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LETTER

An investigation into linearity with cumulative emissions of the climate and carbon cycle response in HadCM3LC

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Abstract

We investigate the extent to which global mean temperature, precipitation, and the carbon cycle are constrained by cumulative carbon emissions throughout four experiments with a fully coupled climate–carbon cycle model. The paired experiments adopt contrasting, idealised approaches to climate change mitigation at different action points this century, with total emissions rising to more than two trillion tonnes of carbon (TtC). For each pair, the contrasting mitigation approaches—capping emissions early versus reducing them to zero a few decades later—cause their cumulative emissions trajectories to diverge initially, then converge, cross, and diverge again. We find that global mean temperature is linear with cumulative emissions across all experiments, although differences of up to 1.5 K exist regionally when the trajectories of total carbon emitted during the course of the two scenarios coincide, for both pairs of experiments. Interestingly, although the oceanic precipitation response scales with cumulative emissions, the global precipitation response does not, due to a decrease in precipitation over land above emissions of around one TtC. Most carbon fluxes are less well constrained by cumulative emissions as they reach two trillion tonnes. The opposing mitigation approaches have different consequences for the Amazon rainforest, which affects the linearity with which the carbon cycle responds to cumulative emissions. The average Transient Climate Response to cumulative carbon Emissions (TCRE) is 1.95 K TtC^{-1}, at the upper end of the Intergovernmental Panel on Climate Change's range of 0.8–2.5 K TtC^{-1}.

1. Introduction

Since the beginning of the industrial age, human activities have resulted in the release to the atmosphere of more than half a trillion tonnes of carbon (TtC) in the form of carbon dioxide (Ciais *et al* 2013), the greenhouse gas primarily responsible for the change in climate the planet has experienced since that time (Myhre *et al* 2013).

An important question which has been addressed through the use of climate models over the last two decades is, at what level should atmospheric CO_2 be limited to in order to prevent dangerous climate change (Hansen *et al* 2008)? Answering this question is difficult due to uncertainty in the amount of warming the planet would undergo following stabilisation of atmospheric CO_2. In the event that a CO_2 concentration target could be agreed upon, the task of deciding how emissions should be controlled in order to achieve the target is itself compounded by the existence of feedbacks between the carbon cycle and climate; the changing climate undermines the capacity of the land surface and the oceans to absorb carbon from the atmosphere, leading to further climate change (Friedlingstein *et al* 2006).

In recent years, a new approach to investigating climate change has emerged: the cumulative carbon emissions framework (e.g. Allen *et al* 2009, Matthews *et al* 2009, Gregory *et al* 2009, Zickfeld *et al* 2012, Gillett *et al* 2013). Such studies have revealed that, to first order, instantaneous global mean temperature is directly proportional to the cumulative carbon emitted to that point (e.g. Matthews *et al* 2009, Zickfeld *et al* 2012) and is largely independent of the emissions

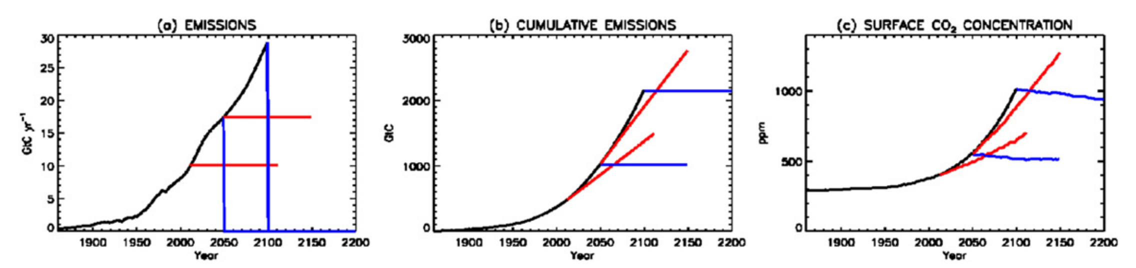

Figure 1. (a) CO_2 emissions scenarios used to drive the four experiments; (b) timeseries of cumulative emissions (GtC) (c) timeseries of atmospheric CO_2 (ppm).

pathway, for cumulative emissions below around two TtC (Matthews *et al* 2009). A simple linear relationship between emissions and global mean warming has therefore been established, with the potential to simplify advice to policymakers. Acknowledging this development, the Intergovernmental Panel on Climate Change (IPCC) for the first time expressed climate change in terms of cumulative carbon emissions in its 5th Assessment Report (AR5), through the transient climate response to cumulative CO_2 emissions (TCRE) metric, defined as the global mean warming per TtC emitted (Collins *et al* 2013, IPPC 2013). More recently, the work of Leduc *et al* (2016) suggests that linearity of surface temperature change with cumulative carbon emissions is also applicable regionally.

In the present study we provide further evidence for the dependence of global mean temperature on cumulative carbon emissions with two pairs of simulations performed with a fully coupled climate–carbon cycle general circulation model, a variant of the third Hadley Centre Climate Model (HadCM3 Gordon *et al* 2000). Each pair of simulations contrasts two idealised approaches to climate change mitigation in the 21st century; early capping of emissions which then proceed at a fixed rate indefinitely, versus increasing emissions under a business-as-usual scenario for a few more decades before they reduce instantaneously to zero. The emissions pathways of each pair diverge initially, before converging, coinciding briefly, and diverging again. We examine the degree to which the climate response and that of the carbon cycle are linear with cumulative emissions, globally and in some cases regionally, and investigate the mechanisms underlying these responses.

2. Methods

We have performed two pairs of simulations with HadCM3LC, the low resolution carbon-cycle variant of HadCM3 (Gordon *et al* 2000). The four experiments were initialised at different action points along a prior simulation driven by the A2 scenario of the Special Report on Emissions Scenarios (SRES) (Nakicenovic and Swart 2000); A2 is a fossil fuel-intensive, business-as-usual scenario. For the first, or 'early' pair, action was taken either in 2012 to cap emissions at 10 GtC yr^{-1}, or delayed until 2050 when emissions were reduced instantaneously to zero, having by then reached 17 GtC yr^{-1}. Hereafter these will be referred to as 2012E10 and 2050E0 respectively. In the second, or 'late' pair, mitigation did not begin until 2050, with emissions then capped at 17 GtC yr^{-1}, or 2100, when emissions dropped instantaneously from 29 GtC yr^{-1} to zero. Hereafter these will be referred to as 2050E17 and 2100E0 respectively. The A2 experiment ran from 2005 to 2100 and was in turn initialised from a historical simulation forced with observations-based CO_2 emissions from 1860 to 2005, following the C4MIP protocol (Friedlingstein *et al* 2006). Cumulative emissions are therefore expressed relative to 1860. Some results from the zeroed emissions experiments have been published in previous studies on the reversibility of climate change (Lowe *et al* 2009) and on the role of the terrestrial carbon cycle in recovery following a halt in emissions (Jones *et al* 2009, 2010).

