Evaluating Time Scales of Carbon Turnover in Temperate Forest Soils with Radiocarbon Data

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Abstract

Time scales of soil organic carbon (SOC) turnover in forests were investigated with soil radiocarbon data. $^{12}C/^{14}C$ ratios were measured by AMS on soil sampled from a deciduous temperate forest in Switzerland during 1969–1995. The resulting $\Delta^{14}C$ values (125–174 ‰) were in line with previously published ^{14}C soil data. We applied ForClim–D, a model of nonliving organic matter decomposition including nine litter and two soil compartments to estimate SOC turnover times for this forest type. ^{14}C aging in woody vegetation was explicitly accounted for. Parameters were calibrated to match radiocarbon ratios (Harkness et al., 1986) observed for forest soils at Meathop Wood, UK. We estimated that roughly 50–94% (best estimate: 49%) of the C flux from decomposing foliar litter, and 11–74% (73%) of that leaving fine root litter are respired as CO_2 at Meathop Wood, the rest being used in the formation of soil humus. Turnover times for the 0–20 cm mineral soil layer ranged from 9–50 years (25 years) for a fast overturning soil compartment comprising 38–74% (68%) of bulk SOC, and 155–10'018 years (3'570 years) for a slowly overturning compartment. For the Swiss site, SOC turnover times were in the same range. Parameter uncertainties were correlated and induced by uncertainties in ^{14}C observations from small scale spatial inhomogeneities and sample preparation and by lack of reliable ^{14}C observations for the "pre-bomb" test period. Model-based estimates of soil organic C turnover derived from ^{14}C data have to be used cautiously since they depend on the underlying model structure: Bypassing litter in ForClim–D overestimated SOC turnover by a factor 2.5. Such an error might remain undetected in studies lacking samples from the late 1960's and early 1970's. Thus, litter C turnover should be included when estimating SOC turnover in temperate forests from ^{14}C data.

Keywords: Carbon cycle, Soil organic carbon dynamics, Forest soils, Bomb ^{14}C, Carbon imbalance, Humus turnover

Introduction

Nonliving organic carbon (NLOC)[1] sequestered in litter and mineral soils of terrestrial systems has been recognized as one of the major pools in the Earth's carbon cycle. The most recent global estimate of Batjes (1996) for soil organic C (SOC) lies at 1505 Pg C. SOC alone, thus, surpasses the C pools of vegetation (610 Pg C) and atmosphere (750 Pg C) by more than a factor of two (Schimel et al., 1995). About 50–60 Pg C yr^{-1} of the global soil respiration flux of 68–77 Pg C yr^{-1} (Schlesinger, 1977; Raich and Schlesinger, 1992; Raich and Potter, 1995) between terrestrial biota and atmosphere can be attributed to the decomposition of NLOC (McGuire et al., 1995; Post et al., 1997). Forest soils sequester 1146 Pg C (Dixon et al., 1994) and forests participate in roughly 70% of the global CO_2 exchange between biosphere and atmosphere (Waring and Schlesinger, 1985). Thus, NLOC decomposition in forests plays a crucial role in the global carbon cycle and changes in the balance between net primary production and decomposition of NLOC have the potential to alter atmospheric CO_2 concentrations.

The chemical and structural heterogeneity of NLOC represents the major problem for assessing its dynamics. Only few ecosystem models have considered this issue, and SOC in particular has often been characterized by a single turnover time (Perruchoud and Fischlin, 1995). However, C isotope studies have proven that this is not admissible for a description of soil organic matter (SOM) dynamics over decades to centuries (Harkness et al., 1986; Balesdent et al., 1988; Trumbore, 1993; Harrison et al., 1993b). Many recent ecosystem models therefore represent SOC by several compartments with different turnover times (Jenkinson and Rayner,

[1]Abbreviations: C: carbon; ^{14}C: radiocarbon; NLOC: Nonliving organic carbon, i.e. C sequestered by litter and soil humus; SOM: Soil organic matter or soil humus; SOC: Soil organic carbon, i.e. C sequestered by soil humus.

1977; Parton et al., 1987; Jenkinson, 1990; Verberne et al., 1990; Potter et al., 1993; Parton et al., 1994) to account for the multitude of SOM transformations during decomposition. Unfortunately, all these models have been originally designed and calibrated for grassland or agricultural, but none for forest soils. Based on radiocarbon studies (Harrison et al., 1993a; Scharpenseel, 1993; Trumbore, 1993) land use clearly affects NLOC dynamics and prohibits an application of one model of NLOC dynamics across distinct soil types without recalibration.

Radiocarbon observations provide information about the dynamics of carbon turnover. Natural ^{14}C is cosmogenically produced in the atmosphere. It partially enters the terrestrial carbon reservoirs where it slowly decays with a mean life time of 8267 years. Over the past 10'000 years before 1950 the natural production rate of ^{14}C has remained constant within $\pm 10\%$ (Trumbore and Druffel, 1995). However, in the late 1950's and early 1960's atmospheric radiocarbon concentration has almost doubled due to atomic bomb tests. After the atomic bomb test ban treaty in 1963, atmospheric ^{14}C declined due to its continuous uptake by the terrestrial and oceanic reservoirs. This transient signal of "bomb-produced" ^{14}C is well suited to investigate time scales of carbon turnover in the range of years to decades.

