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How difficult is it to recover from dangerous levels of global warming?

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Abstract
Climate models provide compelling evidence that if greenhouse gas emissions continue at present rates, then key global temperature thresholds (such as the European Union limit of two degrees of warming since pre-industrial times) are very likely to be crossed in the next few decades. However, there is relatively little attention paid to whether, should a dangerous temperature level be exceeded, it is feasible for the global temperature to then return to safer levels in a usefully short time. We focus on the timescales needed to reduce atmospheric greenhouse gases and associated temperatures back below potentially dangerous thresholds, using a state-of-the-art general circulation model. This analysis is extended with a simple climate model to provide uncertainty bounds. We find that even for very large reductions in emissions, temperature reduction is likely to occur at a low rate. Policy-makers need to consider such very long recovery timescales implicit in the Earth system when formulating future emission pathways that have the potential to 'overshoot' particular atmospheric concentrations of greenhouse gases and, more importantly, related temperature levels that might be considered dangerous.

S Supplementary data are available from stacks.iop.org/ERL/4/014012

Keywords: climate overshoot, climate change, carbon cycle, greenhouse gas emissions, dangerous change, global warming

1. Introduction

Recent research on global climate impacts (IPCC 2007 and Warren 2006) has produced extensive catalogues of the potential impacts expected for different amounts of global warming. Whilst there is no consensus on the precise amount of warming that society ought to consider unacceptable, many scientists and policy-makers have suggested maximum warming targets. For instance, Hansen (2005) suggests that warming should be limited to less than 1 °C above year 2000 temperature, corresponding to around 1.7 °C above pre-industrial levels. The European Union (European Council 2007) has adopted a target of limiting warming to not more than 2 °C above pre-industrial levels. Several authors (e.g. Wigley *et al* 1996) have concluded that major greenhouse gas emission reductions in the near future will be needed in order to avoid realizing such temperature levels. Such cuts will be challenging, both economically and technologically—indeed to achieve climate stabilization, such emissions need to ultimately approach a very small fraction of current levels (e.g. Matthews and Caldeira 2008, House *et al* 2008). Further, the longer the delay in reducing emissions, the larger the cuts expected by future generations to avoid potentially unwelcome change (Kallbekken and Rive 2007). However, rather than declining, recent global emissions have continued to grow and at a significant rate (Raupach *et al* 2007).

This therefore raises the question of whether it is possible to temporarily cross potentially 'dangerous' thresholds of atmospheric greenhouse gas concentrations (notably carbon dioxide, CO_2) or temperature rise before returning quickly to lower safer levels in the future. Such an 'overshoot' policy might be deliberate, or may occur if society is unable to reduce emissions quickly enough to prevent a desired temperature target from being exceeded. The implication is that once at these high levels of change, society would then act to return to lower temperature levels back below this target (Huntingford and Lowe 2007).

Schneider and Mastrandrea (2005) were amongst the first authors to explicitly consider scenarios which involve prescribing a climate change trajectory that crosses a potentially dangerous threshold before reverting back to a lower safer level. They do this by prescribing 'overshoot' scenarios in radiative forcing (thus representing altered levels of atmospheric greenhouse gas concentrations), and then using a simple model to translate this into temperature change. However, they did not derive the associated emissions and the constraints that the natural elements of the Earth system places on reducing atmospheric CO_2 concentrations, and so this raises the issue of scenario feasibility. O'Neill and Oppenheimer (2004), using a simple climate model, suggested it would be possible to reduce atmospheric CO_2 concentrations by around 100 ppm (from maximum levels between 600 and 800 ppm) over approximately 60–80 years, corresponding to a fall in temperature of around 0.5 °C during this time. Recently, Wigley *et al* (2007) did examine a set of concentration overshoot pathways, again with a simple model, and derived both the future temperature rise and most importantly the greenhouse gas emissions that would lead to the given pathways (but without a treatment of climate and carbon cycle uncertainty). They demonstrate that recovery to lower temperatures within a century timescale is difficult. In fact, the overshoot scenario considered by Wigley *et al* (2007) may, for a period, require negative global emissions of CO_2. The linkage between concentration and temperature overshoot trajectories and emissions scenarios depends on key uncertainties of the climate system, especially, climate sensitivity and ocean heat uptake (Knutti *et al* 2005). The linkages also depend on the still uncertain climate–carbon cycle feedback, which has been investigated for 'business-as-usual' scenarios with large increases in emissions (e.g. Cox *et al* 2000, Friedlingstein *et al* 2006). However, there has been much less emphasis on studying this feedback in scenarios with large reductions in future emissions. Matthews and Caldeira (2008) did examine scenarios with large future emissions cuts using a single Earth system model of intermediate complexity (EMIC) and they found that near-zero emissions were required to stabilize atmospheric near-surface temperature. It can be inferred that if extremely low emissions are needed for climate stabilization, then even lower (or zero) emissions will probably correspond to only very slow rates of recovery in global temperature.

