

Global climate change and soil carbon stocks; predictions from two contrasting models for the turnover of organic carbon in soil

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Abstract

Enhanced release of CO₂ to the atmosphere from soil organic carbon as a result of increased temperatures may lead to a positive feedback between climate change and the carbon cycle, resulting in much higher CO₂ levels and accelerated global warming. However, the magnitude of this effect is uncertain and critically dependent on how the decomposition of soil organic C (heterotrophic respiration) responds to changes in climate. Previous studies with the Hadley Centre's coupled climate–carbon cycle general circulation model (GCM) (HadCM3LC) used a simple, single-pool soil carbon model to simulate the response. Here we present results from numerical simulations that use the more sophisticated 'RothC' multipool soil carbon model, driven with the same climate data.

The results show strong similarities in the behaviour of the two models, although RothC tends to simulate slightly smaller changes in global soil carbon stocks for the same forcing. RothC simulates global soil carbon stocks decreasing by 54 Gt C by 2100 in a climate change simulation compared with an 80 Gt C decrease in HadCM3LC. The multipool carbon dynamics of RothC cause it to exhibit a slower magnitude of transient response to both increased organic carbon inputs and changes in climate. We conclude that the projection of a positive feedback between climate and carbon cycle is robust, but the magnitude of the feedback is dependent on the structure of the soil carbon model.

Keywords: soil carbon dynamics, climate change, carbon cycle feedbacks

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Introduction

The possibility of a reduction in global soil carbon under climate warming has been raised by a number of authors (Jenkinson *et al.*, 1991; Schimel *et al.*, 1994; Kirschbaum, 1995). Recently such soil carbon losses have manifested themselves in coupled climate–carbon cycle projections as potentially strong positive feedbacks on climate change (Cox *et al.*, 2000; Friedlingstein *et al.*, 2001; Jones *et al.*, 2003b). Previous studies used the Hadley Centre's coupled climate–carbon cycle general circulation model (GCM) (HadCM3LC, Cox *et al.*, 2001): a version of the Hadley Centre's third generation climate model HadCM3 (Gordon *et al.*, 2000) with lowered ocean horizontal resolution (2.5° × 3.75°) coupled to terrestrial

and ocean carbon cycle models (Top-down Representation of Interactive Follage and Flora Including Dynamics, Cox, (2001), and HadOCC, Palmer & Totterdell (2001), respectively). In these simulations, increased heterotrophic respiration because of rising temperatures during the 21st century exceeded enhanced biospheric uptake as a result of elevated CO₂ levels. As a result, the rate of increase of atmospheric CO₂ and hence the rate of climate change were accelerated.

Soil carbon stocks were found to decrease across most of the globe, even in areas where organic carbon input from vegetation had increased (Jones *et al.*, 2003b), despite the carbon stock having increased steadily during the 20th century. In the studies with HadCM3LC (Cox *et al.*, 2000; Jones *et al.*, 2003b), drying of the Amazon basin as a result of climate change caused a dieback of the Amazon forest and strong reduction in the organic carbon input to the soil, also

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resulting in soil carbon losses in this region, although globally, the Amazon dieback accounted for just 11% of the climate-driven carbon losses (Cox *et al.*, 2004).

The magnitude, timing and even the sign of the feedback between climate and the carbon cycle will depend critically on the response of soil carbon to climate changes (Cox *et al.*, 2001; Friedlingstein *et al.*, 2003; Jones *et al.*, 2003a). This response is highly uncertain (Kirschbaum, 1995; Giardina & Ryan, 2000; Melillo *et al.*, 2002). Although it is possible to a degree to infer the sensitivity of heterotrophic respiration to short-term environmental changes by examining the observed record of atmospheric CO₂ in response to El Niño events and volcanic eruptions (Jones & Cox, 2001; Jones *et al.*, 2003a) or observed monthly weather anomalies (Raich *et al.*, 2002), this does not tell us how soil carbon may behave over the timescale of the next century. The soil carbon response of the IPSL-coupled climate-carbon cycle simulations (Friedlingstein *et al.*, 2001) differs from the HadCM3LC response. Under increased CO₂, but in the absence of climate change, the IPSL model simulates more accumulation of carbon in vegetation than in soil, whereas HadCM3LC simulates more accumulation in soil than vegetation, in agreement with the range of terrestrial biosphere models discussed by Kicklighter *et al.* (1999). This difference is seen as the key contributor to the different magnitude of positive feedback simulated by the two models when climate change is considered (Friedlingstein *et al.*, 2003). It is unlikely that differences in the soil carbon models are the main cause of these differences, but the inclusion in the IPSL model of multiple soil carbon pools with different timescales of decomposition may at least contribute to the difference (Berthelot *et al.*, 2002).

Previous work has considered the use of multipool models of soil carbon (Schimel *et al.*, 1994; McGuire *et al.*, 1995; Post *et al.*, 1996; Trumbore, 2000; Telles *et al.*, 2003). All such studies agree on the importance of the inclusion of multiple pools in order to represent the heterogeneous nature of the different turnover times within the soil. They show that soil carbon stocks are largely determined by the slow turnover pools but the fluxes are determined by the fast turnover pools (Schimel *et al.*, 1994; Trumbore, 2000). Hence, an attempt to simulate transient soil carbon behaviour with a single pool and just one turnover rate is not possible – it will overestimate the rate of response because large changes to the total stock will require changes to the slow pool (Telles *et al.*, 2003). Single-pool models can represent the same equilibrium changes (providing all the pools have the same temperature sensitivity), but not the transient response to them (Schimel *et al.*, 1994; Telles *et al.*, 2003). A recent study

with RothC (Gu *et al.*, 2004) demonstrated that the inclusion of a fast turnover pool of soil carbon could dramatically alter the magnitude of observed respiration in response to seasonal temperature changes because of the seasonality of the organic carbon inputs, a result corroborated experimentally by Yuste *et al.* (2004).

Jenkinson *et al.* (1991) used the RothC model to simulate the impact of global warming on soil carbon, but did not include the impact of changes to organic carbon inputs to the soil. They found substantial release of carbon from the soil in response to the elevated temperatures and postulated that this could cause a positive feedback onto climate. In other experiments, McGuire *et al.* (1995) saw decreased soil carbon caused by increased temperature, but the magnitude is much smaller and they conclude that substantial feedbacks as a result are unlikely. Post *et al.* (1996) took the next step and simulated soil carbon changes for both changing temperature and organic carbon inputs. Using RothC (but a different model of vegetation productivity to HadCM3LC), they found similar decreases in the case of temperature forcing alone. By contrast, in the case of increased organic carbon inputs as a result of increased net primary productivity (NPP), the soil carbon remained a sink (i.e. increased inputs dominated over increased respiration). This demonstrates that soil carbon response is a fine balance between potentially large and competing effects. The studies using HadCM3LC (Cox *et al.*, 2000; Jones *et al.*, 2003b) include the feedback of terrestrial biosphere carbon changes onto atmospheric CO₂ and simulate much greater temperature and CO₂ changes than considered by Post *et al.* (1996).