Radiocarbon studies have been applied to investigate the dynamics of SOC in uncultivated (Harrison et al., 1993b; Harrison et al., 1995), cultivated (Hsieh, 1992; Harrison et al., 1993a; Hsieh, 1993) and forest soils (Harkness et al., 1986; Harrison et al., 1995), and to assess the impact of climate (Tate et al., 1995; Townsend et al., 1995; Trumbore et al., 1996) and CO_2 fertilization on SOC dynamics (Harrison et al., 1993b; Harrison et al., 1995). Typical decay times of SOC found in previous studies are in the range of a few decades for the major soil fraction and up to several millenia for more persistent structures. A depletion of fast overturning humus has been observed for cultivated relative to native soils (Harrison et al., 1993a; Harrison, 1996) whereas for a recovering forest soil increased stocks and turnover of SOC have been reported (Harrison et al., 1995). Increases in $\Delta^{14}C$ levels for native topsoils have been related to increasing temperatures (Tate et al., 1995; Townsend et al., 1995; Trumbore et al., 1996). Soil ^{14}C data has, however, not been utilized to link decomposition processes of litter and humified organic matter in forests to our believe.

We used published time series of soil ^{14}C data (Harkness et al., 1986) in conjunction with ForClim–D (Perruchoud, 1996), a simple multicompartment model of NLOC decomposition to quantify SOC dynamics for undisturbed northern temperate forest soils. NLOC decomposition was formulated as a sequence of first order kinetic reactions with explicit representation of litter and humified organic matter. We represented SOM by two compartments based on the findings of Harrison et al. (1993b), Balesdent et al. (1988), and a comparison with higher order SOC models (Perruchoud, 1996). Independent and new ^{14}C enrichment data from a forest stand near Basel (Switzerland) were used to test for the plausibility of the model, its C turnover time estimates and the distribution of SOC fractions in undisturbed temperate forest soils under current climatic conditions. Also included in this study is an assessment of uncertainty for these model parameter estimates. Finally, ForClim–D was applied to estimate the change in ^{14}C inventory between 1969–1989 and the C sequestration potential of NLOC in cold deciduous forest soils assuming CO_2 fertilization of plants under transient scenarios of atmospheric CO_2 increase.

Materials and Methods

Model Description

We used the decomposition model ForClim–D (Fig. 1) which simulates NLOC dynamics by nine litter and two soil compartments (Perruchoud, 1996). ForClim–D has been developed as part of the forest model FORCLIM (Bugmann, 1996; Fischlin et al., 1995; Bugmann and Cramer, 1998) to simulate species compositions and C–budgets in forests of the temperate zone under current and altered climatic conditions. In this study FORCLIM–D was run independently and driven by litter inputs from the vegetation which were prescribed here by site specific measurements or literature estimates.

FORCLIM–D includes a climate and litter-quality specific model of foliar litter decomposition and discriminates foliar (six litter types), fine root, twig and woody litter. The carbon budget of litter compartment i is given by:

$$\dot{c}_i(t) = -k_i \cdot c_i(t) + I_i, \tag{1}$$

where c_i is the carbon mass, k_i the turnover rate, and I_i the prescribed input flux for compartment i (cf. Tab. 1). Decomposition rates for foliar litter are described as a function of actual evapotranspiration (AET) and lignin content of fresh litter following Meentemeyer (1978):

$$k = 0.083 + 0.0011 \cdot \mathrm{AET} - 0.832 \cdot \mathrm{lignin}. \tag{2}$$

A fraction ξ_i of the C flux from decomposing litter is transferred from litter compartment i to the soil and the rest $(1 - \xi_i)$ is respired as CO_2. The total C flux entering the soil is given by the sum of the humification fractions ξ_i multiplied by the litter decomposition fluxes (i.e., turnover rate k_i times litter mass c_i):

$$u(t) = \sum_i \xi_i \cdot k_i \cdot c_i(t). \tag{3}$$

SOC is described by a fast (c_{fH}) and a slow (c_{sH}) overturning soil humus compartment. They represent rapid (c_{fH}) and recalcitrant (c_{sH}) topsoil carbon and are coupled in series (Fig. 1). Note, that our simulations of NLOC dynamics refer to the 0–20 cm topsoil layers. Carbon inventories in the fast and slow compartment are described by:

$$\dot{c}_{fH}(t) \quad = \quad -a \cdot c_{fH}(t) + u(t), \tag{4}$$

$$\dot{c}_{sH}(t) = -a \cdot s \cdot c_{sH}(t) + \nu \cdot a \cdot c_{fH}(t), \tag{5}$$

where u is the input from the litter compartments, a and $a \cdot s$ denote turnover rates for fast and slow humus and s is the ratio of their turnover times. ν controls the fraction of fast humus that is passed to the slow humus compartment (Fig. 1). Thus, s/ν represents the ratio of C masses in fast and slow compartment at steady state. The ratios of turnover times (s) and carbon inventories (s/ν) were assumed to be constant across soil depth. For calibration, equations of ^{14}C dynamics for litter and SOM were implemented as described in Appendix A.

Site Description and Data

Meathop Wood (Great Britain): This temperate *Quercus* – *Fraxinus* – *Betula* type woodland was used for calibration of ForClim–D. The site has been intensively studied during the International Biological Programme (De Angelis et al., 1981) and is well documented (Bunce, 1968; Harkness et al., 1986; Harrison et al., 1990).

