

# The turnover of organic carbon in subsoils. Part 2. Modelling carbon turnover

D. S. JENKINSON<sup>a,b</sup> & K. COLEMAN<sup>a</sup>

<sup>a</sup>Department of Soil Science, Rothamsted Research, Harpenden AL5 2JQ, and <sup>b</sup>Department of Soil Science, School of Human and Environmental Sciences, University of Reading, Reading RG6 6DW, UK

## Summary

A new model, RothPC-1, is described for the turnover of organic C in the top metre of soil. RothPC-1 is a version of RothC-26.3, an earlier model for the turnover of C in topsoils. In RothPC-1 two extra parameters are used to model turnover in the top metre of soil: one,  $p$ , which moves organic C down the profile by an advective process, and the other,  $s$ , which slows decomposition with depth. RothPC-1 is parameterized and tested using measurements (described in Part 1, this issue) of total organic C and radiocarbon on soil profiles from the Rothamsted long-term field experiments, collected over a period of more than 100 years. RothPC-1 gives fits to measurements of organic C and radiocarbon in the 0–23, 23–46, 46–69 and 69–92 cm layers of soil that are almost all within (or close to) measurement error in two areas of regenerating woodland (Geescroft and Broadbalk Wildernesses) and an area of cultivated land from the Broadbalk Continuous Wheat Experiment. The fits to old grassland (the Park Grass Experiment) are less close. Two other sites that provide the requisite pre- and post-bomb data are also fitted; a prairie Chernozem from Russia and an annual grassland from California. Roth-PC-1 gives a close fit to measurements of organic C and radiocarbon down the Chernozem profile, provided that allowance is made for soil age; with the annual grassland the fit is acceptable in the upper part of the profile, but not in the clay-rich Bt horizon below. Calculations suggest that treating the top metre of soil as a homogeneous unit will greatly overestimate the effects of global warming in accelerating the decomposition of soil C and hence on the enhanced release of CO<sub>2</sub> from soil organic matter; more realistic estimates will be obtained from multi-layer models such as RothPC-1.

## Introduction

In this paper, we use data from Part 1 (this issue, Jenkinson *et al.*, 2008) to develop a model for the turnover of organic C in subsoils. These data came from soil profiles located on the Rothamsted long-term field experiments, sampled over a period of more than a century. Thermonuclear bomb tests between 1955 and 1965 briefly doubled the concentration of radiocarbon in the atmosphere, inadvertently providing a tracer for the turnover of C in soil. A good model for C turnover must be able to account for secular changes in total organic C and in radiocarbon before and after the bomb tests. Bomb radiocarbon is a particularly useful tool for investigating soil fractions with long turnover times (Bruun *et al.*, 2005).

Several processes contribute to the movement of organic matter down the soil profile. Roots move downwards through the

soil, releasing exudates and sloughed-off organic material, ultimately to die and decompose *in situ*. Surface litter is moved downwards by earthworms and similar soil animals. Soluble organic matter descending from the upper soil horizons can be retained by lower horizons, as in podsolization. On the other hand, activities by creatures such as moles or ploughmen sometimes reverse this downward movement.

The first approach to modelling the migration of C down the profile (O'Brien & Stout, 1978; Van Dam *et al.*, 1997), is from the classical diffusion equation (Atkins, 1986)

$$\delta C / \delta t = D \delta^2 C / \delta z^2, \quad (1)$$

where  $C$  is carbon concentration,  $t$  time,  $z$  distance below the surface and  $D$  the diffusion coefficient (distance<sup>2</sup> time<sup>−1</sup>). The second (Nakane, 1978a, b) is based on convective mixing:

$$\delta C / \delta t = -v \delta C / \delta z, \quad (2)$$

where  $v$  is the coefficient of convection (distance  $\text{time}^{-1}$ ), sometimes called advection. Elzein & Balesdent (1995) used a model in which both processes were employed. Baisden *et al.* (2002b) used a three-compartment model, each compartment having a characteristic transfer coefficient that moves C and N downwards by advection. A small portion of material from the oldest compartment is passively returned to the surface. Feng *et al.* (1999) also used advective mixing, but with a single compartment model.

Models for the turnover of organic matter in subsoils must deal with the downward movement of C and with the stability of that C in the deeper soil layers. Some models (Nakane, 1978a; Van Dam *et al.*, 1997) assume that the decomposability of organic C declines with depth; that is, the decomposition rate 'constant' declines with depth. This is clearly in accord with gross respiration rate measurements (Chiba, quoted in Nakane, 1978a; Van Dam *et al.*, 1997; Lomander *et al.*, 1998; Gill *et al.*, 1999 – apart from the surface layer; Certini *et al.*, 2003). However, O'Brien & Stout (1978) and Elzein & Balesdent (1995) were able to model downward movement satisfactorily by assuming that soils contain a passive fraction of great age (and, in O'Brien and Stout's model, of uniform concentration down the profile), plus much younger 'active' fractions that decrease in concentration with depth, but have the same rate constants throughout the profile.

There is strong evidence that many soils contain a passive fraction of great age that is intimately and inextricably bound to clay and/or amorphous iron and aluminium oxides. Wattel-Koekkoek *et al.* (2003) showed that soil surface area, which is related to both clay mineralogy and hydrous oxide content, is negatively correlated to  $^{14}\text{C}$  activity and hence positively to radiocarbon age. Kaiser *et al.* (2002) and Masiello *et al.* (2004) observed that the major mechanism protecting subsoil C from biological attack was its close association with secondary hydrous oxides in the clay fraction. Likewise, Scharpenseel & Becker-Heidmann (1992), Scharpenseel *et al.* (1986), Torn *et al.* (1997) and Paul *et al.* (2001) all found that radiocarbon age was directly related to clay content. Clay-bound passive organic matter is not of course the only stable fraction; many soils contain carbonized materials such as ancient charcoal, coal or 'black carbon' (Schmidt, 2001; Skjemstad, 2001; Dai *et al.*, 2005; Jenkinson *et al.*, 2008).

Much recent work has been directed to calculating the concentration of C down the soil profile from that in the surface layer, often by means of an exponential relationship; see for example Minasny *et al.* (2006) or Hilinski (2001), who extended the CENTURY model to subsoils in this way. These papers are about how much C is held in the soil profile; here we are more concerned with the dynamics of this stock of C.

We have two aims in this paper. First, to use data from the Rothamsted soil profiles to develop a version of the dynamic topsoil model RothC-26.3 that can be applied to subsoils. Our second aim is to test the new model on data taken from the literature, where the necessary measurements (particularly of