HadCM3LC is a fully coupled ocean–atmosphere general circulation model, with both sub-models at 2.5° by 3.75° resolution. The atmosphere has 19 levels in the vertical, the ocean has 20 levels. In HadCM3LC the full carbon cycle is included, with the TRIFFID (Top Down Representation of Foliage and Flora Including Dynamics, Cox 2001) dynamic vegetation model and the Hadley Centre Ocean Carbon Cycle model, HadOCC (Palmer and Totterdell 2001). The model was configured in emissions-driven mode, with atmospheric CO_2 carried as a tracer over the full depth of the atmosphere model. Atmospheric CO_2 is therefore interactive, free to change in response to ocean and land uptake, with prescribed CO_2 emissions providing a source term at the land surface. This configuration allows the emergence of feedbacks between the climate and the carbon cycle. The concentrations of other greenhouse gases and aerosols were fixed throughout, at pre-industrial or early 20th century levels in all experiments. Land use change was not represented in any experiment.

3. Results

3.1. Cumulative emissions and atmospheric CO_2

The four emissions scenarios used to drive the model are shown in figure 1(a) and the resulting cumulative

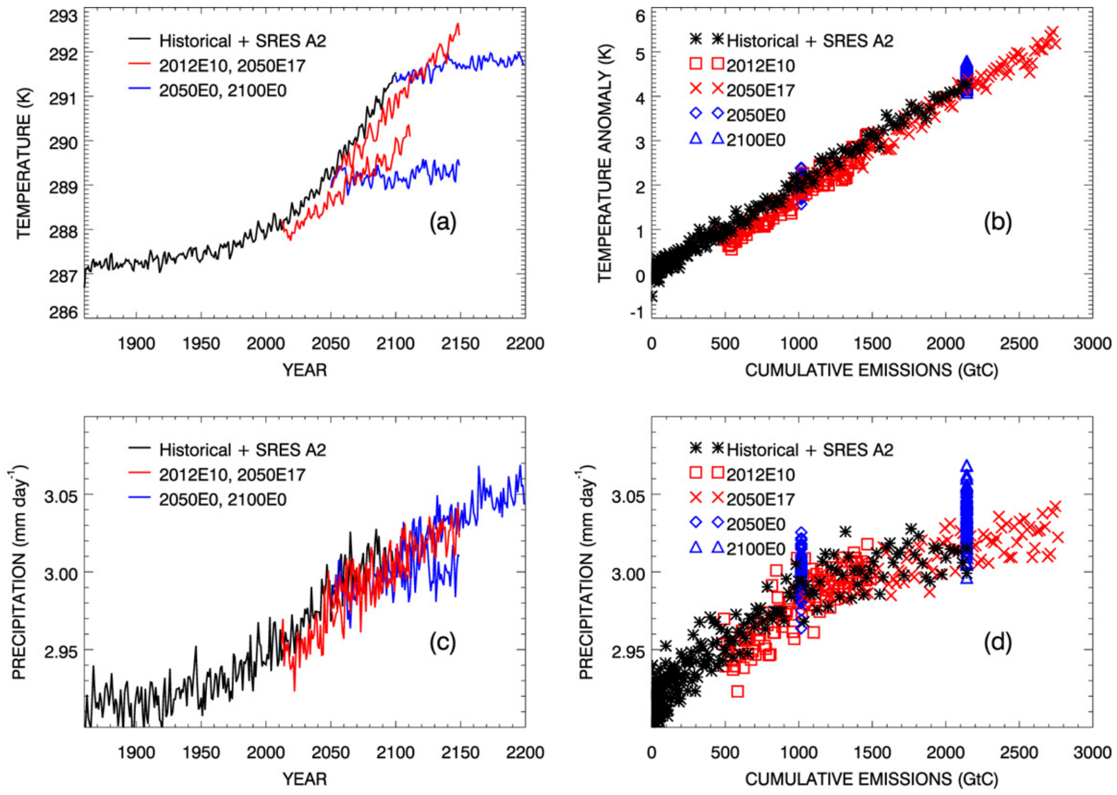

Figure 2. (a) Timeseries of global mean 1.5 m temperature; (b) change in global mean 1.5 m temperature as a function of cumulative emissions; (c) timeseries of global mean precipitation; (d) global mean precipitation as a function of cumulative emissions.

emitted carbon in figure 1(b). In the first pair of experiments, cumulative emissions coincide at 1017 GtC in 2065, whereas the later pair coincide in 2113, at 2142 GtC.

Atmospheric CO_2 increases linearly in 2012E10 and 2050E17 from when emissions are capped, reaching 702 and 1274 ppm respectively (figure 1(c)). In 2050E0 atmospheric CO_2 reduces from 551 to 514 ppm during the century after emissions cease, while in 2100E0 it reduces from 1013 to 937 ppm.

Table 1. Cumulative emissions, change in global mean 1.5 m temperature, and transient climate response to cumulative emissions. The TCRE is calculated as the temperature change averaged over the final decade of each experiment, relative to the 1860–1899 mean, divided by the cumulative CO_2 emitted (TtC) between the mid-points of the same two periods. TCRE is not an appropriate metric for zeroed emissions scenarios so has not been calculated for the E0 experiments.

Experiment	Final year cumulative emissions (GtC)	ΔT	TCRE (K TtC^{-1})
2012E10	1487	2.83	1.98
2050E0	1017	2.13	—
2050E17	2762	5.15	1.93
2100E0	2142	4.65	—

3.2. Global mean climate response

3.2.1. Temperature

The global mean 1.5 m temperature rises steadily following the capping of emissions in 2012E10 and 2050E17, but also continues to rise gradually after emissions cease in 2050E0 and 2100E0 (figure 2(a)), consistent with other studies (e.g. Plattner *et al* 2008, Solomon *et al* 2009, Gillett *et al* 2011, Fröhlicher and Paynter 2015). The cumulative emissions of 2012E10 coincide with those of 2050E0 in year 2065; their decadal mean temperatures centred on that year differ by only 0.05 °C. The equivalent for the later experiments is 0.09 °C, centred on year 2113.

3.2.1.1. TCRE

The TCRE, determined from the preindustrial to the end of each fixed emissions experiment, is 1.98 K TtC^{-1} for 2012E10 and 1.93 K TtC^{-1} for 2050E17. In each case the global mean temperature change is calculated as the preindustrial, averaged over 1860–1899, subtracted from the mean of the final decade of the fixed emissions experiment (table 1). A total of 1487 GtC was emitted by the end of 2012E10, whereas almost twice that (2762 GtC) was emitted by the end of 2050E17, demonstrating that the TCRE is approximately constant to almost 2800 GtC in HadCM3LC.