Litter input observations were taken from the literature (Tab. 1): Data of foliar and woody litterfall were equated with the average of the three year monitoring study of Sykes and Bunce (1970). Estimates of fine root litter input reported in Harrison and Harkness (1993) were not depth-specific. We assumed a non-linear depth distribution with inputs of 0.4 t C ha^{-1}yr^{-1} for the 0–5 cm layer and 1.0 t C ha^{-1}yr^{-1} for the 0–15 cm layer. Fine root litter litter input was set to 1.5 t C ha^{-1}yr^{-1} for the 0–20 cm layer. Biomass and litterfall data of twigs and branches from De Angelis et al. (1981) and Harrison and Harkness (1993) were used to model Δ^{14}C in woody tissue (see Appendix A).

Depth-specific SOC inventories at Meathop Wood required for initialization of soil humus compartments and calculation of ^{14}C in the 0–10 and 0–15 cm soil layer (Tab. 2) were available from Harrison et al. (1990). ^{14}C data for calibration were taken from Harkness et al. (1986) and

Harrison and Harkness (1993). Two soil samples (collection dates 1977, 1982) were excluded for calibration since they do not originate from the original field campaign of Harkness et al. (1986) (A. F. Harrison, Institute of Terrestrial Ecology, Grange-over-Sands, *pers. comm.*) and exceed ^{14}C levels measured in topsoil profiles around the globe.

Haumättli (Switzerland): A forest reserve ("Naturreservat") located 20 km east of Basel on flat terrain in close proximity (about 200 m) to the river Rhine (Kuhn, 1990; Burri, 1996) was used for model testing. This stand is characterized by vegetation dominated by *Fraxinus excelsior* L. with *Acer pseudoplatanus* L. and *Fagus silvatica* L. as secondary species. Soils are classified as Fluvisols.

We determined foliar and total litterfall according to Meentemeyer et al. (1982): Twig litterfall was calculated as the difference between total and foliar litterfall. Fine root litter input was determined via fine root production from total aboveground litterfall C (Nadelhoffer and Raich, 1992). For the required estimate of AET (viz. 595 mm yr^{-1}) simulation results of ForClim at Basel (Bugmann, 1994) were invoked (Tab. 1).

At Haumättli, SOC content and ^{14}C for the 0–5 cm mineral soil layer were determined from archived and recently collected soils. Ungrounded soil material was available for 1969, 1988 (Kuhn, 1990), 1993 (Burri, 1996), and additional samples were collected in 1995 (this study). All samples were obtained by bulking seven soil cores from within two hectares (Burri, 1996). The SOC inventory was calculated by multiplying C concentrations and bulk soil density of the soil fraction < 2 mm (Tab. 3). Accelerated mass spectroscopy (AMS) was selected for measuring isotope ratios of bulk soil. AMS-targets were prepared according to the protocol of Rutberg et al. (1996) and measured at the PSI/ETH facility in Zürich, Switzerland (Tab. 3).

Experimental Design

The model was calibrated for the Meathop Wood site assuming steady state conditions for ^{12}C fluxes in model calibration and model testing. Turnover rates of litter (k_i) and soil humus (a, $a \cdot s$), humification ratios (ξ_i), the humus transfer coefficient (ν), and the prescribed ^{12}C litter inputs from vegetation (I_i) were kept constant in time. This is justified as net carbon storage contributes little to the ^{14}C flux into the terrestrial biota (Joos and Bruno, 1998).

^{14}C simulations were initialized for an atmospheric ^{14}C ratio of $\Delta^{14}\mathrm{C}_A \equiv 0$ ‰. The model was run from 1930 to 1995 forced by the observed atmospheric ^{14}C history for the Northern Hemisphere (Fig. 2) in Enting et al. (1994). Atmospheric ^{14}C ratios ($\Delta^{14}\mathrm{C}_A(t)$) were prescribed in Δ^{14}C notation, which implicitly accounts for fractionation effects. Δ^{14}C in living wood was explicitly modeled by a one compartment model to account for the slower growth and turnover rates of this tissue. On the other hand an instantaneous transfer of the atmospheric Δ^{14}C signal was assumed for foliar and fine root litter (see Appendix A).

For the decomposition rate of twig litter, a value of $k_{tl} = 0.22$ yr^{-1} was adopted in accordance with Pastor and Post (1985) and Bugmann (1994). Decomposition rates for fine root and the six foliar litter classes were tuned until simulated and observed levels of Δ^{14}C as measured on "undecomposed plant rootlets" in the 0–5 cm soil layer (Harrison et al., 1990) and foliar litter in the leaf layer (Harkness et al., 1986) visually matched (Tab. 1). ξ_{wood} was determined using the estimate of Mattson et al. (1987) for CO_2 losses of decomposing fine and coarse woody debris ($\approx 55\%$).

Next, the parameters ξ_{foliar}, $\xi_{f.roots}$, s, and s/ν were determined by the global least square optimization algorithm Sufi (Abbaspour et al., 1997). We used the root mean square error (RMSE) defined by:

$$RMSE := \sqrt{\frac{1}{N}\sum_{i=1}^{N}\left(x_{meas.}(t_i) - x_{sim.}(t_i)\right)^2} \quad (6)$$

as target function where, $x_{meas.}$ and $x_{sim.}$ refer to measured and simulated ^{14}C values in the 0–5 cm soil layer (Tab. 2).

