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Role of vegetation change in future climate under the A1B scenario and a climate stabilisation scenario, using the HadCM3C earth system model

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Abstract

The aim of our study was to use the coupled climate-carbon cycle model HadCM3C to quantify climate impact of ecosystem changes over recent decades and under future scenarios, due to changes in both atmospheric CO₂ and surface albedo. We use two future scenarios – the IPCC SRES A1B scenario, and a climate stabilisation scenario (2C20), allowing us to assess the impact of climate mitigation on results. We performed a pair of simulations under each scenario – one in which vegetation was fixed at the initial state and one in which vegetation changes dynamically in response to climate change, as determined by the interactive vegetation model within HadCM3C.

In our simulations with interactive vegetation, relatively small changes in global vegetation coverage were found, mainly dominated by increases in scrub and needleleaf trees at high latitudes and losses of broadleaf trees and grasses across the Amazon. Globally this led to a loss of terrestrial carbon, mainly from the soil. Global changes in carbon storage were related to the regional losses from the Amazon and gains at high latitude. Regional differences in carbon storage between the two scenarios were largely driven by the balance between warming-enhanced decomposition and altered vegetation growth. Globally, interactive vegetation reduced albedo acting to enhance albedo changes due to climate change. This was mainly related to the darker land surface over high latitudes (due to vegetation expansion, particularly during winter and spring); small increases in albedo occurred over the Amazon. As a result, there was a relatively small impact of vegetation change on most global annual mean climate variables, which was generally greater under A1B than 2C20, with markedly stronger local-to-regional and seasonal impacts. Globally, vegetation change amplified future annual temperature increases by 0.24 and 0.15 K (under A1B and 2C20, respectively) and increased global precipitation, with reductions in precipitation over the Amazon and increases over high latitudes. In general, changes were stronger over land – for example, global temperature changes due to interactive vegetation of 0.43 and 0.28 K under A1B and 2C20, respectively. Regionally, the warming influence of future vegetation

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change in our simulations was driven by the balance between driving factors. For instance, reduced tree cover over the Amazon reduced evaporation (particularly during summer), outweighing the cooling influence of any small albedo changes. In contrast, at high latitudes the warming impact of reduced albedo (particularly during winter and spring) due to increased vegetation cover appears to have offset any cooling due to small evaporation increases.

Climate mitigation generally reduced the impact of vegetation change on future global and regional climate in our simulations. Our study therefore suggests that there is a need to consider both biogeochemical and biophysical effects in climate adaptation and mitigation decision making.

1 Introduction

Significant alterations in future global land use patterns are anticipated, both as a result of anthropogenic changes in land use and management (e.g., Nakicenovic et al., 2000; Moss et al., 2009; Hurtt et al., 2011) and due to climate-driven changes in ecosystem distribution and agricultural suitability. These changes in land use and land cover patterns may have considerable impacts on the climate (Desjardins et al., 2007), both through modifying the physical properties of the land surface (biophysical effects) and by altering the absorption or emission of greenhouse gases (biogeochemical forcings).

1.1 Biogeochemical impacts of land cover change on climate

Biogeochemical forcings due to land cover change include changes in greenhouse gas (GHG) fluxes – for instance changes in soil and vegetation carbon storage, carbon dioxide fluxes, changes in emissions of nitrous oxide and methane (Paustian et al., 2006), and changes in water vapour, which may have a global effect due to their impact on longwave radiation. In addition to these biogeochemical climate impacts driven by changes in land use/cover and management, further impacts will occur as a result

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of the changing future climate (Falloon and Smith 2009; Falloon et al., 2009, 2010, 2011). For instance, changing temperatures, precipitation patterns and carbon dioxide concentrations may alter ecosystem distribution, productivity, carbon storage and GHG fluxes.

5 There is general agreement that future warming will likely negatively affect the uptake of carbon on land (Sitch et al., 2008; Scholze et al., 2006), although more recent studies suggest that nitrogen limitation may reduce or even reverse the sign of the land carbon feedback (Friedlingstein and Prentice, 2010).

1.2 Biophysical impacts of land cover change on climate

10 The biophysical impacts of changes in land cover may include alterations to surface fluxes of radiation (e.g. albedo), heat, moisture (e.g. evaporation) and momentum (e.g. roughness length), which in turn may alter may local and regional climates (Lean and Warrilow 1989; Pielke et al., 1998; Betts, 2001; Feddema et al., 2001; Betts, 2005; Betts et al., 2007; Raddatz, 2007). Previous studies have shown that in climate models the effects of regionalised forcings are spread out through the model domain, particularly
15 zonally, as a result of model dynamics and convection scheme (Forster et al., 2000; Joshi et al., 2003; Shindell and Faluvegi, 2009). However, there is generally a higher response in the region where the forcing is applied (Forster et al., 2000), especially in the Northern Hemisphere extra-tropics (Shindell and Faluvegi, 2009). Changes in
20 albedo and other biophysical changes may therefore also exert climate impacts over broader scales. For example, McCarthy et al. (2012) have demonstrated that the representation of vegetation patterns may impact the location of the Inter Tropical Convergence Zone (ITCZ) and the South Asian monsoon circulation. Overall, anthropogenic land cover change between the potential natural state and the present day is generally
25 considered to have increased global albedo, resulting in an overall cooling effect (Sagan et al., 1979; Brovkin et al., 1999; Betts, 2001; Govindasamy et al., 2001), although there is considerable variation in responses across models (Pitman et al., 2009; De Noblet-Ducoudre et al., 2012). Vegetation dynamics can also enhance the

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low-frequency variability of the biosphere-atmosphere system at timescales from a few years to a century, by slowly modifying the physical characteristics of the land surface (Delire et al., 2011).

Conversion of forest to cropland or pasture reduces the aerodynamic roughness of the landscape and decreases both the capture of precipitation on the canopy and the root extraction of soil moisture. These changes tend to decrease evaporation and hence reduce the fluxes of moisture and latent heat from the surface to the atmosphere, which acts to increase the temperature near the surface (Lean and Rowntree, 1993). Also, a forested landscape generally has a lower surface albedo than open land, particularly in conditions of lying snow when shortwave radiation is trapped by multiple reflections within the forest canopy (Betts and Ball, 1997). Deforestation in mid and high latitude regions can therefore lead to increased shortwave reflection, which provides a cooling influence (Thomas and Rowntree, 1992; Bonan et al., 1992; Douville and Royer, 1997; Lee et al., 2011). The relative importance of these processes depends on local conditions such as the underlying surface albedo and soil moisture availability, and can vary with season and location (Betts, 1999; Betts et al., 2007).

1.3 Regional impacts of land cover change on climate

Significant changes in ecosystem distribution are projected for two key regions: the Amazon and high latitudes. In response to future climate change, most stand-alone dynamic global vegetation models (DGVMs: Sitch et al., 2008) and the coupled climate-carbon cycle General Circulation Models (GCMs) that include DGVMs in their formulation (Cox et al., 2000; Friedlingstein et al., 2006) tend to simulate varying degrees of loss in the tree plant functional type (PFT) cover over the Amazon, and gains in woody cover in the tundra, while some models suggest losses of herbaceous vegetation in the tundra and others gains, or no change.

Most stand-alone and coupled DGVMs simulate a reduction in vegetation carbon over Amazonia and increases in vegetation carbon over tundra ecosystems. There is less agreement in simulated changes in soil carbon stocks – at high latitudes responses

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vary from large and small increases to a strong decrease; some models project strong decreases over Amazonia and others a small increase (Sitch et al., 2008; Friedlingstein et al., 2006).

Overall, deforestation in high-latitude regions may lead to a cooling under present climate due to the dominant effect of increased surface albedo (Bonan et al., 1992; Betts, 1999; Betts, 2001; Bounoua et al., 2002; Davin and De Noblet-Ducoudre, 2010). Conversely, increases in cool-region forest area may have the opposite effect. However, recent studies have questioned whether warming resulting from large-scale mid and high latitude afforestation may be altered by enhanced transpiration (Swann et al., 2010) and water vapour export (Swann et al., 2011), which may trigger further feedbacks and alter circulation patterns. Tropical deforestation is expected to lead to a warming and drying of local climate since the impact of reduced evapotranspiration (Lean and Warrilow, 1989; Betts et al., 2007; Wang and Davidson, 2007; Davin and De Noblet-Ducoudre, 2010) may outweigh the relatively small changes in albedo. Future Amazon forest losses due to climate change may therefore contribute to further warming. In addition, the biogeophysical effects of the projected future climate-driven Amazon forest dieback may also be important locally, acting to further reduce rainfall (Betts et al., 2004). On the other hand, the projected future increases in high latitude forest area and productivity may warm future climate due to reduced surface albedo (Betts, 2000), particularly during winter and spring (Betts et al., 2007).

1.4 Present study

Prior to the Coupled Model Intercomparison Project Phase 5 (CMIP5: Taylor et al., 2012), only Met Office Hadley Centre coupled climate-carbon cycle models included a dynamic vegetation model, and all other assessments of future changes in vegetation were made using offline DGVMs. The coupled climate-carbon cycle model HadCM3C (Murphy et al., 2009; Booth et al., 2011, 2012) includes a dynamic vegetation model and the carbon cycle as fully interactive components of the climate system. Vegetation interacts with the climate both through the carbon cycle and effects on the

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surface energy budget such as surface albedo. As part of the European Union Project, “CARBO-North – Quantifying the carbon budget in Northern Russia: past, present and future” (Kuhry, 2010), the aim of our study was to use HadCM3C to quantify the climate impact of ecosystem changes in Northern Eurasia over recent decades and under future scenarios, due to changes in both biogeochemical and biogeophysical effects. We use two future scenarios – the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emissions Scenarios (SRES) A1B scenario (Nakicenovic et al., 2000), and a climate stabilisation scenario (2C20; May, 2008), allowing us to assess the impact of climate mitigation on results. We performed a pair of simulations under each scenario – one in which vegetation was fixed at the initial state (1860 and 2020, respectively) and one in which vegetation changes dynamically in response to climate change, as determined by the interactive vegetation model within HadCM3C. However, since our simulations were global (and vegetation changes occurred over the whole globe, not just over the CarboNorth study region in North East Russia), we describe global results from our simulations, and investigate impacts over regions where significant changes in vegetation cover occurred (the Amazon and high latitude regions). Significant changes in ecosystem distribution and resulting impacts on climate over these two regions have also been illustrated in previous studies (see Sects. 1.2 and 1.3). Kuhry (2010) describes simulations with the ECHAM5/MPI-OM climate model, using data for vegetation changes from our A1B simulations where only high-latitude vegetation changes were implemented (see discussion).

The aim of our study was to investigate the following key questions:

1. How do differences in vegetation changes under the two future scenarios differ, and affect mitigation advice?
2. How do general aspects of climate change differ between the two scenarios (A1B and 2C20) in the simulations with interactive vegetation?
3. How does the impact of vegetation change on key surface climate variables in our simulations affect mitigation advice?

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There are several key differences between our study and recent related work. For instance, Strengers et al. (2010) studied 20th century climate-vegetation feedbacks (but not future impacts); Jiang et al. (2011) considered future vegetation-climate feedbacks using an equilibrium (not dynamic) vegetation model and not including biogeochemical responses. Swann et al. (2011) assessed imposed mid latitude afforestation impacts on global climate but under a present day climate; Swann et al. (2010) investigated regional climatic impacts of vegetation change in high latitudes under a “control” scenario and did not investigate interactions with future climate change. The European-scale study of Wramneby et al. (2010) investigated climate response to vegetation change under one future emissions scenario. Our study uses two different future climate scenarios, allowing us to assess the dependency of climate impacts due to vegetation change on scenario and forcings. Our study uses a fully-coupled atmosphere ocean GCM with an interactive carbon cycle, whereas the study of Swann et al. (2010) used slab (mixed-layer thermodynamics only) and fixed (sea surface temperature) ocean models. Our simulations are also fully dynamic integrations, in contrast to the time-slice simulations employed by Swann et al. (2010).

