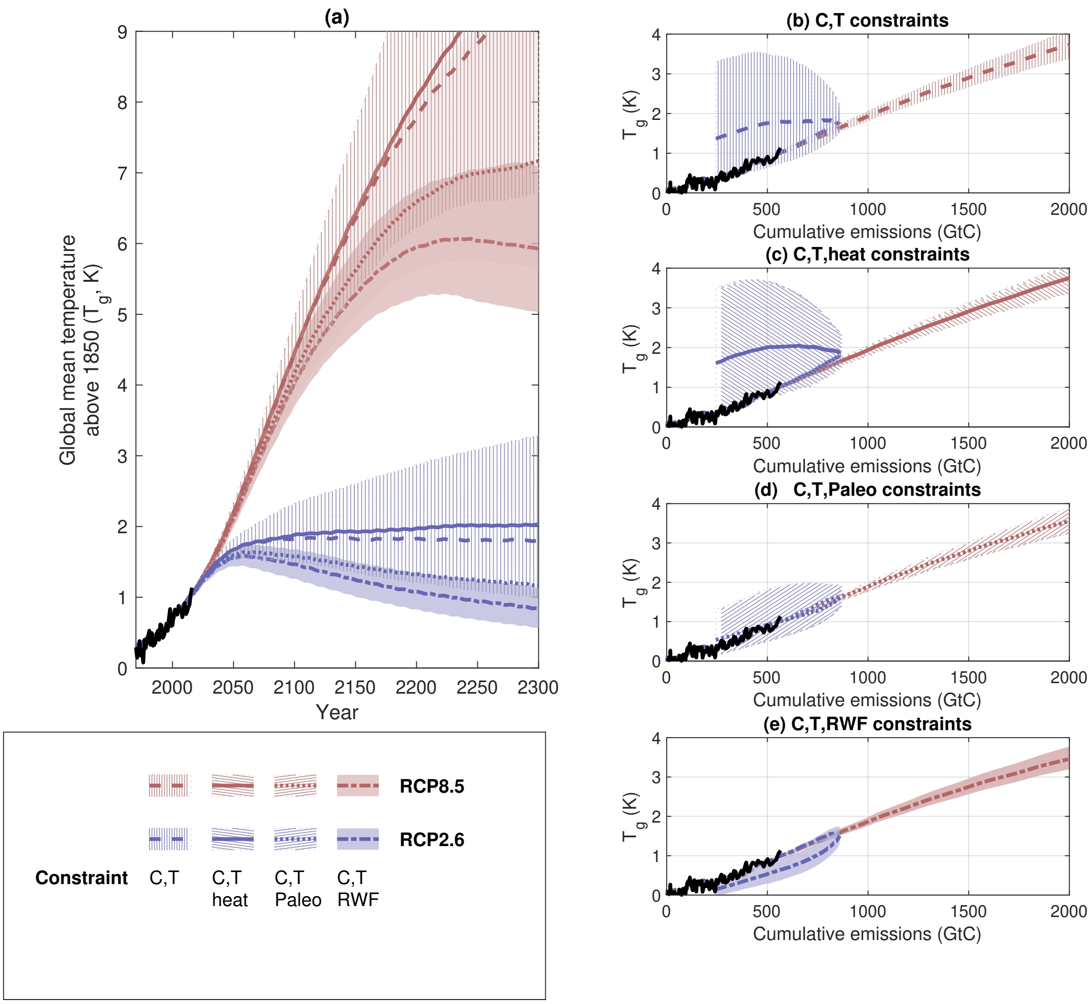

Figure 1. Posterior distributions of future global mean temperature projections constrained by 1850-2016 historical temperatures in a range of scenarios, priors and structural choices as a function of (a) time and (b-e) cumulative emissions of carbon (with 1000 years of climate evolution plotted from 1851-2850). Colored lines represent RCP8.5 (red) and RCP2.6 (blue). (b) and dashed lines in (a) show 2-timescale model posterior constrained using emissions (C) and temperature (T) only, (c) and solid lines in (a) are constrained using C,T and ocean heat content (H), (d) and dot-dash lines in (a) use C,T and RWF. (e) and dotted lines C,T and a paleoclimate prior on ECS. Shaded regions indicate the 10-90th percentile range. Solid black lines show observed HadCRUT values and historical emissions.