Finally, the turnover rate for the fast humus compartment, a, was calculated from total soil carbon inventory data (c_{SOC}) for the 0–5 cm layer (Harrison and Harkness, 1993), the soil C input from litter decomposition (u), and the ratio of fast and slow humus C mass at steady state (s/ν):

$$a = \frac{1}{c_{SOC}} \cdot (1 + \frac{\nu}{s}) \cdot u. \tag{7}$$

The parameters ξ_{foliar}, $\xi_{f.roots}$, ξ_{wood}, s, and s/ν identified for the 0–5 cm layer were also applied to a soil depth of 0–15 cm and 0–20 cm. This means that the ratio of turnover times (s) and C masses (s/ν) was assumed to be the same for all depths. Absolute values of the turnover rates for the 0–15 cm and 0–20 cm layer, however, varied according to Eq. (7) and depth-specific soil inventories (c_{SOC}, see Tab. 2) , and fine root litter inputs (see "Site description and data"). The assumption of homogeneity for the 0–20 cm soil layer was justified by the high biological activity (earthworms) at Meathop Wood (Harrison et al., 1990).

The site Haumättli was used to test the model parameters. Decomposition rates for foliar litter were recalculated according to Eq. (2) with a lignin concentration of 24% (of dry weight organic matter) for mixed foliar litter. AET was assumed to be 595 mm yr^{-1} (see above). Fine root and woody litter decomposition rates and the SOC parameters ξ_i, ν, and s were adopted from Meathop Wood. The turnover rate a for Haumättli was derived via Eq. (7). The required SOC inventory (13.2 t C ha^{-1} for the 0–5 cm layer) was averaged from measured carbon concentrations of samples collected in 1969, 1988, and 1993 (A. Burri, University of Zürich, *pers. comm.*).

Results

Model Calibration

At Meathop Wood, we found best estimates for the turnover times of 11 years for 68% of the SOC and around 1560 years for the rest at 0–5 cm soil depth (Tab. 4). A reasonable agreement between simulated and calculated ^{14}C data was obtained for the 0–15 cm soil layer (Fig. 2) with turnover times of about 23 and 3340 years. Our extrapolation for the top 0–20 cm soil layer which is thought to be affected by environmental changes most directly (Parton et al., 1993; Schimel et al., 1994; McGuire et al., 1995; Melillo et al., 1995), resulted in turnover time estimates of 25 and 3570 years for fast and slow humus respectively.

Next, we investigated uncertainties of estimated parameters (Tab. 5) as resulting from uncertainties in soil radiocarbon observations (0–5 cm layer). Best estimates for the parameters ξ_{foliar}, $\xi_{f.roots}$, s, and s/ν were obtained by a global stepwise search, i.e. model simulations were carried out on a stepwise refined 4-dimensional grid in the parameter space and the root mean square error function ($RMSE$) evaluated for each grid point. The minimum for RMSE was 0.0262 corresponding to a mean deviation of $\pm 26\,^o\!/_{oo}$ in modeled versus measured $\Delta\,^{14}$C. The span of $52^o\!/_{oo}$ is comparable to the range of $74^o\!/_{oo}$ in replicate Δ^{14}C measurements reported by Harrison et al. (1990) for the 0–5 cm soil layer. This variation in replicate samples is much larger than the measurement precision and could be due to small scale inhomogeneities in the soil layer and sample preparation.

Parameter ranges associated with $RMSE \leq 0.030$ were determined to estimate uncertainties in model parameters (Tab. 5). We found that 6–50% (ξ_{foliar}) of foliar litter C and 27–90% ($\xi_{f.roots}$) of fine root litter was humified during decomposition and the rest respired to the atmosphere. The uncertainty range for the humification ratio of fine root litter ($\xi_{f.roots}$) was larger than for foliar litter (ξ_{foliar}) as the latter dominates soil C inputs at the soil surface (Tab. 1). Less than 9% of the fast humus C which amounts to 38–74% of bulk SOC was passed to the slow humus compartment (parameter ν). Turnover times for the fast and slow compartment were 9–50 and 155–10'018 years. A more reliable determination of the relative size of the fast and slow humus compartment and the turnover time of the latter was hampered by the absence of samples for the period prior to the atomic tests.

Parameters for humification ratios, soil reservoir sizes, and turnover times are correlated in ForClim–D (cf. Eq. (3) and Eq. (7)). This can also be seen from the SOC response to a step change in litter input (see Appendix B, Eq. (11)). A numerical analysis of Eq. (11) based on the parameter ranges obtained for $RMSE = 0.03$ (Tab. 5) revealed that the step response of SOC varied by no more than 16% (after $t = 15$ years) relative to the equilibrium response within a few centuries. Thus, the general response of ForClim–D to a decadal scale transient carbon perturbation is better determined by the ^{14}C data than reflected by the uncertainty of individual parameters.

Is it necessary to use a coupled litter-soil model to quantify SOC dynamics via ^{14}C in soils? To answer this question we simulated ^{14}C enrichment in mineral soil bypassing litter decomposition and associated CO_2 mineralization processes, assuming instead an instantaneous, complete transfer of ^{14}C from fresh plant litter to the soil. This was achieved by setting $k_i = 500\,\mathrm{yr}^{-1}$ (instantaneous litter decomposition) and $\xi_i = 1.0$ (no respiration losses from litter) for all litter types. Thereby, soil radiocarbon input fluxes were driven by the atmospheric isotopic ratio except for the input from woody vegetation. Fig. 3 shows the anticipated rise in ^{14}C enrichment and the convergence in the mid 1980's for the "bypassing litter" run compared to our default run for a soil depth of 5 cm. Due to Eq. (3) and Eq. (7) we obtained lower values for the turnover times in this case, viz. 5 (fast humus) and 712 years (slow humus) for a 5 cm soil depth as compared to 11 and 1560 years when litter dynamic was included. Turnover times recalculated for the 0–20 cm soil layer via Eq. (7) were 10 and 1421 years in the "bypassing litter" run. Although these values were still within the uncertainty range estimated above, they were systematically lower than our best estimates. This suggests that the complete litter-soil dynamics should be tracked when comparing modeled versus observed radiocarbon levels. Otherwise, turnover times will be systematically underestimated.