Working Group 1 of the IPCC fourth assessment report provides only limited coverage of this type of 'overshoot' scenario. For instance, Meehl *et al* (2007) reported an experiment by Tsutsui *et al* (2007) that prescribed reductions in atmospheric CO_2 concentrations by around 150 ppm over 100 years, and the associated derived temperature response gave a reduction of around 1 °C. While this used a complex climate model (the community climate system model), driving the model with prescribed CO_2 concentrations again raises the question of whether such rapid reductions are actually feasible in terms of emissions. Meehl *et al* (2007) also reported (in their figure 10.35) the results from five EMICs, and this same experiment set was analysed more fully later in Plattner *et al* (2008). These authors used a scenario with emissions set to zero at 2100, by which time the atmospheric CO_2 concentration had reached between 650 and 700 ppm. Immediately following this extreme emissions reduction, atmospheric CO_2 concentration fell by only 50–100 ppm over 100 years. Further, between 2100 and 2300 the global average surface temperature levelled out in two of the models and began decreasing slightly in the other three. Finally, Solomon *et al* (2009) also used an EMIC to derive long-term CO_2 and temperature response following emissions being set to zero. Their model also attempted to capture expected precipitation changes using a 'pattern scaling' method, although caution should be exercised as this methodology requires verification for long periods of slowly declining temperature.

The implications of this very slow rate of atmospheric CO_2 and temperature reduction for such a drastic emissions cut have considerable implications for the climate change debate. Unfortunately still missing in the numerical experiments performed to date that depict 'overshoot' scenarios is a coupled (interactive) treatment of the climate system and carbon cycle response using the sufficiently complex process representations only found in general circulation models (GCMs). Many also lack an up-to-date treatment of the combined uncertainties in the climate system and its interaction with the carbon cycle.

GCMs capture climate processes on both relatively fine spatial scales (typically a few hundred kilometres) and temporal scales, and remain the mainstay of projections of future climate change. EMICs do represent some, but not all, of the processes explicitly represented in GCMs although often at a lower resolution. Simple climate models, by definition, aggregate all such processes affecting evolution of the Earth system to much larger, often global average, scales. However these processes, such as atmospheric circulation, the hydrological cycle, oceanic functioning and all interactions within the carbon cycle do exhibit significant geographical variation in their perturbations within a changing climate. It is, therefore, prudent to check that spatially aggregated parameterizations in simpler models remain valid when forced by a range of anthropogenic emissions, especially for more novel situations such as 'overshoot' where potential (regional) hysteresis effects may be important. The penalty is that GCMs require a massive computational overhead, and despite advances in computational power, long century timescale simulations still need several months to complete. Hence, only a relatively few long GCM simulations are possible. It is in this context that simple models do remain useful in providing measures of uncertainty given their capability to rapidly make long simulations. However, this should only be done where the simple model has some demonstrated skill in replicating the simulations that exist by more complex GCMs.

In this letter we investigate the issue of recovery time in overshoot scenarios in much greater depth. The first part of our study uses a version of the Hadley Centre GCM to investigate, for the first time, the robustness of previous predictions by simple models and EMICs and explore the mechanisms/processes behind the very slow temperature recovery rates following dramatic reduction of emissions in future (highly idealized) CO_2 scenarios. The second part of our work uses a simple model to estimate uncertainty on our conclusions and also to extend them to a more policy relevant scenario, and where such a scenario also includes the impact of non-CO_2 greenhouse gases and aerosols.

2. Methods

The complex GCM with carbon cycle that we employ here is the HadCM3LC model. The simple climate model we use is a version of the MAGICC model, with parameters representing the aggregated behaviour of a range of more complex climate models (both EMIC and GCMs).

2.1. GCMs

The HadCM3LC model (Cox *et al* 2000) couples a version of the Hadley ocean–atmosphere GCM (HadCM3) (Gordon *et al* 2000, Pope *et al* 2000) to an ocean carbon cycle model (Palmer and Totterdell 2001) and a terrestrial carbon cycle model (Cox 2001). The terrestrial carbon cycle model TRIFFID simulates growth of, and competition between, five vegetation plant functional types: broadleaf trees, needleleaf trees, C_3 grasses, C_4 grasses and shrubs. Stomatal conductance and photosynthesis are calculated using a coupled leaf-level model (Cox *et al* 1998). The ocean carbon cycle model represents four biological components, a single class of phytoplankton, a single class of zooplankton, detritus and nutrient, along with dissolved inorganic carbon and alkalinity. As well as interacting within the biological model, each of these components describing the ocean carbon cycle are advected around the ocean by the modelled oceanic physics. The horizontal resolution of HadCM3LC is 2.5° latitude by 3.75° longitude with 19 levels in the atmosphere and 20 levels in the ocean. Because our focus is on changes over time periods of a few hundred years it is not necessary to include very long timescale features of the global carbon cycle, such as a treatment of rock weathering or carbon sedimentation (e.g. Archer 2005). Before making simulations corresponding to prescribed emission scenarios, HadCM3LC is spun up to achieve an initial stable state representative of pre-industrial climate conditions. Specifically the GCM is forced with prescribed pre-industrial levels of atmospheric CO_2 concentration, and until a situation is reached whereby net land–atmosphere and ocean–atmospheric fluxes of CO_2 are negligible.