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linear extrapolation of temperature change as a function of net top of atmosphere radiative imbalance in an instantaneous CO_2 quadrupling experiment (Gregory et al., 2004)).

However, the Equilibrium Climate Sensitivity (ECS), realized over a multi-century to millennial timescale, is often significantly greater than the Effective Climate Sensitivity (Rugenstein et al., 2016; Knutti et al., 2017) and its value may not be well constrained by observed warming (Proistosescu and Huybers, 2017; Andrews et al., 2018). As such, and it is not apparent that the long-term ECS in a model like Myhre et al. (2013) can be constrained by TCR.

These prior assumptions strongly impact the range of possible behavior under strong negative emissions in RCP2.6. However, under RCP8.5, the ensembles constrained by historical temperatures show a near-linear relationship between cumulative emissions and temperature, irrespective of prior assumptions and constraints used (Figure 1(b-e), red lines), this can be broadly understood by considering that in RCP8.5, radiative forcing continues to increase at current rates and thus long term warming is broadly a function of TCR, which is itself constrained by historical temperature evolution.

1.2 Implications for meeting Paris temperature targets

If we consider a 'high risk' world where ECS (and its relationship to TCR) is not independently constrained, corresponding to subplot (b) in Figure 1, the cumulative emissions framework is not guaranteed to hold under negative emissions, and the concept of an indefinite cumulative carbon budget associated with a temperature target may not be helpful for near-term carbon mitigation planning (results for other prior assumptions are shown in the additional material).

We illustrate this in some idealized cases where 1.5 and 2 degree C climates are achieved post 2100. Scenarios are conducted in 3 phases: before 2020 is the 'historical' period, where emissions follow RCP2.6 (which is broadly consistent with observations before 2020). Between 2020 and 2040, the 'uninformed' period, CO_2 emissions follow one of a range of linear mitigation pathways such that 2040 CO_2 emissions are chosen at random for each scenario, ranging from $0GtC/yr$ to $10GtC/yr$ (our focus here is on low emission futures, and we do not consider here futures where emissions increase post-2020).

Post 2040, in the 'adaptive' period, an emission scenario is calculated iteratively to achieve temperature stabilization at a defined target post-2100, allowing for a temperature overshoot before 2100 with a large but finite lower limit on net negative emissions capacity in line with the largest negative emissions values seen in the integrated assessment literature for 1.5 degree temperature stabilization targets ($-20GtC/yr$, First (2018)). Non-CO_2 gas emissions follow RCP2.6 throughout the simulation in all cases (as such these scenarios cannot be treated as socioeconomically plausible scenarios, rather as idealized illustrations of Earth System Response to a range of forcing pathways).

The temperature trajectories are illustrated in Figure 2(a). Each member of the posterior distribution of possible simple climate models in Figure 1(a,b) is then paired with a random 2020-2050 emissions reduction pathway and then a post-2050 emissions pathway is calculated to optimize for stabilization at 1.5 or 2 degrees post-2100. This framework allows us to consider what would be required for long term stabilization in a model configuration where the cumulative emissions-temperature relationship does not necessarily hold.

The resulting scenarios are idealized, some requiring a very rapid switch to large net-negative values after 2040 in order to stabilize temperatures at 1.5C (Figure 2(b)), and such rapid decarbonization may not be achievable in reality (Sanderson et al.,

