

# Effects of variations in simulated changes in soil carbon contents and dynamics on future climate projections

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## Abstract

Climatic variables have major effects on all components and processes of the global carbon (C) cycle, including soil C contents and dynamics, which in turn have significant feedback effects on the global climate. We have investigated the interactive effects between soil C and projected climatic changes using the Institute of Numerical Mathematics Climate Model (INMCM) climate–C cycle model coupled to three soil organic matter dynamics models [the Lund–Potsdam–Jena (LPJ) soil biogeochemistry, ROMUL and Q models] based on three markedly differing conceptual interpretations of soil organic matter transformation (biochemical, discrete succession and continuous quality, respectively). According to simulations using all these couplings the positive effect of CO<sub>2</sub> fertilization on plant productivity outweighed the negative effects of increased soil temperature on soil C, consequently soils were projected to contain 10–104 Pg more C in 2100 than in the preindustrial period. However, the projected soil respiration rates tended to be higher and additional C storage lower when the LPJ soil biogeochemistry model was used rather than either the ROMUL or Q models. Global temperatures for 2100 predicted by the INMCM coupled to either the ROMUL or Q models were almost identical, but 0.4 °C lower than those predicted by the INMCM coupled to the LPJ soil biogeochemistry model. The differences in global predictions obtained with the ROMUL and Q models were smaller than expected given the fundamental difference in their formulations of the relationship between the quality and temperature sensitivity of soil organic matter decomposition.

**Keywords:** C cycle, climate change, general circulation model, soil C, soil organic matter dynamics model

Received 2 March 2009 and accepted 16 May 2009

## Introduction

It is generally recognized now that terrestrial ecosystems have substantial effects on climate regimes, in addition to being strongly affected by climate. The strength of the feedback processes and mechanisms involved have been investigated in detail using coupled biospheric carbon (C) cycle and climate models (e.g. Cox *et al.*, 2000; Zeng *et al.*, 2004; Friedlingstein *et al.*,

2006). In the terrestrial biosphere the annual flux of C to the land (via net primary production of plants, NPP), and atmosphere (via soil respiration and C emissions due to natural and anthropogenic ecosystem disturbances), are in delicate balance and are almost equal (IPCC, 2001). Hence, it is equally important to study the effects of anticipated global warming on both plant productivity and soil organic C dynamics. Changes in soil C are considered in simulations of future climate by several coupled climate–C cycle models (e.g. Cox *et al.*, 2000; Zeng *et al.*, 2004). However, unlike vegetation models, which are becoming increasingly sophisticated in modern versions of climate models, highly simplified models are still often used to

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describe soil organic matter dynamics (e.g. Jones *et al.*, 2005).

Soil C is a complex variable to model. It is accepted that biochemical, chemical and physical processes are involved in decomposition, but there is little consensus on ways to interpret the changes that molecules in fresh litter undergo during conversion to soil organic matter (Chertov *et al.*, 2007a). We can distinguish three major conceptual approaches. According to the first, biochemical, approach (e.g. Mindermann, 1968) plant residues can be separated into labile and more recalcitrant fractions, each decomposing at a specific rate. Labile fractions are respired quickly while recalcitrant compounds accumulate, forming soil organic matter. The second, discrete succession, approach (e.g. Chertov & Komarov, 1997) regards transformation of the organic debris into humus as a series of morphologically and biochemically distinguishable stages (a fraction cascade), corresponding to the concept of 'humus types' used in forest soil science. Each stage in the transformation is then associated with a certain humus type of increasing recalcitrance. In the third, most theoretical, approach, called Q-theory (from *q* in quality) (Ågren & Bosatta, 1996) any physical or chemical fractionation is disregarded, and instead the changes in a continuous variable – the quality of the organic matter in the soil – is considered. Its main postulate is that decomposition involves a combination of mass and quality loss during which fresh, readily decomposed organic matter, is transformed into recalcitrant humic substances. Conceptually, close to the Q-theory is a general theory of humification (Orlov, 1995) based on a thermodynamic consideration of decomposition, which postulates that, regardless of factors influencing the processes and soil type, only the most thermodynamically stable components, such as humic substances, will remain for prolonged periods during the process of organic matter transformation. Each of the three concepts mentioned have been applied in numerical models of soil organic matter dynamics (as reviewed in Powlson *et al.*, 1996; Chertov *et al.*, 1999; Smith *et al.*, 2008).

It is often argued that the fate of soil C in a changing climate is uncertain (e.g. Smith *et al.*, 2008). Indeed, there are uncertainties regarding many fundamental aspects of decomposition process and related ecosystem dynamics in the current climate, which are currently under investigation (e.g. Davidson *et al.*, 2006; Ågren & Wetterstedt, 2007). However, if we compare results obtained using a range of models of soil organic matter dynamics with different formulations, some of this uncertainty should be incorporated in the range of predictions (e.g. Jones *et al.*, 2005). Hence, the purpose of this study was to obtain climate scenario predictions using a climate–C cycle model coupled to three con-

ceptually different soil organic matter dynamics models and assess the extent to which the choice of model affects the predictions.

## Methods

### Models

*Coupled climate C cycle model.* The Institute of Numerical Mathematics Climate Model (INMCM, Diansky & Volodin, 2002; Volodin & Diansky, 2006) is one of the coupled atmosphere–ocean general circulation models with an interactive C cycle used in intercomparisons of models presented in the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC, 2007a). The model is used to reproduce climate change during the 20th century and to simulate possible climate changes during the 21st–22nd centuries (Volodin & Diansky, 2006). The model has 5° longitude × 4° latitude resolution and includes a standard set of subgrid parameterizations. Heat and water transfer in the soil is simulated explicitly with high vertical resolution (23 vertical layers with higher resolution in the upper soil profile). Specific equations describe solid, liquid water and vapour water movement and phase transformations. Liquid water movement is described by 1D (vertical) equations, no effect of water table on soil moisture is considered, and water flux across the low boundary is added to the subsurface runoff. Complete saturation of mineral soil is not possible and maximum water content is determined by the field capacity of the given soil type.