Many previous studies have investigated the impacts of human-induced land cover change on climate, or of simple prescribed changes in land cover (e.g., Lean and Warilow 1989; Pielke et al., 1998; Feddema et al., 2001; Betts, 2001, 2005; Falloon and Betts, 2006; Betts et al., 2007; Pitman et al., 2009; Swann et al., 2010). In common with Wramneby et al. (2010) climate-driven changes in vegetation in our simulations were determined by an interactive vegetation model although our study was global, in contrast to their European-scale simulations. Because of this, our study does not consider the influence of anthropogenic land use change. However, recent simulations performed with the second version of the Met Office Hadley Centre Global Environmental Model (Earth System – HadGEM2-ES) for the fifth IPCC assessment report do include both anthropogenic land use change and interactive (natural) vegetation changes (Collins et al., 2011; Jones et al., 2011).

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2 Methods

2.1 HadCM3C coupled climate-carbon cycle model

The model used as the basis of this study is a configuration of Version 3 of the Hadley Centre GCM, HadCM3C (Murphy et al., 2009; Booth et al., 2011, 2012), which is a version of the Met Office Unified Model (MetUM). This is a flux adjusted version (using an updated algorithm to that described by Collins et al., 2006) of HadCM3 (Gordon et al., 2000) coupled to the land surface and terrestrial carbon cycle component with interactive vegetation (TRIFFID: Cox et al., 1999; Cox et al., 2000; Cox, 2001) and an ocean carbon cycle (HadOCC: Palmer and Totterdell, 1999). In coupled GCMs with a fully-dynamic ocean component such as HadCM3, flux adjustment involves iterative adjustments to the ocean surface heat and water fluxes which themselves are first calibrated from a preliminary integration relaxed to observed climatological fields, and then applied to subsequent control and climate change simulations (Collins et al., 2006). The need for flux adjustments arise from errors in ocean transport and ocean-atmosphere exchanges. HadCM3C differs from HadCM3LC (Cox et al., 2000; Jones et al., 2003, 2005), and the coupled climate-carbon cycle model submitted to the Coupled Climate Carbon Cycle Model intercomparison (C4MIP), as it is configured to run with the standard (higher) HadCM3 resolution ocean ($1.25^\circ \times 1.25^\circ$) and also contains modelled interactive atmospheric sulphur cycle chemistry and a sulphate aerosol scheme including the direct and first indirect, “cloud albedo”, and aerosol effects (following Jones et al., 2001); note that the second indirect, “cloud lifetime”, effect is excluded). HadCM3C was recently used in the UK Climate Projections Project (UKCP – Murphy et al., 2009) and in the European Union project ENSEMBLES (Van der Linden and Mitchell, 2009).

2.2 MOSES2 land surface scheme and TRIFFID dynamic vegetation model

HadCM3C includes the Met Office Surface Exchange Scheme Version 2 (MOSES2 – Essery et al., 2001, 2003; Smith et al., 2006). MOSES2 (fully described by Essery

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et al., 2003) employs a tiled model of sub-grid heterogeneity, and includes seasonally varying vegetation (Martin et al., 2006). Separate surface temperatures, shortwave and longwave radiative fluxes, sensible and latent heat fluxes, ground heat fluxes, canopy moisture contents, snow masses and snow melt rates are computed for each surface type in a grid-box. These are then aggregated to form a grid-square mean with weightings equal to the fractions of each type in the grid-square. Air temperature, humidity and wind speed on atmospheric model levels above the surface and soil temperatures and moisture contents below the surface are treated as homogeneous across a grid box.

Nine surface types are recognized in MOSES2 (as applied in HadCM3C): broadleaf trees, needleleaf trees, C3 (temperate) grass, C4 (tropical) grass, shrubs, urban, inland water, bare soil and ice. Except for those classified as land-ice, a land grid box can be made up from any mixture of the first eight surface types. Fractions of surface types within each land-surface grid box are read from an ancillary file (as in our FIXVEG experiments) or modelled by the dynamic vegetation model TRIFFID (Cox, 2001), as in our INTVEG experiments. With each type of vegetation is associated a canopy height, a snow-free roughness length and a canopy water capacity which are parameterized functions of the Leaf Area Index (LAI), following Essery et al. (2001). Each vegetated tile also has an exponential root density distribution depending on the plant type. There are four soil layers in MOSES, each with a temperature, and moisture content and four soil layers have thicknesses from the surface downwards, of 0.1, 0.25, 0.65 and 2.0 m.

When the interactive vegetation scheme is included, TRIFFID models the state of the biosphere in terms of the soil carbon, and the structure and coverage of five functional types of plant (PFTs) within each model gridbox (broadleaf tree, needleleaf tree, C3 grass, C4 grass and shrub). Carbon fluxes for each vegetation type are calculated every 30 min as a function of climate and atmospheric CO₂ concentration, from a coupled photosynthesis/stomatal-conductance scheme (Cox et al., 1998, 1999), which utilizes existing models of leaf-level photosynthesis in C3 and C4 plants (Collatz et al., 1991, 1992). The accumulated fluxes are used to update the vegetation and soil carbon every

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10 days. The natural landcover evolves dynamically based on competition between the vegetation types, which is modelled using a Lotka-Volterra approach and a tree-shrub-grass dominance hierarchy. Some agricultural regions are also prescribed, in which grasslands are assumed to be dominant. Carbon lost from the vegetation as a result of local litterfall or large-scale disturbance is transferred into a soil carbon pool, where it is broken down by microorganisms that return CO₂ to the atmosphere. The soil respiration rate is assumed to double for every 10 K of warming (Raich and Schlesinger, 1992), and is also dependent on the soil moisture content (McGuire et al., 1992). Soil C is modelled within TRIFFID using a single pool with a single decay rate and takes no account of input quality. Hence it cannot simulate the dynamics of different classes of soil C. Changes in the biophysical properties of the land surface (Betts et al., 1997), as well as changes in terrestrial carbon (when the interactive carbon cycle is enabled), feed back onto the atmosphere.

2.3 Emissions scenarios and simulations

Table 1 summarizes our simulations, which were based on a run which was spun-up to a perpetual 1860 control state prior to the experiment. This was followed by simulations with forcings that simulate historical conditions and then two future scenarios corresponding to a) the IPCC SRES A1B scenario (Nakicenovic et al., 2000) and b) the “2C20” stabilization scenario specified by May (2008). These scenarios were also used in global and regional climate modelling studies performed under the EU CarboNorth project (Kuhry, 2010). The A1B simulations were all forced by CO₂ emissions, natural and anthropogenic sulphur emissions, prescribed changes in minor greenhouse gas concentrations (Nakicenovic et al., 2000), ozone concentrations (Meehl et al., 2007), solar forcing and also background and large-scale volcanic eruptions for the historical period (Fig. 1 – as used by Stott et al., 2006). For comparison, the setup of the A1B simulation using the “standard” version of HadCM3 (without an interactive carbon cycle) is also summarized in Table 1, and forcings for that simulation are shown in Fig. 1.

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5 May (2008) defined the original 2C20 stabilization scenario, which is broadly based on the “EU target” to limit global average temperature increases to below 2K over preindustrial levels, and to keep global atmospheric CO₂ concentrations below 550 ppmv. Using the ECHAM5/MPI-OM model, May (2008) applied observed concentrations of well-mixed greenhouse gases (GHGs) including CO₂ from 1861–2000, then following the A1B scenario from 2001–2020, and using constant concentrations from 2020 onwards. Ozone and sulphate concentrations used observed values from 1861–2000, A1B values from 2001–2020, from 2021–2036 the levels of A1B for 2100 were reached (i.e. five times faster than A1B), and from 2037–2100 constant 2100 A1B concentrations were applied. The global average temperature change relative to 1861–1990 values, projected by the ECHAM5/MPI-OM model for 2071–2100 was 1.92 K under the 2C20 scenario, compared to 3.47 K for the A1B scenario (May, 2008). For comparison, Fig. 1 also shows the CO₂ and sulphate concentrations, and total forcings used in the Representative Concentration Pathways (RCPs) applied in the IPCC’s forthcoming fifth assessment report (Van Vuuren et al., 2011).

15 The forcings applied in our simulations under the A1B and 2C20 simulations are shown in Fig. 1. Many aspects of the HadCM3C and ECHAM5/MPI-OM A1B and 2C20 simulations were very similar or identical (e.g. 2020 atmospheric concentrations of CH₄ = 2026 ppbv, N₂O = 331 ppbv, CFC-12 = 486 pptv). However, due to the difference in climate model structure and setup some aspects of our A1B and 2C20 simulations varied from those of May (2008). The resolution of the atmosphere and ocean models differ (HadCM3C: 2.5° × 3.75° atmosphere, 1.25° × 1.25° ocean; ECHAM5/MPI-OM: 1.9° × 1.9° atmosphere, 1.5° × 1.5° ocean). HadCM3C additionally includes a fully interactive carbon cycle (and hence uses CO₂ emissions, rather than concentrations) whereas ECHAM5/MPI-OM does not. This, in combination with the carbon cycle-climate feedback in HadCM3C resulted in greater CO₂ concentrations in 2020 (440 ppmv vs. 418 ppmv for HadCM3C and ECHAM5/MPI-OM, respectively – see Fig. 1a). HadCM3C includes interactive sulphur cycle chemistry and therefore uses ammonia, dimethyl sulphate (DMS) and high and low level SO₂ emissions whereas

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ECHAM5/MPI-OM uses concentrations. In our simulations, constant emission values for DMS and ammonia were used throughout (Stott et al., 2006). HadCM3C also includes natural and volcanic sulphate emissions whereas ECHAM5/MPI-OM does not. Because HadCM3C treats CFC-11, CFC-13 and HCFC-22 separately (using 2020 concentrations of 214, 72 and 229 pptv, respectively), but ECHAM5/MPI-OM only uses CFC-11, the latter used a higher CFC-11 concentration to compensate (786 pptv in 2020). In addition, in HadCM3C the forcings for the historic period were applied for 1860–1990 (Stott et al., 2006), rather than 1861–2000 in ECHAM5/MPI-OM.

Since the aim of our study simulations was to assess the role of vegetation change under future climate conditions, we performed two simulations for each future scenario. In one simulation, the vegetation cover in the model was allowed to evolve dynamically as determined by the TRIFFID interactive vegetation model (INTVEG) – A1B-INTVEG and 2C20-INTVEG for the A1B and 2C20 scenarios, respectively. In the other set of simulations, vegetation was fixed at the initial state from the corresponding period in the A1B-INTVEG simulation – in the year 1860 for the A1B simulation (A1B-FIXVEG), and in 2020 for the 2C20 simulation (2C20-FIXVEG). A1B-FIXVEG used annual CO₂ concentrations for the atmosphere and ocean from A1B-INTVEG throughout, while 2C20-FIXVEG used the 2020 values from A1B-INTVEG. Note that because they were driven by CO₂ concentrations (not emissions), the carbon cycle in both the 2C20-INTVEG and FIXVEG simulations and the A1B-FIXVEG run were not coupled to the climate. Effectively this meant that only the A1B-INTVEG considered the impact of changing carbon stocks on the climate; in 2C20 the terrestrial carbon stores changed in response to climate, but did not affect the climate. Note also that in the INTVEG simulations, only the PFT surface fractions (broadleaf trees, needleleaf trees, C3 grass, C4 grass and shrubs) plus bare soil change; the remaining surface fractions remain constant (urban, inland water, and ice). In addition, no anthropogenic land use change is applied in these simulations.