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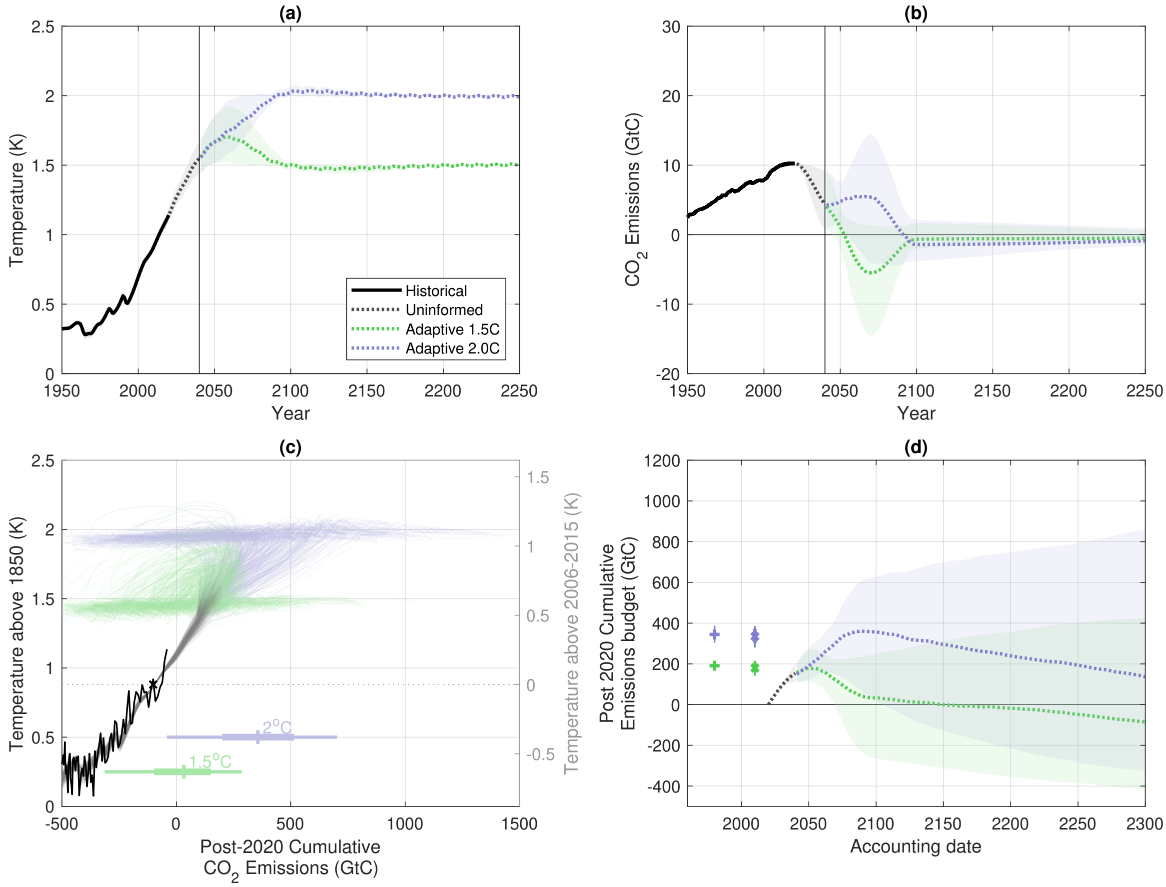

Figure 2. Plots showing idealized pathways to 1.5 or 2.0C temperature stabilization for an ensemble of coupled carbon-climate model configurations. (a) shows the global mean temperature as a function of time for 1.5 and 2.0C stabilization ensembles (b) shows emissions in the historical, uniformed and adaptive stages of the simulation (c) shows the global mean temperatures above pre-industrial/2006-2015 (left/right axis) levels as a function of post-2010 cumulative CO_2 emissions while (d) shows the cumulative carbon emissions total for ensemble members as a function of time. Shaded regions in (a,b,d) indicate 10th-90th percentile range of the ensemble distribution, while dotted lines shown the 50th percentile. Gray/blue/black areas refer to uninformed/adaptive for 2.0C/adaptive for 1.5C respectively. Box/whisker plots in (c) show the long term cumulative carbon budget assessed in 2100 for 1.5 and 2.0C stabilization from 1850-2500. Box/whisker plots in (d) show the TCRE estimate of carbon budget with (median shown by '+') and without (median shown by 'x') non-CO2 gas correction.

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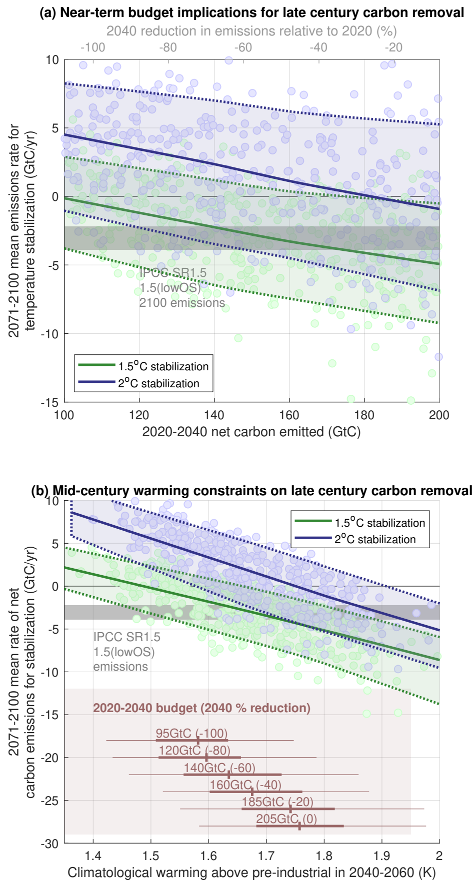