Model Testing and Applications

Simulations with ForClim–D at Haumättli using estimated litter inputs (Tab. 1) and measured SOC inventory for this Swiss site combined with parameters (ξ_{foliar}, $\xi_{f.roots}$, ξ_{wood}, s/ν and s) derived for Meathop Wood underestimated measured ^{14}C enrichment in the 0–5 cm soil layer somewhat. The turnover times for fast and slow humus were 8 and 1178 years. The mismatch between modeled and observed values was within the range of natural variability of ^{14}C observations in soils (Harrison et al., 1990) and was remedied by slight adjustments of the litter input rate and SOC inventory (Fig. 4). Foliar litterfall rates were increased by 10% to 1.32 t C ha^{-1} yr^{-1} which seems justified given the intra-site variability of foliar litterfall of 12% (stand. dev.) for deciduous trees across the Swiss Plateau (Lüscher, 1991). Additionally the SOC inventory was reduced by 10% to 11.9 t C ha^{-1} which seems a reasonable estimate for the combined errors of bulk density and %C concentration (Trumbore et al., 1995). Humification ratios ξ_i, humus transfer ν and the scaling factor s remained unchanged. For these settings, our best estimate of SOM dynamics (0–5 cm soil layer) resulted in turnover times of 7 and 1006 years for fast and slow humus respectively. In summary, turnover of C in soil at the Swiss site and at Meathop Wood were indistinguishable within the uncertainty of the observations. For model applications, we used parameter values derived for the Meathop Wood site.

It has been suggested recently to use observations of the $^{14}C/^{12}C$ ratio of heterotrophically respired CO_2 to determine turnover time scales of terrestrial biospheric C (Dörr and Münnich, 1986; Joos et al., 1997). We simulated the $^{14}C/^{12}C$ ratio in heterotrophically respired CO_2 (Fig. 5, top) to calculate the ^{14}C air-biota disequilibrium as defined by the difference in atmospheric (assimilated) and heterotrophically respired $\Delta^{14}C$ for the 0–20 cm soil layer. The $\Delta^{14}C$ disequilibrium increased initially with a peak at 604.3 ‰ around 1964 (Fig. 5, middle). Thereafter the air-biota disequilibrium decreased monotonously, crossed over in 1982 and reached a value of –48.5 ‰ by 1990.

We also used ForClim–D to assess the bomb radiocarbon inventory in SOC of cold deciduous forests. The modeled radiocarbon inventory of soils, $^{14}c_{NLOC}$, in units of [^{14}C atoms·m^{-2}] was obtained by:

$$^{14}c_{NLOC}(t) = 10^2 \cdot \frac{N_{avog.}}{12} \cdot {}^{14}\mathrm{R}_{stand.} \cdot \sum_j \left(1 + \frac{\Delta^{14}\mathrm{C}_j(t)}{1000}\right) \cdot {}^{12}c_j, \tag{8}$$

where $\Delta^{14}C_j(t)$ is the simulated $^{14}C/^{12}C$ ratio of litter and topsoil in ‰ ($j = foliar$, $f.root$, $woody\,litter$ and SOC), $^{14}\mathrm{R}_{\mathrm{stand.}} = 1.176 \cdot 10^{-12}$ is the standard isotopic ratio of $^{14}C/^{12}C$, $N_{avog.} = 6.022 \cdot 10^{23}$ is the number of atoms per mol and $^{12}c_j$ the C content in t C/ha of litter and soil compartments. NLOC contained $322 \cdot 10^{12}$ atoms·m^{-2} in 1950 (85% in SOC) and increased to attain its maximum in 1983 ($385 \cdot 10^{12}$ atoms·m^{-2}) (Fig. 5, bottom). For the period 1965–1989 NLOC's ^{14}C inventory increased by $33 \cdot 10^{12}\mathrm{atoms} \cdot \mathrm{m}^{-2}$. We obtained a bomb inventory rise of $4.4 \cdot 10^{26}$ atoms for soils between 1965 and 1989 by multiplying the above estimate with the area covered by cold deciduous forests ($13.2 \cdot 10^{12}$ m^2). This is about 5–10% of the estimated increase in the global terrestrial biosphere system.

ForClim–D was used to investigate the soil response for instantaneous changes of SOC inputs. In Fig. 6 SOC dynamics in the 0–20 cm soil layer is shown for a step change in SOC input (δ in Eq. (11), Appendix B) by 10%, 20% and 35%. Changes over the first century were dominated by accumulation of fast humus, while for slow humus the build-up took place on the time scale of a millennium (not shown). Even for the instantaneous litter input changes assumed here, more than 10 years were required to change the SOC inventory by 10% or more for all cases.

Finally, the transient response of NLOC stocks (litter plus 0–20 cm soil humus) to changes in primary productivity as a consequence of a postulated CO_2 fertilization were assessed (Fig. 7 and Tab. 6). We used a logarithmic relationship between the increase in atmospheric CO_2 and litter input, i.e. $\left(1 + \beta \cdot ln\left(\frac{CO_2(t)}{CO_2(t_0)}\right)\right)$, with $\beta = 0.35$. This value was derived from mean total tree dry mass increases for a doubling of CO_2 in controlled-exposure studies (Wullschleger et al., 1995). For transient simulations between 1765–2100, atmospheric CO_2 was prescribed according to observations until 1990 and then following IS92a and S550 (Schimel et al., 1995). Atmospheric CO_2 roughly follows the present trend under the IS92a scenario. According to the S550 pathway, future growth in atmospheric CO_2 is decreasing and atmospheric levels eventually stabilize at 550 ppmv, i.e. about twice the preindustrial level.