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2.4 Analysis approach

With the exception of timeseries, changes in climate variables were analysed using 30 yr means for the time periods in question. Our study focuses mainly on annual timescales although seasonal averages are also included in the main figures for changes in albedo and temperature under the A1B scenario, since marked seasonal differences were noted for those variables. Additional figures for seasonal average changes under the 2C20 scenario and for other variables are shown in the Supplement. Changes were assessed as global area averages, and for two regions where marked changes in terrestrial ecosystems occurred in our simulations – these are the Amazon region (AMZ: 40–70° W, 15° S–5° N) and the Northern high latitude region (HIGHLAT: 45–80° N). These regions were also studied by Jones et al. (2010). In the present study, changes are calculated over these regions both for the whole region (including ocean grid points) and for land grid points only. Changes in climate variables (ΔV) under the A1B and 2C20 scenarios are assessed as the difference between the 2080s (2071–2100) period for each future scenario (V_{future}), and the 1870s (1861–1890) period (V_{historic}) from the historic simulation, as follows:

$$\Delta V = V_{\text{future}} - V_{\text{historic}} \quad (1)$$

By separately calculating the change in climate variables for the INTVEG (ΔV_{INTVEG}) and FIXVEG (ΔV_{FIXVEG}) simulations, the influence of vegetation change (ΔC_{VEG}) is then assessed as:

$$\Delta V_{\text{VEG}} = \Delta V_{\text{INTVEG}} - \Delta V_{\text{FIXVEG}} \quad (2)$$

We also used data from a 240-yr control simulation of HadCM3C with pre-industrial forcings to assess changes in climate variables in our A1B and 2C20 simulations relative to “natural variability” in the control simulation. In the present study, we define natural variability as \pm two standard deviations of the control simulation (Collins et al., 2001; Cowling et al., 2009). Excluding changes within \pm two standard deviations of the

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control simulation is approximately equivalent to a signal-to-noise ratio of 2 (Hawkins and Sutton, 2011). Note that the global and regional average changes shown in the tables are presented as means across the full datasets – changes due to natural variability are not excluded as described above.

5 In order to take account of variations in insolation, surface albedo values (α) were calculated from total downward ($R \downarrow_{\text{tot}}$) and net downward shortwave radiation fluxes ($R \downarrow_{\text{net}}$) at the surface as follows:

$$\alpha = \frac{R \downarrow_{\text{tot}} - R \downarrow_{\text{net}}}{R \downarrow_{\text{tot}}} \quad (3)$$

3 Results

10 3.1 Vegetation changes

Relatively small increases in total vegetation fraction were found under both the A1B and 2C20 scenarios (Fig. 2) – with some increases over Eastern Asia and the Rockies, and decreases over the Amazon. Some decreases in total vegetation fraction were found over Australia under A1B, while increases were found over Western Australia under 2C20. Changes in individual PFTs were generally similar under both the A1B and 2C20 scenarios, but less marked under 2C20 (Fig. 3). Between 1961–1990 and 2071–2100 under A1B, there was a gain in the global area under needleleaf trees (from 6 to 7 %), shrub (from 7 to 10 %) and bare soil (from 25 to 26 %), and a loss of broadleaf trees (from 22 to 21 %), C3 grasses (from 18 to 15 %) and C4 grasses (from 18 to 16 %). Globally, the increase in scrub and needleleaf coverage was most rapid between 2000 and 2100 under A1B, while the rate of change slowed after around 2020–2030 under the 2C20 scenario. Broadleaf tree and C4 grass coverage both increased globally until 2050 and 2000, respectively, and declined thereafter, with less marked changes under 2C20. The global gain in scrub and needleleaf tree area were mostly accounted for by

changes in the high latitude regions, while the global losses of broadleaf tree, C3 and C4 grasses were mostly driven by the changes over the Amazon region.

Changes in PFT composition were much more marked over our two focus regions. During the same time period over the Amazon under A1B, the area under broadleaf trees was strongly reduced (from 64 to 39 %), C3 grass coverage contracted from 3 to 0.1 %, C4 grass coverage was reduced from 30 to 25 % and scrub coverage contracted from 0.5 to 0.03 %. These vegetation types were largely replaced by bare soil, which expanded from 2 % to 36 % of the total area. In contrast, over the HIGHLAT region under A1B, the area of bare soil contracted from 12 to 9 %, along with C3 grass (from 33 to 26 %) and C4 grass (from 6 to 5 %), being replaced by an expansion of broadleaf trees (from 3 to 5 %) needleleaf trees (from 20 to 23 %) and scrub (from 23 to 28 %).

In addition to these changes over the two focus regions, other notable changes in PFT composition occurred. For example, C3 grass coverage was also reduced across the USA, but increased across some parts of Eastern Asia. Increases in C4 grass were found over Western Australia under 2C20, but losses under A1B. Differences in the spatial patterns of change also occurred within our two focus regions. At high latitudes, increases in shrub cover and losses of C3 grass were most marked over Siberia; while some losses of shrub cover occurred over Northeastern Russia and Alaska. Over the Amazon, losses of C3 grass coverage occurred in the east, while gains were found in western parts.

3.2 Terrestrial carbon storage

Averaged over 1861–1890, the global total carbon stocks in our simulations were 1144, 514 and 1658 Pg C for soil, vegetation and total land carbon, respectively. Global total mean carbon stocks increased during the twentieth century until around 2020 under both INTVEG simulations and declined strongly thereafter under A1B (Table 2, Fig. 4) with a slower rate of decline under 2C20 (–42.6 and +14.3 PgC under A1B and 2C20, respectively). Interestingly the initial rate of loss of total carbon stocks under 2C20 from 2020 exceeded that under A1B until around 2070, and under 2C20 there was

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a small net increase in carbon stocks between 1860 and 2100. Changes in soil carbon seem to explain much of the global total carbon trends. For instance, there was a much larger decrease in soil carbon under A1B than under 2C20 (-109.4 and -39.9 PgC under A1B and 2C20, respectively). We found a small increase in vegetation carbon storage under both scenarios, which was smaller under 2C20 and had a decreasing trend during the final decades of the simulation (66.8 and 54.2 PgC under A1B and 2C20, respectively). These changes were equivalent to per-area changes of -0.29 , $+0.45$ and -0.74 kgCm^{-2} for total, vegetation and soil carbon under A1B, or $+0.10$, $+0.37$ and -0.27 kgCm^{-2} under 2C20.

Strong reductions in total carbon storage were found over AMZ under A1B as a result of both soil and vegetation carbon changes (-10.22 , -5.9 and -4.32 kgCm^{-2} for total, vegetation and soil carbon, respectively), although vegetation carbon changes appear to dominate. Under 2C20, much smaller total carbon reductions were found over AMZ, with changes in vegetation carbon also dominating (-5.47 , -3.16 and -2.31 kgCm^{-2} for total, vegetation and soil carbon changes, respectively). Over HIGHLAT, increases in total carbon storage were found under both A1B and 2C20 although there was a decreasing trend in both A1B and 2C20 from around 2060 and a smaller overall increase under A1B (0.91 and 1.08 kgCm^{-2} under A1B and 2C20, respectively). The differences in total carbon changes between A1B and 2C20 appear to be mostly driven by soil carbon changes over HIGHLAT. A reduction in soil carbon was found under A1B (-0.56 kgCm^{-2}), but a small increase under 2C20 (0.09 kgCm^{-2}), while a greater increase in vegetation carbon was found under A1B than 2C20 (1.48 and 0.99 kgCm^{-2} under A1B and 2C20, respectively).

Total annual mean land carbon storage under A1B-INTVEG (Fig. 5a) was reduced over the Amazon, mid USA (>4 kgCm^{-2}), Southern and Western Africa and mid latitude Eurasia from the Mediterranean to China ($\sim 1-4$ kgCm^{-2}). Gains in total carbon storage were found across high latitudes (~ 4 kgCm^{-2}) and over Southern Asia. Much smaller changes were found under 2C20 (Fig. 5b), generally following the regional patterns found under A1B. Differences in total carbon storage changes between A1B and

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2C20 appear to have been mostly driven by changes in soil carbon changes (Fig. 5e, f) – for instance smaller losses of soil C were found across mid-latitude Eurasia under 2C20 compared to those in A1B. Gains in vegetation carbon storage (Fig. 5c, d) were found over high latitudes and Southern Africa, with losses over the Amazon, and smaller changes (but similar patterns) under 2C20 compared to A1B.

3.3 Surface albedo

A very small decline in global mean surface albedo was found under A1B-INTVEG (Table 2), with a much smaller reduction under 2C20 after 2020 (-0.018 and -0.012 under A1B and 2C20). Stronger reductions in albedo were found over HIGHLAT in our INTVEG simulations (-0.06 and -0.04 under A1B and 2C20). A small increase in albedo was found over AMZ under A1B-INTVEG, with a smaller increase under 2C20 (0.01 and 0.002 under A1B and 2C20, respectively).

Interactive vegetation changes reduced albedo globally and over HIGHLAT under both A1B and 2C20, while over AMZ albedo increased under A1B with smaller increases under 2C20. Globally, vegetation change reduced albedo by -0.002 and -0.0006 under A1B and 2C20, respectively (or ~ 10 and 5% of the respective INTVEG changes). Over HIGHLAT, albedo was reduced by vegetation change by -0.01 and -0.005 under A1B and 2C20, or ~ 18 and 13% of the respective INTVEG changes. Over AMZ, small increases in albedo were found: 0.01 and 0.004 for A1B and 2C20 – 100% and 164% of the respective INTVEG changes. Much larger albedo changes due to vegetation change occurred over land-only grid points both globally and for HIGHLAT, often twice the value across the whole domain.

Under A1B-INTVEG, annual mean albedo was reduced over most land areas, particularly at high latitudes where reductions of over 0.15 were found (Fig. 6a). The exception to this was the Amazon, where albedo increased by nearly 0.05 . Changes in albedo due to vegetation changes mostly followed the signal found under A1B-INTVEG – and thus amplified overall albedo changes as a result of climate change (Fig. 6b), except over the Arctic Ocean. The strongest impacts were found under A1B, where albedo

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increased by ~ 0.05 over the Amazon and was reduced at high latitudes by nearly 0.05 . Vegetation change increased albedo over the Arctic ocean more strongly under 2C20 compared to A1B (Fig. 6c, d).

Changes in surface albedo due to vegetation change were generally more marked seasonally, compared to the annual average changes (Fig. 7) although seasonal differences were similar during all seasons over the Amazon. In particular, strong reductions in albedo were seen across mid-high latitudes in winter (December–February) and high latitudes in spring (March–May); in contrast to the annual mean decrease, small increases in albedo were found during summer (June–August) across Siberia. Seasonal changes in temperature under 2C20 (Fig. S1) were largely a diminished pattern of those found under A1B. The main exception was stronger albedo reduction under 2C20 during winter for the far West of Europe and North Eastern Russia, covering a region between approximately 45 to 65° N and 25 to 75° E. The reduction in winter albedo was smaller under A1B for this region, with some increases in albedo in the eastern parts.

3.4 Near surface temperature

Global mean temperatures increased steadily in both of our A1B simulations, while a much slower increase was observed in the 2C20 simulations after 2020. Global mean temperature increases between 1861–1990 and 2071–2100 (Fig. 8), under both the A1B and 2C20 simulations using HadCM3C with interactive vegetation (5.12 and 2.95 K, respectively) were greater than the values found by May (2008) using ECHAM5/MPI-OM (3.47 and 1.92 K, respectively). For the same time period, global mean temperature increases under our simulations with fixed vegetation were smaller than those with interactive vegetation (Table 2: 4.87 and 2.80 K for A1B and 2C20, respectively), but still greater than those found by May (2008). Under the A1B scenario, much larger changes in annual mean temperature for 2071–2100 relative to 1861–1890 were found under our simulations for AMZ and HIGHLAT (Fig. 8b, c): 8.6 K and 8.3 K, respectively, compared to the global mean change of 5.1 K.