Coupled to the INMCM is a Land Surface Model (LSM, Bonan, 1996) used to simulate the productivity of terrestrial vegetation and related CO<sub>2</sub> uptake from the atmosphere with an hourly time step. Photosynthesis is simulated based on the models by Farquhar *et al.* (1980) and Collatz *et al.* (1991) for sun and shaded leaf fractions separately (Bonan, 1996). The model formulation incorporates a CO<sub>2</sub>-fertilization effect in an elevated CO<sub>2</sub> atmosphere. Plant respiration (PLR) is simulated using the growth-maintenance paradigm, but generally the estimated PLR/gross primary production (GPP) ratio is fairly constant. Twelve plant functional types are specified (needleleaf evergreen, needleleaf deciduous, broadleaf evergreen, broadleaf deciduous and tropical seasonal trees; evergreen, deciduous and arctic deciduous shrubs; C3, C4 and arctic grass; crops), each having its own set of biochemical, morphological, physiological, phenological, optical and aerodynamic properties. The proportions of vegetation types and their phenological attributes (leaf indices) are prescribed in the model, using observed seasonal patterns, but the photosynthetic activity of

plants is assumed to occur only during periods with air temperatures above plant type-specific thresholds. The proportions of GPP allocated to different plant compartments and biomass turnover time are constants. In this General Circulation Model formulation evapotranspiration is simulated according to Sellers *et al.* (1986). The ocean C cycle is described by the convection–diffusion equation, with turbulent exchange parameterization at the atmosphere–ocean boundary, an analytical equation for the carbon dioxide–carbonate equilibrium and a very simple parameterization of the ocean biological C cycle (prescribed constant transfer of C from the upper into deep ocean with zero net biological exchange with the atmosphere).

*Soil organic matter dynamics models.* Three soil organic matter dynamics models were subsequently used within the C cycling part of the INMCM model and returned soil CO<sub>2</sub> fluxes to the atmospheric model.

The Lund–Potsdam–Jena (LPJ) soil biogeochemistry model is a simple box model with a framework of biochemical concepts. It is part of the LPJ Dynamic Global Vegetation Model (Sitch *et al.*, 2003), which is often used to describe processes in vegetation and soils in coupled climate–C cycle models (e.g. Friedlingstein *et al.*, 2006). Three pools of soil organic matter are specified: litter, intermediate and slow. Each pool decays exponentially in time at a specific rate, which is maximal for the fresh litter and minimal for the slow pool. Litter that is decayed is distributed in fixed proportions between the atmosphere ( $f_{\text{air}} = 0.7$ ) and intermediate and slow soil pools;  $(1 - f_{\text{air}})f_{\text{inter}}$  ( $f_{\text{inter}} = 0.985$ ) enters the intermediate pool and  $(1 - f_{\text{air}})f_{\text{slow}} = (1 - f_{\text{air}})(1 - f_{\text{inter}})$  the slow pool. Turnover rates for each pool are functions of temperature and soil moisture. Temperature dependence follows the modified Arrhenius relationship (Lloyd & Taylor, 1994). Model coefficients were chosen from the literature (Zaehle *et al.*, 2005). The model is formulated as a system of ordinary differential equations. In this study, stepwise integration according to analytical formulae was used.

ROMUL is a model based on the concept that soil organic matter transformation occurs in a succession of discrete steps. ROMUL (Chertov *et al.*, 2001) is also a compartment model, but the compartments here are specified using the concept of ‘humus type’. The humus in organic and mineral soil horizons is separated. One specific compartment is a humified forest floor, protected from decomposition by the presence of stabilizing humic compounds. Decomposition is treated as proceeding via sequential stages of organic matter transformation mediated by different functional groups of soil organisms. C and N dynamics are

simulated in parallel and are interrelated. The rates of mineralization and the flows of matter between different compartments are described by kinetic coefficients, each of which is specifically dependent on both the organic matter quality (C/N ratio) and the environmental conditions (temperature and moisture). The coefficients used for litter and humified forest floor were derived from long-term laboratory studies (Chertov *et al.*, 2007b; Nadporozhskaya & Chertov, 2007), and the rates of organic matter turnover in the mineral horizon were deduced from data on the activity of specific groups of soil organisms (Chertov & Komarov, 1997; Chertov *et al.*, 2007b). The functions describing the temperature dependence of mineralization are approximations of temperature responses obtained in experimental trials of mass loss under laboratory conditions (Chertov *et al.*, 2007b), which are equivalent to the usage of the  $Q_{10}$  function in the interval 1–20 °C. Temperature response functions become linear above a certain threshold temperature, which is different for each of the compartments. Specific functions are used to describe the temperature and soil moisture dependence of humification and activity of earthworms (Chertov *et al.*, 2001). The ROMUL model is formulated as a system of ordinary differential equations. In this study, the system was solved with an explicit first order method. The model has been evaluated against numerous long-term experimental observations and provides good results for real forest datasets (e.g. Chertov *et al.*, 1997). ROMUL is also a part of the ecosystem model EFIMOD (Chertov *et al.*, 2003; Komarov *et al.*, 2003), which has been applied in site and regional C balance assessments (e.g. Chertov *et al.*, 2006, 2009; Van Oijen *et al.*, 2008).

The Q model is both an analytical and a numerical realization of Q-theory (Ågren & Bosatta, 1996). Soil organic matter is characterized by a distribution  $\rho_c(q, t)$  over a continuous variable quality,  $q$ , which describes its ease of decomposition. Here, we have used the formulation in equation 4.15 presented by Ågren & Bosatta (1996) for  $\rho_c(q, t)$ , in which decomposition is described by two terms: one for the mineralization of organic matter and one that can be compared with the ‘advection’ of C towards lower qualities (humification). A function  $u(q, t)$  describes the microbial utilization rate of the soil organic matter, which is temperature and quality dependent. The thermodynamic equation from Bosatta & Ågren (1999) describes temperature dependency with an Arrhenius-type formulation, but such that utilization of lower qualities is more temperature sensitive. A decomposer efficiency  $e_0$  determines the proportion of C assimilated by decomposers that is lost in mineralization and emitted as CO<sub>2</sub> to the atmosphere. After each cycle of

decomposition system C is returned to the soil, with its quality shifted by, on average,  $\eta_1(q)$  towards lower qualities. This shift is proportional to the square of quality (e.g. Ågren & Bosatta, 2002), with a fixed coefficient of proportionality  $\eta_{12}$ . The model was slightly modified here, by omitting the parameter  $f_c$  ( $g_C g_{\text{biomass}}^{-1}$ ), the C concentration in microbial biomass, and incorporating it instead in  $u(q, t)$ . Equation (4.15) is a partial differential equation, which was solved in this study using an explicit total variation diminishing (TVD) scheme (e.g. Yee, 1987). The numerical solution was tested against model experimental data with an analytical solution written for the sum of annual cohorts (e.g. Ågren & Bosatta, 2002). The Q model has been used as part of ecosystem models in site-specific and regional C balance assessments across Europe (e.g. Ågren & Hyvönen, 2003; Ågren *et al.*, 2007; Van Oijen *et al.*, 2008).