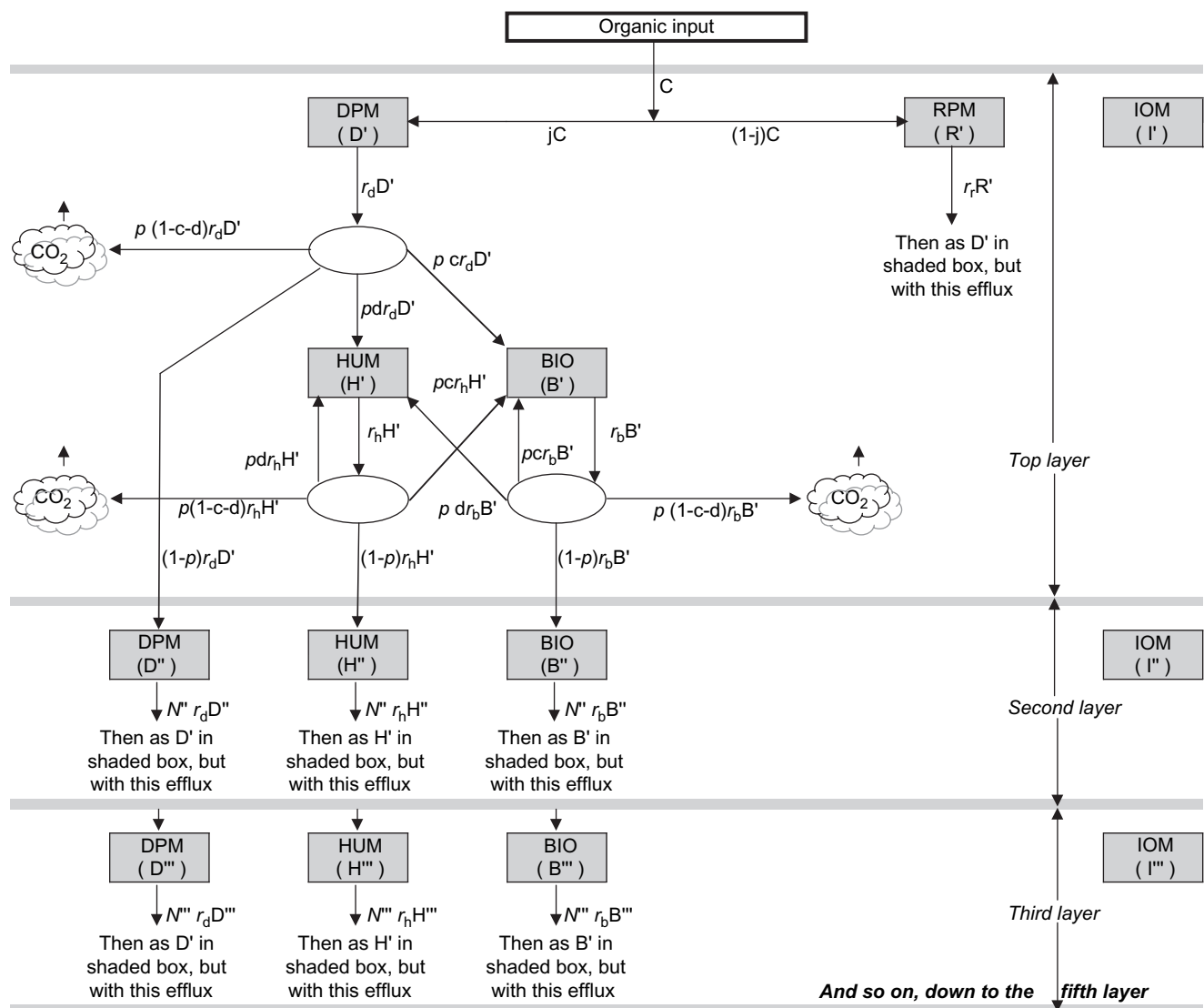
pre- and post-bomb radiocarbon) have been made throughout the top metre of soil (Baisden *et al.*, 2002b; Torn *et al.*, 2002).

### RothPC-1: the subsoil version of RothC-26.3

RothPC-1 is based on the earlier model RothC-26.3 (Coleman & Jenkinson, 1999), which was designed for use in topsoils, like most models for the turnover of C in soil. Smith *et al.* (1997) give a comparative review of nine topsoil models, including RothC. Petersen *et al.* (2005) discuss contemporary models. The structure of RothPC-1 (the PC standing for profile carbon) is set out in Figure 1. Incoming organic C is partitioned between two compartments, decomposable plant material (DPM) and resistant plant material (RPM), each of which decomposes by a first-order process. On decomposition, both DPM and RPM produce  $\text{CO}_2$ , microbial biomass (BIO) and humus (HUM), in the same proportions but at different rates, the rate constant for DPM being greater than that for RPM. Both BIO and HUM decompose by first-order processes, again with different rate constants, to give more  $\text{CO}_2$ , BIO and HUM. Thus the DPM and RPM compartments of RothC-26.3 decompose by Bruun *et al.*'s (2004) non-feedback pathway, the BIO and HUM compartments by their feedback pathway. Each layer of the soil also contains an inert organic matter (IOM) compartment, containing C that was in place before the commencement of any of the model runs and neither decomposes nor moves during any of the model runs. Different layers are permitted different amounts of IOM, but the IOM has the same radiocarbon age (50 000 years, i.e. with a  $\Delta^{14}\text{C}$  of -998) in all of the layers. This age was chosen so that IOM contains virtually no radiocarbon. We visualise the IOM compartment as a receptacle for ancient clay-bound organic matter, but it may also contain ancient carbonized material such as coal. In the model, all five compartments, DPM, RPM, HUM, BIO and IOM, are defined by the amount of C they contain, not the amount of organic matter. Like RothC-26.3, RothPC-1 uses a monthly time step to calculate total organic C ( $\text{t ha}^{-1}$ ), microbial biomass C ( $\text{t ha}^{-1}$ ) and  $\Delta^{14}\text{C}$  (from which a radiocarbon age may be calculated), on a years-to-centuries time scale.

The rate constants in the top layer (expressed for convenience in  $\text{year}^{-1}$ , although the model runs in monthly steps, monthly rates being obtained by dividing the annual rate by 12) are the same as those used in RothC-26.3 for the DPM compartment ( $10 \text{ year}^{-1}$ ), the RPM compartment ( $0.3 \text{ year}^{-1}$ ) and the HUM compartment ( $0.02 \text{ year}^{-1}$ ). The rate constant for the BIO compartment is  $1.5 \text{ year}^{-1}$ , compared with a value of  $0.66 \text{ year}^{-1}$  in RothC-26.3.

RothPC-1 divides the top metre of soil into five layers, of equal thickness, although in this paper only four are used. These four are the historic 9-inch Rothamsted sampling layers, giving layers of 0–23, 23–46, 46–69 and 69–92 cm. The input of organic C (from litterfall, dead roots, etc.) is postulated to enter the top layer alone.



**Figure 1** Structure of RothPC-1. Incoming organic C is split between the DPM compartment (decomposable plant material) and the RPM compartment (resistant plant material), with a fraction  $j$  going to DPM and  $(1-j)$  to RPM. The DPM compartment decomposes with a rate constant  $r_d$  month<sup>-1</sup>, the RPM at  $r_r$  month<sup>-1</sup>. Most soil organic C is in the HUM (humus) compartment, with a rate constant  $r_h$  month<sup>-1</sup>. The microbial biomass C compartment (BIO) has a rate constant  $r_b$  month<sup>-1</sup>. All four rate 'constants' are adjusted (see text) for temperature, vegetational cover and soil water content. The IOM (inert organic matter) compartment contains C that is totally resistant to decomposition. A letter in parentheses inside a shaded box represents the total stock of C in that box, not the quantity formed each month. Primes indicate layers; thus (H') is the quantity of HUM C in the surface layer, (H'') in the second layer. The factor  $N$ , applied to all except the IOM compartment, slows decomposition in the lower layers (see text); in the top layer it is 1 and not shown. An oval represents the quantity of C mobilized each month from a particular compartment. Thus, a quantity  $r_h H'$  is mobilized each month from the HUM compartment; of this, a part  $p$  stays in the same layer, some ( $pc r_h H'$ ) as newly-formed BIO, some ( $pd r_h H'$ ) as newly-formed HUM, and some ( $p(1-c-d) r_h H'$ ) is liberated from that layer as  $\text{CO}_2$ -C and then lost to the atmosphere. A quantity  $(1-p) r_h H'$  of HUM moves to the layer below.

This input is assumed to have the same radiocarbon content as the atmosphere at the time it enters the soil; the radiocarbon content of the atmosphere at the appropriate date is taken from an internal table, just as in RothC-26.3. Any lag between between photosynthesis, which creates plant tissue with the current radiocarbon composition of the atmosphere, and the time when this same plant tissue dies and starts to decompose, is ignored.