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The impact of interactive vegetation on global mean temperature changes in our simulations was relatively small (approximately 4.8 and 5.2 % of the A1B-INTVEG and 2C20-INTVEG temperature increases, respectively), and greater under the A1B scenario (0.24 K) than under the 2C20 scenario (0.15 K). Regionally, the absolute impact of interactive vegetation under the A1B simulations on temperature was greater than the global value for both the AMZ (0.56 K or 6.6 % of the A1B-INTVEG increase) and HIGHLAT (0.35 K or 4.25 % of the A1B-INTVEG increase) regions, although the relative impact for HIGHLAT was smaller than the global effect. Under the 2C20 scenario, the absolute and relative impact of interactive vegetation on regional temperature increases was greater than the global changes for both HIGHLAT (0.26 K or 5.6 % of the 2C20-INTVEG increase) and AMZ (0.3 K or 6.4 % of the 2C20-INTVEG increase). Changes in temperature in the A1B and 2C20 simulations with both fixed and interactive vegetation generally followed very similar trends (Fig. 8). This was the case globally and for AMZ and HIGHLAT, with the only discernible impact of interactive vegetation appearing in the final decades over HIGHLAT, where temperatures in both of the INTVEG simulations were greater than those under the FIXVEG simulations. As for albedo, land-only grid points showed much greater temperature increases due to interactive vegetation, both globally and regionally (particularly for HIGHLAT).

Under our A1B-INTVEG simulation, annual mean warming was stronger over land than ocean, and strongest at high latitudes and over the Amazon, exceeding 10 K in some locations (Fig. 9a). Interactive vegetation led to annual mean warming over almost all land areas (Fig. 9b, c), with a weaker signal under 2C20 than under A1B. The most pronounced warming due to interactive vegetation (around 1K) was found over high latitudes (particularly Northeastern Russia and Siberia) and the Amazon under A1B, where the strongest changes in vegetation composition occurred. Interactive vegetation led to relative cooling (i.e. less warming) over the Arctic ocean under both 2C20 and A1B (~2 K; Fig. 9b,c). Compared to the vegetation effect under A1B, the vegetation changes under the 2C20 scenario induced greater cooling over the Arctic

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Ocean, less warming over Siberia, Australia and the Amazon, but more warming over the Atlantic, the Great Lakes and parts of Europe (Fig. 9d).

In response to vegetation change, larger seasonal changes in temperature were found compared to the annual mean changes (Fig. 10). This was particularly the case during boreal spring (March-May), when warming exceeding 1.5 K was found across Siberia, parts of Northeastern Russia, Alaska and Northern Canada, and the Eastern Amazon. As for albedo, seasonal changes in temperature under 2C20 (Fig. S2) were largely a diminished pattern of those found under A1B. The main exception was stronger warming during winter and summer for the far West of Europe and North Eastern Russia, between approximately 45 to 65° N and 25 to 75° E. Little warming was found for this region under A1B, but in contrast warming of over 1.5 K was found under 2C20.

Vegetation changes for this region were largely similar between the two scenarios. The differences in temperature changes due to interactive vegetation between the A1B and 2C20 scenarios for this region therefore appear to be driven by other factors (see Sects. 3.3, 3.5 and 3.6). These may include the greater winter precipitation increase under A1B enhancing snow cover, thus increasing albedo and leading to a greater relative temperature increase. In summer, greater reductions in precipitation and evaporation were found under 2C20, which may have contributed to the larger relative temperature increases compared to A1B.

3.5 Precipitation

Global mean precipitation increased steadily in all of our A1B and 2C20 simulations (Table 2). The precipitation changes for 2071–2100 relative to 1861–1890 were: 0.13 and 0.13 mm d⁻¹ for A1B-INTVEG and 2C20-INTVEG, respectively. A strong reduction in annual mean precipitation over AMZ was found under A1B-INTVEG, with a much smaller response under 2C20 (-2.52 and -1.11 mm d⁻¹ under A1B and 2C20, respectively). Precipitation increases over HIGHLAT were similar under both A1B and 2C20

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with a slightly smaller increase under 2C20 (0.29 and 0.21 mm d^{-1} under A1B and 2C20, respectively).

The impact of interactive vegetation on precipitation changes in our simulations varied with region and scenario, although little impact was discernible between precipitation trends in the INTVEG and FIXVEG simulations. Globally, interactive vegetation led to a relatively strong increase in precipitation in both the A1B (0.02 mm d^{-1} or 12.5 % of the A1B-INTVEG increase, for 2071–2100 relative to 1861–1890) and 2C20 simulations (0.02 mm d^{-1} or 14.4 % of the 2C20-INTVEG increase). In contrast, over the HIGHLAT region, interactive vegetation led to a stronger increase in precipitation in the 2C20 simulation for the same time period (0.02 mm d^{-1} or 11 % of the 2C20-INTVEG increase) compared to almost no impact under the A1B simulation (0.0004 mm d^{-1} or 0.1 % of the A1B-INTVEG increase). Strong decreases in precipitation over AMZ were found in our simulations as a result of interactive vegetation in both the A1B and 2C20 simulations (-0.35 and -0.16 mm d^{-1} , or 14 and 14 % of the A1B-INTVEG and 2C20-INTVEG decreases, respectively). In other words, the loss of Amazon forest cover led to a further decrease in precipitation. Much of the changes in total precipitation in our simulations were accounted for by changes in convective precipitation (not shown). Smaller changes in precipitation due to vegetation change were found over land-only grid points globally, with larger increases over HIGHLAT.

Under A1B, we found increases in annual mean precipitation over high latitudes, east and south Asia, and east Africa, with large decreases over Northern South America, Southern Africa and Southeast Asia (Fig. 11a). Generally, only small patchy signals in annual mean precipitation were found as a result of interactive vegetation, with smaller responses under 2C20 than A1B (Fig. 11b, c). The most pronounced response was the drying over the Amazon basin under A1B (~ -1.5 mm d^{-1}), with a much weaker signal in 2C20. An increase in precipitation was also found over the Congo basin, with a stronger signal under A1B than 2C20. Precipitation changes due to vegetation change under 2C20 were mostly a diminished signal of the A1B changes (Fig. 11d). Local changes in precipitation were dominated by changes in convective precipitation, for

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both the A1B and 2C20-INTVEG simulations and for the impact of vegetation change under A1B and 2C20 (not shown).

Seasonal patterns of change in precipitation due to the vegetation effect were largely similar to the annual mean changes. In response to vegetation change, larger seasonal changes in precipitation were found compared to the annual mean changes (Figs. S3 and S4) for some regions, most notably during winter and spring which dominated the annual decreases found over the Amazon. For the far West of Europe and North Eastern Russia (approximately 45 to 65° N and 25 to 75° E), there was a greater winter precipitation increase under A1B and a weaker reduction in summer precipitation, compared to 2C20.

3.6 Evaporation and latent heat flux

Between 1861–1890 and 2071–2100, global mean evaporation increased by 0.13 mm d^{-1} in both our A1B and 2C20 INTVEG simulations, equivalent to increases in latent heat flux of 3.78 and 3.63 W m^{-2} under A1B and 2C20, respectively (Table 2, Fig. 12). Compared to the global mean change, a slightly smaller increase in evaporation was found for HIGHLAT (0.10 and 0.10 mm d^{-1} for A1B and 2C20, respectively, equivalent to latent heat flux changes of 2.95 and 2.86 W m^{-2}). In contrast to the global increases in evaporation, large decreases were found over AMZ with a much smaller decrease under 2C20 compared to A1B (-1.37 and -0.58 mm d^{-1} or latent heat flux changes of -39.6 and -16.92 W m^{-2} under A1B and 2C20, respectively).

The impact of interactive vegetation changes in global annual mean evaporation changes was relatively small. A slight increase was found under both A1B and 2C20 (0.02 and 0.02 mm d^{-1} or latent heat flux changes of 0.47 and 0.52 W m^{-2} under A1B and 2C20, respectively). These changes are equivalent to approximately 12.3 and 14.2% of the A1B and 2C20-INTVEG increases, respectively. A small decrease in evaporation resulted from vegetation changes for AMZ under A1B, with a small increase under 2C20 (-0.02 and 0.001 mm d^{-1} or latent heat flux changes of -0.57 and 0.04 W m^{-2} under A1B and 2C20, respectively). This is equivalent to ~ 1 and -0.25 % of

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the A1B and 2C20 INTVEG changes, respectively. Interactive vegetation led to a small increase in evaporation over HIGHLAT under both simulations, with a smaller increase under A1B than 2C20 (0.004 and 0.007 mm d^{-1} or latent heat flux changes of 0.14 and 0.20 W m^{-2} under A1B and 2C20, respectively). These changes are equivalent to ~ 4 and 7% of the A1B and 2C20 INTVEG changes, respectively. Over AMZ, larger reductions in evaporation were found over land-only grid points in response to vegetation change.

Under A1B-INTVEG, we found large reductions in total annual evaporation over the Amazon and Southern Africa, smaller reductions over Southern Europe and mid USA, and increases in evaporation at high latitudes and in Central Asia (Fig. 12a). As for precipitation, small patchy signals in annual mean evaporation were generally found as a result of interactive vegetation (Fig. 12b, c). Under both A1B and 2C20, some small increases in evaporation were found over high latitudes. Increases in evaporation were found under both A1B and 2C20 in the Western Amazon (stronger under A1B), while decreases over the Eastern Amazon were stronger under A1B. Interactive vegetation led to a reduction in evaporation over the Arctic ocean under both A1B and 2C20 (Fig. 12d). As for precipitation, evaporation changes due to vegetation change under 2C20 were mostly a diminished signal of the A1B changes (Fig. 12d).

As for precipitation, changes in seasonal patterns of evaporation due to the vegetation effect were largely similar to the annual mean changes. In response to vegetation change, larger seasonal evaporation changes were found compared to the annual mean changes (Figs. S5 and S6) for some regions. For instance, the annual increases in evaporation over the Amazon were dominated by spring and summer changes while the increases over high latitudes mostly occurred during winter and spring. For the far West of Europe and North Eastern Russia (approximately 45 to 65° N and 25 to 75° E), there was a greater summer evaporation increase under 2C20 compared to A1B.

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4 Discussion

4.1 Climate simulations

The ECHAM5/MPI-OM A1B simulation of May (2008) gave a global average warming of 3.03 K for 2071–2100 relative to 1980–1999, which is larger than the average (2.80 K) across the AR4 models. This is consistent with the larger radiative forcing for a doubling of CO₂ in ECHAM5/MPI-OM compared to the average of the IPCC's fourth assessment report (AR4) models (4.01 and 3.75 W m⁻², respectively; May, 2008). The equivalent value for the standard version of HadCM3 (without an interactive carbon cycle) is 3.81 (Meehl et al., 2007). HadCM3 (without a carbon cycle) simulates global average warming of 2.72 for 2071–2100 relative to 1980–1999 under the A1B scenario, while the equivalent value for our HadCM3C INTVEG simulation is 4.33 K. Therefore, although the sensitivity of the standard HadCM3 model to CO₂ forcing is less than that of ECHAM5/MPI-OM, including the carbon cycle (and the associated increase in atmospheric CO₂ concentrations for the same emissions pathway) in HadCM3C significantly increases the models response to the same CO₂ concentration forcing, compared to ECHAM5/MPI-OM. This explains the greater temperature increases found in our A1B and 2C20 simulations compared to the results of May (2008).

Relatively strong increases in precipitation were found under our 2C20 simulation, compared to the A1B simulations. The reason for this is likely to be the sharp reduction in sulphate aerosols over the years 2021–2036 which was applied in the 2C20 scenario. In the simulations of May (2008), although the general increase in global mean precipitation under the 2C20 scenario was strong with respect to the magnitude of simulated global warming, the absolute value of global mean precipitation by 2100 in 2C20 remained below that under the A1B scenario. May (2008) also performed a simulation similar to 2C20 but where GHG concentrations were stabilised in 2030 and the stratospheric ozone concentrations and sulphate load used the 2030 values from the A1B scenario (2C30A). In 2C30A, the higher anthropogenic sulphate load generally weakened future changes in climate, particularly for precipitation, with the most pronounced

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impacts in the Northern Hemisphere and the tropics where the main sources of anthropogenic sulphate aerosols are located. Conversely, May (2008) observed a weaker increase in precipitation in his simulations with a higher anthropogenic sulphate load, consistent with a strengthened precipitation increase due to the removal of the anthropogenic sulphate load in our simulations, and as found by Brasseur and Roekner (2005).