#### Model parameterization

The LPJ soil biogeochemistry model has previously been used in global applications. Therefore, no calibration of this model was done. One coefficient, the 'true humus' mineralization rate, was calibrated in the ROMUL model. C/N ratios in litters of each of the plant functional types were selected from a database published by Rodin & Bazilevich (1965). The Q model was parameterized as follows: the activation energy,  $A = 6500 \text{ K}$ , used in the thermodynamic equation describing the temperature dependence of microbial growth rates was chosen to correspond to  $Q_{10}$  2.0–2.4.  $e_0 = 0.25$  was selected as in previous applications of the Q model (Ågren & Bosatta, 2002). The remaining model parameters,  $u_0$  (the base growth rate in  $u(q, t)$ ) and  $\eta_{12}$ , were calibrated using values selected from the range reported by Hyvönen *et al.* (2005). The two parameters ( $u_0$  and  $\eta_{12}$ ) were adjusted simultaneously to keep the difference in degree of humification between different biomes close to observed differences (Orlov, 1995), giving  $\eta_{12} = 0.07$  and  $u_0 = 4.0 \times 10^8 \text{ year}^{-1}$ . Currently, all litter, regardless of plant functional types, is assigned the initial quality  $q_0 = 1.0$ , in accordance with the viewpoint of Orlov (1995), that in most ecosystems, except planted coniferous monocultures with no ground vegetation, different plant species produce litter with a wide range of qualities, but soil fauna largely homogenize the litter.

Models were calibrated using the area density of soil C in major biomes or soil types. The data for the model calibration came both from a global database (Zinke *et al.*, 1998) and the Russian Soil Database (Stolbovoi & Savin, 2002). The latter includes data obtained from samples of all soil types in Russia, which are thought

to be reliable, because samples have been taken from a very wide range of biomes (from tundra to steppe and deserts) and the sample sizes are very large (Stolbovoi & Savin, 2002). Additional information, such as humic to fulvic acid concentration ratios (indicators of soil organic matter quality), are also widely available in the published literature (e.g. Orlov, 1995).

#### Modelling protocol

The LPJ soil biogeochemistry, ROMUL and Q models were used on their own with litter and soil climate outputs from the INMCM to produce a steady-state distribution of soil C over the globe with preindustrial climate and NPP. A numerical solution with a monthly time step was chosen for the 10 000-year spin-up of the models. The results from the numerical solutions were compared with the analytical steady-state solutions. Analytical and numerical solutions converged in experiments with LPJ biogeochemistry and ROMUL models. The steady-state soil C storage produced by the analytical solution of the Q model were higher than the numerical estimates, indicating that 10 000 years was insufficient to reach steady-state (cf. Ågren *et al.*, 2007). However, the change in soil C storage simulated by the Q model is very slow during the period following the 10 000-year spin-up and is negligible in the 200-year timescale studied here.

Each of the studied models was then fully coupled to the climate–C cycle model. The model was run for the period 1860–2100 with no other changing external forcing except for anthropogenic  $\text{CO}_2$  emissions, which were taken from the IPCC-SRES A1B scenario (IPCC-SRES, 2000). The effects of initial conditions were studied with the Q model in a model experiment in which the initial soil C storage was decreased by 30% relative to the reference case.

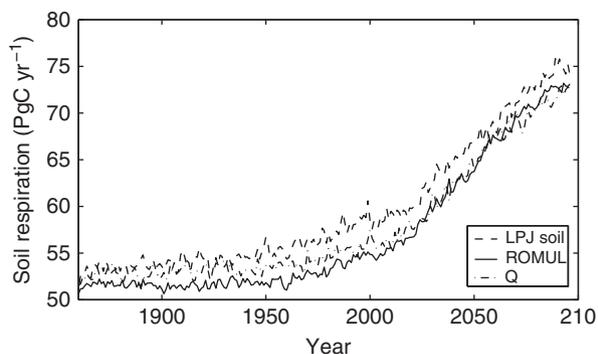
#### Analytical procedure to test the effect of model parameterization

Performing a comprehensive sensitivity study with the fully coupled climate–C cycle model is not a straightforward task at the moment, because multiple Monte Carlo runs require significant computer resources. As a simplification, we calculated analytically steady-state C stores with the LPJ biochemical soil and Q models, then used the returned values to investigate the extent to which model parameterization can affect the results (Appendix A). Temperatures corresponding to the globally average air temperatures for 1860 and 2100 were used and the litter input in 1860 was chosen to produce the steady-state soil C storage equal to the globally average C density in soil in 1860. The litter input in

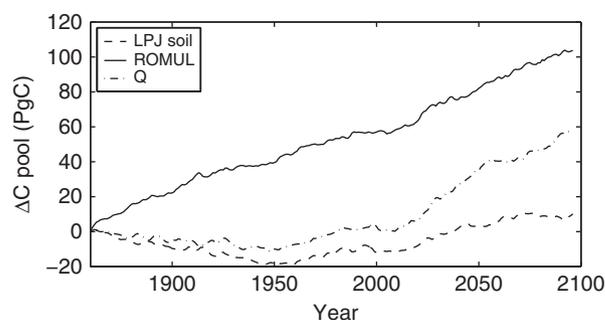
2100 was either higher than, or equal to, that in 1860. The higher litter input is as much higher (in percent) as the litter input produced by a coupled climate C cycle model is higher in 2100 compared with 1860. The analytical model does not represent transient behaviour and was only used to investigate qualitatively the relationship between model parameters and model results. We varied values of model parameters and studied the resulting differences between the two steady-state solutions (Appendix A).