Figure 3. Plots showing (a) the relationship between mid-century cumulative carbon budgets and (b) mid-century warming and associated likelihoods of long term carbon removal requirements for temperature stabilization. (a) shows the ensemble relationship between the net carbon emitted between 2020 and 2040 (uninformed period in Figure 1) and the associated range of possible carbon removal required later in the century in the adaptive phase for 1.5C (green) and 2.0C (blue) stabilization. Filled circles represent an individual ensemble member, while shaded blue/green areas represent a moving estimate of the 10-90th percentile range of the 2.0C/1.5C distribution (solid blue/green lines are 2.0/1.5C median. (b) shows 2050-2100 allowable carbon budget as a function of 2050 warming above pre-industrial levels. Dots and shading show ensemble distribution as in (a). Horizontal box/whisker plots show 10th,25th,50th,75th and 90th percentiles of 2050 warming consistent with labeled 2020-2040 carbon budgets and the associated percentage reduction in 2040 emissions relative to 2020. Gray bar shows the range of reference 2100 net carbon budgets considered for end of century 1.5 degree overshoot scenarios in the IPCC spacial report on 1.5 degrees (First, 2018)).

2016), but we can learn some useful properties of the system response by studying the relationships between near term and long term emissions commitments.

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The range of long-term trajectories for temperature stabilization is diverse (Figure 2(c)), allowing for large positive or negative fluxes over the following centuries in some cases as global mean temperatures remain stable (by construction). This implies that although in nearly all cases, temperatures have stabilized by 2100 (Figure 2(a)), the cumulative carbon budget plume allows for a 1.5C(2.0C) budget of -250 to 200GtC (75 to 650GtC) by 2100, a budget which continues to grow more uncertain over the centuries which follow (Figure 2(d)). This is in contrast to the indefinite cumulative carbon budget for 1.5 or 2 degrees calculated from assumed TCRE - which is relatively tightly constrained as 160-200GtC (300-380GtC) for 1.5C (2.0C) stabilization after corrections for present day warming due to non-CO2 gases (Figure 2(d)).

This large uncertainty in the face of long term stabilization scenarios draws into question the utility of an indefinite carbon budget, hence we can consider to what degree we can constrain future response using a definite budget with a 2020-2050 timeframe (Figure 3). Firstly, even in the face of response timescale uncertainty, there is a linear relationship between 2020-2040 budgets and associated late century carbon removal rates required for stabilization (Figure 3(a)).

For example, if a late century net carbon emission of -2.9 GtC/yr is assumed for late century (corresponding to the central estimate of 1.5 degree, low overshoot stabilization from the IPCC Special Report on 1.5C warming (First, 2018), a 50 percent chance of 1.5 degrees requires a 2020-2040 budget of 150GtC, which would require a 60 percent cut in emissions from present day levels by 2040. A 75 percent chance of meeting the target would require a 2020-2040 budget of 100GtC - requiring just over 100 percent cut in carbon emissions by 2040.

Here again, the choice of prior constraint on model parameters has an important effect. If the Paleoclimate(or RWF) constraint on ECS is used as in Figure 1(e, or d), a 75 percent chance of 1.5 degrees given the -2.9GtC/yr late century removal rate would allow a 160GtC(or 220GtC) budget from 2020-2040 (see Additional Material Figure S11(c,or d).

However, in all cases, by mid 21st century, observed warming will provide a good indication of the degree of negative emissions required for stabilization - as the average realized warming in 2040-2060 provides quite a strong constraint on budgets for the latter half of the century (Figure 3(b)). The degree of possible mid-century warming can be reduced by minimizing the 2020-2040 carbon budget, but there still exists uncertainty due to the degree of thermal inertia in the system as greenhouse gas concentrations stabilize.