2.2. Simple climate model

The MAGICC climate model has been used extensively to make climate projections, including use in the IPCC third and fourth assessments (Cubasch *et al* 2001, Meehl *et al* 2007). It has previously been tuned to credibly reproduce the global mean temperature results of seven atmosphere–ocean GCMs (AOGCMs) (Raper and Cubasch 1996, Raper *et al* 2001, Cubasch *et al* 2001). The carbon cycle (Wigley 1993) in MAGICC can emulate the Bern-CC and ISAM model results over the range of the SRES scenarios. In our analysis we use a version of the MAGICC simple climate model to simulate the global average near-surface warming and its uncertainty for each of our new emissions trajectories. The three specific parameters that we varied are the climate sensitivity (defined as the equilibrium global mean temperature increase for a doubling of atmospheric CO_2), the ocean mixing rate (which determines how quickly the warming at the surface is diffused throughout the ocean), and a new climate–carbon cycle feedback factor (which either amplifies or suppresses the temperature dependent climate–carbon cycle feedbacks already prescribed in version 4 of MAGICC). Warren *et al* (2009) show that there are particular combinations of these three parameters that enable MAGICC to closely emulate the atmospheric CO_2 change and global average surface warming of the 11 models (forced for SRES A2 emissions scenario) that took part in the C4MIP study (Friedlingstein *et al* 2006).

In our analysis, uncertainty was incorporated into our simulations by sampling parameter values from range estimates of the three parameters discussed above. For climate sensitivity we draw on a widely-used probability distribution of Murphy *et al* (2004), which has a modal value of approximately 3.2 °C. A log normal distribution of ocean mixing rates, with a modal value of 2.04 $cm^2\ s^{-1}$, was derived following Wigley and Raper (2001) based on a fit to the AOGCMs employed by Cubasch *et al* (2001). The climate–carbon cycle amplification parameter follows a normal distribution whose mean and standard deviation was based on the fitting of MAGICC to the C4MIP ensemble. The three parameter values were selected by dividing the uncertainty ranges into nine bins, each of which is associated with a probability from the distributions. A probability estimate for each possible triplet of parameters (729 sets of values) was then derived by multiplying together these individual probabilities. For our new ‘overshoot’ scenarios, all 729 possible simulations were performed for each scenario.

2.3. Experimental design

We used four scenarios in this work, all following historic emission estimates up to year 2000 followed by the SRES A2 emissions (Nakicenovic and Swart 2000) until at least 2010.

The first three scenarios comprised a set of highly idealized emission reduction experiments. In experiments named 2012E0, 2050E0 and 2100E0 the emissions followed SRES A2 values until being set to zero for the next 100 years at years 2012, 2050 or 2100. This set of emissions was supplied to both the complex model, HadCM3LC, and to the probabilistic version of the MAGICC model. While society is almost certainly not going to reduce CO_2 emissions to zero on such short timescales, the purpose of these experiments is to show the overall constraints on concentration and temperature reduction inherent in the climate system. For simplicity, non-CO_2 greenhouse gases and aerosol forcings were omitted from these three scenarios. The simulations with MAGICC were extended to year 2500 (with emissions remaining at zero) but computational constraints meant this was not possible with HadCM3LC. As both models overlap for at least part of the scenarios, this allows us to estimate whether the simple model has credibility in the zero emissions section of the scenario several decades after it deviates from the original scenario (i.e. SRES A2). This model equivalence for these new emission scenarios cannot be assumed as climate sensitivity estimates used in the simple model were derived from an experiment in which atmospheric CO_2 concentration was doubled, whilst the carbon cycle climate feedback amplification parameter was derived from model comparisons with 'business-as-usual' SRES A2 emissions.

Our fourth emissions scenario was less idealized and additionally included forcing from other greenhouse gases, the non-CO_2 Kyoto gases, the Montreal protocol gases and atmospheric pollutants including sulfate aerosol particles. This scenario was only used with the simple climate model. Called 2010P3, emissions deviated from SRES A2 at 2010, peaking five years later in 2015, and then reducing at an increasing rate until a long-term compound rate of reduction of 3% per annum is reached. This compound rate was then applied for the rest of the numerical experiment, which, for consistency with the other scenarios ends in 2500. The reduction in CO_2 equivalent emissions[5] in 2010P3 are around 47% of the 1990 value by 2050. Sulfate aerosol 'emissions' are linearly reduced to zero from the projected 2010 value, and over a period of 50 years. As a sensitivity experiment we have also performed variants of the 2010P3 experiment in which sulfate emissions were reduced over a shorter 25 year period or a longer 100 year period.

3. Results and discussion

3.1. General circulation model simulations

When the CO_2 component of the SRES A2 emissions scenario is used to force the HadCM3LC GCM for a period up to the end of the 21st century, then the atmospheric concentration of carbon dioxide is projected to exceed 1000 ppm at 2100. Following the emissions being set to zero in the 2012E0 experiment (at a CO_2 concentration around 404 ppm), the 2050E0 experiment (at a CO_2 concentration around 556 ppm) and the 2100E0 experiment, HadCM3LC simulates very low rates of decline in atmospheric CO_2 concentration. Mean (regressed) rates of change for the following hundred years are predicted as -0.2 ppm y^{-1}, -0.4 ppm y^{-1} and -0.75 ppm y^{-1}, respectively (see figure 1 upper panel). The magnitudes of these rates of reduction are all considerably smaller than that of the current rate of increase in atmospheric CO_2 concentrations (known to be around $+2$ ppm y^{-1}) and less than 12% of the magnitude of the HadCM3LC modelled 21st century average rate of increase for the SRES A2 simulation.