3.2.2. Precipitation

The rise in precipitation following cessation of emissions in 2050E0 and 2100E0 (figure 2(c)) is consistent with other studies using HadCM3 (Wu *et al* 2010), and also with results from other models (Gillett *et al* 2011, Zickfeld *et al* 2012). Across all experiments with

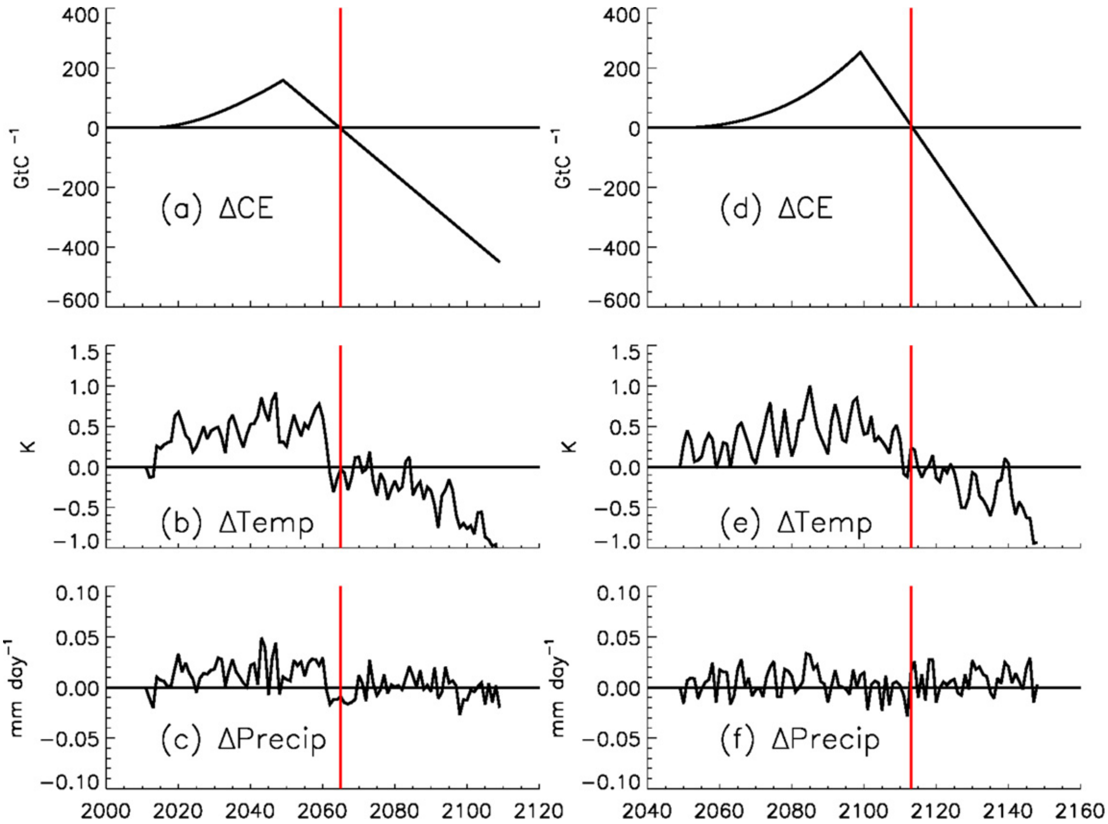

Figure 3. Timeseries of the 2050E0–2012E10 difference in variable V, $\Delta V_{2050\mathrm{E0}-2012\mathrm{E10}}$, where V is (a) cumulative emissions; (b) global mean temperature; (c) global mean precipitation; and the equivalent for the later experiments, $\Delta V_{2100\mathrm{E0}-2050\mathrm{E17}}$, where V is (d) cumulative emissions; (e) global mean temperature; (e) global mean precipitation. The vertical line indicates the year in which the cumulative emissions coincide.

non-zero emissions, the global mean precipitation initially increases as cumulative emissions rise, but begins to level out beyond around 1500 GtC (figure 2(d)). This is discussed further in the section 4.1.

3.3. Correlation of inter-experiment differences in cumulative emissions and global means and totals

Figure 3(a) shows a timeseries of the difference in cumulative emissions (CE) between 2012E10 and 2050E0, i.e. $\mathrm{CE}_{2050\mathrm{E0}} - \mathrm{CE}_{2012\mathrm{E10}}$, or $\Delta\mathrm{CE}_{2050\mathrm{E0}-2012\mathrm{E10}}$. The two simulations diverge from 2012 onwards when emissions are capped in 2012E10, whereas they continue to follow the A2 scenario in 2050E0; as a result, $\Delta\mathrm{CE}_{2050\mathrm{E0}-2012\mathrm{E10}}$ grows. When emissions cease in 2050E0, $\Delta\mathrm{CE}_{2050\mathrm{E0}-2012\mathrm{E10}}$ starts to reduce, reaching zero when cumulative emissions in the two coincide and becoming negative when $\mathrm{CE}_{2012\mathrm{E10}}$ exceeds $\mathrm{CE}_{2050\mathrm{E0}}$. The equivalent for the later experiments is shown in figure 3(d).

Also shown in figure 3 are timeseries of differences in temperature (T) and precipitation (P) i.e. $\Delta V_{2050\mathrm{E0}-2012\mathrm{E10}} = V_{2050\mathrm{E0}} - V_{2012\mathrm{E10}}$ and $\Delta V_{2100\mathrm{E0}-2050\mathrm{E17}} = V_{2100\mathrm{E0}} - V_{2050\mathrm{E17}}$ for $V = T$ and $V = P$. Figure 4 shows timeseries of differences in carbon fluxes and stores, i.e. gross primary productivity (ΔGPP), soil respiration (ΔSResp), plant respiration (ΔPResp), plant litter (ΔLitter), ocean carbon uptake (ΔOcnUptake), vegetation carbon (ΔVegC), and soil carbon (ΔSoilC).

If a variable scales linearly with cumulative emissions, then ΔV would be around its greatest when ΔCE is greatest, and close to zero when ΔCE is zero; the coefficient of correlation between ΔV and ΔCE would therefore be positive and close to one. Table 2 records the correlation coefficient between ΔCE and ΔV for all variables listed above, as well as for net primary productivity (ΔNPP) and net ecosystem productivity (ΔNEP). The correlation coefficient used is the Pearson product-moment coefficient, ρ. No allowance has been made for any lagged response when calculating the correlations.