The strong relationship between mid-century warming and late century carbon removal requirements for 1.5 or 2.0C stabilization occurs because 2040-2060 warming can be potentially decreased either by fortuity (with a small value of real-world equilibrium climate sensitivity) or by action (by minimizing near-term emissions), both of which reduce late century net carbon removal requirements. Conversely, high climate sensitivity or slow decarbonization would both result in greater mid-century warming and greater necessity for negative emissions deployment.

2 Discussion

Recent climate policy discussions have been framed in the context of a carbon budget, an allowable net total of cumulative emissions which are consistent with a desired limit on planetary warming (Allen et al., 2009; Millar et al., 2016). Nuances in the estimation of this budget have been noted relating to bias correction of existing models (Millar et al., 2017a), the

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compensation for the effects of non-CO2 anthropogenic emissions (Rogelj et al., 2015a) and the need for additional carbon fluxes for temperature stabilization after net-zero emissions have been achieved (Rogelj et al., 2016b; Jones et al., in review).

But in the current framework, these factors are deemed to be corrections to the TCRE-computed carbon budgets (Rogelj et al., 2019a), and value of TCRE informed by a combination of model response historical records of global surface temperatures (Gillett et al., 2013; Steinacher and Joos, 2016) form the basis for published model estimates on carbon budgets for temperature stabilization (Matthews et al., 2017a, a).

Limitations in the applicability of the TCRE relationship due to the response timescales of the Earth System have been noted before (Rogelj et al., 2019a) in terms of the discrepancy between "Threshold Avoidance Budgets" and "Threshold Exceedance Budgets" (Rogelj et al., 2016b) which differ due to the lag of peak temperatures after net-zero emissions have been achieved as slower timescale components of the system equilibrate. But, the scale of these effects is generally assumed to be small - on the order of 1-2 decades (Ricke and Caldeira, 2014; Zickfeld and Herrington, 2015)

However, as we have seen, models can be constrained to follow TCRE-like behaviour. The parameters of the FAIR (Millar et al., 2017c; Smith et al., 2018) simple climate model, for example, are constrained using a prior on RWF. This constraint arises due to an observed relationship between *Effective* Climate Sensitivity and TCR, and is thus likely overly constraining on possible model behavior consistent with state of art GCMs (see Additional Material section S1).

Similarly, in the MAGICC model (Meinshausen et al., 2011a), non-stationary feedbacks are represented in two ways - using an allowance for an oceanic surface and and land surface feedback strengths, as well as having forcing dependent feedback strengths. However, ECS values calculated using MAGICC when calibrated as an emulator of CMIP GCM simulations remain very close to the Effective Climate Sensitivities of the target model (Meinshausen et al., 2011a) - even though in some cases we know that the true ECS realized in millennial time-frames is significantly greater than the EffCS value (Rugenstein et al., 2019). This requires further research, but is possibly explained by the consensus that multiple feedback timescales arise from warming patterns associated with shallow and deep ocean warming (Li et al., 2013; Geoffroy et al., 2013). Representing feedbacks as a function of the warming of the ocean *surface* warming is therefore a strong structural assumption which may not capture this effect.

Indeed, recent work has made clear that the long timescale response of the Earth system is not well constrained by past observations (Proistosescu and Huybers, 2017; Andrews et al., 2018), drawing into question whether recent transient warming is able to constrain Equilibrium Climate Sensitivity (Otto et al., 2013) or the Realized Warming Fraction (Millar et al., 2015). In the absence of these constraints, we cannot rule out without additional data that the slow timescale response of the Earth System associated with deep ocean warming may lead to a world which exhibits a (relatively) low TCR but a high ECS realized over centuries or millennia (Rugenstein et al., 2019) which, as we show here, may complicate the use of an indefinite carbon budget for temperature targets.

Other sources of information which may yet resolve the uncertainty. Independent information to constrain ECS from paleoclimate (Royer et al., 2011) or process understanding (Sherwood et al., 2014; Zhai et al., 2015; Tian, 2015; Tan et al., 2016; Cox et al., 2018) may help constrain the potential for temperature hysteresis. But many constraints to date have considered only *effective* climate sensitivity (Gregory et al., 2004) - whereas it is increasingly clear that both the timescale and amplitude

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of climate feedbacks need to be constrained in order to understand Earth System response to future forcing pathways (Armour et al., 2013). Such avenues could and should be explored further.