The associated projected future global average surface temperatures from the HadCM3LC experiments are shown in figure 1 (lower panel). By the time emissions are set to zero in the 2050E0 experiment, simulated temperature has already exceeded 2 °C above pre-industrial levels. Thereafter there is actually a slight trend of continued warming (around 0.2 °C/century) implying temperatures would remain at more than 2 °C for at least a century, and possibly much longer.

We analyse the simulated mechanisms in HadCM3LC of the removal of atmospheric CO_2 for the three century long model runs following cessation of emissions. In the 2012E0 experiment CO_2 was removed from the atmosphere over the subsequent century by a combination of uptake by the land and ocean in a ratio of approx 1 : 2 (23 GtC and 54 GtC respectively, as shown in figure 2). The relatively small degree of climate change that had occurred by this time meant that the terrestrial biosphere was still acting as a carbon sink (due to increased CO_2 fertilization of plant growth outweighing any temperature-increased plant or soil respiration). For the ocean, the increased atmospheric CO_2 concentration was driving modest absorption. However in 2050E0 and 2100E0 the removal of CO_2 from the atmosphere in the century after the emissions were zeroed was driven by a quite different balance of processes. The higher levels of CO_2 drove increasing uptake by the ocean, which more than offset any potential reduction due to increasing ocean temperatures (higher temperatures can

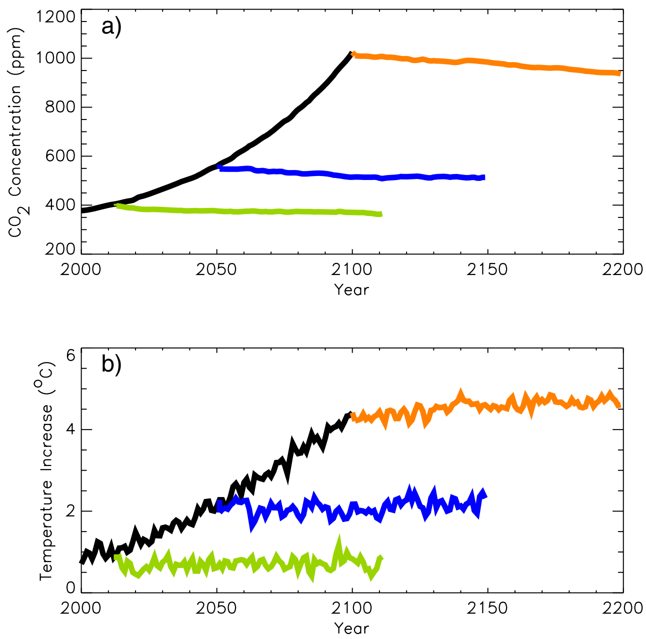

Figure 1. Projected atmospheric CO_2 concentration (upper panel) and temperature increase above pre-industrial levels (lower panel) simulated with the HadCM3LC model. The black curves show the SRES A2 (CO_2 only) emissions forcing reference case. The green, blue and orange curves show the mitigation experiments in which emissions were zeroed at years 2012 (scenario '2012E0'), 2050 ('2050E0') or 2100 ('2100E0').

[5] The CO_2 equivalent emissions were estimated for the combined emissions of CO_2, CH_4 and N_2O using global warming potentials of 1, 21 and 310, respectively.

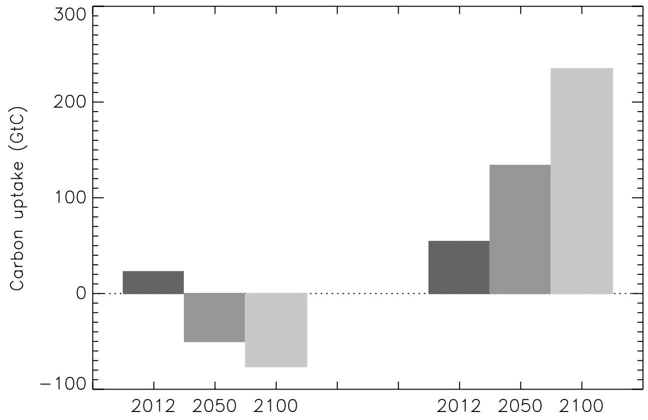

Figure 2. Cumulative natural carbon uptake split by land (left-hand bars) and ocean (right-hand bars) in the 100 years following cessation of emissions in the 2012E0, 2050E0 and 2100E0 experiments. Negative values imply a natural source of CO_2 in to the atmosphere.