Soil carbon is modelled within HadCM3LC using a single pool with a single decay rate and takes no account of input quality. Hence, it is not able to simulate the dynamics of the different classes of soil carbon. An important question to answer is: how would the inclusion of multipool carbon dynamics affect the simulation of a positive feedback between climate change and the carbon cycle? Our aim in this paper is to test this by using RothC (Jenkinson, 1990; Coleman & Jenkinson, 1999), a well tested multi-compartment model of soil organic carbon dynamics, driven by the climate output from HadCM3LC.

Model description

The two soil carbon models used in this study differ in several ways. The HadCM3LC soil carbon model is based on a single pool of soil carbon with a single first-order decay rate dependent on temperature and soil moisture. A decay rate of 0.15 year⁻¹ is predicted at

25 °C and optimal soil moisture, which is dependent on soil hydrological properties (Cox, 2001). This rate is then modified for other temperature and moisture conditions. The specific respiration is modified by a 'Q₁₀' function of soil temperature (Raich & Schlesinger, 1992):

$$F_{\text{HadCM3LC}} = Q_{10}^{(T_s - 10)/10}$$

The model's value of $Q_{10} = 2.0$ is constant in time and space and has been found to produce a good fit to the observed global mean sensitivity of respiration to temperature on an interannual timescale (Jones & Cox, 2001). The respiration rate is also modified by a function of soil moisture (McGuire *et al.*, 1992) which is low for very dry or wet soils and peaks at an optimal soil moisture which is dependent on soil type. The optimal soil moisture is defined as a function of the saturation and wilting points for each soil.

The RothC soil carbon model (Jenkinson, 1990; Coleman & Jenkinson, 1999) has four active soil carbon pools: decomposable plant material (DPM), resistant plant material (RPM), microbial biomass (BIO) and humified organic matter (HUM), with associated first-order decay rates of 10.0, 0.3, 0.66 and 0.02 year⁻¹ at 9.25 °C (for comparison, the decay rate of HadCM3 at this temperature is 0.052 year⁻¹). Organic carbon inputs are apportioned between the two input pools, DPM and RPM, in a ratio that depends on vegetation type. Inputs from arable crops or grassland are assumed to be more readily decomposable (a higher proportion of DPM), whereas a higher proportion of tree input is assumed to be more resistant to decomposition (lower proportion of DPM). Each of the four pools decays at its own rate, releasing some CO₂ to the atmosphere and also feeding C into the BIO and HUM pools. The rate of decay for each pool is modified by a function, F_{RothC} , of monthly mean air temperature, T_a :

$$F_{\text{RothC}} = \frac{47.9}{1 + e^{\left(\frac{106}{T_a + 18.3}\right)}}$$

This function exhibits lower sensitivity to temperature changes at high temperatures than at low temperatures (i.e. it has a lower effective Q_{10} value at higher temperatures than at lower temperatures, see Fig. 1). This behaviour is widely believed to be more realistic than a constant Q_{10} for all temperatures (Lloyd & Taylor, 1994; Reichstein *et al.*, 2003). The decomposition rate is also modified by a function of soil moisture deficit (SMD), where SMD is inferred from the difference between precipitation and open pan evaporation. Decomposition is retarded in dry soils but the rate modifier is equal to unity when SMD is below 44% of the maximum deficit, which is calculated as a function of the soil's clay content (Coleman & Jenkin-

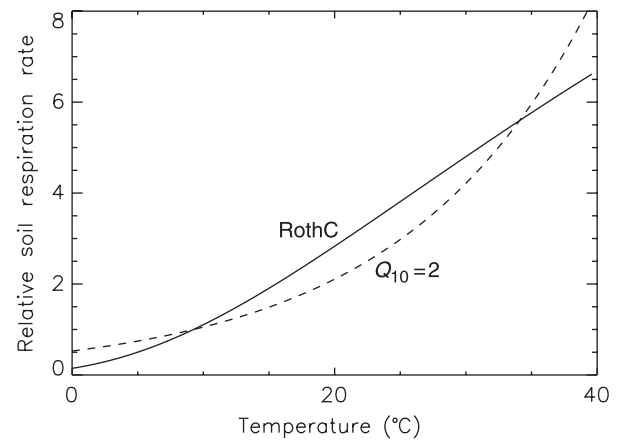


Fig. 1 Temperature sensitivity of specific soil respiration rate assumed in TRIFFID with $Q_{10} = 2$ (function of soil temperature, F_{HadCM3LC} : dashed line), and RothC (function of air temperature, F_{RothC} : solid line). This factor is used to scale the specific respiration rate at a reference temperature.

son, 1999). A further modification is made depending on whether vegetation is present: the rate is reduced where vegetation is growing (Sommers *et al.*, 1981; Sparling *et al.*, 1982; Jenkinson *et al.*, 1987). RothC can include the effects of land-management practices (such as the application of farmyard manure), but this aspect is not considered in this study. In addition to the four active pools mentioned above, RothC also contains a pool of carbon assumed to be inert organic matter (IOM), often about 10% of total soil carbon. This inert pool is essential to model the radiocarbon age of soil organic matter, but is unnecessary for the present study.

The different forms of temperature sensitivity of the two models is in addition to the fact that HadCM3LC uses the soil temperature of the upper 10 cm of soil whereas RothC uses surface air temperature. As a result of the higher variability of air temperature (i.e. soil temperature is damped and varies more slowly) effective Q_{10} values for air temperature are lower. Raich & Potter (1995) suggest a Q_{10} value of 2.0 for soil temperature that is roughly equivalent to 1.7 for air temperature. This difference should be born in mind when considering Fig. 1. Zhuang *et al.* (2003) also state that consideration of soil temperature rather than air temperature can have an impact on soil carbon evolution, although the effect is strongest for the seasonal cycle of respiration rather than long-term trends.

In both models, carbon input to the soil from vegetation turnover is treated as a single organic carbon flux: above/below-ground carbon inputs are not separated by either model.

A high clay content has been found to inhibit soil respiration (Schimel *et al.*, 1994; Post *et al.*, 1996; Telles *et al.*, 2003). Although both models include the effect of soil texture on soil moisture (and hence indirectly on respiration), only RothC takes into account the direct effect of soil texture on respiration (Coleman & Jenkinson, 1999).

Both models were equilibrated at an initial state using the climate from the first decade (1860s) of the climate change simulation. Estimated global totals of soil carbon to a depth of 1 m (excluding the inert organic matter pool) are about 1350 GtC (Prentice *et al.*, 2001) comprising 300 GtC of detritus and 1050 GtC of modified soil carbon. Both models simulate slightly lower values than this: HadCM3LC simulates a total of 1180 GtC (in a single pool), and RothC 950 GtC of which 220 GtC is detritus (DPM and RPM) and 730 GtC is BIO and HUM. The global mean NPP is 61 GtC yr⁻¹ – very close to recent best estimates of present day NPP of about 60 GtC yr⁻¹ (Prentice *et al.*, 2001).