Clearly, the models used here are idealizations. Emission rates and rates of change are not constrained by technological or societal limitations, and only CO2 pathways are modified from the RCP2.6 scenario - and so results are only illustrative of how the Earth System might respond to different hypothetical pathways. Finding pathways for technology and policy which can actually achieve these pathways is a question for Integrated Assessment Models. However, the present standard approach of producing scenarios through forward-looking solvers (O'Neill et al., 2016) is unable to capture the risk highlighted here associated with actors who act today with imperfect knowledge about future technology and Earth System response.

The pulse response model of the type used here is also a simplification of global response, albeit a commonly used one (Joos et al., 2013) - which resolves the degrees of freedom in the range of responses exhibited in physical Earth System Models. However, the model only resolves the central tendency of the long term equilibration of the Earth System to a forcing change, without any estimate of climate variability. The real-world climatological temperature in 2040-2060 would be subject to internal variability (Kay et al., 2015), but such variation in annual mean temperatures is of the order 0.1-0.2C (Rogelj et al., 2017), and decadal average deviations from climatology of global mean temperature due to internal variability are of order 0.1C (Dai et al., 2015), which implies that by 2060, observed mid-century warming will be of some value in constraining negative emissions requirements later in the century which spans nearly 0.7C over the ensemble range (Figure 3(b)).

In summary, even in the presence of large uncertainty on long term response to emissions, near-term climate policy can be well posed through the use of a time-limited net carbon budget, or equivalently, a near-term commitment for a percentage reduction in emissions by a certain date (Sachs et al., 2016; Kaya et al., 2019). Such a framework allows near-term emissions reduction requirements to broadly be considered separately from the negative emission fluxes required for temperature stabilization, the feasibility of which remains deeply uncertain (Fuss et al., 2014; Anderson and Peters, 2016), and does not require waiting for peak warming to occur (Rogelj et al., 2019b) in order to inform the required scale of negative emissions capacity (especially in the theoretical case where peak warming occurs significantly after net zero emissions are reached).

Observed warming over the coming decades will provide additional information on our commitments to implement negative emissions infrastructure for temperature stabilization - commitments which may or may not prove feasible to realize. But a near-term budget would provide decision-makers with the tools to assess the risk of failure to meet temperature targets as a function of clearly defined targets for near-term decarbonization.

Data availability. CMIP5 and CMIP6 data are available through a distributed data archive developed and operated by the Earth System Grid Federation (ESGF).

Code and data availability. Code for this study is available on Github at https://github.com/benmsanderson/matlab_pulse

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Appendix: Methods

A1 Simple Climate Model Implementation

The temperature portion of the code allows for 2 representative temperatures, each with an equilibration timescale d_j (for 2 timescales, j following Myhre et al. (2013); Millar et al. (2017c)), producing a simple model for temperature and radiation response to a step change in forcing:

$$P(t) = F_{4xCO2} \sum_{n=1}^{3} q_n(1 - exp(-t/d_n))$$

$$R(t) = F_{4xCO2} \sum_{n=1}^{3} r_n(exp(-t/d_n)), \quad \text{(A1)}$$

where $P(t)$ is the annual global mean temperature and $R(t)$ is the net top-of atmosphere radiative imbalance (used only for the calculation of Effective Climate Sensitivity), and F_{4xCO2} is the instantaneous global mean radiative forcing associated with a quadrupling of CO_2, taken here to be $3.7Wm^{-2}$(Myhre et al., 2013).

Constraining thermal parameters from historical temperatures and emissions requires a consideration of both the carbon cycle as well as other relevant climate forcers. MCMC optimization of even a simple model of this form requires 10^7 or more calculations, so a very rapid model is required for computational tractability. This study employs a fast pulse-response model to represent the response of surface global mean surface temperatures to emissions. The model is implemented as a digital filter in MATLAB (see attached code) - allowing efficient computation and enabling Markov-Chain Monte Carlo parameter estimation for the physical parameters.