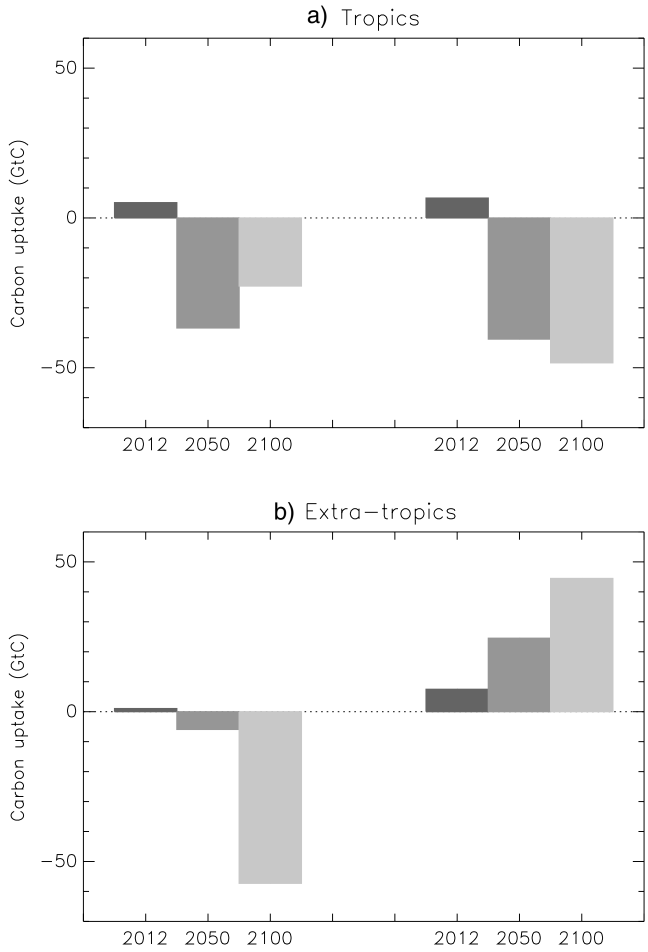

Figure 3. Cumulative carbon uptake by the land biosphere in the 100 years following cessation of emissions, split into soil carbon storage (left-hand bars) and increase in vegetation biomass (right-hand bars). Tropics (30°S–30°N; top panel), extra-tropics (outside 30°S–30°N; bottom panel). Negative values imply a natural source of CO_2 into the atmosphere.

reduce the solubility of carbon dioxide into the mixed layer and uptake due to mixing of carbon into the deeper ocean, reducing potential 'draw-down' of atmospheric CO_2 (Prentice *et al* 2001, Sarmiento *et al* 1998). However what is noteworthy for these two scenarios is that the warmer global temperature (accompanied by precipitation changes) meant that, averaged globally, the terrestrial biosphere was unable to act as a sink of carbon. There is a modelled net terrestrial loss of 50 GtC and 76 GtC respectively over the 100 years following a cessation of emissions.

More details of this terrestrial vegetation and soil carbon exchange during the zero emissions periods are shown in figure 3. In the 2012E0 case both the tropical and extra-tropical land regions continue to absorb carbon. In the 2050E0 case there is an extra-tropical sink due to enhanced boreal forest growth, but this is smaller than the tropical carbon loss from both vegetation biomass and soil organic material (respiration increase for this region is larger than any CO_2-fertilisation effect). In 2100E0 carbon loss from northern soils also now outweighs biomass gain and hence the extra-tropics as well as the tropics are predicted to be a carbon source.

A particular example of regional carbon cycle changes, which is present in both the 2050E0 and 2100E0 experiments is the simulated die-back of the Amazonian forest. This continues long after the emissions have been set to zero, and can be described as an 'ecosystem change commitment' (Jones *et al* 2009). The mechanisms for Amazon die-back in HadCM3LC are discussed in Betts *et al* (2004) and Cox *et al* (2004), and are found to be forced by movement in the Atlantic inter-tropical convergence zone (ITCZ) (Good *et al* 2009).

Here we have, for the first time using the most complex class of Earth system model, a coupled climate–carbon cycle GCM, demonstrated that only very low rates of temperature reduction follow even massive reductions in emissions. This is in agreement with the general conclusions from the work of, for example, Plattner *et al* (2008), Matthews and Caldeira (2008) and Solomon *et al* (2009). However, this could not have been assumed *a priori* because many detailed long-term carbon cycle changes and their meteorological forcings (as highlighted by the discussion of both the tropics/extra-tropics split in terrestrial carbon store behaviour, and projected Amazonian 'die-back') might not be present in EMICs or simple climate models.

We can make a direct comparison against one of the recent Solomon *et al* (2009) scenarios (the zeroing of emissions from 550 ppm) to our 2050E0 experiment (which zeroes emissions from a concentration of 556 ppm). In the former, the EMIC-derived CO_2 concentration falls by around 100 ppm in the 100 years following emission reductions, compared to a 40 ppm drop in our GCM simulation. During the same period the temperature in the Solomon *et al* (2009) experiment falls by around 0.1–0.2 °C but in our experiment it actually rises by 0.2 °C. However, whilst individual GCMs and EMICs allow us to investigate the processes that govern the slow climate response following a reduction in emissions, an ensemble of models is needed to span the range of plausible response times. In section 3.2 we attempt to provide uncertainty bounds on the time global average temperature is likely to remain above various target temperatures. We also extend the results to an emission scenario that has a higher level of likely policy realism.