#### *Effects of vegetational cover, temperature and soil water content on the rate constants*

The rate constant for each compartment (except IOM) is modified by a multiplier for vegetational cover, a multiplier for temperature and a multiplier for soil water content. When the soil is vegetated, decomposition is slowed in all compartments (except

IOM) by the retainment factor, just as in RothC-26.3. This retainment factor is applied to the top layer alone and has the same value (0.6) as in RothC-26.3. Decomposition rates in all compartments and in all five layers (shown in italics on Figure 1) are modified for temperature, using monthly mean air temperatures, exactly as in RothC-26.3. The rate constant for each compartment (except IOM) is multiplied by a quantity  $A$ , given by

$$A = 47.9 / (1 + \exp(106 / (tm + 18.3))), \quad (3)$$

where  $tm$  is the monthly mean air temperature. It is applied to all model compartments alike – a contentious oversimplification, as pointed out by Knorr *et al.* (2005) and by Davidson & Janssens (2006). This function was set (Jenkinson *et al.*, 1987) by measuring the rate of decomposition of the same  $^{14}\text{C}$ -labelled plant material in the same soil in two contrasting climates. One was at Rothamsted, in England (10 year incubation in the field; mean annual temperature 8.9°C), and the other was at the International Institute of Tropical Agriculture Ibadan, in Nigeria (5 year incubation in the field; mean annual temperature 26.1°C; for details see Ayanaba & Jenkinson, 1990). The decomposition curves could be almost perfectly superimposed if one Nigerian year was equivalent to four English years. In these experiments with  $^{14}\text{C}$ -labelled plant material, decomposition was dominated by the HUM compartment, which makes up by far the largest part of the residual labelled plant C, once the initial months of decomposition are over. The assumption in RothPC-1 that monthly air temperature applies to all five layers is acceptable in the humid tropics, but will be increasingly wrong in higher latitudes, where the swing between seasons is larger and there is a marked seasonal lag in temperature down the profile.

Decomposition rates are modified for the water content of each layer as follows.

The maximum soil moisture deficit (Max SMD) for each layer is first calculated from the expression

$$\text{Max SMD} = -[20 + 1.3(\% \text{ clay}) - 0.01(\% \text{ clay})^2], \quad (4)$$

just as in RothC-26.3. This is for a layer 23 cm in thickness, with % clay being the clay content of the specified layer.

When the site is vegetated and monthly potential evapotranspiration exceeds rainfall, each layer, starting from the top and moving down the profile, dries out until it reaches the Max SMD for that layer. When the soil wets up again, each layer fills up to its Field Capacity, starting from the top, and moving, layer by layer, down the profile. The rules are different when the soil is not vegetated, for example during the period between harvest and the emergence of the next crop. Then the Max SMD for the top 23 cm (as given by Equation (4)) is multiplied by 0.556 (just as in RothC-26.3) to give Bare SMD. The top layer is not allowed to dry out further than Bare SMD. If the SMD is already less than Bare SMD, the top layer cannot dry out any further. Lower layers of bare soil do not dry out any further; although they can already be dry, say just before harvest.

The rate constants for the DPM, RPM, BIO and HUM compartments are then modified for the water content of each layer, exactly as in RothC-26.3. Thus the rate constant is multiplied by a factor that stays at 1 between Field Capacity and 0.444 of Max SMD as the soil dries out. This factor then falls linearly from 1 at 0.444 of Max SMD to 0.2 at Max SMD as the soil continues to dry, but cannot decrease further than this, no matter how dry the soil becomes. RothPC-1 should be used with great caution on desert soils. This is because RothPC-1 (and, for that matter, RothC-26.3) is primed to start in the first month after monthly precipitation exceeds monthly evapotranspiration. In very dry places, monthly evapotranspiration is greater than monthly precipitation throughout the year and the model then runs with the moisture rate multiplier set at 0.2 throughout. Nor should RothPC-1 be used if decomposition is restricted throughout the year by water-logging. It should therefore not be used on undrained peat.

#### *Influence of clay on decomposition*

In RothC-26.3 clay content (as % clay) does not directly alter the rate of decomposition, as for example, does temperature. Instead, clay has two indirect effects on decomposition: (i) it influences the quantity of water that the soil can hold and (ii) it affects the ratio ( $\text{CO}_2\text{-C evolved}$ )/(newly formed (BIO + HUM) retained in the soil) whenever organic matter undergoes decomposition (Coleman & Jenkinson, 1999). Similar calculations are made in RothPC-1 for each layer, using the appropriate clay content of that layer (for details see below).

#### *Rate constants and soil depth*

The rate constants of the DPM, RPM, BIO and HUM compartment all decrease with depth. This is done by multiplying the RothC-26.3 rate constant for a particular compartment in the layer in question by a factor  $N$  calculated from the sigmoid relationship:

$$N = (-1 / (1 + (\exp(-s(F - f))))) / (-1 / (1 + (\exp(-s(-f))))) \quad (5)$$

where  $s$  is a constant, in  $\text{cm}^{-1}$ ,  $F$  the distance, in cm, from the middle of the top layer to the middle of the layer in question, and  $f$  is the distance, in cm, from the surface to the middle of the top layer. For the Rothamsted data used in this paper,  $f$  is always 11.5 cm. For the top layer,  $F$  is zero, so  $N$  becomes 1. The constant  $s$ , which slows decomposition as depth increases, was set during fitting (see below).

#### *Downward movement of organic C*

Carbon moves downwards from the DPM, RPM, HUM and BIO compartments (see Figure 1). Upward movement by inversion ploughing or bioturbation is ignored. At the end of each month, the amount of C 'mobilized' during that month from

each compartment is calculated. The term ‘mobilized’ covers both C subjected to microbial attack and C moved to the layer below. Consider, for example, the HUM compartment in the top layer, containing  $H'$  t C ha<sup>-1</sup>. The quantity mobilized in a month is  $r_h H'$ , where  $r_h$  is the rate constant, with dimensions years<sup>-1</sup>, modified as necessary by temperature, plant cover and water content. Of this  $r_h H'$ , a part  $p$  stays in the top layer, some ( $pc r_h H'$ ) as newly-formed HUM, some ( $pd r_h H'$ ) as newly-formed BIO and some ( $p(1-c-d) r_h H'$ ) is liberated as CO<sub>2</sub>-C and then lost to the atmosphere (Figure 1). The terms  $p$ ,  $c$  and  $d$  are dimensionless fractions. A quantity  $(1-p) r_h H'$  of HUM moves to the layer below. Similar calculations (see Figure 1) are done every month for organic C in each compartment and in each layer, allowing of course for the clay content of the particular layer.

RothPC-1 is internally adjusted so that the quantities of newly-synthesized humus and biomass formed during mobilization from any topsoil compartment are the same as the corresponding quantities formed each month in RothC-26.3. This is done as follows.

In RothC-26.3, monthly decomposition in any compartment gives CO<sub>2</sub>, BIO and HUM (where BIO denotes newly-formed biomass and HUM newly-formed HUM) in proportions given by the equation (Coleman & Jenkinson, 1999):

$$\begin{aligned} (\text{CO}_2\text{-C})/(\text{BIO} + \text{HUM}) \\ = 1.67(1.85 + 1.60\exp(-0.0786(\%\text{clay}))) = Z. \end{aligned} \quad (6)$$

But, again taking the HUM compartment as an example,

$$r_h H' = \text{CO}_2\text{-C} + \text{BIO} + \text{HUM}. \quad (7)$$

Substituting and rearranging,

$$(\text{BIO} + \text{HUM}) = r_h H'/(Z + 1). \quad (8)$$

Taking the HUM compartment as an example, Figure 1 shows that in RothPC-1 the ratio between the quantity of CO<sub>2</sub>-C

formed in the top layer at the end of one month and the quantities of BIO and HUM formed is given by:

$$\begin{aligned} (\text{CO}_2\text{-C})/(\text{BIO} + \text{HUM}) = [(p r_h H') \\ - (pc r_h H' + pd r_h H')]/(pc r_h H' + pd r_h H'). \end{aligned} \quad (9)$$

But we wish to make the quantity of new biomass and HUM formed in the top layer each month the same in RothC-26.3 and RothPC-1, so, for RothPC-1:

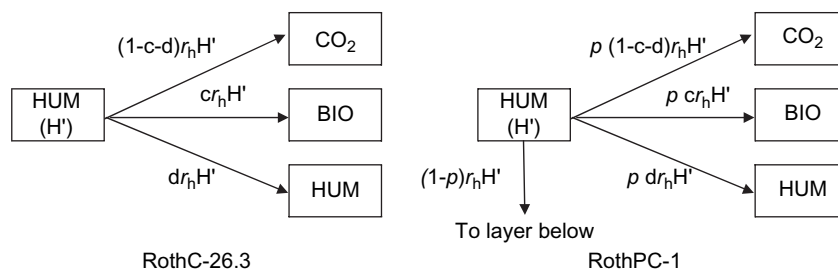
$$\begin{aligned} (\text{CO}_2\text{-C})/(\text{BIO} + \text{HUM}) = [(p r_h H') \\ - (r_h H'/(Z + 1))]/(r_h H'/(Z + 1)), \end{aligned} \quad (10)$$

$$= p(Z + 1) - 1. \quad (11)$$

In essence, C is moved downwards, at the expense of CO<sub>2</sub> released by the topsoil (Figure 2). Thus, if  $p$  is 1, no C moves down to the layer below and RothPC-1 becomes like RothC-26.3; as  $p$  becomes less than 1, more C moves down the profile, at the expense of CO<sub>2</sub>. The parameter  $p$  can only vary between a maximum of 1 and a minimum that depends on the clay content of the layer in question. Thus the minimum is 0.19 for a clay content of 10%, 0.22 for 20% clay and 0.24 for 50%. The model should not be used below the minimum value of  $p$  for a particular clay content; to do so implies that all of the C mobilized in a month moves downward and none is available for microbial attack – a somewhat implausible assumption.