3.3.1. Correlation of ΔTemperature with ΔCE

As figure 3(b) shows, $\Delta T_{2050\mathrm{E0}-2012\mathrm{E10}}$ is largest very close to the peak in $\Delta\mathrm{CE}_{2050\mathrm{E0}-2012\mathrm{E10}}$, and is almost zero (0.03 K) when $\Delta\mathrm{CE}_{2050\mathrm{E0}-2012\mathrm{E10}}$ is zero, yielding a very high correlation coefficient of +0.90. For the later pair of experiments, the correlation is almost as strong at +0.85 as ΔT grows and declines in phase with ΔCE.

3.3.2. Correlation of ΔPrecipitation with ΔCE

Figure 2(c) shows that global mean precipitation is rising at similar rates in 2050E17 and 2100E0 and

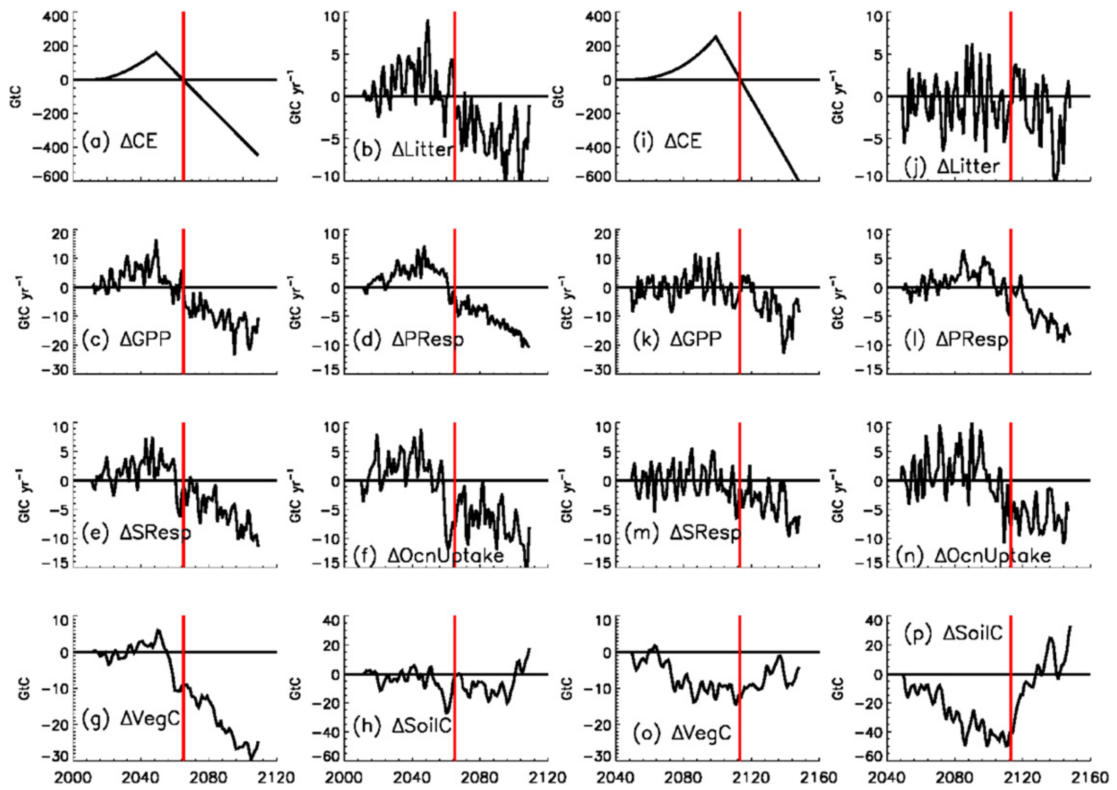

Figure 4. Timeseries of the 2050E0–2012E10 difference in variable V, $\Delta V_{2050\text{E0}-2012\text{E10}}$, where V is (a) cumulative emissions; (b) plant litter; (c) GPP; (d) plant respiration ; (e) soil respiration; (f) ocean carbon uptake; (g) vegetation carbon store; (h) soil carbon store; and the equivalent for the later two experiments, $\Delta V_{2100\text{E0}-2050\text{E17}}$, where V is (i) cumulative emissions ; (j) plant litter; (k) GPP; (l) plant respiration; (m) soil respiration; (n) ocean carbon uptake; (o) vegetation carbon store; (p) soil carbon store. The vertical line indicates the year in which the cumulative emissions coincide.

Table 2. Pearson coefficient of correlation, ρ, between the difference in cumulative emissions and the difference in the global quantity listed (first column): differences are calculated as 2050E0–2012E10 (third column) and 2100E0–2050E17 (fourth column). **Bold** indicates a correlation of magnitude 0.6 or more, ***bold italic*** of more than 0.8.

	Correlation of ΔV with ΔCE	
Variable, V	2050E0–2012E10	2100E17–2050E0
1.5 m Temperature	***+0.90***	***+0.85***
Precipitation	+0.45	−0.11
GPP	***+0.87***	**+0.65**
NPP	**+0.66**	+0.18
NEP	−0.15	−0.19
Soil respiration	***+0.88***	**+0.63**
Plant respiration	***+0.94***	***+0.89***
Plant litter	**+0.76**	+0.29
Vegetation carbon	***+0.96***	−0.24
Soil carbon	−0.01	***−0.85***
Ocean carbon uptake	**+0.74**	+0.55
Atmospheric CO_2	***+1.00***	***+1.00***

has high interannual variability. As a result, $\Delta P_{2100\text{E0}-2050\text{E17}}$ (figure 3(f)) shows no discernible trend, so correlation with $\Delta\text{CE}_{2100\text{E0}-2050\text{E17}}$ is negligible ($\rho = -0.11$). Since precipitation in 2050E0 rises more slowly than that in 2012E10 (figure 2(c)), a trend in $\Delta P_{2050\text{E0}-2012\text{E10}}$ is detectable, with a moderate correlation with $\Delta\text{CE}_{2100\text{E0}-2050\text{E17}}$ ($\rho = 0.47$). However, as figure 2(d) suggests, precipitation is much less well constrained by cumulative emissions than temperature.