Experimental design

The two soil carbon models were used in experiments to assess the sensitivity of soil carbon to climate change with and without multipool soil carbon dynamics. Output climate forcing data and plant carbon inputs from a climate change simulation were used to drive the two models (see Fig. 2). In the original coupled climate carbon cycle experiments (Cox *et al.*, 2000; Jones

et al., 2003b), accumulation and release of soil carbon directly affected the atmospheric CO₂ concentration and hence climate. In the present work, the output of the climate model was used to drive both of the soil carbon models in an ‘off-line’ manner – in other words, there was no feedback between soil carbon changes and climate. Performing off-line runs in this way makes it possible to vary or hold constant some of the components of the forcing data to examine which factors are more important in determining the evolution of soil carbon.

The offline runs were performed using monthly mean output from the GCM averaged over a decade for each month. Hence, there is no interannual variability in these experiments. A time step of 1 month was used successfully by Raich & Potter (1995) and Reichstein *et al.* (2003). The model simulates naturally occurring changes in vegetation, but disturbance because of agriculture is held fixed at present-day levels and no attempt is made to include the effects of changes in anthropogenic land use.

The fully coupled and off-line versions of HadCM3LC were compared to ensure that the offline methodology could correctly recreate the fully coupled results. Figure 3 compares the global total and the regional pattern of soil carbon changes from the fully coupled run and the offline run. It shows the averaged monthly forcing in the off-line version of the model can reproduce the behaviour seen when it is a fully integrated component of the climate model.

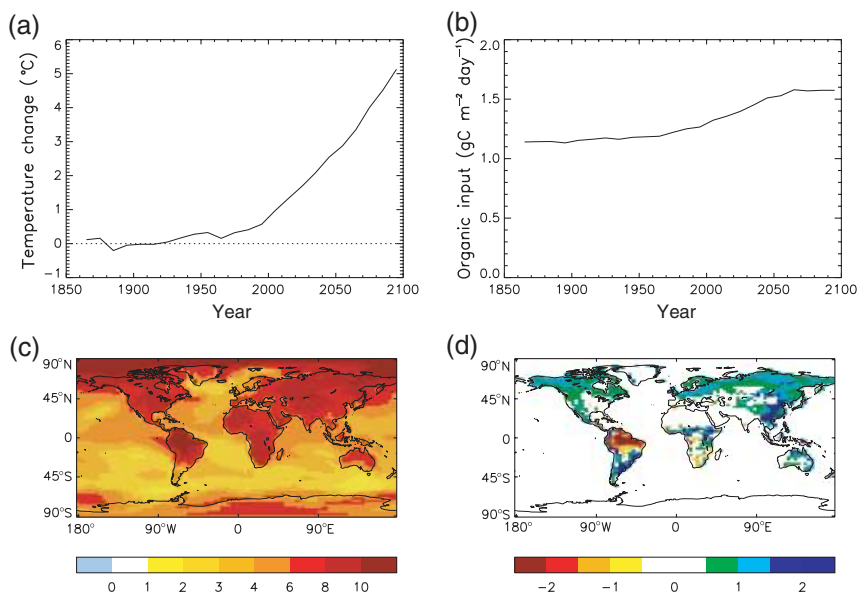


Fig. 2 Evolution of the forcing data during the course of the simulation. (a) Changes in global mean surface air temperature; (b) global mean rate of organic carbon input to the soil (in gC m⁻² day⁻¹); (c) map of surface air temperature changes (K): 1860 – 2100; (d) map of changes in organic carbon input (gC m⁻² day⁻¹): 1860–2100.

Further experiments were performed for a number of points representative of regions that exhibit different soil carbon behaviour. Temperature, soil moisture and organic carbon inputs were varied separately for these points. The points chosen, and the reason for the choice, are summarized in Table 1. The resulting simulation of soil carbon shows how each model responds to the separate temperature, moisture and vegetation input forcing.

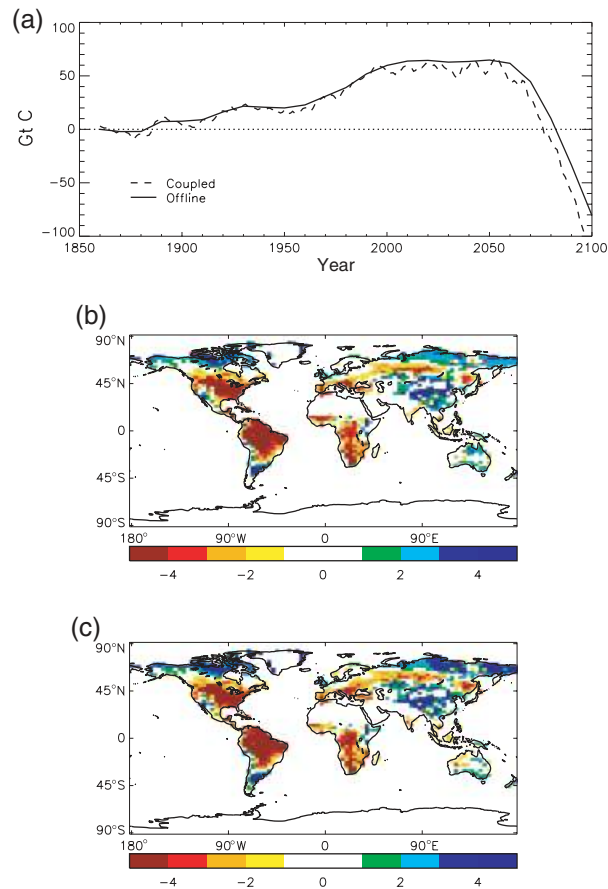


Fig. 3 Soil carbon changes for the fully coupled HadCM3LC experiment and the off-line version driven by climate forcing from the coupled experiment. (a) Global total changes in GtC for the two runs. Off-line: solid line; coupled: dashed line. (b) Changes in the coupled run (1860–2100) in kgC m^{-2} , (c) as (b) but for the off-line run.