The carbon scheme is a simple pulse dissipation model, with four atmospheric carbon pools R_i (where $i = 0..3$, following Myhre et al. (2013)) with dissipation timescales τ_i as detailed in Table A1. Each unit pulse of emissions is allocated to each of the four pools with a fraction a_i:

$$\frac{dR_i}{dt} = a_i E(t) - \frac{R_i}{\tau_i}, \quad \text{(A2)}$$

for which the solution for a unit emissions pulse $\delta(t)$ can be written:

$$R_i(t) = a_i(1 - e^{-t/\tau_i}). \quad \text{(A3)}$$

A generic emissions time-series $E(t)$ can then be expressed as a sum of discrete pulses, allowing the corresponding carbon pools $C_i(t)$ to be expressed as a sum of pulse-responses $R_i(t)$

$$C_i(t) = \int_0^t \frac{dE(t')}{dt} R_i(t - t')dt'. \quad \text{(A4)}$$

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Atmospheric CO_2 concentrations C are calculated as the sum of the four pools $C(t) = C_0 + \sum_i C_i$(t), and are converted into a radiative forcing estimate assuming the standard logarithmic relationship:

$$F(t) = 5.4ln\left(\frac{C(t)}{C_0}\right) + f_r F_{ext}(t) \tag{A5}$$

following Myhre et al. (2013), and all other forcings (aerosols, and non-CO2 greenhouse gases) are combined into a single term $F_{ext}(t)$ using global mean RCP values from Meinshausen et al. (2011b). Uncertainty in the amplitude of non-CO_2 forcings is simply represented simply by an uncertainty factor f_r, which is also optimized in the course of the MCMC calibration (Table A1). The thermal response is calculated by expressing the derivative of the forcing timeseries $F(t)$ as a series of step functions and using the CO2 quadrupling response T_p from equation A1 to calculate the integrated thermal response.

$$T(t) = \int_0^t \frac{\frac{dF}{dt}(t')}{F_{4xCO2}} T_p(t-t')dt'. \tag{A6}$$

This is again performed in a computationally efficient manner using MATLAB's 'filter' function.

A1.1 Model Optimization

The earth system configuration of the pulse model has time-series inputs emissions of CO_2, along with radiative estimates from Meinshausen et al. (2011b) of non-CO2 forcing agents. We optimize the thermal model parameters for 2 timescales $[\mathbf{q},\mathbf{d},\mathbf{r}]$, the carbon dissipation parameters $[\mathbf{a},\tau]$ for 4 pools and the non-CO2 forcing factor f_r. Optimization, as for the 4xCO2 case is conducted with the Goodman and Weare (2010) MCMC implementation, using flat initial parameter distributions as shown in Table A1, 200 walkers and 50,000 iterations for each optimization. Cost functions are computed for global mean temperature (T), global CO_2 concentrations (C), Shallow Ocean Heat Content (H) and Deep Ocean Heat Content (D):

$$E_T = \sum_t \left(\frac{(T(t) - T_{GCM}(t))}{\sqrt{2}\sigma_T}\right)^2$$
$$E_C = \sum_t \left(\frac{(C(t) - C_{GCM}(t))}{\sqrt{2}\sigma_C}\right)^2, E_H = \sum_t \left(\frac{(H(t) - H_{GCM}(t))}{\sqrt{2}\sigma_H}\right)^2, E_D = \sum_t \left(\frac{(D(t) - D_{GCM}(t))}{\sqrt{2}\sigma_D}\right)^2, \tag{A7}$$

where σ_T is defined as for the abrupt-CO2 case as the standard deviation of HadCRUT 1850-1950 values. For σ_C, we lack an unforced standard deviation estimate - so a normalization constant of $\sigma_C = 0.3ppm$ was chosen empirically to produce a ±1 ppmv range in 2016 observed concentrations in the posterior distribution. Shallow and Deep Ocean heat is taken as the 0-300m and 300m+ heat content respectively in Zanna et al. (2019), with σ_H and σ_D taken as 1850-1950 standard deviations from the same dataset.