As the quantity of C moving down each month is directly proportional to the concentration of HUM in a specified layer, RothPC-1 is a variant of the convective mixing model (Equation (2)). Because all C inputs are postulated to enter the top layer, it might be thought that direct inputs to lower layers from (say) roots are ignored. This is not so: each month some material from each of the DPM and RPM compartments is mobilized and moved downwards (see Figure 1), in effect mimicking direct inputs of fresh organic material to lower layers.

RothC-26.3 was originally calibrated using long-term field data for topsoils (Jenkinson & Rayner, 1977; Jenkinson *et al.*, 1987). Many people have used RothC-26.3 and its variants to



**Figure 2** Differences between RothC-26.3 and RothPC-1 in the way the same quantity of organic C that is mobilized each month from an individual compartment (here illustrated by the HUM compartment) is channelled. In both models,  $c$  is the fraction of the HUM decomposed in one month that goes to new BIO,  $d$  the fraction that goes to new HUM. The values of  $c$  and  $d$  are the same in both models and are calculated from Equation 6 and from the relationship  $c/d = 0.852$ .

investigate the dynamics of soil organic C; they will get virtually the same answers for topsoils if they use RothPC-1. The differences between the two models in the rate constant for the BIO compartment will have little effect on generated values for total organic C, because biomass C is commonly only 1–2% of the total organic C in topsoils. The two models generate identical outputs for topsoil if the BIO rate constant in RothC-26.3 is changed from 0.66 to 1.5 year<sup>-1</sup>.

Another advantage of the procedure used to move C down the profile is that movement is assumed to be subject to temperature, soil water content, etc., exactly as is decomposition in RothC-26.3. This is plausible – for example, downward movement is less likely in a dry soil or in a very cold soil – although whether the assumption is correct remains to be seen.

#### Fitting RothPC-1 to the data

For the four Rothamsted sites, rainfall, air temperature and evaporation are 30-year mean monthly values for Rothamsted, as given by Coleman & Jenkinson (1995). The clay contents of the four layers on the Broadbalk and Park Grass sites are taken to be 23, 30, 50 and 49%, starting from the top; for the Geescroft site, 21, 43, 55 and 63% (see Jenkinson, 1971). Arable soils under winter wheat are deemed to be bare in August, September, October and November; thereafter vegetated throughout the year. Grassland and woodland soils are taken to be continuously vegetated. For soils reverting to woodland, the DPM/RPM ratio is taken to be 1.44, the usual value for arable crops and grassland, for the first 25 years and 0.25 thereafter, as grasses and herbs died out and trees became dominant (Jenkinson *et al.*, 1992).

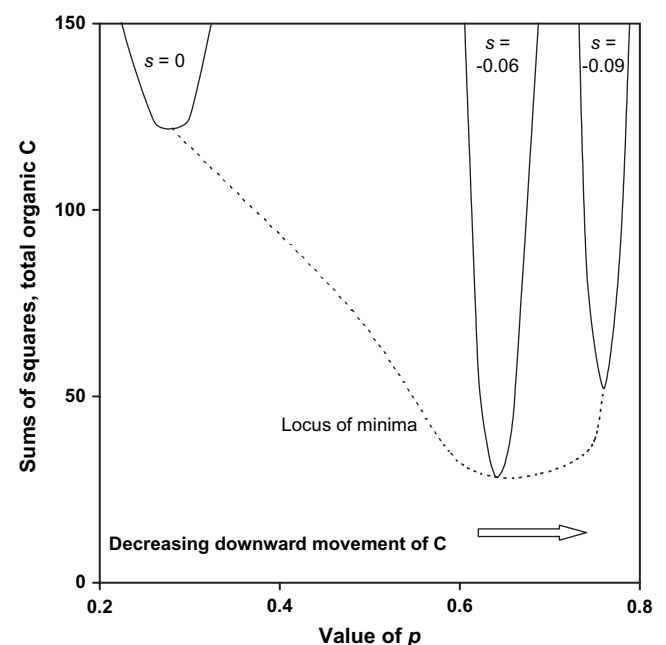
The model was first run to equilibrium (taken as a run of 10 000 years) by iteratively adjusting the annual input of organic C (C in Figure 1) and the size of the IOM compartment (I' in Figure 1) to give the measured total organic C and radiocarbon contents of the 0–23 cm layer of the earliest soil (pre-bomb) sample. Using the value of IOM thus found, the annual input of C was iteratively adjusted to give the best fit to measured total organic C in the 0–23 cm layer for all of the following samplings, best fit being defined as that giving least squares of differences between measurement and prediction. Note that the data used in this fitting did not come exclusively from Jenkinson *et al.* (2008); some came from Jenkinson *et al.* (1992).

The next stage was to obtain the optimal values of the parameters  $p$  and  $s$  for each site, using the measured values of % modern for each layer to calculate the IOM content of that layer. The IOM content of a particular layer is assumed to be always the same, whenever sampled. An arbitrary IOM concentration was first set for each of the three lower layers. The model was then run for a specified value of  $p$  and a specified value of  $s$  to give the total organic C contents and the radiocarbon contents (as % modern) of all four horizons for each of the sampling dates. The modelled radiocarbon content of each layer was averaged over all the sampling dates. The average measured radiocarbon content (again as % modern) of

that layer was also averaged over all the sampling dates. The IOM content of that layer was then adjusted iteratively until the measured and modelled values of % modern were the same. The procedure was repeated for each of the three lower layers. IOM thus calculated was added to the calculated total organic C and the initial arbitrary value subtracted. The differences between the total organic C thus calculated and the corresponding measured values were squared. The squares for the three lower horizons (i.e. excluding the 0–23 cm layer) at all the different sampling times were added, to give a sum of squares,  $Q$ . The parameter  $p$  was varied systematically, each time readjusting IOM as described above, until a minimum value of  $Q$  was found for the specified value of  $s$ . The parameter  $s$  was then varied systematically and the whole procedure repeated until a minimum value of  $Q$  for both  $p$  and  $s$  was established. Figure 3 illustrates this process for the Geescroft Wilderness site. The best fit values of  $s$  (shown as the dotted line on Figure 3) improved as  $s$  moved from 0 (i.e.  $N = 1$  and the rate constants are not affected by depth) to a flat minimum at about  $-0.06$  to  $-0.07$  and then rose again. Best fits were very sensitive to  $p$ , particularly as  $p$  tended towards 1.