Results

Global

The results from the off-line runs of HadCM3LC and RothC are qualitatively similar. Both models simulate an accumulation of soil carbon during the 20th century followed by a more rapid release of soil carbon during the 21st century (Fig. 4). It is this rapid release that causes the transition of the terrestrial biosphere from carbon *sink* to *source* and is responsible for the positive feedback between climate and the carbon cycle. Global totals and a latitudinal breakdown are given for 1860, 2000 and 2100 in Table 2. In HadCM3LC, soil carbon accumulates faster than in RothC during the 20th century, reaching 60 GtC higher than its preindustrial level by 2000. RothC accumulates 32 GtC during the same period. This accumulation is the result of increased plant carbon input, which itself is largely caused by CO_2 fertilization of vegetation as CO_2 levels rise during the century but climate changes are relatively modest. It should be remembered that these simulations do not take account of changing land use: current land use was used in all simulations. After 2000 both models simulate a period of several decades where enhanced respiration caused by rising global temperatures compensates for increased plant carbon input. Soil carbon stocks change little during the first half of the 21st century. After 2060, increased respiration as a result of the increase in temperature begins to dominate and both models simulate a rapid decrease of soil carbon as CO_2 is released to the atmosphere. Over the 21st century HadCM3LC simulates a loss of 140 GtC, finishing 80 GtC below its preindustrial level. RothC simulates a loss of 86 GtC over this period, finishing 54 GtC below its preindustrial level.

The different carbon pools within RothC have similar long-term trends despite their very different timescales of response (Fig. 5). All four pools show an increase during the 20th century and rapid decrease during the later half of the 21st century. The majority of the total soil carbon changes are caused by the changes to the large soil carbon pools: HUM (29 GtC) and RPM (23 GtC). BIO and DPM contribute just 2 and 0.3 GtC, respectively, to the total changes during the course of the simulation.

Table 1 Location of points and reason for their choice

Point number	Location	Latitude	Longitude	Reason for choice
1	Siberia	105°E	72°N	Increase in soil carbon
2	Saskatchewan, Canada	105°W	55°N	Moderate decrease in soil carbon
3	UK	0°E	52°N	Small increase or neutral response
4	Manaus, Brazil	60°W	30°S	Dramatic decrease in soil carbon

In both soil carbon models, warmer temperatures tend to increase decomposition rates. As temperature is increasing across the globe, this means that soil carbon everywhere would experience faster decay if other environmental conditions were to remain constant. Temperature change is one of the two dominant forcings of the stock of soil carbon in these experiments. The other important factor is changes in plant carbon input because of changes in vegetation. Changes in these two factors between 1860 and 2100 are shown in Fig. 2. The upper panels show the evolution of the global mean temperature and carbon inputs and the lower panels show the regional distribution of the changes between 1860 and 2100. In HadCM3LC, all of the organic carbon input goes into the single-soil carbon pool. In RothC it is split between the DPM and RPM pools dependent on the type of overlying vegetation. Organic carbon inputs increase in many places, especially in the temperate and boreal zones where increases in temperature are beneficial to vegetation that experiences a longer growing season. Elevated CO₂ also contributes to enhanced productivity. However, in the tropics and in particular in Amazonia, the very high temperatures and increasingly dry conditions lead to a decrease in the vegetation cover (Betts *et al.*, 2004;

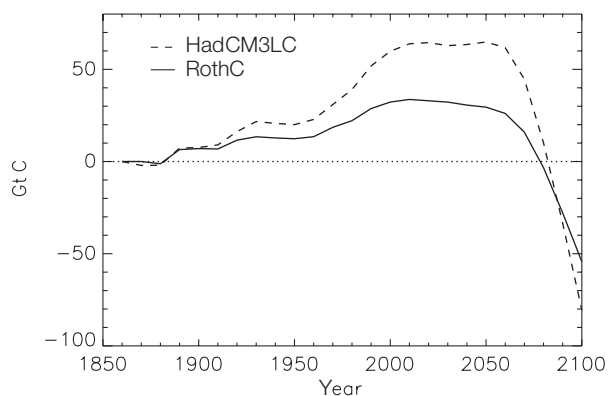


Fig. 4 Changes in global soil carbon amount (GtC) for the offline runs of HadCM3LC (dashed line) and RothC (solid line).

Cowling *et al.*, 2004; Cox *et al.*, 2004) and hence a large reduction in the carbon input to the soil.

Regional patterns of soil carbon and soil carbon change during the experiment are shown in Fig. 6. The models simulate very similar patterns, although the magnitudes of changes in RothC are smaller. The amounts of soil carbon within latitudinal bands are listed in Table 2. The latitudinal bands from 30°S to 30°N and from 30°N to 50°N follow the same pattern as the global response, with increased soil carbon up to 2000 and a decrease thereafter, and a greater magnitude of response in HadCM3LC than RothC. North of 50°N, HadCM3LC simulates roughly constant soil carbon amounts after 2000 whilst RothC simulates a decrease. This may be because of the different dependencies on soil moisture – in HadCM3LC respiration decreases with increasing soil moisture after an optimal level, but in RothC continues at high rates. South of 30°S both models simulate a slight increase in the small soil carbon stock throughout the entire period.

Response at the four selected points

The response of each model to the three forcings (i.e. temperature, moisture and organic carbon input) was investigated at four points that were representative of regions characterized by differing soil carbon behaviour (see Table 1). The off-line experiments were repeated but allowing only one of the forcings to change at a time with the other forcings held constant at their initial values. Hence, we separate the direct effects of temperature and soil moisture on heterotrophic respiration from indirect effects via changes in vegetation productivity.

Point 1, in Siberia, was chosen because this region showed a net increase in soil carbon during the model experiments. Figure 7 shows the soil carbon evolution for three cases: (a) all forcings changing (i.e. the original experiment), (b) temperature changing as before but with all other forcings held constant, and (c) organic carbon inputs changing but other forcings held constant. The large increase in carbon input at this point

Table 2 Global and latitudinal totals of soil carbon (in GtC) at 1860, 2000 and 2100 in the off-line runs

	Global total (GtC)		Latitudinal totals (GtC)							
	H	R	90–50°N		50–30°N		30°N–30°S		30–90°S	
	H	R	H	R	H	R	H	R	H	R
1860	1180	951	478	414	301	253	365	249	37	35
2000	1240	984	499	424	323	265	380	260	38	35
2100	1100	897	498	403	283	239	271	218	48	38

H columns are values from HadCM3LC and R columns from RothC.