## Results

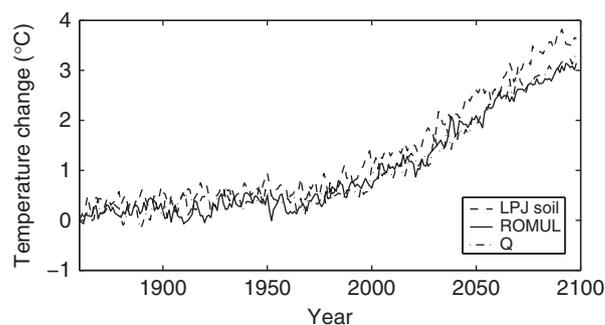
Predicted global fluxes of C from the soil to the atmosphere (heterotrophic respiration) were highest with the LPJ soil biogeochemistry model (Fig. 1), while the ROMUL and Q models predicted lower fluxes, which did not differ much. The difference in fluxes of C from the soil between the LPJ biogeochemistry model and the other two models was highest during the period 1960–2000, when it amounted to  $4.0 \text{ Pg C yr}^{-1}$  and diminished to  $1.5 \text{ Pg C yr}^{-1}$  at the end of the studied period in the year 2100 (Fig. 1). The  $\text{CO}_2$ -fertilization effect is incorporated in the model and higher productivity of terrestrial plants in the higher  $\text{CO}_2$  atmosphere compensates for the loss of soil C resulting from increased temperature. Losses and gains of C are nearly balanced according to the LPJ soil biogeochemistry model and soils globally accumulate only  $10 \text{ Pg C}$  between 1860 and 2100 (Fig. 2). According to the ROMUL and Q models, soils accumulate more C, 104 and  $60 \text{ Pg}$ , respectively (Fig. 2). The higher the proportion of C (as  $\text{CO}_2$ ) in the atmosphere relative to that in other reservoirs, the higher the productivity of terrestrial plants and C accumulation in living biomass. Hence, the higher soil respiration in the LPJ soil biogeochemistry model is partially compensated by increases in vegetation C.



**Fig. 1** Global annual soil respiration ( $\text{Pg C yr}^{-1}$ ) in 1860–2100 according to the Lund–Potsdam–Jena (LPJ), ROMUL and Q models when coupled to the climate carbon (C) cycle model Institute of Numerical Mathematics Climate Model (INMCM).



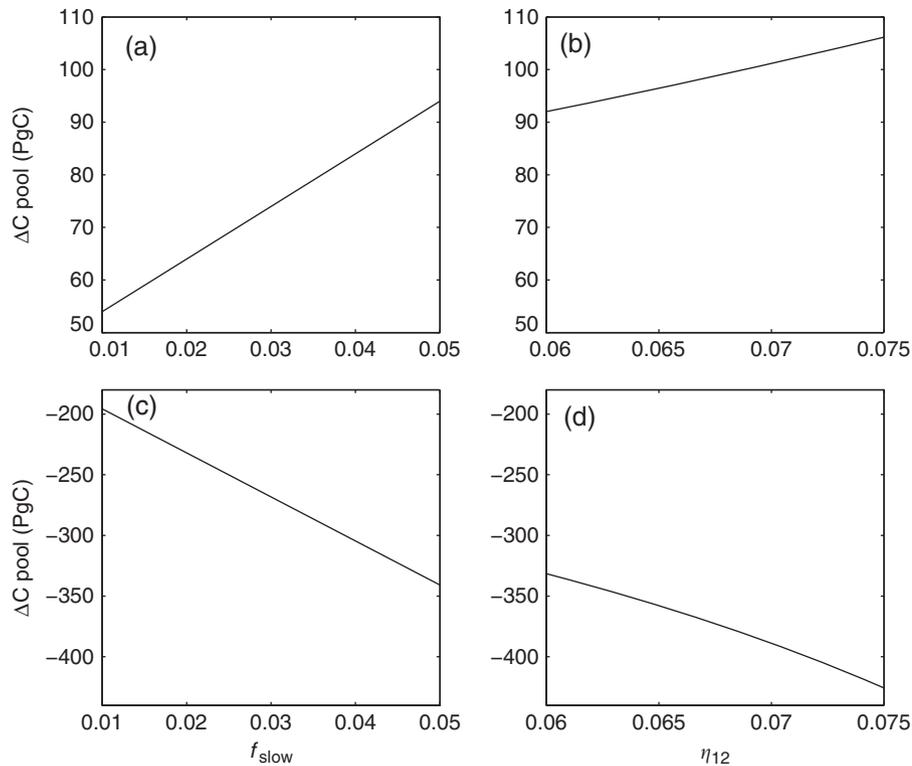
**Fig. 2** Change in soil carbon (C) storage from 1860 to 2100 ( $\text{Pg C}$ ) according to the Lund–Potsdam–Jena (LPJ), ROMUL and Q models when coupled to the climate C cycle model Institute of Numerical Mathematics Climate Model (INMCM).



**Fig. 3** Global average air temperature change ( $^{\circ}\text{C}$ ) in 1860–2100 simulated with the climate carbon (C) cycle model Institute of Numerical Mathematics Climate Model (INMCM) and the Lund–Potsdam–Jena (LPJ), ROMUL, or Q models describing soil organic matter dynamics.

The projected atmospheric concentration in 2100 was highest when the LPJ soil biogeochemistry model was used (789 ppm, compared with 745 and 736 ppm, respectively, when the Q and ROMUL models were used). The differences in  $\text{CO}_2$  concentration are reflected in differences in the modelled global temperature, which is  $0.4^{\circ}\text{C}$  higher in 2100 with the LPJ soil biochemistry model than with the other two models, which produce almost equal results (Fig. 3). The differences in temperature between the LPJ soil biogeochemistry model and the other two models are highest from 2020 to the end of the simulation period.