#### Statistics

A range of statistical tests was used to help assess the fit between measurement and model prediction. These tests were applied to



**Figure 3** Sum of squares for the differences between measured and modelled values of total organic C (expressed as t ha<sup>-1</sup>), for varying values of  $s$  and  $p$ . The measurements are from the 23–46, 46–69 and 69–92 cm layers of the 1883, 1904, 1965 and 1999 samplings of Geescroft Wilderness, excluding the top layer. See text for details of the fitting process and the model settings.

the lower three layers alone, as data from the 0–23 cm layer had already been used to set annual input and IOM. In any case, this paper is about the validation of a subsoil model, RothPC-1. A  $t$ -test (as used by Sutherland *et al.*, 1986) was applied separately to each of the three layers. The Mean Sum of Squares was calculated for each profile, excluding the 0–23 cm layer. Whitmore's (1991) lack of fit statistic LOFIT was calculated and used to obtain the variance ratio  $F$  ( $= \text{MSLOFIT}/\text{MSE}$ ), where MSLOFIT is the mean squares due to lack of fit and MSE is the mean squares due to error in the measurements (Smith *et al.*, 1997). If  $F$  thus calculated was greater than the critical 5% value, the discrepancy between measurement and prediction was too large to be explained by errors in the measurement alone; the measurements are sufficiently precise for further improvements to the model to be worthwhile. If  $F$  is less than the critical 5% value, nothing is to be gained from a better model. In pooling data from the 23–46, 46–69 and 69–92 cm layers to obtain  $F$  in this way, we assume that the variances are similar in the different layers and the data for each layer are independent and normally distributed. Measurements made on three different layers from the same sampling hole (as ours were) cannot be regarded as truly independent, so conclusions based on  $F$  should not be pressed too far.

#### *Modelled and measured total C and radiocarbon in the Rothamsted sites*

Table 1 shows the optimized values for  $p$  and  $s$  and the resulting values of total organic C, radiocarbon and IOM for the 1999 sampling of the four Rothamsted sites. It should be noted that the statistical tests in Table 1 are based on the 1999 sampling alone, because this was the only sampling with error measurements. However, the values of  $p$ ,  $s$  and IOM in Table 1 were set using all the data for the 23–46, 46–69 and 69–92 cm layers from all the sampling dates, so that they are not necessarily the best for 1999 alone.

In all four sites the top three layers contained about 4 t IOM C ha<sup>-1</sup>, with the bottom layer being more variable. For Geescroft Wilderness, the model gave a fit ( $F = 3.0$ ; Table 1) for the total organic C measurements that was less than the 5% level of significance for errors in the measurements themselves ( $F = 3.8$ ; Table 1). The modelled fit for radiocarbon ( $F = 5.9$ ) was greater than the 5% level of significance for the measurements ( $F = 5.1$ ), so that the model lack of fit was marginally greater than that of the measurements.

Total C and radiocarbon are plotted against time for Geescroft Wilderness (our best site; see Jenkinson *et al.*, 2008) in Figure 4(a,b). The model consistently overshoots the total C measurements taken in 1965 and undershoots those taken in 1999. Differences in sampling procedure may be responsible, although a more likely explanation is that the annual returns of C to the soil increased during reversion to woodland, whereas the model input is set at 2.76 t C ha<sup>-1</sup> year<sup>-1</sup> throughout the period 1883–1999. Figure 4b reveals more serious

problems – at least for the 23–46 cm layer. The radiocarbon content of this layer is modelled too old in 1883 and too young in 1999. There was a similar discrepancy in Broadbalk Wilderness (not shown). We have no explanation. The model shows that the deeper the soil layer, the longer it takes for bomb radiocarbon to attain its maximum value (Figure 4b). The models used by Wang *et al.* (1999) and by Telles *et al.* (2003) showed similar lags, but for shallower soil layers.

There was a close relationship between modelled and measured cumulative total organic C down the profile of Geescroft Wilderness (Figure 5a), which held for both the 1883 and 1999 samplings. A similar relationship held for radiocarbon, here expressed in units of mass (Figure 5b).

For Broadbalk Wilderness and Broadbalk Arable, both  $t$  and  $F$ -tests indicate that the model predictions for total organic C and radiocarbon were within the errors in the actual measurements, so that there is no gain to be had from a better model. It should be noted that the  $F$ -test in Table 1 is insensitive for the Broadbalk Wilderness and Broadbalk Arable radiocarbon data, with only two replicate measurements per layer; for the same reason we do not give the  $t$  values.

The discrepancies between measured and modelled data are greater for Park Grass than for the other three sites (Table 1); in particular, the modelled value for total organic C in the 69–92 cm layer of Park Grass is too small, showing up as a very large  $F$  value. The best fit was for a  $p$  of 0.82 and an  $s$  of  $-0.07$ . In calculating this fit, the measured organic C contents of the 0–23 cm layer were corrected for coal C as described by Jenkinson *et al.* (2008); the lower layers were assumed to be coal-free. Aggregating the Park Grass measurements made on different fertilizer and liming treatments (as in Table 1) could have been a mistake, although attempts to fit the fertilized treatment (plots 14/2c and 14/2d) by itself made little improvement. We have modelled Park Grass assuming it was never cultivated, despite the marks of medieval ploughing. Park Grass has certainly been under grass since the beginning of the 18th century, although this may not have been long enough for organic C to recover after the disturbance caused by ancient cropping. There is evidence (see Jenkinson *et al.*, 2008, Table 3; this issue) that organic C has been accumulating throughout the profile over the last 100 years.

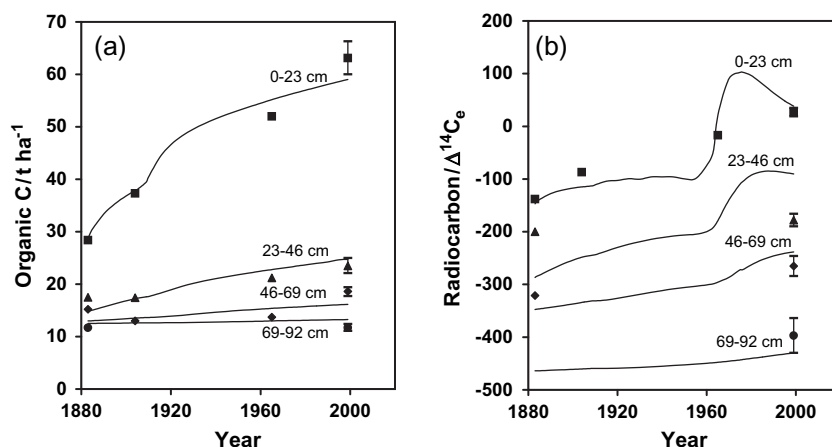
Errors will be introduced by the assumption in RothPC-1 that there is no lag between the uptake of atmospheric CO<sub>2</sub> by vegetation, with its characteristic  $\Delta^{14}\text{C}$  value for a particular year, and the start of decomposition of this plant C in the soil. This assumption is not far wrong on the Broadbalk Arable site, where the crop is annual winter wheat and the lag will be less than a year. In the old grassland site (Park Grass) the lag will be a little greater, with some plant roots living for a year or two (Jenkinson *et al.*, 2004). On the calcareous site under regenerating deciduous woodland (Broadbalk Wilderness) there is no carry-over of surface litter from year-to-year because of earthworm activity, although there will be a lag of a few years between the formation and death of fine roots. The