Due to the accelerated increase of atmospheric CO_2 in the recent past, the increase of modeled NLOC stocks between 1770–1960 and 1960–2000 equaled each other (i.e. 2.6% relative to 1770 AD). For the next century simulated NLOC accumulation diverged under the two CO_2 scenarios. We obtained an increase in NLOC of 14.2% (relative to 1770) by 2100 for S550. Annual NLOC accumulation rates did not change between the past decade and the end of the next century under this CO_2 scenario. For the IS92a scenario, NLOC accumulation rates increased further during the course of the next century leading to an enhancement of NLOC stocks of 19.9% by 2100 (Fig. 7).

These figures were used to estimate the potential of C sequestration of cold deciduous forest soils (Tab. 6). Based on an analysis of regional SOC inventories (Tab. 2.4 in Perruchoud

(1996)), we assumed NLOC densities of 50–70 t C ha^{-1} for litter and soil down to 20 cm depth and an area extent of $13.2 \cdot 10^{12} \cdot m^{-2}$ for this vegetation type (Dai and Fung, 1993; King et al., 1997). For the 1980's we obtained an NLOC accumulation rate of 0.04–0.06 Gt C yr^{-1} in cold deciduous forest soils. These soils cover ≈11% of the global terrestrial surface (deserts and ice neglected, King et al. (1997)). If the postulated fertilization is indeed real and if the carbon dynamics of forest soils would be the same for non-forest soils, then global NLOC accumulation would be of the order of 0.4–0.6 Gt C yr^{-1} for the 1980's. This corresponds to about 30–45% of the global C "imbalance" (1.3 Gt C yr^{-1}) for the last decade (Schimel et al., 1996a).

Discussion and Conclusions

There is general agreement between the findings from previous studies (Harrison et al., 1993a; Harrison et al., 1993b; Harrison et al., 1995; Townsend et al., 1995; Trumbore et al., 1996) and our quantification of SOC dynamics for Meathop Wood (Tab. 4): The agreement of turnover times and SOC distribution is particularly good for estimates obtained by a global compilation of ^{14}C enrichment data from native forest and grassland soils (Harrison et al., 1993a; Harrison et al., 1993b). Variations in ^{14}C at the stand/plot level have been related to land use (Harrison et al., 1993a; Harrison, 1996; Harrison et al., 1995) and climate (Tate et al., 1995; Townsend et al., 1995; Trumbore et al., 1996). Trumbore et al. (1996) have investigated SOC dynamics in temperate forest soils under coniferous vegetation. For soils with comparable mean annual air temperature (MAT) ("Shaver", "Corbett": MAT ≈ 8° Celsius; Meathop Wood: MAT = 7.8° Celsius) they obtained estimates of SOC turnover times of 35–55 yr for a maximum soil depth of 20–23 cm. This is slightly higher than our turnover estimate of 25 years for fast humus accounting for 68% of topsoil C, but comparable within the general uncertainty of the ^{14}C method (Tab. 5).

As pointed out by Aber et al. (1990) "very few studies have dealt with the role of litter decay in the formation of stable forest soil organic matter". However, this process eventually determines how plant response affects C storage in the soil. We have estimated from soil ^{14}C observations how much of the foliar and fine root litter C is respired to the atmosphere and how much is converted to SOC (Tab. 5). Our results suggest that more than 50% of the decomposing foliar litter leaves the ground as CO_2. The respective fraction for fine root litter is about three quarters, but is highly uncertain (Tab. 5). A better estimate for the fine root litter humification ratio could probably have been obtained if more ^{14}C data from soil layers at greater depth (5–15 cm) had been available, as the influence of foliar litter is smaller relative to fine root litter. Within the soil compartment, we find that more than 90% of the C that is overturning on a decadal time scale is respired to the atmosphere; the rest is stable or very slowly decaying material.

The model was applied to estimate the potential of forest soils to sequester anthropogenic C under increasing atmospheric CO_2. Two assumptions were made: (1) atmospheric and climatic changes have not changed the respiration rates for NLOC, and (2) primary productivity and litter production have increased in pace with atmospheric CO_2. Under these settings cold deciduous forests potentially accumulated 0.04–0.06 Gt C yr^{-1} in litter and soils during the 1980's. Although turnover and thus response times of soils in the boreal and the tropical zone differ, a tentative spatial extrapolation yielded a global potential NLOC accumulation of 0.4–0.6 Gt C yr^{-1} for the 1980's. This number would explain about 30–45% of the "imbalance" in the contemporary C budget and is comparable to SOC accumulation estimates of Post et al. (1997) (0.7 Gt C yr^{-1}, globally) and Harrison et al. (1993b), Harrison (1996) (0.5–0.7 Gt C yr^{-1}, non-wetland soils).

The assessment of the soils potential for storing anthropogenic carbon is simplistic for following reasons: First, a long-term CO_2 fertilization effect for mature forests ecosystems is highly uncertain (Kirschbaum and Fischlin, 1996; Melillo et al., 1996). Second, the response of species assemblages (Bazzaz et al., 1993) and C stocks (Schimel et al., 1994) to increases of atmospheric CO_2 under multiple resource limitations (light, nutrient, water) was not considered here. Third, it is not clear how plants growing under elevated CO_2 affect soil processes. NLOC accumulation is one possibility, but increased fine root production and root rhizodeposition could also stimulate microbial activity and deteriorate SOC levels (Norby et al., 1992; Körner and Arnone III, 1992; Rouhier et al., 1994).