The reasons for the difference in soil respiration rates returned by the LPJ soil biochemistry model and the other two models were investigated with the help of the analytical model. Based on a previous sensitivity study (Zaehle *et al.*, 2005), three factors were selected as probable major causal factors: the litter turnover rate, fraction of decomposing litter going to the atmosphere and proportion of decomposed litter allocated to the slow pool ( $f_{\text{slow}}$ ). As shown in Fig. 4, even slight changes in the proportion of decomposed litter allocated to the



**Fig. 4** The difference in carbon (C) storage (PgC) between 1860 and 2100 estimated from an analytical expression as a function of the humification parameter. (a) and (c) The Lund–Potsdam–Jena (LPJ) soil biochemistry model,  $f_{\text{slow}}$  is the fraction of decomposed litter going into the slow pool. (b) and (d) The Q model,  $\eta_{12}$  is a parameter describing the humification rate. (a) and (b) Litter production is increased in 2100, (c) and (d) litter production is the same in 2100 and 1860. Model parameters are as elsewhere in this study.

slow rather than intermediate pool by the LPJ soil biochemistry model can change the amount of C stored. The amount of C additionally stored in soil under increased litter inputs is linearly related to the proportion of decomposed litter allocated to the slow pool (Fig. 4a). On the other hand, if it is assumed that the litter input remains constant then the higher the proportion of decomposing litter going into the slow pool the higher the loss of C from the soil (Fig. 4c). The other two studied factors were found to be less important (not shown). The responses simulated by the Q model are similar when the analogous parameter, the humification rate  $\eta_{12}$ , is changed. In the Q model soil C decreases when the humification rate and temperature is increased, because more soil C is allocated to lower qualities that are more sensitive to temperature changes (Fig. 4d). However, when litter inputs are also increased the higher humification rate also increases the input of litter to the lower qualities and this more than outweighs the increased temperature sensitivity (Fig. 4b). Experiments with the analytical model solution have revealed the high sensitivity of the Q model to the initial litter decomposition rate ( $u_0$ ) while the LPJ soil biogeochemistry model sensitivity to the litter turnover rate

( $k_{\text{lit}}$ ) is low (not shown). ROMUL is also not sensitive to the initial litter decomposition rate (Komarov, 2007).

The simulated values of soil respiration from the Q model runs with different initial soil C pools converged in the climate model after about 10 years, indicating that model parameterization may have stronger effects than initial conditions on the model's performance.

All three models were reasonably successful in simulating the geographical distribution of soil C storage over the globe (Fig. 5). Soil C storage is high throughout much of the boreal and temperate zone, with values exceeding  $15 \text{ kg m}^{-2}$  found in Siberia, Northern Europe and Canada (Fig. 5). According to the Q model the area of high soil C storage is shifted towards the regions of high productivity in the temperate zone, and boreal maximum is not as pronounced as in the LPJ soil biogeochemistry and ROMUL model simulations. Outside of boreal and temperate zones, soil C content in vegetated areas typically ranges between 5 and  $15 \text{ kg m}^{-2}$  while semiarid or arid regions are characterized by soil C contents of  $<5 \text{ kg m}^{-2}$  (Fig. 5). The area density of soil C in major biomes (Table 1) is within the range presented in the literature (Zinke *et al.*, 1998; Stolbovoi & Savin, 2002). However, there are two major

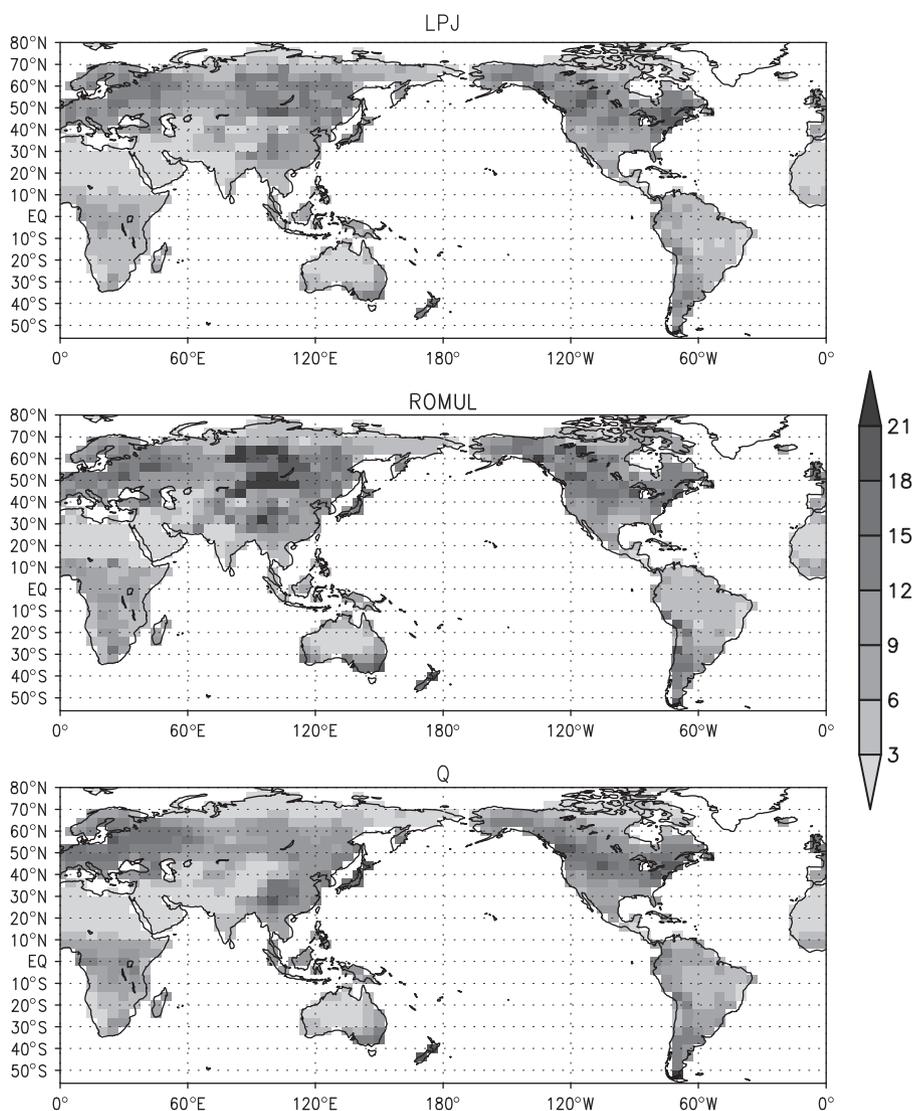


Fig. 5 Total carbon (C) density in the soil ( $\text{kg C m}^{-2}$ ) in 1860 according to Lund–Potsdam–Jena (LPJ), ROMUL and Q models.