3.2. Uncertainty analysis and policy relevant scenario

In this section we make use of large ensembles of MAGICC simulations for each emissions profile of interest. An initial task is to establish whether MAGICC has credibility at representing more complex models in scenarios with rapid reductions in emissions. The range of temperatures (relative to pre-industrial) during the period 2050–2150 projected by HadCM3LC for scenario 2050E0 is 1.65–2.5 °C. The spread is caused by natural variability in the GCM and the small upward trend in temperature. If we select the particular MAGICC version that has the best fit to HadCM3LC in the SRES A2 scenario we find this projects a temperature rise of 2.15 °C for 2100 for 2050E0 emissions. Because the MAGICC result lies within the bounds of the projections from HadCM3LC decades after the cut to zero emissions, we can have some confidence that the simple model can replicate the complex GCM in aggressive mitigation scenarios even though its parameters were derived from experiments that used very different emissions pathways.

Having established at a basic level that the MAGICC model has some predictive skill for scenarios with significant mitigation, we consider the probability estimates from our full ensemble of simple models as the warming evolves over time. The three cumulative distributions in figure 4 describe the probability of various amount of future warming in the 2050E0 scenario for years 2050, 2100 and 2200. By 2100 the median warming is around 1.8 °C, and the probability of exceeding 2 °C is around 40%. The probability of exceeding 3 °C by this time is a little less than 5%. A particularly interesting feature of this plot is that the cumulative distribution functions cross over each other. This implies that as time passes following the cut to zero emissions, the probability of exceeding the higher temperatures (>2.5 °C) actually increases as some model variants continue to warm. However, the probability of exceeding lower temperatures (<2 °C) becomes less as time progresses. This is because there are some simple model variants (typically those with lower climate sensitivity and faster response time) that have already reached their peak temperature and begun to cool. The cumulative distribution functions for the warming to 2100 for the 2050E0, 2100E0 and 2010P3 scenarios are compared in the supplementary material, available from stacks.iop.org/ERL/4/014012.

Because figure 4 shows that some of our simulations exhibit temperature peaking behaviour, we now focus on the possible length of time for which the global average temperature overshoots various temperature thresholds. These thresholds are selected as 1.7 °C (based on Hansen 2005), 2 °C (based on the EU target) and 3 °C above the pre-industrial levels. Using the ensemble of MAGICC simulations and associated probabilities for each run, the left-hand panels of figure 5 show the evolution over time of the probability of exceeding the various temperature targets. The results from the idealized CO_2-only scenarios (2050E0 and 2100E0) are shown alongside those from the more policy relevant multi-gas 2010P3 experiment. The right-hand panels of figure 5 shows the cumulative probability of being over various temperature thresholds for a given length of time. For the 2010P3 experiment there is around a 55% chance that the temperature

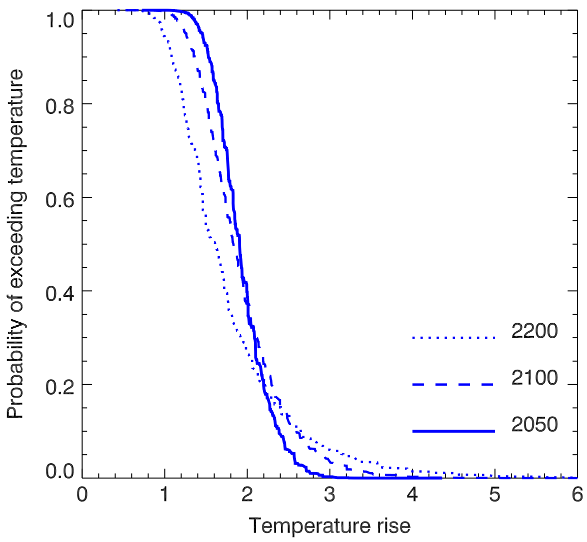

Figure 4. Cumulative probability of exceeding particular temperature levels at 2050, 2100 and 2200 in the 2050E0 scenario.

peaks above 2 °C, and around a 30% (20%) chance that it stays above 2 °C for at least 100 (200) years.

A further interesting feature of figure 5, seen most clearly in the time spent over 1.7 or 2 °C, is that whilst the probability of staying over this temperature target for less than a century is greater in the 2010P3 case than in the 2050E0 case, the reverse is true for overshoot times of more than a century. Initially, this seems counter intuitive because 2010P3 still has emissions beyond 2050, albeit at a diminishing rate. The main reason for this behaviour is that although the 2010P3 simulation has a higher radiative forcing earlier in the simulation (before 2050) this is due to it having a large non-CO_2 greenhouse gas composition. Following 2050, due to the shorter atmospheric lifetimes of these gases, radiative forcing in the 2010P3 falls more quickly despite these simulations having non-zero emissions beyond year 2050. A smaller secondary reason is that there are slight differences in carbon uptake due to different carbon cycle climate feedback in the different warming and fertilization pathways taken by the two experiments earlier in the simulations. The consequence of these combined effects is that the probability of being over the 2 °C target for more than 400 years is only around 6% for the multi-gas 2010P3 scenario but is almost three times bigger for 2050E0, despite 2010P3 having a greater probability of going over 2 °C earlier in the simulation.

3.3. Sensitivity of results to aerosol forcing scenario

One of our key assumptions in the 2010P3 scenario is the rate at which sulfate emissions are reduced following the start of mitigation. The short lifetime of sulfate aerosols in the atmosphere means that the radiative forcing associated with sulfates reduces to zero over approximately the same period of time as the linear decrease down to zero in sulfate emissions. Here we examine the sensitivity of our results to the rate of sulfate emission reduction by repeating 2010P3, but using aerosol reduction times of 25 and 100 years compared to the initial reduction time of 50 years (figure 6 and supplementary material, available from stacks.iop.org/ERL/4/014012).