Note, that turnover times for NLOC and its high spatial variability make it unlikely that increases in NLOC stocks would be detected by field monitoring studies within less than a few decades. Moreover Huntington et al. (1988) have concluded that a massive sampling is required particularly in heterogeneous landscapes to obtain statistically reasonable estimates of C changes in topsoils. While the step response of SOC for increased litter input already took more than a decade (Fig. 6), gradual changes e.g. by CO_2 or nitrogen fertilization would even take longer to become visible. Models, therefore remain a promising way to assess the NLOC dynamics at regional or higher scales.

An important but difficult aspect of model building is assessing the uncertainty of model structure and model parameters. The age distribution of SOC is viewed to cover a continuum of time scales ranging from years to millenia. Our soil representation was by two compartments only with turnover times of about 25 years ($1/a$) and 3600 years ($1/(a \cdot s)$) governing the age distribution of SOC by two exponential functions with characteristic rate parameters a and $a \cdot s$ (cf. Eq. (11)). A representation of SOC by two boxes is justified by the following arguments: On one hand, the uncertainty of available ^{14}C data does not allow us to reasonably determine the parameter values for a model with an improved resolution. On the other hand, a two box representation can roughly explain the slow long-term accretion rate of humus observed during soil formation (Schlesinger, 1990) and the large potential of soils to respond to changes in C input fluxes as induced by climatic and atmospheric concentration changes on time scales of a few decades (Townsend et al., 1995).

Related to the uncertainties in model structure is the neglect of an explicit description of ^{14}C turnover in litter and vegetation in analyses of soil radiocarbon data. Rafter and Stout (1970) and Harkness et al. (1986) have observed higher than atmospheric levels of Δ^{14}C in the litter layer of forest soils for the "post-bomb" period. This proves that soil ^{14}C dynamics is not directly driven by the atmospheric ^{14}C signal (Fig. 3). Simulations at Meathop Wood revealed that calculated turnover times are in fact sensitive to the representation of litter decomposition. A direct, undelayed transfer of the atmospheric ^{14}C signal into the soil compartment lowered turnover times by a factor 2.5 relative to our standard calibration. Bypassing the breakdown and mineralization of forest litter and transferring the ^{14}C signal from the vegetation directly to the soil had the effect of a premature ^{14}C enrichment in mineral soil (see Fig. 3, 1960–1980 AD). This would not be detected in ^{14}C simulations of SOC without litter representation if soil samples covering the period 1965–1975 AD were missing. We conclude that the dynamics of the complete vegetation-litter-soil system should be explicitly described when determining turnover times of SOC from ^{14}C data in order to avoid a systematic bias.

We estimated the range of possible parameter values within the two box representation of SOC as coupled to nine litter pools. Individual parameter values of the soil submodel varied over a wide range while still matching soil ^{14}C observations within their uncertainty. Parameter uncertainties turned out to be correlated: Litter humification ratios (ξ_{foliar}, $\xi_{f.root}$) were high when the fast humus fraction (s/ν) was low and vice versa. Highly asymmetric distributions of the range around the best estimates were found for other parameters. Since the individual parameters are not independent in ForClim–D, the overall response of SOC to a decadal scale forcing is much better constrained by ^{14}C data than uncertainties of individual parameters might suggest.

Are there other ways to utilize radiocarbon observations for a better understanding of terrestrial C turnover? Dörr and Münnich (1986) have demonstrated the usefulness of $\Delta^{14}C$ observations for quantifying the age of heterotrophically respired C by measurements of soil CO_2 samples taken during the early eighties. Dörr and Münnich (1986) have found that $\Delta^{14}C$ in soil CO_2 are close to atmospheric levels in summer when respiration fluxes are largest. In winter, root respiration and decomposition of young material is reduced relative to the decomposition of more refractory SOM and observed $\Delta^{14}C$ in soil CO_2 are about 50–100 ‰ lower at a grassland and a beech spruce forest site. The annual averaged observed $\Delta^{14}C$ in respired CO_2 is close to atmospheric levels. Our results are in agreement with these observations. Using contemporary radiocarbon information could also lead to an improved quantification of SOC dynamics via the ^{14}C air-biota disequilibrium (Fig. 5, middle). Joos et al. (1997) have found in a global modeling study a spatial variability between assimilated and respired C of several tens of ‰ $\Delta^{14}C$ which can potentially be detected. The advantage of measuring ^{14}C concentrations in heterotrophically respired CO_2 is the reduction of spatial variability by air mixing in comparison to soil ^{14}C.

After the bomb test ban treaties were set in place in the early sixties, the total radiocarbon content of the atmosphere-biosphere-ocean system should have remained roughly constant. This budget constancy additionally constrains C cycle models that include the three relatively rapidly exchanging reservoirs atmosphere, biosphere, and ocean (Hesshaimer et al., 1994; Joos, 1994). Until now the biospheric change in radiocarbon inventory has been assessed by simple global box models of the terrestrial biota. In this study, ForClim–D was used to interpolate soil ^{14}C data to estimate in a consistent way the increase in local ^{14}C inventories. Required in the future for testing our understanding of the global ^{14}C budget are estimates of global terrestrial ^{14}C storage derived by applying models that match local ^{14}C observations.