May (2008) found that changes in global mean precipitation are not a linear function of the change in global mean temperature, as noted for the IPCC AR4 models (Meehl et al., 2007). In the AR4 models, the strongest (weakest) changes in global mean precipitation scaled with the corresponding changes in global mean temperature (“hydrological sensitivity”) occur for the scenario with the weakest (strongest) global warming. In his experiments with reduced GHG concentrations, May (2008) also found stronger relative increases in precipitation. In addition, the hydrological sensitivity is greater for aerosol forcing than for GHG forcing (Feichter et al., 2004).

4.2 Changes in vegetation types and terrestrial carbon storage

Key shifts in vegetation type in our simulations included increases in needleleaf and broadleaf tree cover over high latitudes, loss of broadleaf tree cover over the Amazon and replacement with C4 grass, and losses of C3 grass in high latitudes with replacement by shrubs, particularly over Eastern Siberia. These changes are in broad agreement with recent studies of future ecosystem distribution, which generally suggest replacement of herbaceous vegetation with trees at high latitudes and varying degrees of loss of the Amazon forest (Scholze et al., 2006; Sitch et al., 2008); although Jiang et al. (2011) find tropical forest expansion. For comparison, Hurtt et al. (2011) estimate that between 1500 and 2005, the global fractional area of cropland increased from 2 to 10 %; pasture from 2 to 22 %; primary vegetation decreased from 94 % to 34 %; secondary vegetation increased from 0 to 20 % and urban area increased from 0 to 0.4 %. Our changes in natural vegetation types were much smaller – generally of the order of 1–2 % of the global land area for any one PFT.

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A small overall increase in total terrestrial carbon storage (14 PgC) was found in our 2C20 simulation, in contrast to the decrease found for A1B (−43 PgC). This is consistent with findings from recent studies, which suggest the conversion of the present day land carbon sink into a source during the 21st century for stronger warming (Scholze et al., 2006). Similarly, all of the HadCM3C ensemble members studied by Booth et al. (2011, 2012) find a reduction in land carbon uptake under the A1B scenario. In agreement with Booth et al. (2011, 2012) and Friedlingstein et al. (2006), much of the difference in total carbon changes between our simulations was driven by changes in soil carbon (soil: −109 and −40 PgC; vegetation: 67 and 54, for A1B and 2C20, respectively). The C4MIP simulations (using the A2 scenario) find a wide spread in future global carbon stock changes (Friedlingstein et al., 2006) – between 1901–1931 and 2071–2100 the range of changes were 22 to 708, −16 to 348, and 38 to 475 PgC for total, soil and vegetation carbon, respectively. The range of changes between the 2090s and the present day across the HadCM3C ensemble of Booth et al. (2011) were −647 to 210 PgC for soil, −83 to 183 PgC for vegetation, and −632 to 393 PgC for total carbon. Our future total carbon changes under A1B are therefore within the range found by Booth et al. (2011) but outside that of the C4MIP models; differences between these studies again appear to be largely explained by differing soil responses.

Regionally, total carbon storage was reduced over the Amazon and increased over high latitudes. This is in broad agreement with recent studies (Sitch et al., 2008; Qian et al., 2009; Devaraju et al., 2011) – for instance the majority of C4MIP models locate reduction of land carbon uptake in the tropics and find an increase in high latitudes (Friedlingstein et al., 2006). The HadCM3C ensemble used by Booth et al. (2011) also found robust vegetation carbon increases in Northern high latitudes, and higher altitude lower latitude regions along with smaller regions of vegetation loss in Central America, Northern Brazil, the Kalahari and Crimea. Robust soil carbon losses were found in Central North America, Central America, Northern South America, continental Europe and Southern Africa; robust soil carbon gains were only found in the extreme North and for isolated high altitude regions.

Across the C4MIP models at Northern high latitudes ($>60^{\circ}\text{N}$), Qian et al. (2009) find an increase in total carbon stocks of $38 \pm 20\text{PgC}$ over 1901 levels by 2100 of which $17 \pm 8\text{PgC}$ comes from vegetation and $21 \pm 16\text{PgC}$ from soil (increases of 43 % and 8 % of the vegetation and soil pools, respectively). Qian et al. (2009) found that both CO_2 fertilisation and warming enhanced vegetation growth, and although the intense warming over the region enhanced decomposition, soil carbon storage continued to increase in 21st century due to increased litterfall. Our values (for land $>45^{\circ}\text{N}$) were similar for total carbon changes (38PgC), larger for vegetation carbon changes (61PgC) and of the opposite sign for soil carbon changes (-63PgC). The reason for the difference between our simulated changes and those of Qian et al. (2009) is likely to be mainly the difference in averaging area – Fig. 5a shows strong losses of soil carbon around 45°N and gains north of around 60°N . Kuhry (2010) describe simulations using the LPJ-GUESS DGVM for a region of Northeast European Russia, under the A1B scenario. In their simulations, the region was predicted to lose carbon due to future climate change in contrast to our findings, though the exact amount was strongly dependent the rate of forest disturbance and treeline advance. Carbon release from permafrost is also not included in the current simulations; under the RCP 8.5 scenario with the HadGEM2-ES earth system model, Burke et al. (2012) found a range of permafrost carbon release of $50\text{--}270\text{PgC}$ by 2100. Falloon et al. (2011) provide a more detailed discussion on the robustness of soil carbon changes for the Amazon region.

Booth et al. (2011) note that ecosystems will continue to respond to climate change for decades or centuries after climate has stopped changing (Jones et al., 2009). As a result, carbon losses in the short term due to tropical ecosystem loss may be compensated for in the longer term by slower increases in high latitudes (Jones et al., 2010). This also appears to be the case in our simulations and may explain the differences in carbon changes between the A1B and 2C20 scenario. For example, compared to A1B the smaller total carbon loss over the Amazon under the 2C20 stabilisation scenario is balanced against the stronger high latitude gain (driven in the short-term by enhanced

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growth, and areal expansion in the longer-term – Jones et al., 2010) resulting in an overall global gain in contrast to the global loss under A1B.

4.3 Impacts of vegetation change on future climate

Although there are few directly comparable studies to ours, qualitative comparisons can be made with relevant global and regional scale studies. In general, the impacts of vegetation change on future annual global climate in our study were small, particularly for the biophysical aspects. In addition, the strongest impacts were generally located in the regions of strongest vegetation change, with little evidence of impacts over broader areas. This is perhaps not surprising since the area changes in vegetation in our study were small by comparison to those in other studies.

Strengers et al. (2010) assessed twentieth century global climate-vegetation feedbacks using a coupled vegetation-climate model (not including a coupled carbon cycle). In agreement with our findings, anthropogenic land use change had a stronger effect on climate than natural vegetation responses, although the extent of natural vegetation changes over the twentieth century was small in comparison to our projected future changes and those of Scholze et al. (2006) and Sitch et al. (2008). In the study of Strengers et al. (2012), albedo was found to be an important driver, although evapotranspiration and cloud formation were equally important especially in the tropics. In their natural vegetation experiments, small changes in tree cover (mainly increases), led to reductions in albedo and increases in temperature, particularly over Siberia.

Jiang et al. (2011) used an atmosphere GCM (AGCM) and a coupled AGCM-terrestrial (equilibrium) biosphere model driven by an ensemble of AR4 model outputs under the A2 scenario to assess the impact of future vegetation change on global climate. Key differences in their changes in future vegetation distribution include the lack of Amazon forest dieback and stronger tropical forest increases in general. Their study also found a small impact of vegetation on climate change globally, but significant regional impacts. A warming of 0.1–1 K was found over continental Eurasia east of 60° E, mainly driven by albedo changes, and there was a reduction in precipitation over the

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Amazon. However unlike our study, their simulations did not include the impact of CO₂ on vegetation stomatal conductance. An additional factor not considered in our study, Strengers et al. (2010) or Jiang et al. (2011) may be vegetation down-regulation under elevated CO₂ which could reduce photosynthetic activity and leaf area index, increasing evapotranspiration, leading to an additional cooling effect which could reduce the overall warming impact of vegetation change (Bounoua et al., 2010).

The global-scale study of Swann et al. (2011) found that prescribed large scale afforestation in Northern Hemisphere mid latitudes (45 to 60° N) under the present day climate warmed the Northern Hemisphere and altered the Hadley circulation leading to a northward displacement of tropical rain bands. In general, they found small impacts of mid latitude afforestation on global temperature and CO₂ concentrations, but significant regional impacts, broadly agreeing with our findings. The greatest warming was found in water-limited regions (in contrast to our findings), and precipitation decreased over the Amazon and increased over the Sahel and Sahara. Equilibrium water vapour content increased over Northern Hemisphere mid latitudes by 6.9%, enhancing downward longwave radiation, larger than the changes in our simulations. Swann et al. (2011) showed that evapotranspiration increases outweighed precipitation increases in their simulations, indicating significant regional export of water vapour, which was not the case in our simulations. As noted above, vegetation changes in our study were much less extreme, and tree expansion was mainly located north of 45° N, which may explain the weaker climate changes observed in our simulations. Terrestrial carbon uptake in their study was 270 Pg; for comparison our study found an uptake of 38 and 45 PgC for the HIGHLAT region under the A1B and 2C20 simulations, respectively, although our estimates additionally include the impact of future climate change and the location and extent of change also differ. In Swann et al. (2011), the additional warming from vegetation change caused a loss of soil carbon at high latitudes (and elsewhere). We also found a loss of soil carbon and high latitudes under A1B but a small gain under 2C20, although our assessments additionally include future climate drivers. Overall, Swann

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et al. (2011) estimated that biogeochemical and biophysical feedbacks led to changes in global temperature of -0.4 to 0.1 °C.

Swann et al. (2010) investigated the regional impact of prescribed afforestation with deciduous trees at Northern Hemisphere high latitudes, finding that the top-of-atmosphere radiative imbalance from enhanced transpiration was up to 1.5 times greater than the albedo forcing. In their study, greenhouse warming from increases in atmospheric water vapour content melted sea ice, triggering a positive feedback via ocean albedo and evaporation. In our simulations, expansion of the needleleaf tree PFT was more widespread geographically than that of broadleaf trees, with increases of 2–3% in fractional coverage of the HIGHLAT area for both types. In addition, our climate-driven increases in forest area were much more modest than those of Swann et al. (2010). When fully leafed out, broadleaf trees have twice the albedo and 50–80% greater evapotranspiration rates than needleleaf trees (Swann et al., 2010), which, in combination with the less extreme afforestation rate may explain the much weaker change in evapotranspiration in our simulations. In addition, larger impacts on vegetation distribution may be anticipated following a longer period of adjustment to the new climate (Jones et al., 2009, 2010; Booth et al., 2011), so potentially larger impacts on evapotranspiration may occur in the longer-term. Our simulations also considered future, rather than present climate as in the experiments of Swann et al. (2010) and used different models, which will further alter our results.

Wramneby et al. (2010) used a regional climate model coupled to a DGVM to assess the impacts of changing vegetation patterns on future climate over Europe. Impacts were categorized into three “hotspots”: the Scandinavian mountains where reduced albedo from snow masking enhanced winter warming; Central Europe where a negative evapotranspiration feedback via stimulation of photosynthesis and plant growth due to elevated CO₂ concentrations mitigated warming; and Southern Europe where increased summer dryness reduced plant growth, reducing evapotranspiration and leading to a positive (warming) feedback. They found climate feedbacks over Europe to be small compared to the radiative forcing of increased CO₂ concentrations, but noted

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significant local, regional and seasonal effects. Although the patterns of winter warming (and albedo change) found in our simulations were similar to those of Wrarnbeby et al. (2010), we found warming in summer across most of Europe in contrast to the cooling found in their study for Central and Northern Europe.