3.3.3. Correlation of Δ(carbon fluxes and stores) and ΔCE

For 2012E10 and 2050E0, the differences in carbon fluxes between the two experiments generally correlate well with the difference in cumulative emissions

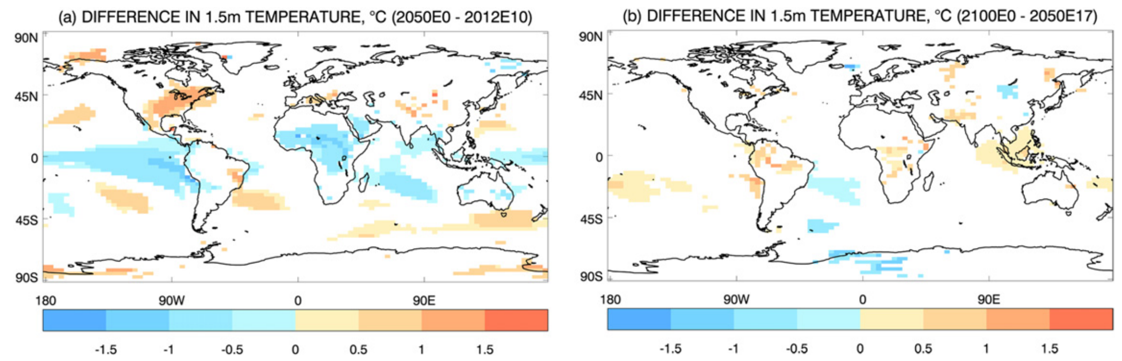

Figure 5. Difference in 1.5 m temperature, (a) 2050E0–2012E10, and (b) 2100E0–2050E17. Ten year average centred on the year at which cumulative emissions in both experiments of each pair were equal. Regions in which the differences are not statistically significant ($p < .04$) are masked out.

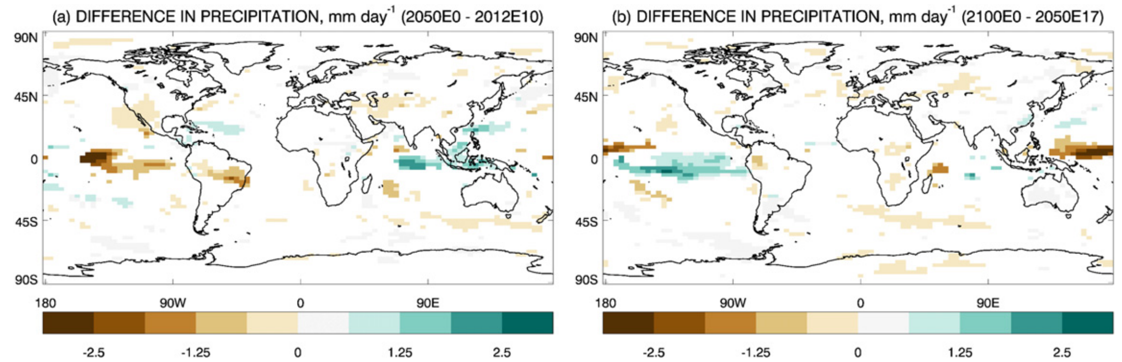

Figure 6. Difference in precipitation, mm d^{-1} (a) 2050E0–2012E10 and (b) 2100E0–2050E17. Ten year average centred on the year at which cumulative emissions in both experiments of each pair were equal. Regions in which the differences are not statistically significant ($p < .04$) are masked out.

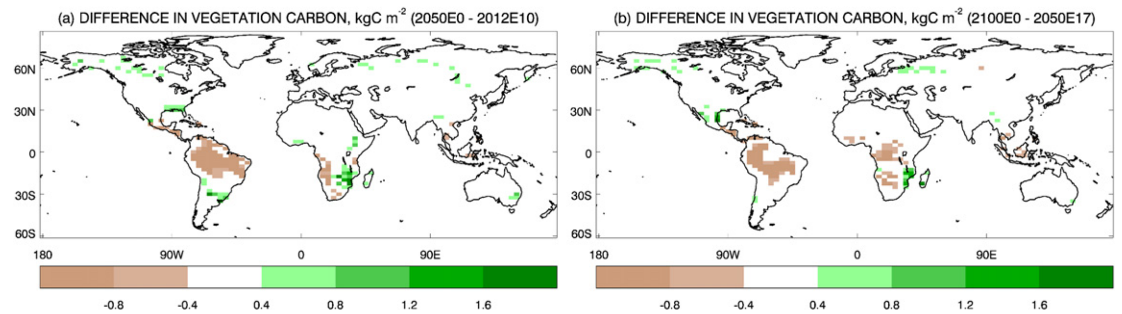

Figure 7. Difference in vegetation carbon, kgC m^{-2} (a) 2050E0–2012E10 and (b) 2100E0–2050E17. Ten year averages centred on the year at which cumulative emissions in both experiments of each pair were equal.

(figure 4, table 2); the same is true of the vegetation carbon store, but there is no correlation in the case of soil carbon.

The difference in carbon fluxes between 2050E17 and 2100E0 correlate less well with $\Delta CE_{2100E0-2050E17}$. As well as the environmental drivers, the fluxes depend also on the store sizes, and as cumulative emissions approach and exceed 2 trillion tonnes changes in the carbon store sizes start to become important. This is discussed further in section 4.4.

3.4. Regional comparison

The concept of the linear dependence of climate change on cumulative carbon emissions is a global one, though a recent study (Leduc *et al* 2016) utilising output from twelve CMIP5 models suggests that regional climate change is also approximately linear with cumulative emissions. So we now turn our attention to regional differences in temperature, precipitation, and land carbon stores in our pairs of simulations when the cumulative emissions of both

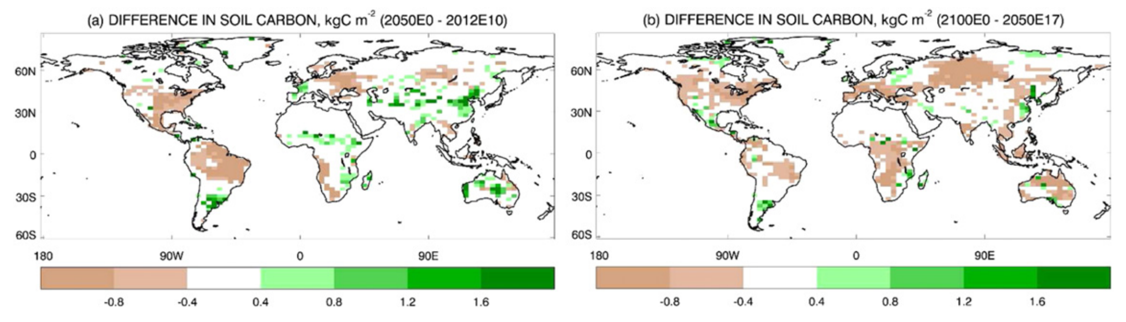

Figure 8. Difference in soil carbon, kgC m^{-2} (a) 2050E0–2012E10 and (b) 2100E0–2050E17. Ten year averages centred on the year at which cumulative emissions in both experiments of each pair were equal.

coincide and their global mean temperatures are comparable. Figures 5–8 show maps of differences in decadal mean temperature, precipitation and vegetation and soil carbon respectively, centred on the year in which their cumulative carbon emissions coincide.