problems, which all three models share. The first is that the climate model does not take into consideration peatlands and temporally water-logged soils. These soil types store most of the organic C and, if present, greatly modify the area density of soil C. Discrepancies between the modelled and observed values are most pronounced for subarctic regions and the West Siberian plains, where peatlands occupy significant proportions of the area. In these areas models underestimate soil C storage. The second problem concerns the Chernozem soils, grassland soils that, according to observations, store great amounts of C (up to 40 and 20  $\text{kg C m}^{-2}$  on average; Orlov, 1995), although the soils are not water logged. The models do indicate that C accumulates in an area north of the grassland zone and south of the temperate forest. However, this area is shifted in Eurasia northwards from the observed maximum of soil C

accumulation, and is less pronounced than the observed peak (on average Chernozems should store twice as much C as Podzols).

The most striking between-model differences in predictions regarding changes in soil C storage from 1860 to 2100 is that the LPJ soil biogeochemistry model simulates losses of C across the entire boreal zone, while the ROMUL and Q models predict (moderate) losses of C only in Central Siberia and, according to ROMUL, in the north of the boreal zone in North America (Fig. 6). Soils in the subarctic regions gain C according to all three models. The ROMUL model predicts an increase in soil C storage in the temperate zone, while the LPJ soil biogeochemistry model and the Q model simulate loss of C in this zone. In the tropics the LPJ soil biogeochemistry model predicts stronger uptake of C than the other two models.

**Table 1** Soil carbon characteristics for the major biomes calculated using the three models

Biome	LPJ_soilC			ROMUL			Q model		
	$m_C^*$ (Pg C)	$\rho_C^\dagger$ (kg C m <sup>-2</sup> )	HD $\ddagger$ $C_{slow}/$ ( $C_{lit} + C_{inter}$ ) (unitless)	$m_C$ (Pg C)	$\rho_C$ (kg C m <sup>-2</sup> )	HD $H/$ ( $L + F$ ) (unitless)	$m_C$ (Pg C)	$\rho_C$ (kg C m <sup>-2</sup> )	HD $\S$ $C_H/C_F$ (unitless)
Tropical forest	87	5.6	0.36	87	5.6	3.9	125	8.0	5.53
	105	6.7	0.30	98	6.3	3.7	128	8.2	5.37
Temperate forest	31	12.1	0.36	33	12.8	5.3	38	15.0	5.73
	27	10.5	0.41	35	13.8	5.9	38	14.8	5.76
Boreal forest	262	11.8	0.36	298	13.4	4.3	247	11.1	5.64
	226	10.1	0.42	311	14.0	4.6	264	11.9	5.28
Tundra	34	3.5	0.36	52	5.3	2.7	23	2.4	5.47
	40	4.1	0.29	60	6.2	2.2	36	3.7	4.23
Desert	47	1.8	0.36	77	3.0	8.9	47	1.8	5.65
	48	1.8	0.35	87	3.4	8.5	50	1.9	5.51
Savannah	155	5.1	0.36	214	7.1	7.2	200	6.6	5.76
	186	6.1	0.30	248	8.2	6.7	217	7.2	5.49
Steppe	120	9.7	0.36	157	12.7	8.4	131	10.6	5.61
	117	9.4	0.37	171	13.8	8.8	138	11.1	5.38
Cultivated areas	86	7.3	0.36	100	8.5	6.7	115	9.7	5.67
	87	7.4	0.35	110	9.3	7.4	115	9.7	5.54
Total	862	6.4	0.36	1063	7.9	5.6	979	7.23	5.65
	874	6.5	0.35	1171	8.7	5.6	1039	7.67	5.37

\*Total mass of soil C.

†Area density of soil C.

‡Humification degree.

§The 'threshold quality', which divides the spectrum of quality into two intervals, was determined so that the ratio of the integral over the spectrum below this threshold ( $C_H$ ) and the integral over the spectrum above this threshold ( $C_F$ ) were approximately equal to HD calculated with the ROMUL model.

Upper and lower values are for 1860 and 2100, respectively.

L, litter; F, humified forest floor; H, humus of mineral soil.

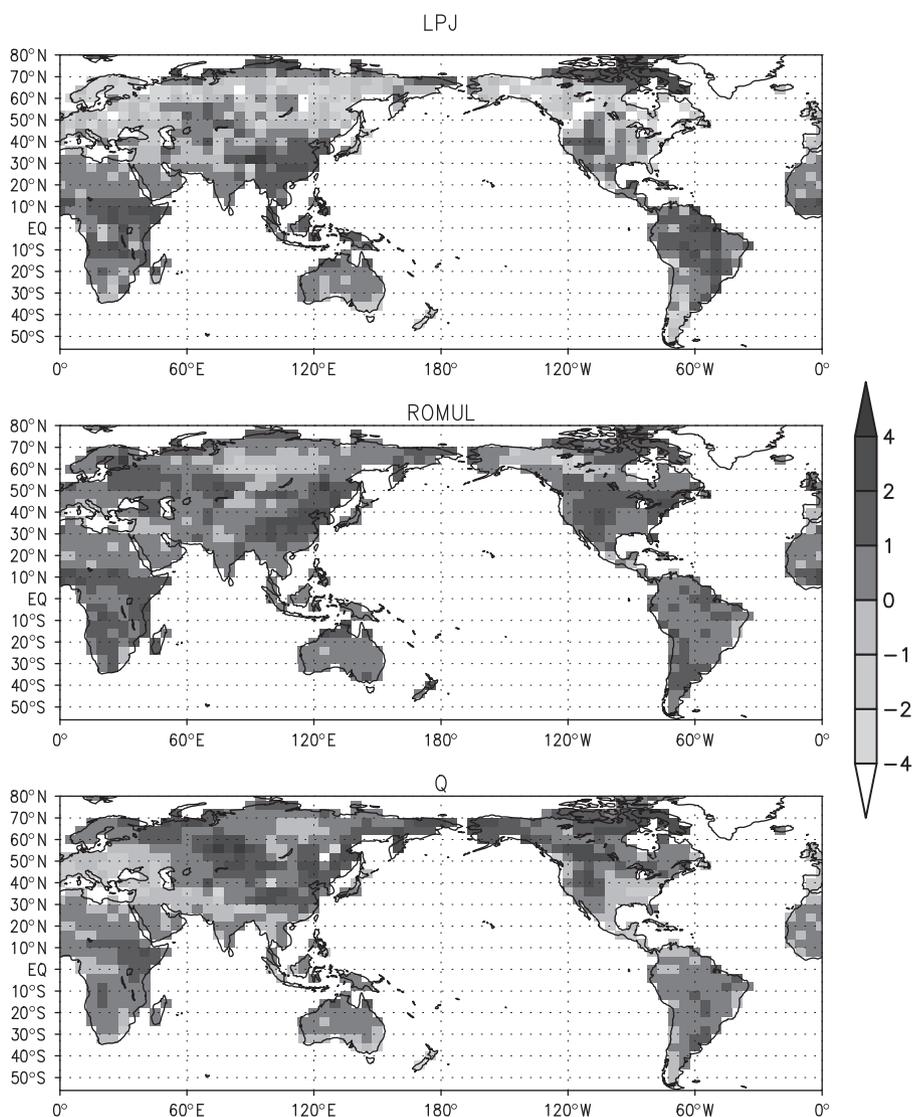
According to both the LPJ biogeochemistry and ROMUL models the humification degree (or degree of soil organic matter recalcitrance) increases in temperate and boreal zones and decreases in tundra and tropical zones (Table 1). According to the Q model the humification degree decreases everywhere except in the temperate forest zone, and most strongly in the tundra zone (Table 1). A decrease in humification degree means that C is mainly lost from the recalcitrant fractions of organic matter or gained mainly by the fractions of high decomposability. A decrease of humification degree simulated almost everywhere in the globe by the Q model is most probably because high-quality organic matter (e.g. fresh litter) is less temperature-sensitive than low-quality organic matter.