Kuhry (2010) reports ECHAM5-MPI/OM-JSBACH simulations which applied the vegetation changes from our HadCM3C simulations, but only changing vegetation patterns for the CarboNorth project study region of Northeast European Russia while all other land points were kept constant. Temperature increases of approximately 0.3K per decade were found for most of the Arctic, whereas in the region where vegetation actually was changed, the temperature trend was roughly twice as large. This implies that the transition from tundra to taiga in the pan-Arctic domain has wider implications at least for the Northern Hemisphere and may additionally enhance the temperature increase by almost 20 %. This result corroborates findings by Dethloff et al. (2006), who found a large sensitivity over large parts of the Northern Hemisphere to slight changes in the albedo.

There is significant uncertainty in the impact of vegetation change on future climate. Potential sources of uncertainty include: uncertainties in future land use patterns (and how they interact with natural vegetation); differing vegetation model responses to future climate, uncertainties in future projected climate itself (Friedlingstein et al., 2006; Scholze et al., 2006; Denman et al., 2007; Fischlin et al., 2007; Sitch et al., 2008), and resulting uncertainties in biophysical and biogeochemical impacts on climate. In the latter sense, recent assessments of the impact of contemporary anthropogenic land cover change have found considerable variation in responses across models (Pitman et al., 2009; De Noblet-Ducoudre et al., 2012).

4.4 Implications for climate mitigation and adaptation

In our simulations, the effect of climate mitigation can be assessed by comparing changes under the 2C20 and A1B scenarios. For instance, comparing the INTVEG simulations, mitigation avoided over 2K of warming relative to pre-industrial climate

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(over 40% of the unmitigated increase), with almost twice this impact regionally. In contrast, because of the relatively strong reduction in sulphate aerosol concentrations, mitigation had a small impact on global precipitation trends resulting in a relatively strong increase.

Climate mitigation generally led to smaller gains in high latitude tree cover on the one hand, but a smaller loss of the Amazon forest on the other. Mitigation reduced the global loss of terrestrial carbon (mainly due to soil effects). As noted previously, mitigation reduced the total carbon loss over the Amazon which was balanced against the stronger high latitude increase resulting in an overall global gain in contrast to the global loss under business-as-usual. By 2100, high latitude total carbon gains were similar under both scenarios, but still changing, with the loss from soil generally outweighing the gain in vegetation storage. This was presumably because the greater warming under the A1B scenario benefits high latitude vegetation growth but also increases decomposition losses from soils. Overall, this results in a smaller declining trend in total carbon globally (and for high latitudes) under the climate stabilisation scenario at the end of the 21st Century.

Mitigation generally moderated the impact of vegetation change on future global and regional climate – for instance, by acting to reduce the (global and Amazon) decline in albedo, and reduce the increase in temperature. In addition, while the reduction in evaporation over the Amazon was reduced, the increase over high latitudes was stronger.

5 Conclusions

In our simulation under the 2C20 climate stabilisation scenario, atmospheric CO₂ concentrations were stabilised leading to a much smaller and slower global temperature increase by 2100, compared to the simulations under A1B emissions, avoiding over 2 K of warming relative to pre-industrial climate. Larger impacts were found regionally – compared to A1B, 2C20 avoids approximately 3.6 K and 3.9 K of warming over

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high latitudes and the Amazon, respectively (for 2071–2100 relative to 1861–1890). An increase in global mean precipitation was found under both scenarios, with similar increases under the 2C20 scenario compared to A1B, likely due to the sharp reduction in sulphate aerosols implied under 2C20. Regionally, the strong reduction in precipitation found under A1B over the Amazon was moderated under 2C20 while the increase found over high latitudes under A1B was slightly weakened under 2C20. Similar increases in global mean evaporation were found under our 2C20 and A1B simulations, with a reduction over the Amazon (weaker in 2C20) and an increase over high latitudes.

In our simulations with interactive vegetation, only small changes in total vegetation fraction were found – mostly a reduction over the Amazon, and increases over East Asia and the Southern Rockies. Key shifts in vegetation type included increases in needleleaf and broadleaf tree cover over high latitudes, loss of broadleaf tree cover over the Amazon and replacement with C4 grass, and losses of C3 grass in high latitudes with replacement by shrubs, particularly over Eastern Siberia. The global reductions in albedo found in our simulations were greater under A1B than 2C20, with the most significant reductions occurring over high latitudes and increases over the Amazon.

A small overall increase in total terrestrial carbon storage was found in our 2C20 simulation, in contrast to the decrease found for A1B, with much of the difference being driven by changes in soil carbon. Regionally, total carbon storage was reduced over the Amazon and increased over high latitudes. Relative to A1B, 2C20 reduced the total carbon loss over the Amazon, but amplified the carbon gain over high latitudes. Since carbon storage did not change in our simulations with fixed vegetation, these changes represent the impact of interactive vegetation on carbon storage. Globally, and over high latitudes, interactive vegetation amplified reductions in albedo and increases in albedo over the Amazon, with smaller impacts under 2C20 than under A1B.

Relatively small global impacts of interactive vegetation on most annual mean surface climate variables were found under both A1B and 2C20, with generally smaller impacts under 2C20. For instance, the slight increase in global mean temperature found under A1B (0.24 K) was greatly reduced under 2C20 (0.15 K). Larger impacts were

observed regionally and seasonally and over land. For instance, additional warming of ~ 1 K was found over Siberia and over the Amazon under A1B, with smaller changes under 2C20; over land only grid points, global temperature changes due to interactive vegetation were 0.43 and 0.28 K under A1B and 2C20, respectively.

The main regional impact of interactive vegetation change on precipitation was the strong reduction found over the Amazon (under A1B), which was much smaller under 2C20. Precipitation changes due to interactive vegetation changes were mostly dominated by changes in convective precipitation. Under A1B, some small increases in evaporation were found over high latitudes and the Western Amazon, with reductions over the Eastern Amazon, with much smaller impacts under 2C20 (and no widespread reduction over the Eastern Amazon). Over the Amazon, the warming influence of vegetation change in our simulations was driven by the loss of tree cover and increase in bare soil, which led to a reduction in (summer) evaporation (and latent heat flux), offsetting the slight increase in albedo. At high latitudes, warming due to vegetation change resulted from increasing vegetation cover (grasses, trees and shrubs), where the reduced albedo (particularly during winter and spring) appears to have offset any small evaporation increases.

Climate mitigation generally reduced the impact of vegetation change on future global and regional climate in our simulations. Our study therefore suggests that there is a need to consider both changes in the future land carbon cycle (due to their impact on climate) and vegetation changes (since they will also affect climate via both biogeochemical and biophysical effects) in climate adaptation and mitigation decision making. While the impacts of vegetation change on future climate are relatively small globally, larger impacts may result at local to regional scales and on seasonal timescales with implications for adaptation planning. In addition since encouraging forestation and avoiding deforestation form part of the current portfolio of climate mitigation measures, there is a need to better understand future climate vulnerabilities (or opportunities) for forests and land carbon sinks in order to avoid unintended consequences and risks.

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Table 1. Climate simulations.

Time period			CO ₂ concentrations	Other GHGs	Carbon cycle	Vegetation distribution
A1B	1859–2100	(prescribed)	Prescribed according to Nakicenovic et al. (2000)	A1B	Off	Fixed
A1B (C cycle)	1859–2100	INTVEG	From model carbon cycle using A1B emissions	A1B	Fully interactive	TRIFFID Dynamic vegetation model
		FIXVEG	From A1B-INTVEG	A1B	Off	From 1860 A1B-INTVEG
2C20	2020–2100	INTVEG	2020 A1B-INTVEG	2C20	Uncoupled land carbon cycle	TRIFFID Dynamic vegetation model
		FIXVEG	2020 A1B-INTVEG	2C20	Off	From 2020 A1B-INTVEG

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Table 2. Changes in selected climate and ecosystem variables.

Change in ecosystem /climate variable		A1B (2080s*-1870s**)		2C20 (2080s-1870s)		Impact of interactive vegetation***	
		INTVEG	FIXVEG	INTVEG	FIXVEG	A1B	2C20
Soil carbon stocks (PgC) / concentration (kgCm ⁻²)	Global	-109 /-0.74	-	-40/-0.27	-	-	-
	High latitude [†]	-23/-0.56	-	4/0.09	-	-	-
	Amazon [‡]	-30/ -4.32	-	-16/-2.31	-	-	-
Vegetation carbon stocks (PgC)/concentration (kgCm ⁻²)	Global	67/0.45	-	54/0.37	-	-	-
	High latitude [†]	61/1.48	-	41/0.99	-	-	-
	Amazon [‡]	-41/-5.90	-	-22/-3.16	-	-	-
Total carbon stocks (PgC) / concentration (kgCm ⁻²)	Global	-43/-0.29	-	14/0.10	-	-	-
	High latitude [†]	38/0.91	-	45/1.08	-	-	-
	Amazon [‡]	-71/-10.22	-	-22/-5.47	-	-	-
Average albedo	Global	-0.018	-0.016	-0.012	-0.011	-0.002	-0.006
	Global land	-0.017	-0.010	-0.012	-0.008	-0.006	-0.003
	High latitude [†]	-0.062	-0.051	-0.041	-0.036	-0.011	-0.005
	High latitude land	-0.047	-0.028	-0.032	-0.022	-0.019	-0.01
	Amazon [‡]	0.010	-0.00003	0.002	-0.002	0.01	0.004
	Amazon land	0.010	~0	0.003	-0.002	0.01	0.005
Average temperature (K)	Global	5.12	4.87	2.96	2.80	0.24	0.15
	Global land	7.45	7.01	4.07	3.79	0.43	0.28
	High latitude [†]	8.33	7.97	4.77	4.51	0.35	0.26
	High latitude land	8.83	8.24	4.85	4.40	0.58	0.44
	Amazon [‡]	8.59	8.02	4.71	4.41	0.56	0.30
	Amazon land	9.52	8.88	5.17	4.84	0.64	0.33
Average total precipitation (mm d ⁻¹)	Global	0.13	0.11	0.13	0.11	0.02	0.02
	Global land	-0.08	-0.08	0.02	~0	-0.005	0.002
	High latitude [†]	0.29	0.29	0.21	0.19	0.0004	0.02
	High latitude land	0.20	0.19	0.18	0.16	0.006	0.02
	Amazon [‡]	-2.52	-2.18	-1.11	-0.96	-0.35	-0.16
	Amazon land	-2.55	-2.33	-1.11	-1.03	-0.32	-0.08
Average evaporation (total surface moisture flux, mm d ⁻¹)/latent heat flux (Wm ⁻²)	Global	0.13/3.78	0.12/3.31	0.13/3.63	0.11/3.12	0.02/0.47	0.02/0.52
	Global land	-0.25/-7.18	-0.26/-7.60	-0.05/-1.52	-0.07/-2.00	0.01/0.42	0.02/0.48
	High latitude [†]	0.10/2.95	0.10/2.81	0.10/2.85	0.09/2.66	0.004/0.14	0.007/0.20
	High latitude land	-0.02/-0.78	-0.03/-0.89	0.05/1.45	0.05/1.48	0.003/0.11	-0.001/-0.03
	Amazon [‡]	-1.37/-39.60	-1.35/-39.03	-0.58/-16.92	-0.59/-16.97	-0.02/-0.57	0.001/0.04
	Amazon land	-1.68/-48.77	-1.64/-47.37	-0.76/-21.98	-0.75/-21.66	-0.05/-1.40	-0.01/-0.32

* 2071–2100.

** 1861–1890.

*** INTVEG (2080s–1870s)-FIXVEG (2080s–1870s).

† HIGHLAT: 45–80° N.

‡ AMZ: 40–70° W, 15° S-5° N.