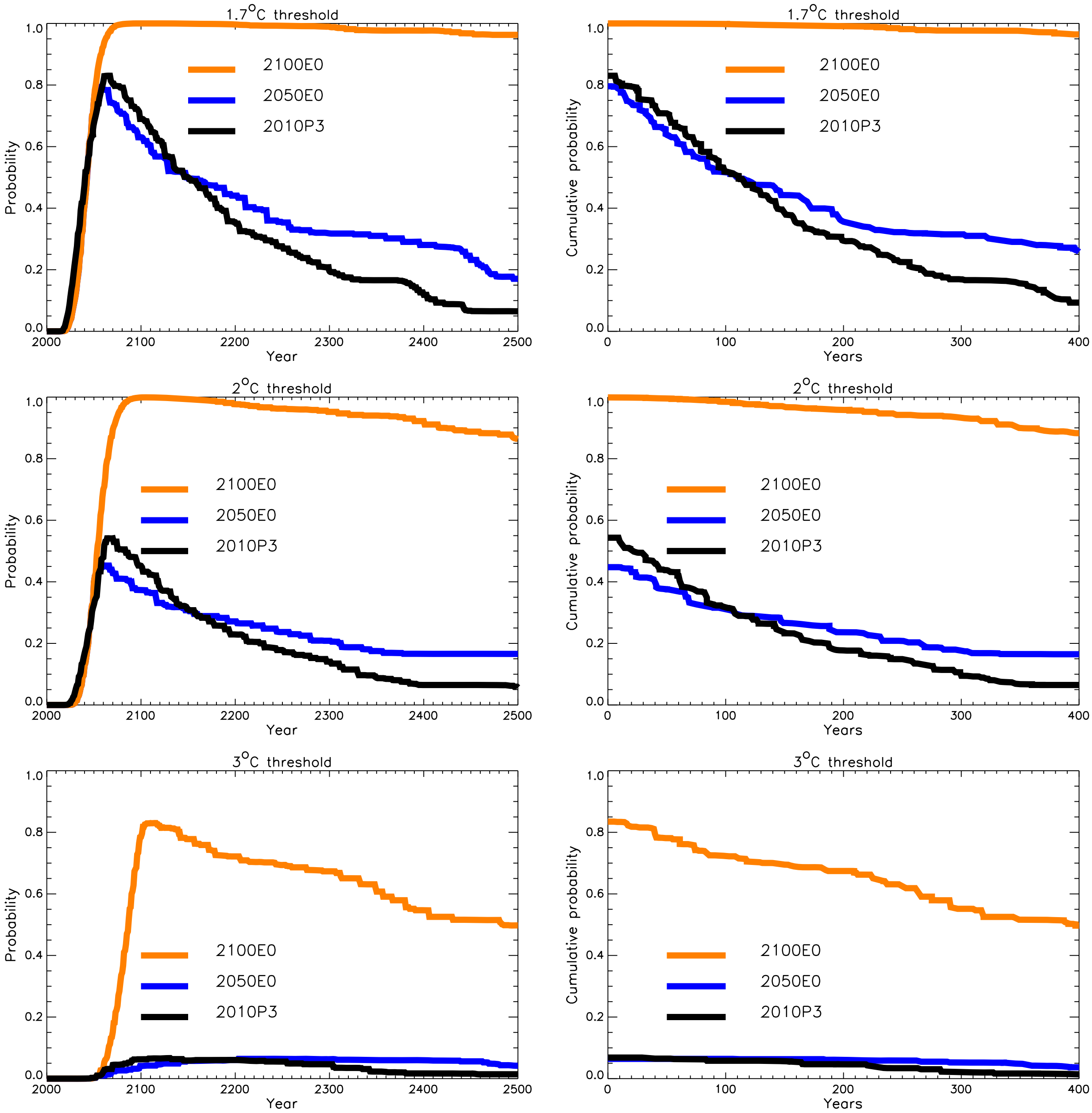

Figure 5. The probability of exceeding three different temperature thresholds. The left-hand panels show the probability of being over the temperature target at times during the simulation. The right-hand panels show the cumulative probability of exceeding the target for a range of time periods. In the right-hand panels the initial year (0 years) value indicates the likelihood that the temperature target is exceeded at any point in the experiment. Results for the 2010P3, 2050E0 and 2100E0 scenarios are presented.

Sulfate aerosol forcing is negative and reductions in sulfate emissions from year 2010 tend to cause a direct warming. At the slowest rate of sulfate emission reduction (i.e. over 100 years) the probability of global average temperature being warmer than pre-industrial by 2 °C or more at any time in the simulation is around 25%. Halving the sulfate emissions reduction time to 50 years increases the chance of exceeding 2 °C to around 55%. A further halving of the sulfate emissions reduction time only increases the chance of going over 2 °C by a further 10%. In terms of the 50th percentile warming at 2100 (i.e. for the probability of exceeding temperature to be 50% in figure 6, top panel), the range across the different aerosol scenarios is around 0.4 °C. Further, the choice of sulfate emission reduction time also alters the time spent overshooting 2 °C (figure 6, lower panel), with the probability of being over the 2 °C threshold for a given period of time higher for the shorter sulfate aerosol reduction times. In the most rapid aerosol reduction case the probability of being over the threshold has decreased by half over a century, whereas in the least rapid aerosol reduction scenario between 150 and 200 years elapse before the probability of being over 2 °C is halved.