**Table 1** Organic C and radiocarbon in four soil profiles, as measured in 1999 and as calculated by RothPC-1

Site	Total organic C					IOM-C		Radiocarbon		
	<i>p</i>	<i>s</i>	Layer /cm	Measured /t ha <sup>-1</sup> ± SE	Calculated /t ha <sup>-1</sup>	Statistical tests <i>t</i>	Calculated /t ha <sup>-1</sup>	Measured / % modern ± SE	Calculated / % modern	Statistical tests <i>t</i>
Geescroft Wilderness	0.64	-0.06	0-23	63.1 ± 3.15 (8) <sup>a</sup>	59.0		3.93	102.7 ± 0.82 (4) <sup>a</sup>	103.8	
[Input pre-1883: 1.39 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			23-46	23.5 ± 1.43 (8)	24.7	-0.84	4.05	82.2 ± 1.16 (4)	91.0	-7.58
[Input post-1883: 2.76 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			46-69	18.6 ± 0.87 (8)	16.1	2.80	4.08	73.5 ± 1.94 (4)	76.2	-1.39
			69-92	11.7 ± 0.68 (8)	13.3	-2.22	4.57	60.3 ± 3.33 (4)	57.0	1.01
<i>t at 0.05</i>						2.37				3.18
Mean sum squares <sup>b</sup>						10.0				95
<i>F<sup>c</sup></i>						<b>3.0 (3.8)<sup>d</sup></b>				<b>5.9 (5.1)<sup>d</sup></b>
Broadbalk Wilderness	0.72	-0.08	0-23	78.0 ± 4.53 (4)	74.5		3.93	104.4 ± 0.58 (2)	105.9	
[Input pre-1881: 1.35 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			23-46	25.6 ± 1.18 (4)	27.6	-1.68	3.99	84.7 ± 2.67 (2)	93.4	-3.26
[Input post-1881: 3.65 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			46-69	19.4 ± 1.76 (4)	19.4	0.00	4.77	73.3 ± 4.14 (2)	74.2	-0.21
			69-92	17.3 ± 1.96 (4)	13.6	1.89	2.00	67.1 ± 4.22 (2)	59.9	1.71
<i>t at 0.05</i>						3.18				
Mean sum squares <sup>b</sup>						17.9				128
<i>F<sup>c</sup></i>						<b>2.1 (5.1)</b>				<b>3.1 (15.4)</b>
Broadbalk Arable	0.49	-0.04	0-23	33.4 ± 1.67 (6)	32.9		3.93	92.3 ± 3.93 (2)	96.1	
[Input pre-1843: 1.35 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			23-46	19.3 ± 0.55 (6)	18.1	2.29	4.33	77.1 ± 3.28 (2)	80.9	-1.16
[Input post-1843: 1.67 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			46-69	15.3 ± 0.58 (6)	15.1	0.33	5.15	68.8 ± 2.15 (2)	66.8	0.92
			69-92	13.9 ± 0.93 (6)	13.0	1.00	5.71	57.5 ± 1.41 (2)	53.6	2.73
<i>t at 0.05</i>						2.57				
Mean sum squares <sup>b</sup>						2.4				33
<i>F<sup>c</sup></i>						<b>1.6 (4.2)</b>				<b>1.9 (15.4)</b>
Park Grass <sup>e</sup>	0.82	-0.07	0-23	75.2 <sup>f</sup> ± 2.33 (6)	75.2		3.93			
[Input pre-1856: 3.20 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			23-46	25.7 ± 0.88 (6)	23.7	2.32	3.63	86.4 ± 1.17 (6)	87.1	-0.59
[Input post-1856: 3.20 t C ha <sup>-1</sup> year <sup>-1</sup> ] <sup>g</sup>			46-69	16.2 ± 0.50 (6)	14.5	3.40	4.23	75.0 ± 1.16 (6)	67.3	6.58
			69-92	12.8 ± 0.85 (6)	8.4	5.08	2.68	60.3 ± 2.59 (6)	52.1	3.15
<i>t at 0.05</i>						2.57				2.57
Mean sum squares <sup>b</sup>						25.8				126
<i>F<sup>c</sup></i>						<b>14.7 (4.2)</b>				<b>13.4 (4.2)</b>

<sup>a</sup>Number of determinations in parentheses.<sup>b</sup>Mean sum squares of differences between modelled and measured values: lower three layers only.<sup>c</sup>*F* = (mean squares due to lack of fit)/(mean squares due to error) = (MSLOFIT)/(MSE). Lower three layers only.<sup>d</sup>Values in parentheses are for *F* at the 5% level, using the appropriate *df* for MSLOFIT and MSE. If the calculated value of *F* (in bold) is greater than that in parentheses, the measurements are sufficiently precise for further improvement to the model to be worthwhile.<sup>e</sup>Mean of plots 3b, 3d, 14/2c and 14/2c.<sup>f</sup>Coal-free basis - see text.<sup>g</sup>Carbon input to top layer.





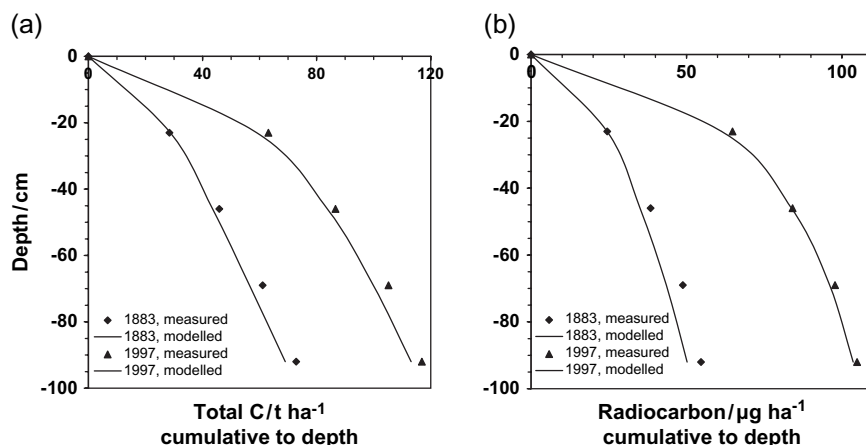
**Figure 4** (a) Changes in organic C down the Geescroft Wilderness profile over the period 1883–1999, as modelled (lines, using  $p = 0.64$  and  $s = -0.06$ ) and as measured. See text for the model settings. Bars show standard errors. (b) Changes in radiocarbon ( $\Delta^{14}\text{C}_e$ ) down the Geescroft Wilderness profile over the period 1883–1999, as modelled (lines) and as measured.

most serious error will be in Geescroft Wilderness, which is also deciduous, but where the soil is too acid for the larger earthworms and a litter layer has accumulated. However, in our work, this litter layer was incorporated into the 0–23 cm mineral soil layer prior to analysis. The modelled radiocarbon content (expressed as % modern) of the 0–23 cm layer sampled in 1999 from Geescroft Wilderness was 103.8% (see Table 1); had there been a 5-year lag it would have been 105.0%, a difference of only 1.2%. The corresponding figures for the 69–92 cm layer were 57.0 and 56.9%. Introducing an arbitrary 5-year lag thus made the fit between measured and modelled radiocarbon in Table 1 marginally worse. Needless to say, a 5-year lag has a greater effect in one of the more dynamic fractions than in soil organic C as a whole. For the same 0–23 cm layer collected in 1999 from Geescroft Wilderness, modelled % modern in the DPM compartment was 110.1%; with a 5-year lag, 113.1%. Although lag effects can be important in litter and in thin soil layers (Hahn & Buchmann, 2004; Torn *et al.*, 2005), they are small for total radiocarbon in the thick mineral soil layers used in this paper.

#### Downward movement of C at Rothamsted

The values of  $p$  in Table 1 indicate that a greater proportion of the incoming organic matter moves downwards over a year in Geescroft Wilderness ( $p = 0.64$ ) than in Broadbalk Wilderness ( $p = 0.72$ ). This is in accord with the pH of the two sites: downward movement of dissolved organic C is more likely in the strongly acid Geescroft (pH 4.5 in the 0–23 cm layer in 1999) than in the calcareous Broadbalk (pH 7.7). The small value of  $p$  in Broadbalk Arable ( $p = 0.49$ ) is consistent with some soil mixing between the top two layers (in the 1980s ploughing occasionally went a little below 23 cm) and deep rooting (*c.* 1.2 m) by the winter wheat. Organic C is less prone to moves downward in Park Grass ( $p = 0.82$ ) than in any of the other three sites ( $p$  ranging from 0.72 to 0.49). Grasslands are generally more shallow-rooted than temperate deciduous forest (Jobbágy & Jackson, 2000), so that the direct input of C from roots dying *in situ* will be less in the lower horizons.

Table 1 gives optimal values of  $p$  (0.72) and  $s$  (−0.08) for Broadbalk Wilderness that are very different to the corresponding values for Broadbalk Arable (0.49 and −0.04, respectively). Yet in 1881 the starting values for the Wilderness



**Figure 5** (a) Measured and modelled organic C down the Geescroft Wilderness profile, cumulative basis. (b) Measured and modelled radiocarbon down the Geescroft Wilderness profile, cumulative basis, assuming that 'modern' radiocarbon contains  $10^{-12} \text{ g } ^{14}\text{C g}^{-1}$  total C.