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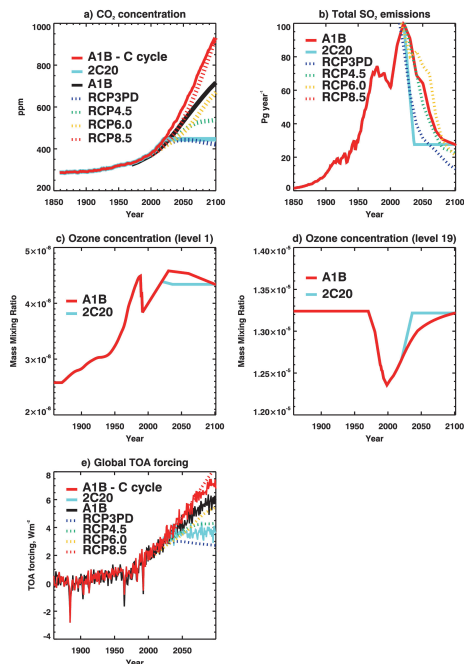


Fig. 1. Global average timeseries of forcings applied in the HadCM3C climate model simulations for the A1B and 2C20 scenarios: **(a)** CO₂ concentration; **(b)** total SO₂ emissions; **(c)** ozone concentration (lowest atmospheric level); **(d)** ozone concentration (top atmospheric level); **(e)** top of the atmosphere radiative forcing. Values for the Representative Concentration Pathways (RCPs) applied in the IPCC’s fifth assessment report are also shown in **(a)**, **(b)** and **(e)**. The A1B – C cycle simulation corresponds to our INTVEG and FIXVEG simulations; A1B alone is the standard “prescribed” concentration scenario.

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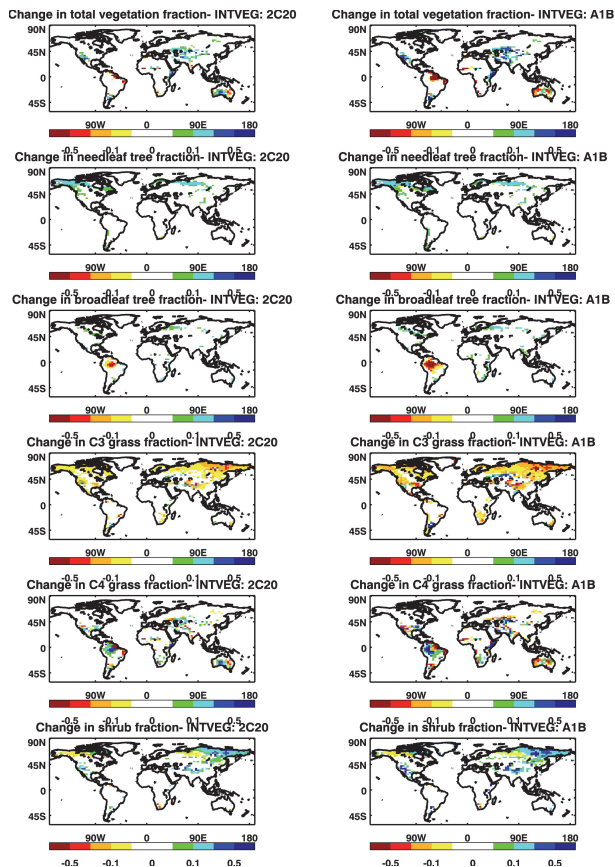


Fig. 2. Changes in vegetation fractions under the A1B and 2C20 INTVEG simulations.

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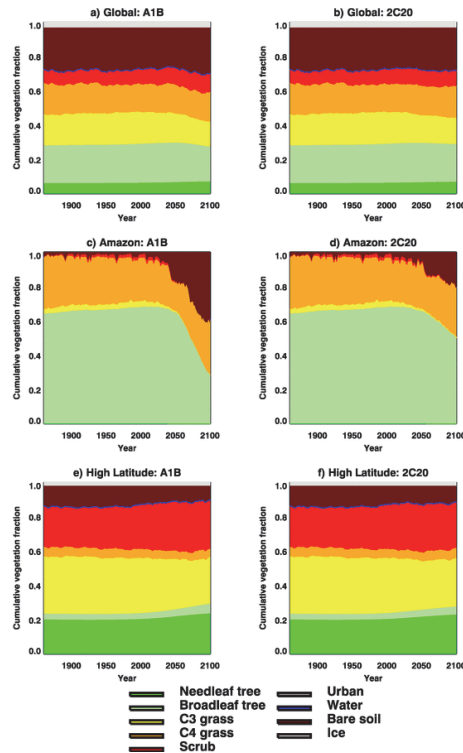


Fig. 3. Timeseries of relative contribution of surface types in HadCM3C simulations under the A1B scenario with interactive vegetation: **(a)** global mean; **(b)** Amazon region, and **(c)** high latitude region. Regions are defined in Table 2.

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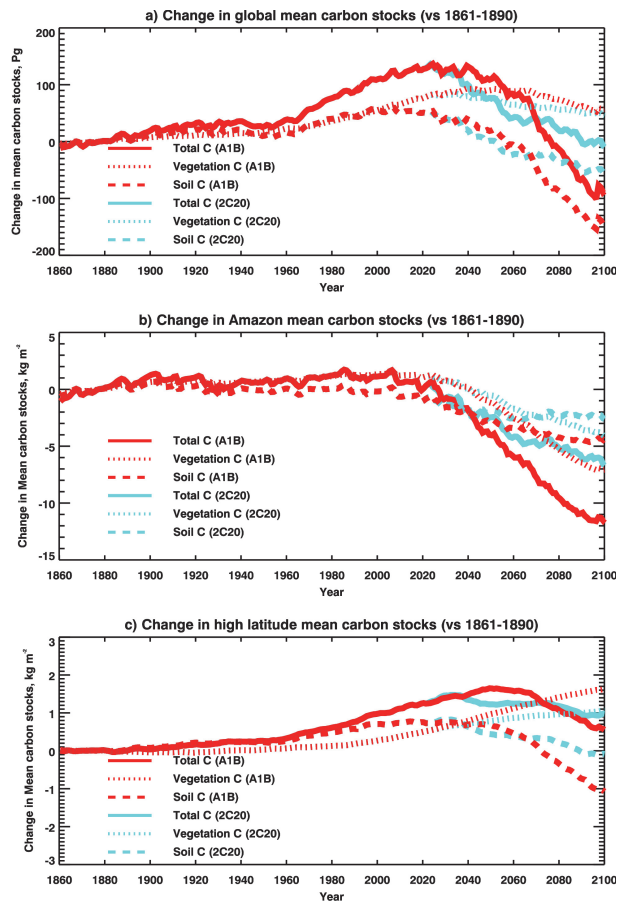


Fig. 4. Timeseries of changes in terrestrial carbon storage in HadCM3C simulations under A1B and 2C20 scenarios with interactive and fixed vegetation: **(a)** global mean; **(b)** Amazon region, and **(c)** high latitude region. Regions are defined in Table 2.

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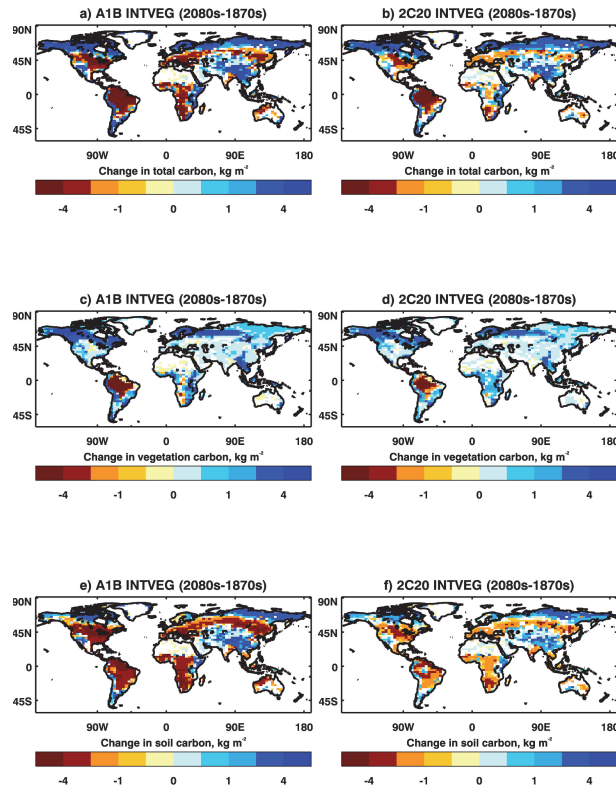


Fig. 5. Change in thirty year mean carbon storage in HadCM3C simulations with interactive vegetation (2080s–1860s): total carbon **(a)** A1B scenario, **(b)** 2C20 scenario; vegetation carbon **(c)** A1B scenario, **(d)** 2C20 scenario; soil carbon **(e)** A1B scenario, **(f)** 2C20 scenario. Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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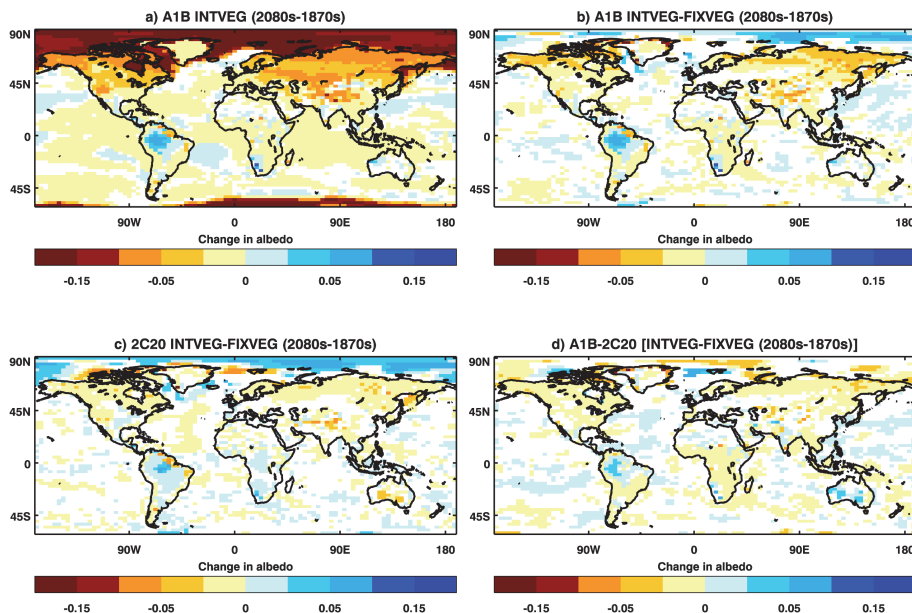


Fig. 6. Change in thirty year mean albedo in HadCM3C simulations (2080s–1860s): **(a)** A1B scenario with interactive vegetation; **(b)** difference between A1B simulations with interactive and fixed vegetation; **(c)** difference between 2C20 simulations with interactive and fixed vegetation; **(d)** difference in the impact of interactive vegetation between A1B and 2C20 simulations. Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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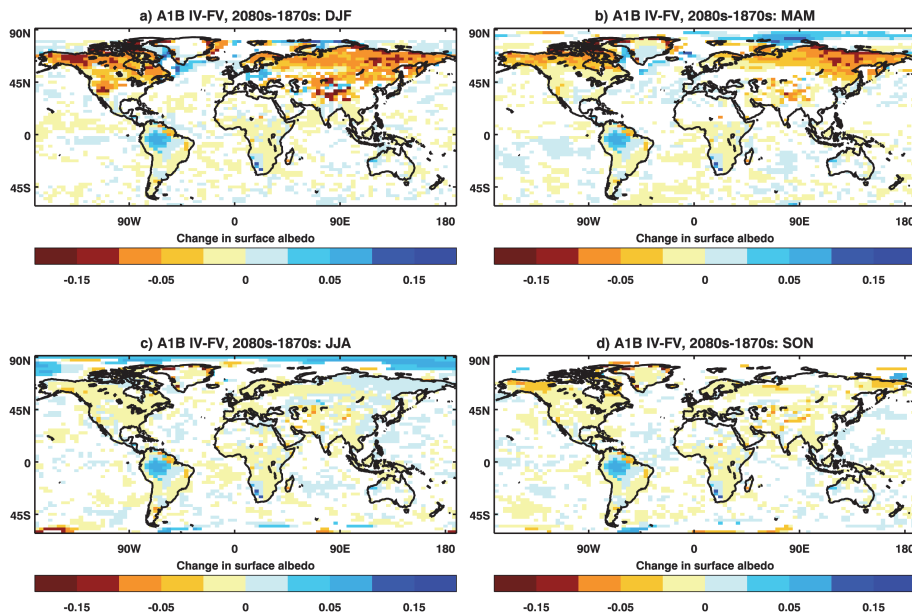