## Discussion

In this study the ROMUL and Q models were applied globally for the first time. Despite some limitations, both models are in principle applicable to the global

scale and can provide competitive options as parts of coupled climate–C cycle models, replacing more simplified box models. With a current formulation of the coupled climate–C cycle model the difference in climate predictions found when different soil organic matter dynamics models were used is 0.4 °C; comparable with modelled temperature differences due to feedback to global warming from the interactive C cycle (e.g. Zeng *et al.*, 2004).

The simple box model (LPJ) produced results with stronger global temperature increases due to larger increases in soil respiration in warmer climates than the other two models. However, a different parameterization, with no change in model formulation, could probably reduce the disagreement between the LPJ soil biogeochemistry model and the other two models. Increasing the parameter  $f_{slow}$  (the fraction of decomposed litter going to the slow pool) should change the results so that the increase in soil respiration simulated by the LPJ soil biogeochemistry model would be lower. It should be noted that a parameterization from other



**Fig. 6** The differences between total carbon (C) density in the soil in 2100 and 1860 ( $\text{kg C m}^{-2}$ ) according to Lund–Potsdam–Jena (LPJ), ROMUL, Q models when coupled to the climate C cycle model Institute of Numerical Mathematics Climate Model (INMCM).

applications was used for the LPJ soil biogeochemistry model, whereas the ROMUL and Q models were specifically parameterized for this application.

The overall response of soil C to global warming strongly depends on two major, counteracting factors. Increased NPP as a result of  $\text{CO}_2$ -fertilization increases soil C storage, while higher soil temperatures and hence soil respiration rates decrease it. Our model results provide predictions of the net effects of these opposing factors. Our model predictions of the  $\text{CO}_2$ -fertilization effect are within the range of other global terrestrial vegetation model predictions (Friedlingstein *et al.*, 2006), but those do not take into account a number of confounding factors that may compensate for the effect of increased productivity (e.g. Moorcroft, 2006).

If NPP stays constant or increases less than currently predicted the currently parameterized LPJ soil biogeochemistry model might simulate smaller losses of C from the soil than the currently parameterized Q model (Fig. 4c and d).

Model derived C budget during 1980s and 1990s shows that the terrestrial biosphere was a sink of C (Table 2). The strength of this sink was within the range estimated by IPCC but lower than the mean values given in the report (IPCC, 2007b). The difference between the model results and the IPCC estimates may be due to the fact that the model does not take into account vegetation dynamics and such an important ecosystem process as disturbance by fire that has decadal dynamics (e.g. Bond-Lamberty *et al.*, 2007).

**Table 2** Residual terrestrial carbon sink ( $\text{PgC yr}^{-1}$ ) during 1980s and 1990s estimated by the models as compared with that of IPCC (IPCC, 2007b); negative fluxes are losses from the atmosphere

	1980s	1990s
LPJ_soilC	-0.6	-0.9
ROMUL	-0.8	-1.0
Q model	-0.5	-0.9
IPCC estimate	-1.7 (-3.4 to 0.2)	-2.6 (-4.3 to -0.9)

The temperature dependence of litter and soil organic matter mineralization has important effects on the strength of the feedback-effects of soil warming on atmospheric  $\text{CO}_2$ , and thus climate. Hence, these processes are subjects of intense scientific debate, although there is little consensus on many pertinent aspects (e.g. Davidson *et al.*, 2006; Kirschbaum, 2006). The models included in this study agree that the relative increase in the rate of decomposition when the temperature is increased is highest at low temperatures. The LPJ soil biochemistry and Q models use Arrhenius-type equations (Lloyd & Taylor, 1994) to account for this effect, while in ROMUL the temperature interval is split into regions of exponential (below a certain threshold) and linear (above it) temperature responses. However, only the LPJ soil biogeochemistry model predicts losses of C in the boreal region. All three models disagree on the relationship between temperature sensitivity and soil organic matter quality. In the LPJ model the temperature response is the same for all organic matter pools. In ROMUL the temperature response of mineralization of all compartments is the same in temperature intervals below  $20^\circ\text{C}$ , but above this threshold litter is more temperature-sensitive than more recalcitrant soil organic matter. In the Q model high-quality organic matter (e.g. fresh litter) is less temperature-sensitive than low-quality organic matter. However, despite these fundamental differences in formulation the ROMUL and Q models' predictions of global soil respiration in a warmer climate are nearly identical, indicating that differences in the temperature sensitivity of the decomposition of substrates of differing lability may not be critical factors.