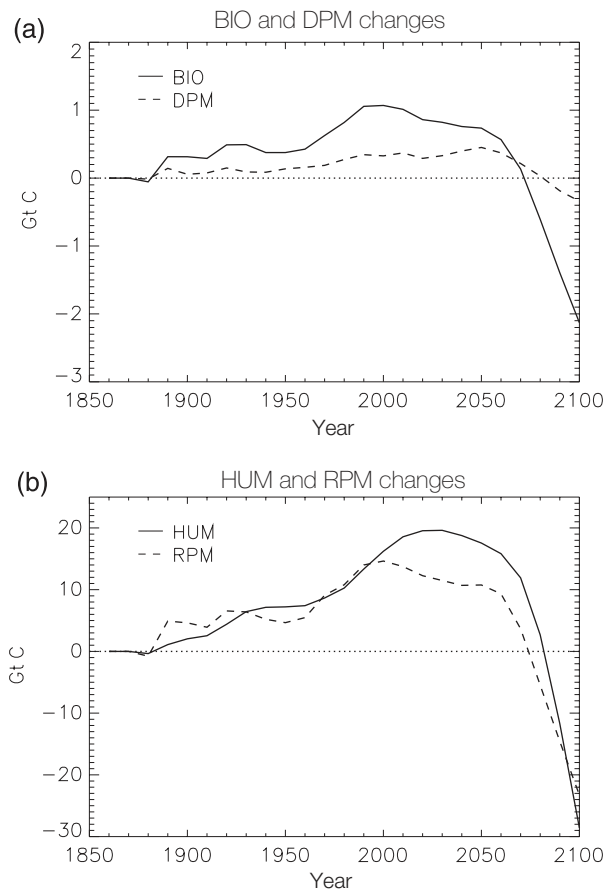


Fig. 5 Changes in global carbon (Gt C) in each of the four RothC soil carbon pools. (a) decomposable plant material (DPM) (dashed line) and microbial biomass (BIO) (solid line). (b) resistant plant material (RPM) (dashed line) and humified organic matter (HUM) (solid line). Note the different scales for each panel. DPM, BIO, RPM and HUM have 1860 values of 7, 20, 212 and 713 Gt C, respectively.

(caused by increased plant productivity because of warmer temperatures, elevated CO₂ and longer growing season) causes a large increase in soil carbon in both models. When only carbon inputs are varied (Fig. 7c), the response of HadCM3LC (13.2 kg C m⁻²) is much greater than that of RothC (6.6 kg C m⁻²), accumulating about twice as much soil carbon. The response of both models to changes in temperature (Fig. 7b) is also similar: both simulate a decrease in soil carbon when temperature alone is changed (−4.4 kg C m⁻² in HadCM3LC and −3.1 kg C m⁻² in RothC). The effect of including moisture changes alone is relatively small, with RothC simulating an increase of 0.15 kg C m⁻² and HadCM3LC a decrease of 0.6 kg C m⁻² (data not shown). The net effect of all the forcings at this point (Fig. 7a) is to increase soil carbon (by 3.7 kg C m⁻² in HadCM3LC and 1.4 kg C m⁻² in RothC) – i.e. in this case the organic carbon input dominates in both models.

At point 2, in Canada (Fig. 8), the temperature and carbon-input forcings act in the same sense as those in Siberia but the resulting soil carbon response is of the opposite sign. With C input forcing alone (Fig. 8c) there is an increase in soil carbon (7.6 kg C m⁻² in HadCM3LC and 4.4 kg C m⁻² in RothC). Again, HadCM3LC shows greater sensitivity than RothC to the carbon input increases. With temperature forcing alone (Fig. 8b) there was a decrease in soil C because of higher temperatures (−5.8 kg C m⁻² in HadCM3LC and −4.7 kg C m⁻² in RothC). The effect of including only moisture changes is again relatively small, with RothC simulating an increase of 0.13 kg C m⁻² and HadCM3LC no change (data not shown). However, in contrast to point 1, the net effect is a decrease in soil carbon – i.e. the temperature-driven soil carbon reductions dominate in this region resulting in changes of −0.66 kg C m⁻² in HadCM3LC and −1.6 kg C m⁻² in RothC. Again, both models are in agreement over the sign of the response, but HadCM3LC accumulates more soil carbon before the reduction begins.

Point 3 is sited in South-eastern UK. In the case of C input forcing alone (Fig. 9c), increasing organic carbon inputs caused an increase in soil carbon (9.0 kg C m⁻² in HadCM3LC and 4.3 kg C m⁻² in RothC). As with the other two sites, temperature forcing alone (Fig. 9b) caused a decrease (−5.1 kg C m⁻² in HadCM3LC and −3.1 kg C m⁻² in RothC). The effect of including moisture changes alone is again relatively small, with RothC simulating no change and HadCM3LC an increase of 0.33 kg C m⁻² (data not shown). In this case, the models differ slightly in their net response when all forcings are applied (Fig. 9a). The greater sensitivity of HadCM3LC to increased carbon input gives a net increase in soil carbon of 2.7 kg C m⁻² while RothC shows a more neutral response (0.4 kg C m⁻²).

At point 4, in Amazonia (Fig. 10), both temperature and plant carbon input act to decrease soil carbon. In HadCM3LC, the decrease was 3.7 kg C m⁻² for temperature forcing alone and 5.4 kg C m⁻² for the plant carbon input forcing alone. With RothC, the corresponding figures were 1.4 and 3.7 kg C m⁻², respectively. At this point, unlike the other three, moisture is important. In the moisture-only experiments both models simulate an increase in soil carbon (by 12.8 kg C m⁻² in HadCM3LC and 0.9 kg C m⁻² in RothC, data not shown). When all forcings are applied, the pronounced drying dominates over the increased temperature and leads to a marked reduction in specific respiration rates even though the climate is much warmer. However, the dieback of the vegetation and associated reduction of organic carbon inputs to near zero are sufficient to dominate over the reduced