Fig. 7. Change in thirty year seasonal mean albedo in HadCM3C simulations scenario (2080s–1860s) showing difference between A1B simulations with interactive and fixed vegetation, for **(a)** winter (December–February), **(b)** spring (March–May), **(c)** summer (June–August) and **(d)** autumn (September–November). Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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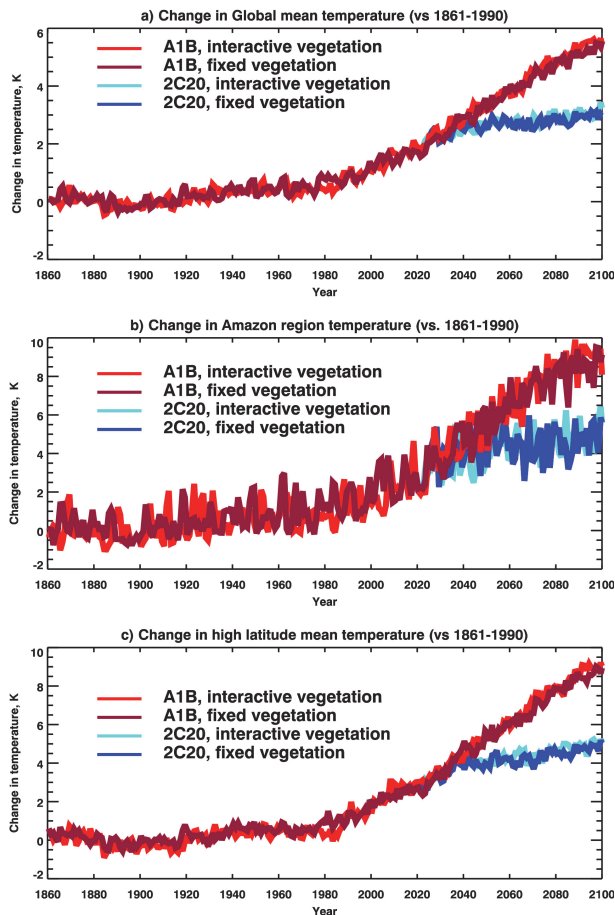


Fig. 8. Timeseries of changes in annual average temperature in HadCM3C simulations under A1B and 2C20 scenarios with interactive and fixed vegetation: **(a)** global mean; **(b)** Amazon region, and **(c)** high latitude region. Regions are defined in Table 2.

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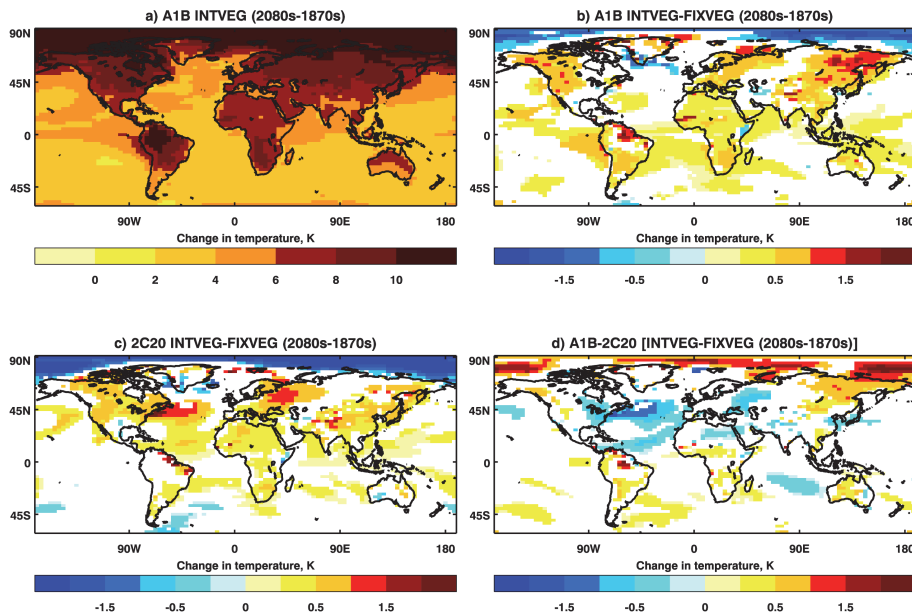


Fig. 9. Change in thirty year mean temperatures in HadCM3C simulations (2080s–1860s): **(a)** A1B scenario with interactive vegetation; **(b)** difference between A1B simulations with interactive and fixed vegetation; **(c)** difference between 2C20 simulations with interactive and fixed vegetation; **(d)** difference in the impact of interactive vegetation between A1B and 2C20 simulations. Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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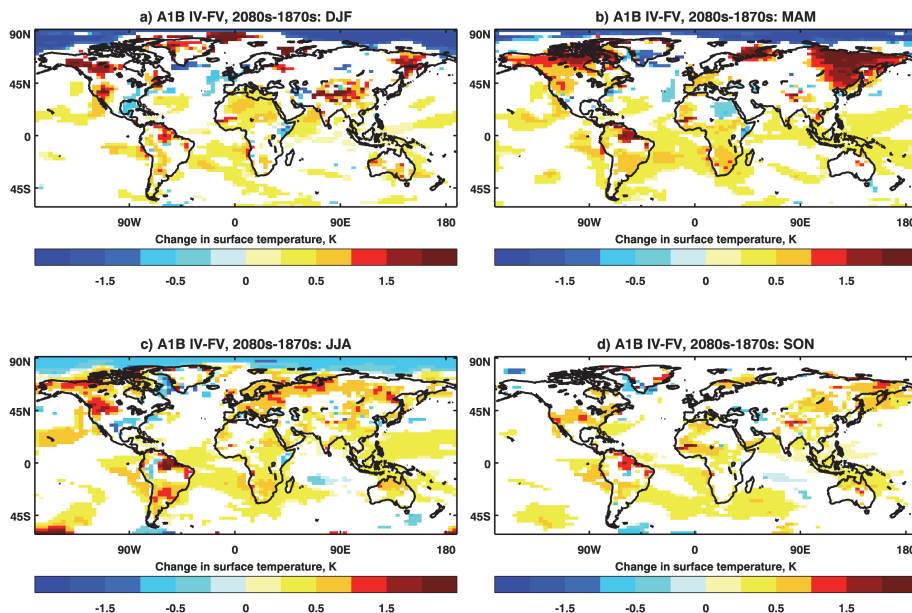


Fig. 10. Change in thirty year seasonal mean temperatures in HadCM3C simulations scenario (2080s–1870s) showing difference between A1B simulations with interactive and fixed vegetation, for **(a)** winter (December–February), **(b)** spring (March–May), **(c)** summer (June–August) and **(d)** autumn (September–November). Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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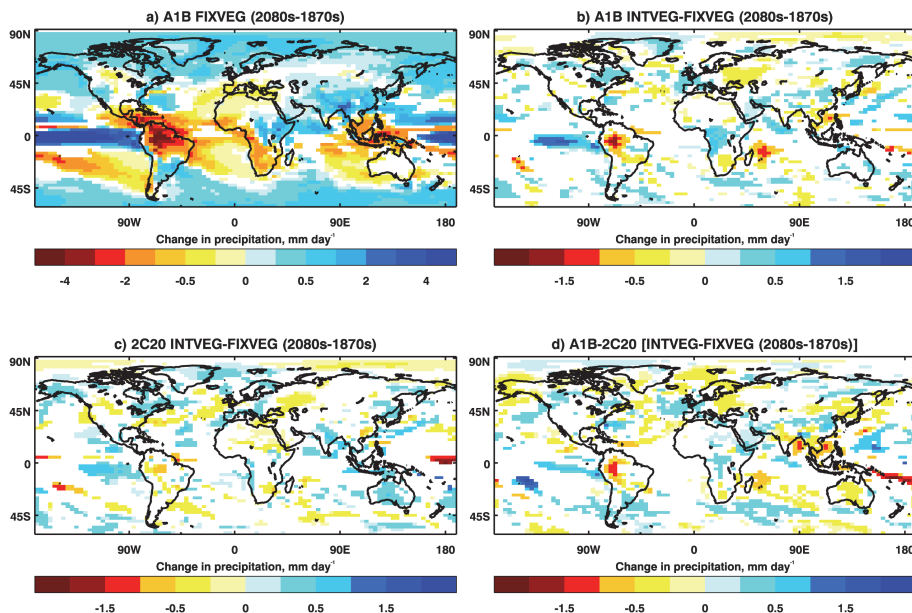


Fig. 11. Change in thirty year mean precipitation in HadCM3C simulations (2080s–1870s): **(a)** A1B scenario with interactive vegetation; **(b)** difference between A1B simulations with interactive and fixed vegetation; **(c)** difference between 2C20 simulations with interactive and fixed vegetation; **(d)** difference in the impact of interactive vegetation between A1B and 2C20 simulations. Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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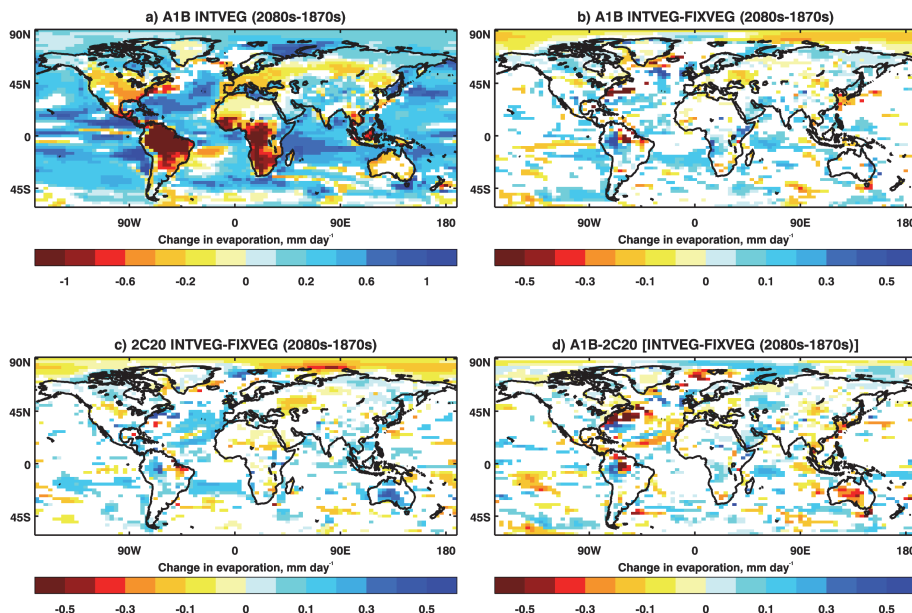


Fig. 12. Change in thirty year mean evaporation in HadCM3C simulations (2080s–1870s): **(a)** A1B scenario with interactive vegetation; **(b)** difference between A1B simulations with interactive and fixed vegetation; **(c)** difference between 2C20 simulations with interactive and fixed vegetation; **(d)** difference in the impact of interactive vegetation between A1B and 2C20 simulations. Changes smaller than two standard deviations from the control simulation are masked out (white areas).

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