3.4.1. Temperature

Figure 5(a) shows the (2050E0–2012E10) difference in 1.5 m temperature at the time of coincident cumulative emissions and figure 5(b) shows the equivalent for (2100E0–2050E17). To assess statistical significance we used a Monte-Carlo approach, at each gridbox comparing the decadal mean inter-experiment difference with the spread of one hundred randomly sampled decadal-mean differences from the HadCM3LC pre-industrial control simulation; gridboxes where the inter-experiment difference did not meet the significance threshold of 0.04 are masked out in figure 5. Although global mean temperatures are comparable at this time, the eastern part of North America is cooler by 0.5–1.5 K in 2012E10 than in 2050E0, whereas Central Africa is warmer in 2012E10 by a similar amount. In the later pair a smaller proportion of the land surface shows statistically significant differences, although there are regions where 2050E17 is 1 K cooler than 2100E0.

3.4.2. Precipitation

Figure 6(a) shows the statistically significant spatial differences between 2012E10 and 2050E0, averaged over the decade centred on the year in which the cumulative emissions coincide. Significance was determined by the Monte-Carlo approach outlined above (section 3.4.1). The largest areas of significant differences in rainfall occur over the oceans, although sizeable regions exist over land. Up to 2.5 mm d^{-1} less rain falls over eastern Brazil in 2050E0 than in 2012E10, while the reverse is true in parts of South East Asia. The differences between the later experiments are smaller in magnitude over land at less than 1 mm d^{-1}, with southern South America and much of Australia being the largest regions of statistically significant differences in rainfall.

3.4.3. Vegetation and soil carbon

When cumulative carbon emissions coincide, in both pairs of experiments the greatest differences in vegetation carbon occur in the Amazon (figure 7). In each case, capping emissions early rather than stopping them altogether decades later preserves more of the rainforest, though in both of the later simulations the Amazon is considerably depleted by this stage (see figure 12(b) and the discussion). For soil carbon the differences are more widespread, particularly so in the case of the later experiments (figure 8(b)).

4. Discussion

4.1. Global temperature and precipitation

Our results are consistent with earlier studies (Allen *et al* 2009, Matthews *et al* 2009, Gillett *et al* 2013, Krasting *et al* 2014) which demonstrate the proportionality of global mean temperature change with cumulative carbon emissions. The TCRE is within 1.95 ± 0.03 K TtC^{-1} over the final decade of 2012E10 and 2050E17, so is constant out to cumulative emissions of more than 2700 GtC and sits towards the upper end of the IPCC's range of 0.8–2.5 K TtC^{-1}. A recent study with an Earth System Model of intermediate complexity found that the TCRE begins to reduce as cumulative emissions reach multiple trillions of tonnes of carbon (Herrington and Zickfeld 2014), while Krasting *et al* (2014) found the TCRE to be sensitive to very high or low rates of emissions with a full complexity ESM. It remains to be seen how the TCRE in HadCM3LC would evolve under similarly high cumulative emissions or with comparable emissions rates. The temperature response of the zeroed emissions experiments is consistent with Fröhlicher and Paynter (2015) who demonstrated that continued warming following the cessation of emissions is a robust feature of fifteen CMIP5 Earth System Models.

Comparatively few studies have investigated the linearity of precipitation with cumulative emissions. One that does (Zickfeld *et al* 2012), in which the Canadian Earth System Model (CanESM1) is forced

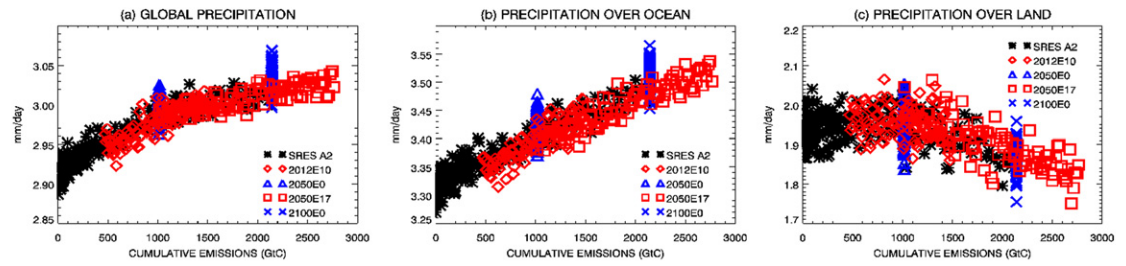

Figure 9. Precipitation (mm d^{-1}) averaged (a) globally (b) over oceans only and (c) over land only, versus cumulative carbon emissions (GtC).

with a range of CO_2 scenarios, finds global mean precipitation to be linearly related to cumulative emissions. This is not true for HadCM3LC (figure 9(a)), but precipitation averaged only over the ocean is, in fact, very close to linear (figure 9(b)). Precipitation over land only shows a marked downward trend beyond emissions of around 1000 GtC and this, therefore, is the primary cause of the deviation from linearity with cumulative emissions of global mean precipitation. This trend in precipitation over land is driven by changes in the Tropics, primarily by a reduction in rainfall over Amazonia as CO_2 continues to rise and the planet warms, which has been observed in HadCM3LC previously (Cox *et al* 2004).

4.2. Spatial patterns of change in temperature and precipitation

Leduc *et al* (2016) demonstrate that regional temperature change is approximately linear in cumulative carbon emissions. Figure 5 of the present study shows that in both sets of experiments with HadCM3LC, regions exist where the surface temperature evolves differently according to the emissions pathway employed. Explaining the evolution of the spatial differences in temperature and precipitation between the individual experiments of each pair when their cumulative emissions coincide (figures 5 and 6) is not straightforward. Of the early pair, one experiment undergoes two large El Nino events during the decade averaged to create the figures; the influence of this in figure 5(a) is apparent. For the later pair, there is no clear physical difference between the two to explain the spatial differences in warming and precipitation. Differences in albedo, for example, do not correlate with those in temperature, and the same is true of the Bowen ratio, describing the partitioning of energy between sensible and latent heat. The spatial patterns of figures 5 and 6 are likely, therefore, to be the result of a combination of factors, with a contribution from albedo differences due to changes at the land surface, but also from localised biophysical feedbacks between vegetation and the atmosphere, as well as larger scale differences in atmospheric and ocean circulation. An additional suite of sensitivity experiments would likely be required to fully account for these results.