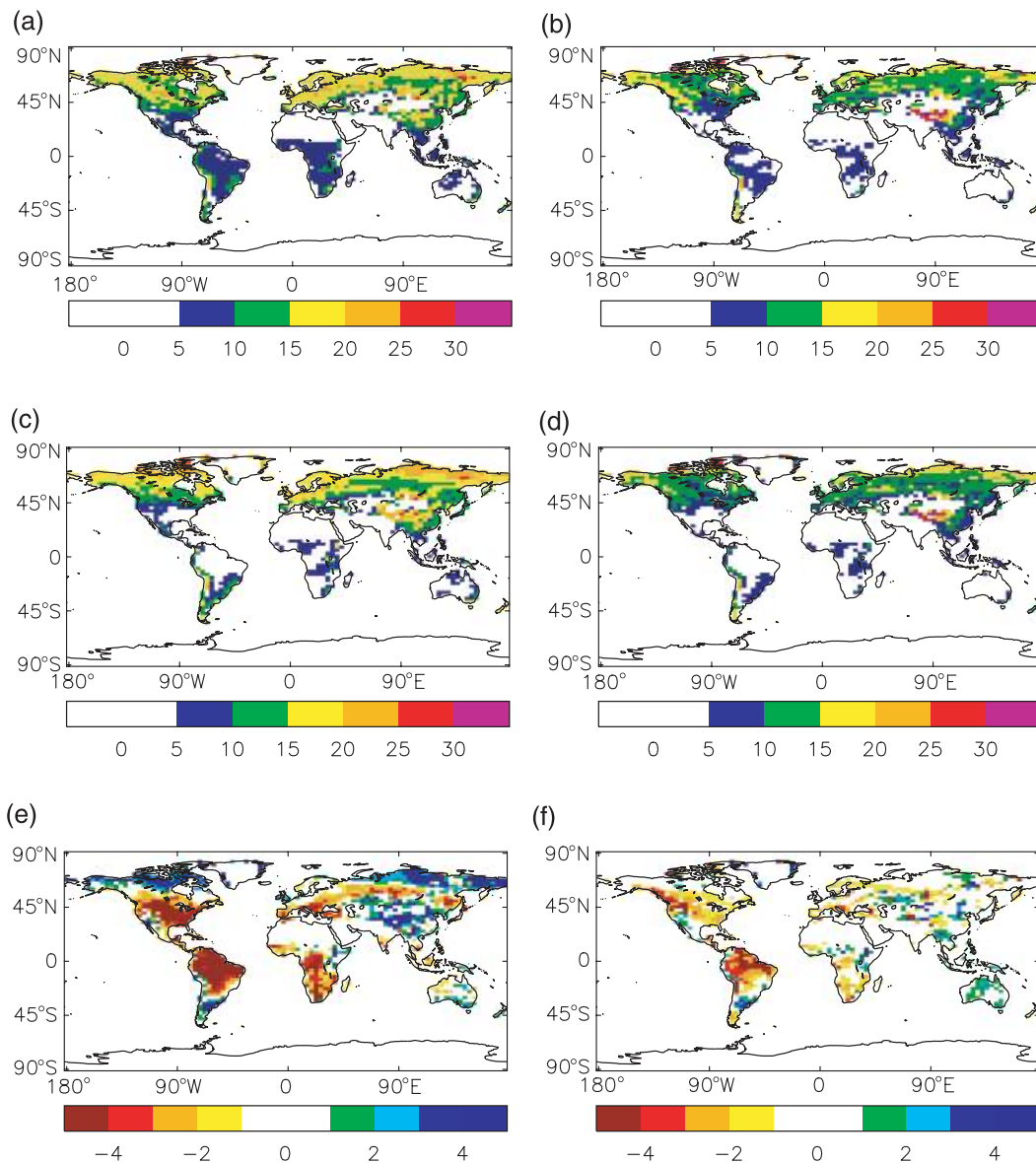


Fig. 6 Soil carbon amounts and changes (in kg C m^{-2}) for the off-line runs of HadCM3LC (lefthand column) and RothC (righthand column). Panels (a) and (b) show soil carbon amounts for 1860, (c) and (d) for 2100. Panels (e) and (f) show changes in soil carbon over the period (1860–2100).

respiration so that soil carbon decreases in both model simulations (by 5.4 kg C m^{-2} for HadCM3LC and by 4.0 kg C m^{-2} for RothC).

Discussion

Comparison of the two models

The response to organic carbon inputs is the biggest contribution to the differences between the two models. The more damped transient response of soil C when simulated by RothC, as compared with HadCM3LC, is

a result of the multipool structure of RothC. Organic carbon input in HadCM3LC goes entirely into the single carbon pool where it decays with a turnover time in common with the rest of the soil carbon. In RothC, the carbon input first enters the DPM and RPM pools both of which decay much faster than the HadCM3LC pool. Therefore, fresh inputs decay quicker in RothC than in HadCM3LC and so RothC simulates slower carbon increase for the same increase of organic carbon input. Conversely, once carbon is in the HUM pool of RothC (which is about 75% of total soil carbon and has a turnover time of 50 years) it decays slower than the

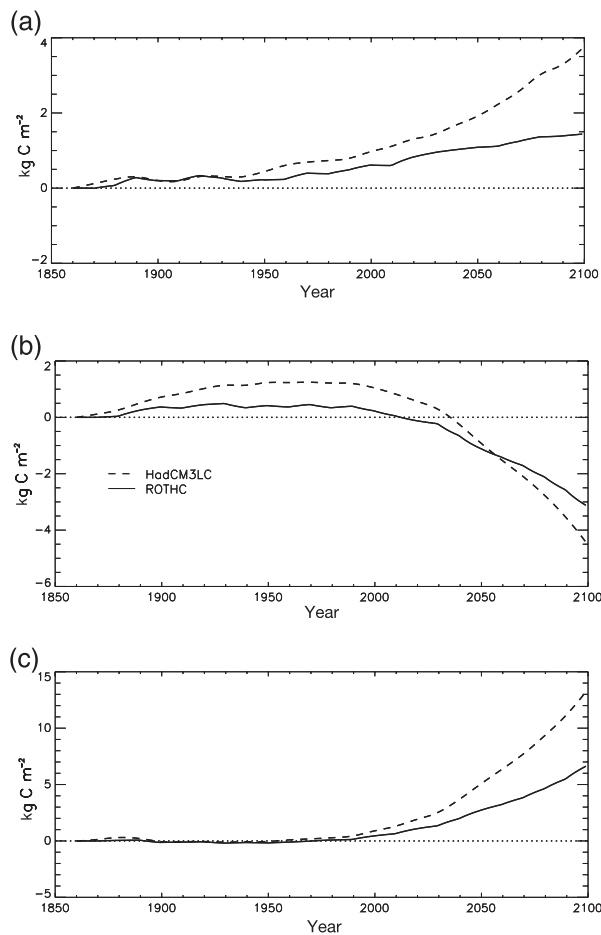


Fig. 7 Soil carbon content (in kg C m^{-2}) for point 1 in Siberia (105°E , 72°N) for the off-line runs of HadCM3LC (dashed line) and RothC (solid line). (a) Results from the fully forced experiment. (b) Results from the experiment where all forcing was held constant except for temperature. (c) Results from the experiment where all forcing was held constant except for organic carbon inputs.

single pool in the HadCM3LC. Thus, RothC simulates a slower loss of soil carbon for the same temperature increase.

The greater difference between the models' responses to carbon inputs than to temperature changes indicates that the impact of climate change on the productivity of vegetation and hence on the organic inputs to soil carbon is at least as important for determining the future soil carbon evolution as the sensitivity of the soil carbon itself to climate changes.