In the 'C, T constraint' case, optimization is conducted using $-E_T$ and $-E_C$ as log likelihoods in the MCMC optimizer, with parameter boundaries as listed in Table A1. The 'C, T, Heat constraint' case uses the sum of $-E_T$, $-E_C$, E_D and $-E_H$ cost functions. The 'C,T, paleo' case is implemented using the likely value and upper bound on Earth System Sensitivity from Goodman and Weare (2010) fit the median and 90th percentile of a gamma distribution for equilibrium. The 'C,T, RWF'

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constraint is implemented using a log-normal prior on Transient Climate Response with 5–95 percentiles of 1.0–2.5 K as in Millar et al. (2017c), and a Gaussian prior on RWF (the ratio between LTE and TCR) with mean 0.6, and 5th and 9th percentiles of 0.45 and 0.75.

A1.2 Adaptive scenario design

We propose an ensemble of simulations which achieve post-2100 stabilization at the 1.5 and 2.0C levels referred to in the Paris Agreement (United Nations, 2015). Each ensemble member uses a single parameter set draw from the posterior distribution of models calculated during the MCMC constraint of model parameter space in Section A1.1. Emissions follow RCP2.6 from 1850 until 2020, after which CO_2 emissions are by a 'pchip' spline which is fixed at a number of points, the first of which are 2010 and 2020 RCP2.6 emissions - ensuring a smooth transition from the RCP time-series to the post-2020 timeseries. An 'uninformed' emissions trajectory takes place from 2020 to 2040, where emissions evolve from RCP2.6 2020 levels (10.26GtC/yr) to a 2040 emissions level drawn randomly from a uniform distribution with bounds at $0GtC/yr$ and $10GtC/yr$.

Post 2050, the emissions are defined by an 'adaptive' phase - with 3 time points (the first, tp_1 in the range 2060-2100, the second (tp_2) in the range 2101-2300 and the third tp_3 fixed at the end of the simulation in 2764. Each time point is associated with an emissions rate $Ep_{1,2,3}$ which are each weakly constrained to lie in the range -40 to +10 GtC/yr. Optimization uses MATLAB's fmincon algorithm to find optimal values of $tp_{1,2}$ and $Ep_{1,2,3}$, where the model is run iteratively for a given physical parameter set to find a solution which minimizes the RMSE from the desired annual mean global mean temperature timeseries target (1.5 or 2.0C, in this case) over the date range 2100-2500.

Author contributions. The author performed all analysis and writing for this project

Competing interests. The author declares no competing interests

Acknowledgements. This work is funded by the French National Research Agency, project number ANR-17-MPGA-0016. Benjamin Sanderson is an affiliate scientist with the National Center for Atmospheric Research, sponsored by the National Science Foundation.

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Long name	Symbol	Default	Min	Max
Geological re-absorption fraction	a_0	0.26	0.1	.3
Deep ocean invasion/equilibration fraction	a_1	0.14	0.1	.3
Biospheric uptake/ocean thermocline invasion fraction	a_2	0.22	0.1	.3
Rapid Biospheric uptake/ocean thermocline invasion fraction*	a_3	n/a	n/a	n/a
Geological re-absorption timescale $(years)$**	τ_0	10^6	10^6	10^6
Deep ocean invasion/equilibration timescale $(years)$	τ_1	200	200	1000
Biospheric uptake/ocean thermocline invasion timescale $(years)$	τ_2	40	40	100
Rapid biospheric uptake/ocean mixed-layer invasion timescale $(years)$	τ_3	1	1	10
Thermal equilibration of deep ocean Sensitivity (KWm^{-2})	q_1	0	0	10*
Thermal adjustment of upper ocean Sensitivity (KWm^{-2})	q_2	0	0	10
Thermal equilibration of deep ocean timescale $(years)$	d_1	239	80	3000
Thermal adjustment of upper ocean timescale $(years)$	d_2	30	1	40
Fraction of forcing in deep ocean response	r_1	0	0.33	0.5
Fraction of forcing in upper ocean response	r_2	0	0.33	0.5
Non-CO2 Forcing ratio	f_r	0.7	1	1.3

Table A1. A table showing default model parameter values and minimum and maximum values used in model optimization. *deep ocean thermal response is limited to zero for 2 timescale model. *a_3 is calculated as the $1 - \sum_{i=1:3}(a_i)$

. **following Millar et al. (2017c), deep ocean carbon uptake timescale is not included in the optimization (the timescale is effectively infinite: sufficiently longer than the scenarios considered here for the a_3 pool to not absorb significant carbon).

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