(when it was fenced off) must have been close to those now found for Broadbalk Arable. It follows that  $p$  increases (i.e. a smaller fraction of the incoming C moves downward) when ploughing ceases and the site reverts to woodland. At the same time,  $s$  becomes more negative; that is, the restriction of decomposition with depth becomes greater.

### Measured and modelled soil microbial biomass

Castellazzi *et al.* (2004) measured soil microbial biomass C in the same samples taken in 1999 from Geescroft and Broadbalk Wildernesses that are used in this paper. We used RothPC-1 to model biomass in these two sites, using an arbitrary rate constant ( $1.5 \text{ year}^{-1}$ ) for the BIO compartment; subsequent work (not reported) showed that  $1.3 \text{ year}^{-1}$  would have been a better choice. Although of the right order, agreement is poor, particularly for Geescroft Wilderness, in which the modelled values are too large, except for the topsoil (Table 2). The fit for Broadbalk Wilderness is closer for the 23–46 and 46–69 cm layers, but not for the other layers. Geescroft is strongly acid (mean pH of the 0–23 cm layer in 1999 was 4.5), with a marked mor layer and no earthworms. Broadbalk Wilderness is a slightly alkaline (mean pH 7.7) mull site, with abundant earthworm casts and no carryover of litter from year to year, so it is perhaps not surprising that the same biomass parameters cannot fit both Wildernesses.

Although not activated in the present work, RothPC-1 contains two biomass compartments that can be independently tuned. Undoubtedly, much better fits than those in Table 2 could be obtained by activating the two compartments and adjusting their rate constants separately for Geescroft and Broadbalk Wildernesses. However, we feel that fine-tuning of the biomass parameters in RothPC-1 should await more data on changes in biomass concentration down a wider range of soil profiles. But even this limited incursion into biomass fitting

ws that it is necessary to slow decomposition with depth if RothPC-1 is to generate realistic quantities of biomass C in the lower soil horizons. If this is not done, the biomass in the bottom layer is more than a magnitude too small.

### Fitting RothPC-1 to other data sets with pre- and post-bomb measurements

There are very few data sets in which pre- and post-bomb carbon and radiocarbon have been measured in sufficient detail down a soil profile to provide a critical testbed for RothPC-1. One comes from the work of Torn *et al.* (2002), in which profiles taken from a Russian Chernozem in 1900 (or thereabouts) and again from the same site in 1997 were radiocarbon dated. The site is in the Kamennaya Steppe National Park, with a mean annual temperature of  $6.4^\circ\text{C}$  and an annual precipitation of  $\sim 420 \text{ mm}$  at the beginning of the 20th century, rising to  $520 \text{ mm}$  at the end. The detailed inputs of monthly mean temperature, monthly precipitation and monthly potential evaporation that are needed to run RothPC-1 were calculated as the means for the three nearest of Müller's (1982) EURASIAN meteorological stations in the same climatic zone (his sites 67, 77 and 81). The mean annual temperature for these three sites is  $6.6^\circ\text{C}$ , mean annual precipitation  $453 \text{ mm}$  and mean annual potential evaporation  $651 \text{ mm}$ . The soil was developed on deep calcareous loess, thought to have been deposited some 4500–6000 years ago during the mid-Holocene. We tested the model on data from four layers (0–23, 23–46, 46–69 and 69–92 cm), calculated by interpolation from Torn *et al.*'s own sampling layers.

The optimal values for  $p$  (0.38) and  $s$  ( $-0.08$ ) were obtained by tuning the model to the organic C and radiocarbon data for the 23–46 and 46–69 cm layers in the 1900 and 1997 samplings. Using these values, the model gave a close fit to all the measurements save those from the 69–92 cm layer (Figure 6). Even though the model had been initiated in the usual way, by running it for 10 000 years, the 69–92 cm layer had still not reached equilibrium. In 1997 this layer contained  $24 \text{ t C ha}^{-1}$ , but the model predicted only  $10 \text{ t C ha}^{-1}$ . If the model is right, the site is too cold and too dry (for much of the year) for the 69–92 cm layer to have reached equilibrium, even if soil formation had gone on for 10 000 years, as was not the case with this particular profile. Apart from the bottom layer, this profile appears to contain vastly more IOM than our Rothamsted profiles; according to the model the 0–23 cm layer contained  $11.5 \text{ t IOM C ha}^{-1}$ , the 23–46 cm layer  $21.8 \text{ t}$ , the 46–69 cm layer  $21.3 \text{ t}$  and the 69–92 cm layer  $1.6 \text{ t}$ . The reason for the low IOM content of the bottom layer is that the model moves down so little C into this layer that very little IOM-C is needed to give the right radiocarbon content. It is likely that the loess contained much IOM on arrival, but we have no idea how this was distributed down the profile. This Kamennaya Chernozem is in no way unusual – a Chernozem from the Kursk region contained almost exactly the same quantity of carbon in the top metre of the profile (Mikhailova & Post, 2006).

**Table 2** Modelled and measured biomass C in the Geescroft and Broadbalk Wilderness profiles

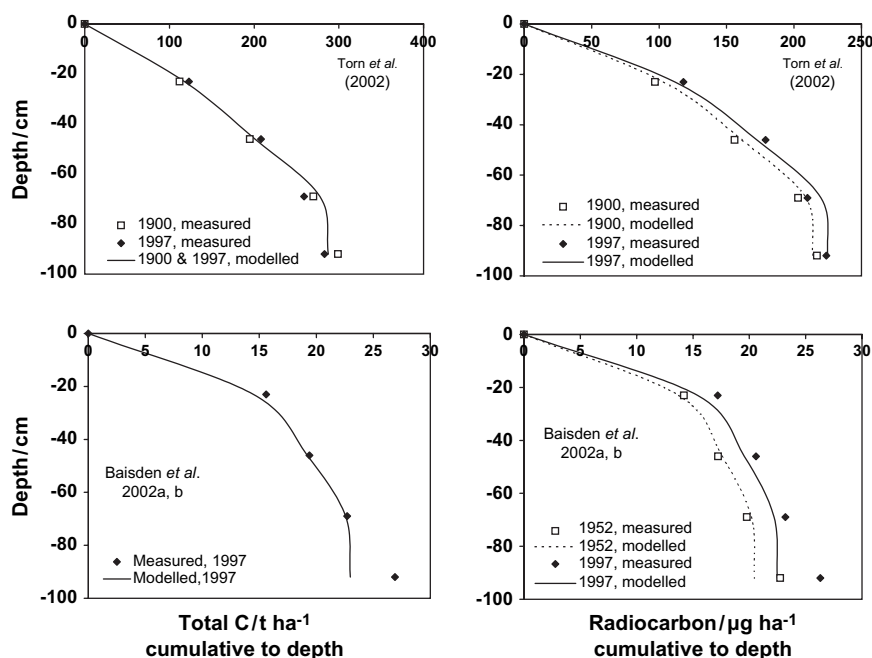
Site	Layer	Measured biomass	Modelled biomass <sup>a</sup>
	/cm	/t C ha <sup>-1</sup>	/t C ha <sup>-1</sup>
Geescroft Wilderness <sup>b</sup>	0–23	$0.61^d \pm 0.073$	0.57
	23–46	$0.15 \pm 0.039$	0.29
	46–69	$0.12 \pm 0.027$	0.22
	69–91	$0.04 \pm 0.025$	0.17
Broadbalk Wilderness <sup>c</sup>	0–23	$1.23 \pm 0.075$	0.76
	23–46	$0.31 \pm 0.008$	0.38
	46–69	$0.25 \pm 0.020$	0.35
	69–91	$0.10 \pm 0.031$	0.29

<sup>a</sup>Modelled with rate constant BIOF = rate constant BIOS =  $1.5 \text{ year}^{-1}$ .

<sup>b</sup>Modelled with  $p = 0.64$  and  $s = -0.06$ .

<sup>c</sup>Modelled with  $p = 0.72$  and  $s = -0.08$ .

<sup>d</sup>Includes litter layer.