Because changes in the environment bring about smaller transient changes in soil carbon stocks in RothC than HadCM3LC, the strength of the feedback between climate and the carbon cycle would be weakened somewhat if RothC were to be run interactively within the coupled climate carbon cycle experiments. How-

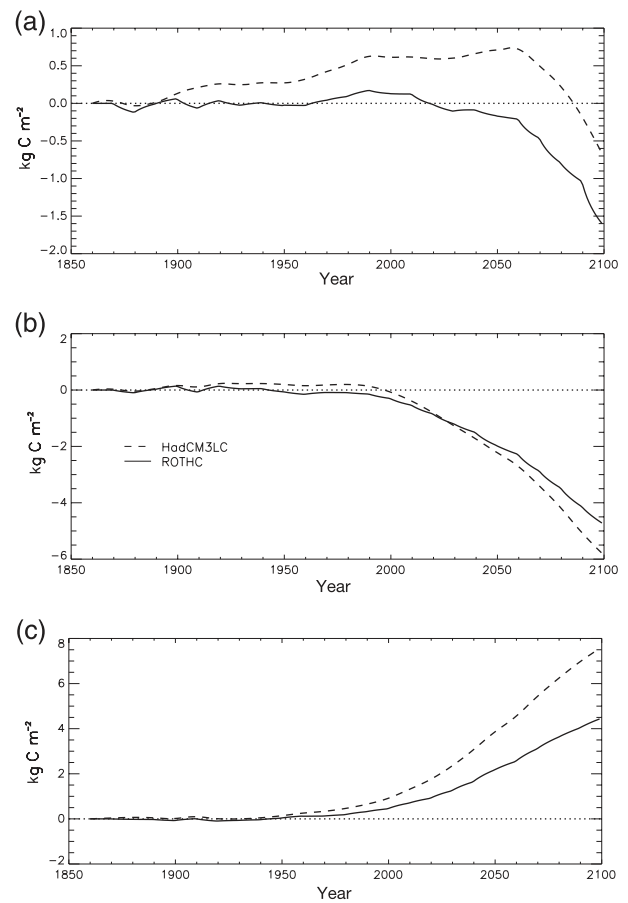


Fig. 8 As for Fig. 7, but for point 2 in Saskatchewan, Canada (105°W , 55°N).

ever, the feedback would still be strong and positive and result in substantially elevated CO_2 levels and accelerated global warming.

Our results are similar, but not identical, to those of Berthelot *et al.* (2002) who found decreased amounts of soil carbon in the tropics because of decreased carbon input while soil drying tended to compensate for the increased temperatures. In the northern regions, Berthelot *et al.* (2002) claim that warming dominates and that changes in organic carbon input are relatively small. Our results agree broadly with this, but do show increased carbon inputs in the north (possibly because of the inclusion of vegetation dynamics in HadCM3LC), which in some regions may be sufficient to overcome the increased respiration from the warmer temperatures. The balance between temperature, moisture and carbon-input forcing is further complicated by the fact that forcing from changing organic carbon inputs controls the total flux of carbon into the soil regardless of the amount of soil carbon present whereas temperature and

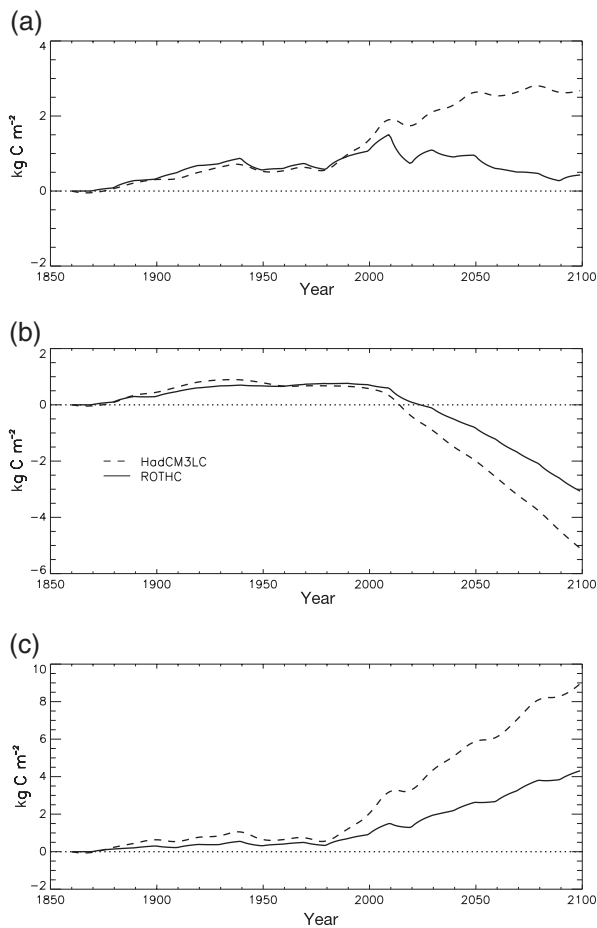


Fig. 9 As for Fig. 7, but for point 3 in South-eastern UK (0°E, 52°N).

moisture forcing affect the *specific* respiration rate (i.e. the *absolute* respiration is also dependent on the amount of soil carbon present). Hence, the balance between the forcings will depend on the initial stock of soil carbon.

In the global total, there is a net decrease in soil carbon (Fig. 4), despite the increase during the 20th century. This increase is because organic carbon inputs increase quickly in response to enhanced productivity caused by CO₂ fertilization. The slight warming is also beneficial for much of the vegetation and precipitation patterns have not changed markedly. However, by the end of the 21st century very high temperatures dominate, causing increased decomposition. In addition, there are regional reductions in carbon input because of locally drier conditions in Amazonia and, to a lesser extent, in tropical Africa. CO₂ fertilization begins to saturate at high concentrations and so the large CO₂ increase does not continue to stimulate big increases in plant carbon inputs.

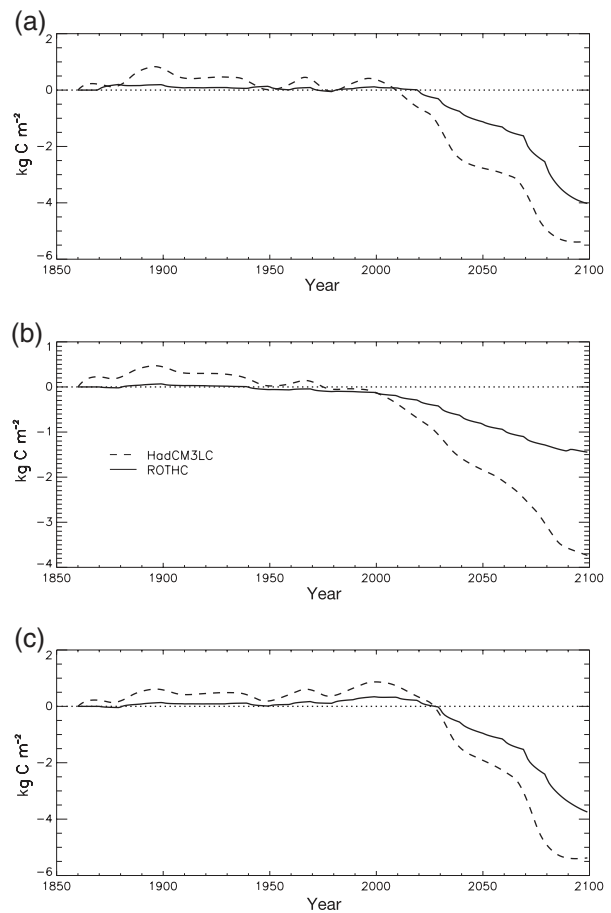


Fig. 10 As for Fig. 7, but for point 4 in Manaus, Brazil (60°W, 3°S).