4. Conclusions

Previous work using simple climate models, or climate–carbon cycle models of intermediate complexity, have demonstrated

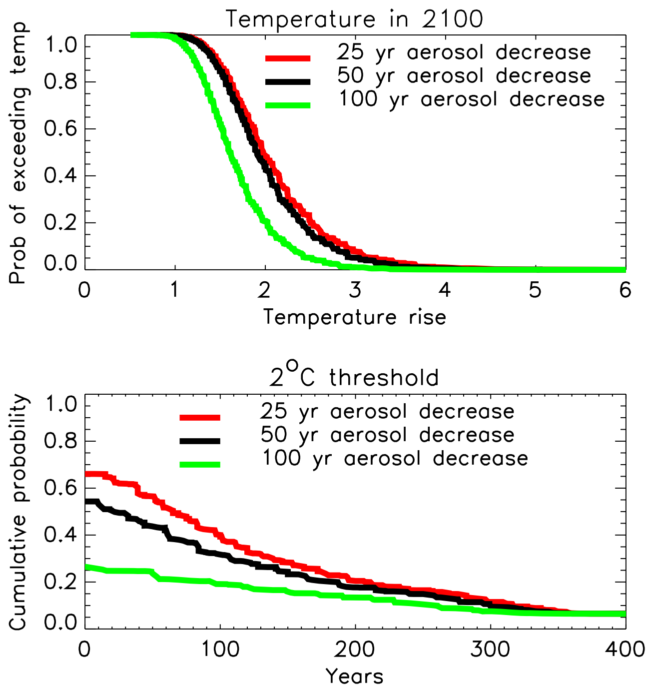

Figure 6. The upper panel shows the cumulative probability of exceeding a range of temperatures at 2100. The lower panel shows the probability of time spent exceeding a 2 °C threshold. The results are present for three versions of the 2010P3 scenario with aerosol reduction times of 25, 50 (the base case) and 100 years.

the potentially long timescales for global temperatures to decline, even following large reductions in carbon dioxide emissions. This is a result of particular importance to policy makers, and especially if unwelcome or dangerous levels of climate change are realized—it is necessary to know that it may not be easy to return back to safer levels of global warming. In this letter, our first key outcome is that we have increased the robustness of this result by attaining it using the most complex class of Earth system model, a GCM, complete with spatially resolved carbon cycle. Indeed, our analysis with a particular coupled carbon cycle GCM shows that, following rapid decreases in emissions, even slower CO_2 and temperature decreases may result, compared to those previously published. We recommend that our analysis be extended in future using additional GCMs.

Next we used large ensembles of simulations with a simple coupled climate–carbon cycle model, with each ensemble member given a different parameterization and an associated probability based on the existing knowledge base. This has allowed probability estimates to be made for the amount of time for which the global surface temperature might exceed critical warming thresholds. We found that for a multi-gas emissions scenario that peaks emissions in 2015 before adjusting to a long-term reduction rate of 3% per year, there is around a 55% probability of exceeding a 2 °C target above pre-industrial levels. Possibly of more importance is that we find a 30% probability that we would remain above this warming level for at least 100 years, and a 10% probability that the 2 °C threshold may be exceeded for up to 300 years. This particular scenario has a reduction in greenhouse gas emissions approaching 50% of the 1990 values by 2050, which we note is similar to the G8 statement in 2008 to consider ‘the goal of achieving at least 50% reduction of global emissions by 2050’. House *et al* (2008) showed that further emissions reductions beyond 2050 would also be needed to limit long-term temperature increases and we include this ongoing emissions reduction here too.

It is noteworthy that although our GCM simulations suggest very little decline in global temperature during the century following a 2 °C warming, the MAGICC simulations show there are values of key climate parameters that do allow a faster recovery. These more optimistic possibilities do, nevertheless, represent recovery timescales that are possibly long for society to deal with. This remaining uncertainty provides extra motivation for narrowing the uncertainty bounds on climate sensitivity, ocean heat diffusivity and the climate–carbon cycle feedback. Our results also imply that we need to focus future research not only on thresholds of dangerous climate change but also on quantifying the resiliency of Earth system components (such as the Greenland ice sheet, major ecosystems or the thermohaline circulation) to temporarily exceeding critical thresholds for a range of different periods. In adaptation cost benefit analysis, it may be appropriate to consider the extra costs of temporarily adapting to levels above an eventual desired target temperature, with analyses of the type we present here providing estimates of timescales above such targets. That such temporary resilience is important to assess was shown in earlier work, e.g. Schneider and Mastrandrea (2005) and O’Neill and Oppenheimer (2004). Here we have extended their concepts to the most complex of climate models and, by using in parallel large ensembles of simulations with a simpler climate model, we have been able to place the ‘overshoot’ issue into a probabilistic framework of direct use for future planning within the climate change debate.

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