**Figure 6** Top two diagrams. Measured and modelled organic C and radiocarbon down a chernozem profile from Russia, sampled in (about) 1900 and again in 1997 by Torn *et al.* (2002). The optimal value of the parameter moving C down the profile ( $p$ ) was 0.38; that slowing decomposition with depth ( $s$ ) was  $-0.08$ . Annual input to the top layer was set at  $2.45 \text{ t C ha}^{-1} \text{ year}^{-1}$ ; for other model settings see text. Bottom two diagrams. Measured and modelled organic C and radiocarbon down a Californian soil profile, sampled in 1952 and 1997 (Baisden *et al.*, 2002b). The optimal value of the parameter moving C down the profile ( $p$ ) was 0.9; that slowing decomposition with depth ( $s$ ) was  $-0.08$ . Annual input to the top layer was set at  $0.768 \text{ t C ha}^{-1} \text{ year}^{-1}$ ; for other model settings see text.

A second paired pre- and post-bomb site came from annual grassland in California (Baisden *et al.*, 2002a,b). We selected their Turlock Lake site from a chronosequence of four alluvial terrace sites in the eastern San Joaquin Valley, with a mean annual temperature of  $16^\circ\text{C}$  and a mean annual precipitation of 300 mm, falling almost entirely in winter. The monthly mean temperature, rainfall and evaporation data needed to run the model were taken from Müller's (1982) North America weather site No 85 (Fresno). The soil contained 5–8% clay in the top 62 cm, increasing sharply to some 40% in the 76–110 cm layer (Bt horizon). It was formed on glacial outwash from the Sierra Nevada, originally deposited some 0.6 M years ago. The soil had been sampled in 1952 and again in 1997. Baisden *et al.* (2002b) had normalized their layer thicknesses to a constant bulk density of  $1.65 \text{ g cm}^{-3}$ . We plotted their total C and radiocarbon data cumulatively to depth (as in Figure 5) and from this obtained total C and radiocarbon contents in the 0–23, 23–46, 46–69 and 69–92 cm layers.

Figure 6 shows the fit between modelled and measured total C and modelled and measured radiocarbon for the 1952 and 1997 samplings. All the fits were tolerable for the top three layers, but poor for the bottom layer. The model predicted that the lowest layer should contain only  $0.4 \text{ t C ha}^{-1}$ , but in fact this layer contained  $4.2 \text{ t C ha}^{-1}$  in 1997, with a  $\Delta^{14}\text{C}$  of  $-262$  (corresponding to a radiocarbon age of 2440 years). Increasing the IOM-C content of the 74–114 cm layer to 3.8 (i.e.  $4.2-0.4$ )  $\text{t C ha}^{-1}$  will of course allow the model to fit the organic C data. However, the radiocarbon content will then be far too small, because the IOM compartment in RothPC-1 is postulated to have a  $\Delta^{14}\text{C}$  of  $-998$ . Relatively young organic C must have been trapped on its downward path by this clay-rich layer, a view already proposed by Baisden *et al.* (2002b).

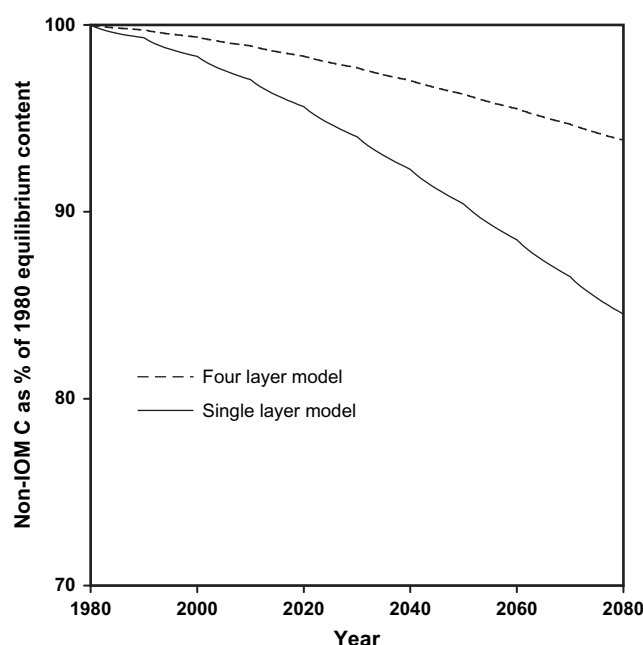
The Chernozem, developed under cold continental conditions, contained some 10 times more C and radiocarbon than the Californian annual grassland, developed under a Mediterranean climate. Yet, setting aside the lowest horizon, RothPC-1 can give an acceptable fit to both.

#### How warming affects modelled C stock in a profile treated as a single entity or in layers

Most attempts to model the effect of global warming on the stock of organic C held in terrestrial soil treat the top metre as a homogeneous unit (see, for example, Jones *et al.*, 2005). We decided to use RothPC-1 to see how a single-layer C model responds to warming, compared with a multi-layer model. We first ran the model for 10 000 years under Rothamsted climatic conditions to give the equilibrium C content of a soil in 1980. A subsequent increase in mean annual temperature of  $0.25^\circ\text{C}$  per decade over the next 100 years caused the stock of C held within the top 92 cm to fall by 16% in the single-layer version, compared with a fall of 6% when the same profile was divided into four equal layers (Figure 7; the legend specifies how the model was run). This calculation strongly suggests that treating the top metre of soil as a homogeneous unit will greatly overestimate the effects of global warming in accelerating the decomposition of soil C and hence on the enhanced release of  $\text{CO}_2$  from soil organic matter.

#### Conclusions

Our work shows that the topsoil model RothC-26.3 can be extended to the top 92 cm of soil by the addition of two extra



**Figure 7** Modelling the stock of C in an arable soil subjected to an increase of 2.5°C over the period 1980–2080. Two alternatives are compared: treating the 0–92 cm soil profile as a single layer, or divided into four equal 0–23 cm layers. RothPC-1 was first run for 10 000 years to give the equilibrium content in 1980, using the usual Rothamsted weather data; thereafter the mean annual temperature increased in steps, by 0.25°C per decade. The soil profile was assumed to contain no IOM. The annual input was 1.67 t C ha<sup>-1</sup> year<sup>-1</sup> before and after 1980, a value typical of an arable soil at Rothamsted. The clay contents of the layers in the four-layer test were 23, 30, 50 and 49% (as in Geescroft Wilderness); the clay content in the single layer test was 38% (the mean for the four layers). The DPM/RPM ratio of the incoming organic C was taken to be 1.44; the rate constants for the different compartments had the usual RothPC-1 values. The parameters  $p$  and  $s$  were taken to be 0.63 and  $-0.06$ . The soil was vegetated throughout the year, except for August, September, October and November, when it was bare.

parameters, one ( $p$ ) that moves organic C down the profile and the other ( $s$ ) that slows decomposition with depth. A single parameter moving C downwards is insufficient – the fit between measured and modelled data becomes poor when  $s$  is set at zero; that is, when decomposition rates are the same throughout the profile. Because topsoil C is younger than subsoil C, it is dominated by the relatively recent history of the soil, whereas subsoil C reflects events far into the past. It is therefore intrinsically easier to model topsoil C, and the problems in modelling ancient subsoil C are vividly illustrated by our attempts to calculate subsoil C turnover in a Russian Chernozem and a Californian annual grassland. Some of the assumptions in the model, in particular the assumption that temperatures throughout the profile are the same as the monthly mean air temperature, are gross oversimplifications. What we have done is to assume, in the absence of experimen-

tal data, that monthly air temperature and soil moisture content influence decomposition throughout the profile in the same way as in the well-tested topsoil model RothC.26-1. This simplification cannot be totally wrong, otherwise we would not have been able to fit data on measured total organic C and radiocarbon from profiles at Rothamsted, central Russia and California, places with very different climates. Despite the oversimplifications implicit in RothPC-1, we think that this model can generate estimates of C turnover in the top metre of soil – a reservoir that contains roughly twice as much C as the atmosphere – that are better than those given by models that treats the top metre as a single entity. We hope that the new model will be useful in understanding how this reservoir responds to climate change.

### Acknowledgements

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