Soil carbon changes and feedbacks on climate

The study of Jenkinson *et al.* (1991), which also used the RothC model to simulate the impact of global warming on soil carbon, estimated a release of about 100 GtC over period 1990 to 2050, assuming constant organic carbon inputs to the soil while temperature increased at the rate of 0.5 °C per decade. The results from our experiment (where the global mean temperature is also increasing at about 0.5 °C per decade) indicate that the release over the same period will be less than 10 GtC but accelerating to reach about 70 GtC by the end of the century (see Fig. 4). The main reason for the difference is almost certainly the increase in carbon inputs resulting from CO₂ fertilization of plant growth in this study.

There are significant sources of uncertainty that may affect the magnitude of the soil carbon feedback on climate change. Similar experiments performed by Post *et al.* (1996) gave different results. Although increased temperatures caused a decrease in soil carbon stocks, increased vegetation productivity was sufficient to

counter this via increased organic carbon input to the soil. The temperature effect reduced the soil carbon uptake but did not produce a sink-to-source transition in their model, possibly because of the use of a different vegetation model, or the assumption of a smaller increase in atmospheric CO₂ and hence climate change. This demonstrates the large uncertainties that result from competition between two large but opposing factors.

A limitation of our experiments is the omission of the effect of nutrient limitation on organic carbon inputs to the soil. It is widely accepted that nutrient cycles within the terrestrial biosphere will have a significant impact on its response to environmental changes (Schimel *et al.*, 1994; Clein *et al.*, 2002; Zhuang *et al.*, 2003), although the nature of this impact is very uncertain. It is possible that future nutrient limitation will reduce the CO₂ fertilization effect and so limit the increases in NPP and carbon inputs to the soil. However, it is also possible that increased nitrogen availability (either from anthropogenic deposition, Kicklighter *et al.* (1999), or released by enhanced soil decomposition, Schimel *et al.*, 1994; Zhuang *et al.*, 2003) may further enhance productivity and hence carbon inputs to the soil. Studies with models that include coupled C/N cycles (Schimel *et al.*, 1994; Clein *et al.*, 2002) suggest that such models may be expected to show less sensitivity to changes in climate and carbon inputs.

Regional climate responses, such as the drying in the Amazon region simulated by HadCM3 (Betts *et al.*, 2004; Cox *et al.*, 2004) and subsequent dieback of the tropical forest, are also uncertain and may significantly affect both local and global carbon stocks. Uncertainty in the response of NPP to climate and CO₂ changes along with uncertainty in whether increased uptake of carbon will be stored in the vegetation or soils (Kicklighter *et al.*, 1999) will affect the strength of the feedback between climate and soil carbon. Also, any changes to the above/below-ground allocation of carbon as biomes respond to changing climate (Post *et al.*, 1996; Davidson *et al.*, 2002) will have an influence on future soil carbon stocks.

Thawing of high-latitude permafrost regions may result in a large source of carbon to the atmosphere (Goulden *et al.*, 1998). Recent observations suggest that such thawing is beginning to take place and that the active layer is increasing in depth (Oelke *et al.*, 2004). A warmer climate and changes to soil drainage because of permafrost melting may have a large impact on the carbon stored in high-latitude peatlands. Drying out of peatland areas has been shown to increase respiration (Bubier *et al.*, 2003; Lafleur *et al.*, 2003) and may become a significant contribution to the climate-carbon cycle feedback (Schimel *et al.*, 1994). The climate impact on permafrost is included in HadCM3LC that simulates

the frozen/unfrozen fraction of the soil. To some extent, the large effective Q₁₀ values at low temperature that result from the temperature rate modifier of RothC may also account for this release, but it is possible that the nature of the temperature response of heterotrophic respiration changes across the freeze-thaw boundary (Michaelson & Ping, 2003; Zhuang *et al.*, 2003). However, we recognize that both models, and indeed almost all other soil carbon models, have been designed and tested primarily for mineral soils. There remains an outstanding need for improved representation of carbon turnover in organic soils, although this is outside the scope of the current study.

Giardina & Ryan (2000) suggest soil respiration is governed in the long-term by substrate availability and organic input quality and will therefore not respond strongly to increases in temperature. They postulate that a small pool of rapidly decaying soil carbon will quickly become exhausted, and that the majority of soil carbon is not sensitive to changes in temperature. However, Davidson *et al.* (2000) have refuted the conclusions of Giardina & Ryan. Different sensitivities to temperature for the different soil carbon pools of the model have not been investigated here. Our results depend upon the structure of the models, which assume that decomposition of all soil C pools (except the 'inert' pool in RothC) is temperature dependent.

Conclusions

Our results show no fundamental change to the nature of the positive feedback between climate and the carbon cycle when a more sophisticated multipool soil carbon model is included in the simulation. Increases in global temperature still result in enhanced respiration rates and hence decreased soil carbon contents. Globally, this effect dominates over increases in organic carbon input to the soil as a result of increased vegetation growth in many parts of the world. However, the magnitude of the feedback is dependent on the modelled soil carbon dynamics.

When both the original HadCM3LC soil carbon model (with a single-soil carbon pool) and RothC (with four soil carbon pools) were driven with forcing data from the same climate change experiment, both models simulated a small increase in soil carbon during the 20th century followed by a rapid decrease in soil carbon during the latter part of the 21st century. The temporal evolution of global totals and regional patterns of change are similar between the two models, with RothC generally exhibiting a smaller magnitude of response: it simulates slower soil carbon accumulation during the 20th century and slower release of soil carbon during the 21st century.

Individual points characteristic of regions with differing soil carbon behaviour were examined. There was generally competition between increased input to the stock of soil carbon from increased organic carbon inputs and increased respiration because of elevated temperatures. The balance between these two opposing factors varied regionally, but these variations were simulated similarly in the two models. In each case HadCM3LC simulated a larger magnitude of change in response to the individual temperature or carbon input forcing. It simulated a greater rate of accumulation of carbon in response to increased organic carbon input and a slightly greater rate of release in response to elevated temperatures. At three of the points examined the direct effects of moisture changes on heterotrophic respiration were less important than those of temperature or carbon inputs. However, moisture had an important indirect effect because of the influence it has on vegetation productivity and hence organic carbon inputs to the soil; this was particularly pronounced at the point in Amazonia.

The high level of agreement between these two models that have different soil carbon dynamics and different sensitivities to temperature and moisture increases confidence in the simulations of the climate carbon cycle feedback.

Hence, while the long-term response of heterotrophic respiration to changes in temperature and moisture is still uncertain, these simulations demonstrate that a more sophisticated representation of multipool soil carbon dynamics modifies, but does not fundamentally change, the conclusions of previous work with a simple soil carbon model. However, such improvements to the treatment of soil carbon are critical if we are to be able to accurately and quantitatively address questions regarding the impact of land-management practices and future climate changes on terrestrial carbon sinks.

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