4.3. Mechanisms underlying linearity of global mean temperature with cumulative emissions

Recent research into the mechanisms underlying the linearity of global mean temperature change with cumulative emissions points towards a combination of factors dominated by the ocean. Matthews *et al* (2009) suggested that the ocean's dual role of absorbing both heat and carbon might be important, and this has been borne out by more recent studies (Goodwin *et al* 2015, MacDougall and Friedlingstein 2015). For a fixed emissions scenario such as 2012E10 and 2050E17, the analytical framework of MacDougall and Friedlingstein (2015) explains the linearity of TCRE to around 2000 GtC through the combination of a number of factors; a fixed ocean-borne fraction of emitted carbon leads to a stable airborne-fraction; radiative forcing (RF) from increasingly high levels of atmospheric CO_2 begins to saturate, and this is offset by a decline in ocean heat uptake.

The fixed emissions experiments of the present study lead to a linear TCRE even out to cumulative emissions of more than 2700 GtC, although the underlying mechanisms are subtly different for 2012E10 and 2050E17. Figure 10 combines the cumulative airborne fraction, atmospheric CO_2, ocean heat content and RF due to CO_2 for both experiments. For 2012E10 the cumulative airborne fraction of emitted carbon ranges from 0.48 to 0.58—so by the end of the 100 year experiment, 58% of all carbon emitted as carbon dioxide since 1860 remains in the atmosphere (ignoring recycling through the biosphere). At the start of the 2050E17 experiment the cumulative airborne fraction is already 0.56, and rises to 0.74 by the end. The atmospheric CO_2 concentration of 2012E10 reaches at most 700 ppm, while 2050E17 rises to 1274 ppm after 100 years. The logarithmic nature of the relationship between RF and CO_2 becomes apparent only in the higher emissions experiment as the increase in RF begins to tail off halfway through, as CO_2 continues to rise. However, the ocean heat content continues to grow at an increasing rate throughout both experiments. For HadCM3LC, therefore, a linear TCRE appears not to require a reduction in the ability of the ocean to absorb heat under high CO_2 to match the reduction in RF; any significant drop in

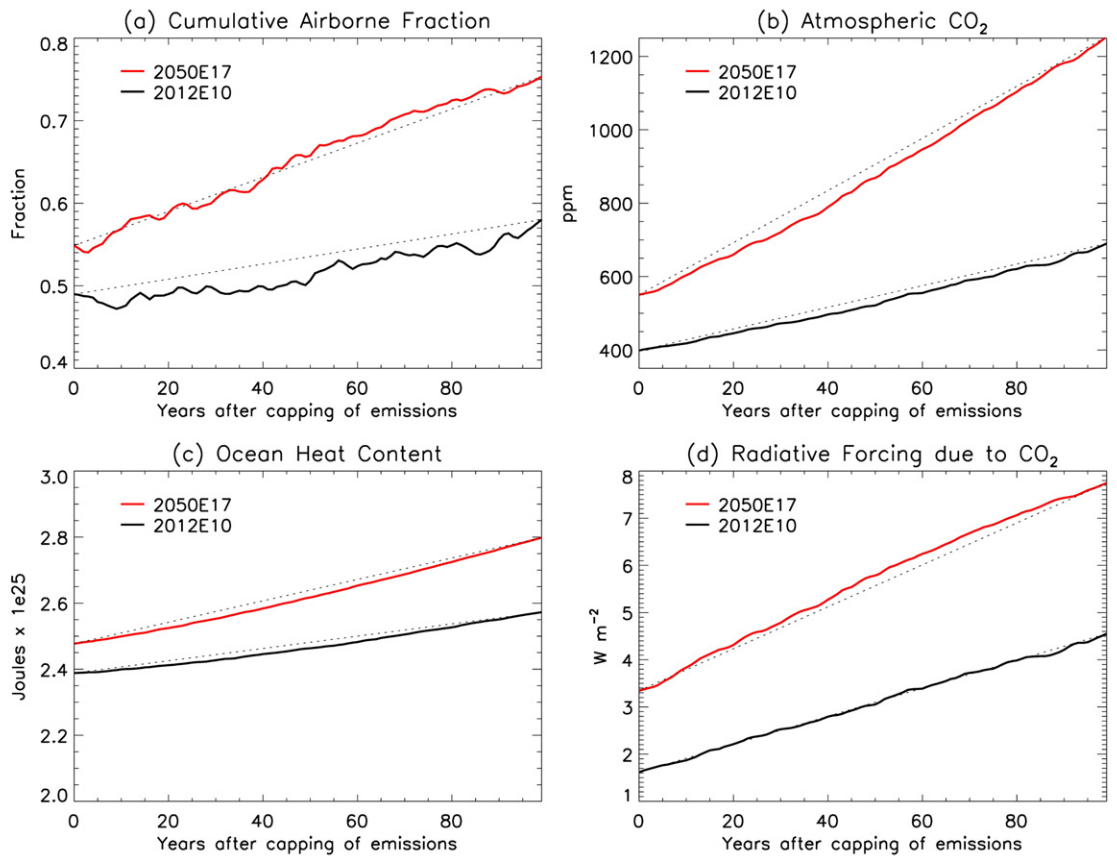

Figure 10. (a) Cumulative airborne fraction of all carbon emitted as carbon dioxide since 1860; (b) atmospheric CO_2 concentration (ppm); (c) ocean heat content over the whole depth of the ocean; (d) radiative forcing due to CO_2 relative to pre-industrial levels, calculated as 5.35 ln (C/C_0) where C is the simulated CO_2 concentration and C_0 is the preindustrial value (averaged over 1860–1899 in this case). Results for the hundred years following the capping of emissions in 2050E17 and 2012E10 are shown on the same axes for ease of comparison. The faint dotted lines link the first and last years to illustrate the deviation from linearity of the timeseries.

ocean heat uptake would result in an upward trend in TCRE.

4.4. Evolution of the cumulative airborne fraction

The cumulative airborne fraction changes over time as the fractions of total emitted carbon stored by the ocean and land change. Figure 11 shows how the land, ocean and atmosphere stores change, (a) and (b) in absolute terms, and (c) and (d) as a fraction of total emitted carbon. Carbon uptake by the ocean remains strong throughout, at a stable/increasing fraction of total emitted carbon in the fixed/zeroed emissions experiments respectively. By the end of all simulations the land has started to re-emit to the atmosphere some or all of the carbon emissions it had previously absorbed; this is not compensated for by increased ocean uptake, so the experiments with continuing emissions see an increase in the fraction of total emitted carbon stored in the atmosphere.