An important variable to understand when attempting to predict climate changes is  $\text{CO}_2$  production by heterotrophic respiration. For this reason the mineralization rate and its relationship to temperature is being intensively investigated. However, even our simple sensitivity study shows that the humification rate is as important as the mineralization rate, thus its relationship to temperature and other factors also warrant further investigation.

The differences between our three models in soil C responses to changing climate and plant productivity are small in relation to results from other coupled climate-C cycle models, which predict that soils may lose as much as  $150 \text{ PgC}$  globally (Hadley model, in Zeng *et al.*, 2004) or accumulate as much as  $170 \text{ PgC}$  (IPSL model, in Zeng *et al.*, 2004) during the period 1860–2100. This variability reflects many factors, including (in addition to soil processes) the models' responses to increases in  $\text{CO}_2$  concentration, the strength of  $\text{CO}_2$ -fertilization effects and other factors influencing terrestrial productivity. For instance, Jones *et al.* (2005) reported that the Hadley model predicts a smaller decrease in global soil C when coupled to the RothC soil organic matter dynamics model (Jenkinson, 1990) than a version utilizing a single pool equation to describe soil C turnover.

There is still considerable scope for further development of models to improve simulations of the geographical variations in the area density of soil C. Notably, for modelling the subarctic and boreal regions it is important to account for effects of water-logging, which are currently ignored by the models. Changes in rates of soil respiration in these regions are likely to occur if relatively labile C, which is currently protected from decomposition by high soil moisture, is exposed to drier conditions (e.g. Carrasco *et al.*, 2006). Another area for which refinement is required is the Chernozem region, where the soil C storage is underestimated (although ROMUL simulates Chernozem soils reasonably well, if all conditions are appropriate; Chertov & Komarov, 1997), possibly due to inaccuracies in either the soil climate simulations or failure to account adequately for fine root productivity and turnover. Other factors that none of the models explicitly take into account are bonds between organic matter and mineral particles, decoupling between the mineralization and polymerization of humic compounds, solubility and mobility (e.g. Orlov, 1995).

Several suggestions for improving soil organic matter dynamics models have been recently published. Notably, Ågren & Wetterstedt (2007) have suggested that substrate supply, diffusion and uptake could be considered separately. Each factor influencing substrate availability could be specified in the model, and Michaelis–Menten kinetic models could replace accepted linear models of the decomposer growth rate (Davidson *et al.*, 2006). Such models are more difficult to parameterize, but developments in laboratory and field methods should foster development of such 'new-generation' models (Smith *et al.*, 2008). It should also be noted that there are a number of important factors that should ideally be considered in any thorough attempt to evaluate the role of soils in a changing

climate, such as nutrient cycling and C–N interactions (e.g. Comins & McMurtrie, 1993; Ågren *et al.*, 2001 and references therein), long-term or inherited soil history (e.g. Zimov *et al.*, 2006), ecosystem dynamics (e.g. Liski *et al.*, 1998; Chertov *et al.*, 2009) and management practises (e.g. Mikhailov *et al.*, 2004; Smith *et al.*, 2008; Chertov *et al.*, 2009). When the soil organic matter dynamics models are further developed, they should also be tested directly in a climate model, rather than drawing indirect conclusions regarding their likely implications for global climate predictions. In our study the range of uncertainty of future global temperature estimates due to discrepancies in predictions of soil respiration by three different soil organic matter dynamics models was  $<0.5^{\circ}\text{C}$ .

### Acknowledgements

We are grateful to two anonymous referees for their constructed comments. This work was supported by RFBR grants 06-05-64331, 06-05-64246 and 09-04-01209.

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## Appendix A

### Analytical equations for testing the effects of model parameterization

The differences between steady-state solutions corresponding to two different temperatures can be calculated as follows.

#### The LPJ soil biogeochemistry model

$$\Delta C = L_{02} \left( \frac{1}{k_{\text{lit}}(T_2)} + \frac{(1 - f_{\text{air}})f_{\text{inter}}}{k_{\text{inter}}(T_2)} + \frac{(1 - f_{\text{air}})f_{\text{slow}}}{k_{\text{slow}}(T_2)} \right) - L_{01} \left( \frac{1}{k_{\text{lit}}(T_1)} + \frac{(1 - f_{\text{air}})f_{\text{inter}}}{k_{\text{inter}}(T_1)} + \frac{(1 - f_{\text{air}})f_{\text{slow}}}{k_{\text{slow}}(T_1)} \right),$$

where  $T_1$  and  $T_2$  ( $^{\circ}\text{C}$ ) are the first and the second temperatures,  $L_{01}$  and  $L_{02}$  ( $\text{kg Cm}^{-2}\text{yr}^{-1}$ ) are the first

and the second rates of litter input,  $f_{\text{air}}$  is the fraction of decomposed litter that goes directly into the atmosphere,  $f_{\text{inter}}$  and  $f_{\text{slow}}$  are fractions of remaining decomposed litter that go into intermediate and slow pools, respectively, and  $k_{\text{lit}}$ ,  $k_{\text{inter}}$  and  $k_{\text{slow}}$   $\text{year}^{-1}$  are turnover rates of the litter, intermediate and slow pools, respectively.

#### Q model

$$\Delta C = \frac{L_{02}}{u_0 e^{-A/q_0 T_2}} \frac{e_0}{1 - e_0 - \eta_{12} e_0 A/T_2} - \frac{L_{01}}{u_0 e^{-A/q_0 T_1}} \frac{e_0}{1 - e_0 - \eta_{12} e_0 A/T_1},$$

where  $T_1$  and  $T_2$  (K) are the first and the second temperatures, respectively,  $L_{01}$  and  $L_{02}$  ( $\text{kg Cm}^{-2}\text{yr}^{-1}$ ) are the first and the second rates of litter input, respectively,  $u_0$   $\text{year}^{-1}$  is a parameter for microbial growth rate,  $A$  (K) is the activation energy for a step in the liberation of a C atom with quality  $q$ ,  $q_0$  is initial litter quality,  $e_0$  is microbial efficiency and  $\eta_{12}$  is a shift in quality parameter.