This land surface response is driven by the strong climate–carbon cycle feedback for which HadCM3LC first became known (Cox *et al* 2000). In both 2050E17 and 2100E0 the global soil carbon stores reduce significantly (figure 12(c)) as losses due to respiration outpace inputs from litter over much of the land surface. Vegetation differences are driven by changes in Amazonia (figure 12(b)). Before emissions reduce to zero in 2100E0, the Amazon rainforest crosses a threshold triggering a rapid decline in broadleaf tree extent which does not stabilise until many decades after emissions cease. The capping of emissions in 2050E17 delays this decline by a couple of decades and reduces its severity, but does not prevent it from occurring. As a consequence, $\Delta \mathrm{VegC}_{2100\mathrm{E}0-2050\mathrm{E}17}$ is dominated by this tipping point response which overrides any proportionality with cumulative emissions that might otherwise have existed. The emissions cap in 2050E17 also delays the sharp decline in soil carbon by around two decades.

So, while the carbon fluxes are primarily driven by temperature and CO_2, both of which are close to linear with cumulative emissions, significant changes to the store sizes begin to undermine the extent to which the fluxes are themselves constrained by cumulative emissions. For that reason the differences in carbon fluxes between the second pair of experiments correlates less well with the difference in cumulative emissions than is true of the first pair (table 2). The carbon stores change in line with the net balance of opposing fluxes integrated over time; there is no reason to suppose this

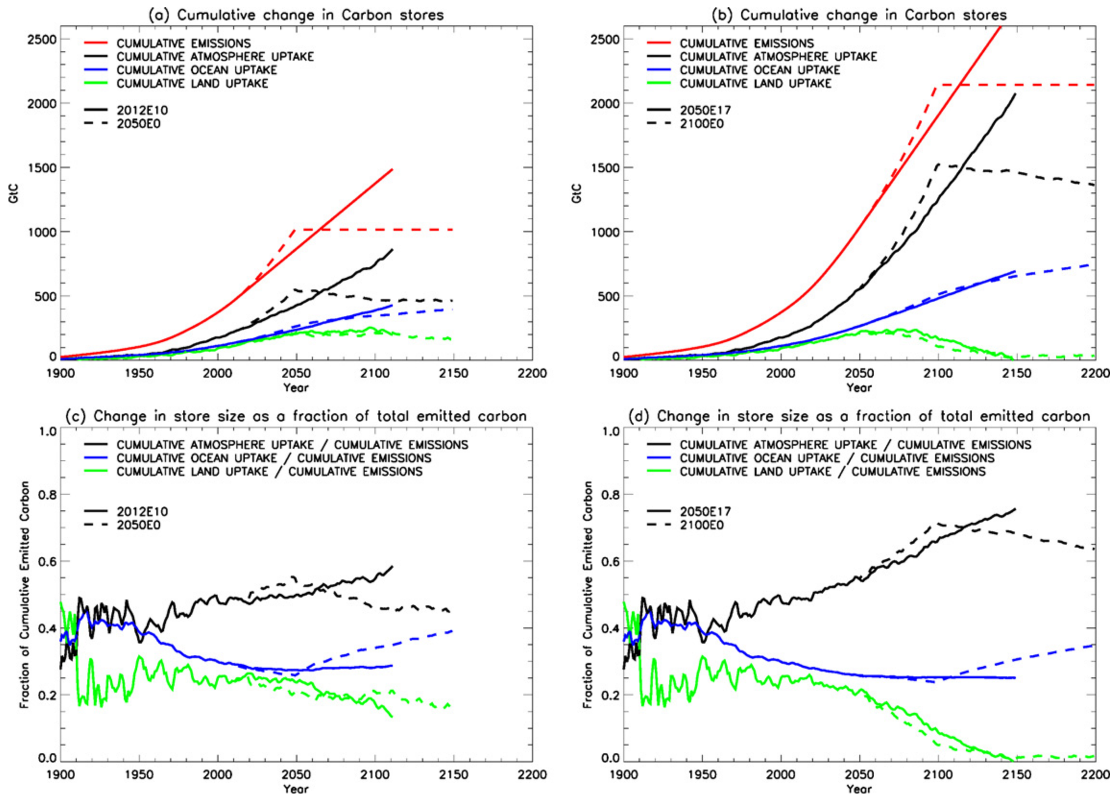

Figure 11. Cumulative emissions (red curve) and cumulative uptake of carbon by atmosphere (black), ocean (blue) and land (green) for (a) 2012E10 and 2050E0 and (b) 2050E17 and 2100E0. The cumulative uptake by atmosphere, ocean and land stores expressed as a fraction of cumulative emitted carbon are shown for (c) 2012E10 and 2050E0 and (d) 2050E17 and 2100E0.

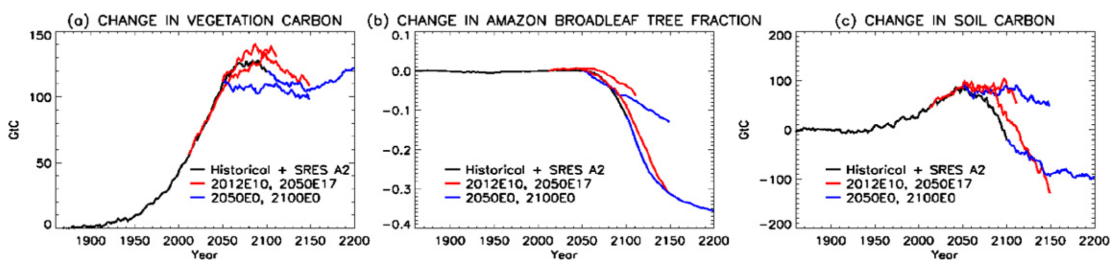

Figure 12. Timeseries of (a) change in global vegetation carbon content (b) change in broadleaf tree fraction of Amazonia, defined as 80W–35W, 30S–10N (c) change in global soil carbon store.

would evolve linearly with cumulative emissions, so it is perhaps less surprising that soil carbon is not constrained by cumulative emitted carbon in the early experiments, than that vegetation carbon is.

5. Conclusions

We have investigated the behaviour of HadCM3LC within the context of the cumulative emissions framework through the use of two pairs of experiments which follow contrasting CO_2 emissions pathways, but whose cumulative emissions coincide briefly. Global mean temperature is linear with cumulative emissions, while global mean precipitation deviates from linearity due to a decrease in precipitation over land, primarily Amazonia, beyond emissions of around one TtC. When the cumulative emissions of each pair of experiments coincide and their global mean temperatures are comparable, regional temperature differences of up to 1.5 K exist. This perhaps implies a greater emissions pathway dependence in HadCM3LC than seen in the experiments of Leduc *et al* (2016); additional simulations would be required to examine this further. The response of the Amazon rainforest to the changing climate causes significant reduction in vegetation and soil carbon stores in the later pair of experiments, in which emissions exceed 2 TtC, undermining the extent to which their carbon fluxes scale with cumulative emissions. The TCRE of HadCM3LC, at 1.95 K TtC^{-1}, is towards the upper end of the range given in the IPCC's fifth assessment report.

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