Soils continue to be one of the great unknowns in the context of global C cycling due to their high spatial variability. Radiocarbon observations offer a good possibility to improve our understanding of the global C cycling in different ways: Radiocarbon is e.g. used in standard tests to assess the time scales of surface-to-deep mixing in ocean C cycle models by comparison of modeled and observed oceanic ^{14}C distributions. Radiocarbon observations also offer an opportunity to improve our understanding of the C dynamics in terrestrial biota as shown in this study. However, to track terrestrial carbon dynamics and to constrain terrestrial ecosystem models the utilization of ^{14}C data remains to be improved.

Appendix A: ^{14}C equations for NLOC dynamics

The following equation system was implemented to identify rate parameters controlling the dynamics of NLOC by ^{14}C enrichment data[2]:

foliar litter

$$^{14}\dot{c}_{flt_i}(t) = {}^{12}\mathcal{F}_{V,flt_i} \cdot {}^{14}\mathrm{R}_{stand.} \cdot \left(1 + \frac{\Delta^{14}\mathrm{C}_A(t)}{1000}\right) - (k_{flt_i} + \lambda) \cdot {}^{14}c_{flt_i}(t), \quad \text{i=}\{1,\ldots 6\}$$

fine root litter

$$^{14}\dot{c}_{frl}(t) = {}^{12}\mathcal{F}_{V,frl} \cdot {}^{14}\mathrm{R}_{stand.} \cdot \left(1 + \frac{\Delta^{14}\mathrm{C}_A(t)}{1000}\right) - (k_{frl} + \lambda) \cdot {}^{14}c_{frl}(t),$$

woody litter

$$^{14}\dot{c}_{wl}(t) = \frac{^{12}\mathcal{F}_{V,wl}}{^{12}c_{wt}} \cdot {}^{14}c_{wt}(t) - (k_{wl} + \lambda) \cdot {}^{14}c_{wl}(t), \text{with}$$

woody vegetation

$$^{14}\dot{c}_{wt}(t) = {}^{12}\mathcal{F}_{V,wl} \cdot {}^{14}\mathrm{R}_{stand.} \cdot \left(1 + \frac{\Delta^{14}\mathrm{C}_A(t)}{1000}\right) - \left(\frac{^{12}\mathcal{F}_{V,wl}}{^{12}c_{wt}} + \lambda\right) \cdot {}^{14}c_{wt}(t),$$

fast humus

$$^{14}\dot{c}_{fH}(t) = -(a + \lambda) \cdot {}^{14}c_{fH}(t) + \sum_{i=flt_i, frl, wl} \xi_i \cdot k_i \cdot {}^{14}c_i(t),$$

slow humus

$$^{14}\dot{c}_{sH}(t) = -(a \cdot s + \lambda) \cdot {}^{14}c_{sH}(t) + a \cdot \nu \cdot {}^{14}c_{fH}(t).$$

[2]For a complete derivation of the model equations refer to p.67 in Perruchoud (1996).

$^{14}c_j$ and $^{12}c_j$ denote the mass of ^{14}C (fractionation corrected) and ^{12}C of foliar ($j = flt_i$), fine root ($j = frl$), and woody litter ($j = wl$), woody vegetation ($j = wt$), fast ($j = fH$), and slow humus ($j = sH$). For the litter compartment j, $^{12}\mathcal{F}_{V,j}$ denotes its steady state carbon input flux from vegetation, k_j its decay rate and ξ_j the corresponding humification ratio (i.e. the fraction of decomposing litter which is not lost to the atmosphere as CO_2). λ is the radioactive decay rate of ^{14}C (8267yr^{-1}) and $^{14}\mathrm{R}_{stand.}$ the ^{14}C standard ratio ($\approx 10^{-12}$).

Atmospheric ^{14}C concentrations($\Delta^{14}\mathrm{C}_A(t)$) are prescribed in Δ^{14}C notation, thereby implicitly accounting for fractionation effects. ^{14}C enrichment in living wood ($^{14}c_{wl}$) is explicitly modeled to account for the slower growth and turnover rates of these tissues. On the other hand an instantaneous transfer of the atmospheric ^{14}C signal is assumed for foliar ($^{14}c_{flt_i}$) and fine root litter ($^{14}c_{frl}$).

For comparison of simulated and measured ^{14}C enrichment data in leaf litter and mineral soil, simulated values were expressed as weighted averages and converted to Δ^{14}C units consistently with the above equations:

$$
\begin{aligned}
\Delta^{14}\mathrm{C}_{leaf\,litter,n}(t) &= \left(\frac{^{14}R_{leaf\,litter,n}(t)}{^{14}\mathrm{R}_{\mathrm{stand.}}} - 1\right) \cdot 1000\ ‰, \quad \text{with} \\
^{14}R_{leaf\,litter,n}(t) &= \frac{\sum_{\forall flt}\ {}^{14}c_i(t)}{\sum_{\forall flt}\ {}^{12}c_i(t)}, \quad \text{and} \\
\Delta^{14}\mathrm{C}_{humus,n}(t) &= \left(\frac{^{14}R_{humus,n}(t)}{^{14}\mathrm{R}_{\mathrm{stand.}}} - 1\right) \cdot 1000\ ‰, \quad \text{with} \\
^{14}R_{humus,n}(t) &= \frac{^{14}c_{fH}(t) + {}^{14}c_{sH}(t)}{^{12}c_{fH} + {}^{12}c_{sH}}.
\end{aligned}
$$

Net carbon fluxes between atmosphere and biota were neglected as they have a marginal impact on modeled Δ^{14}C values. C stocks of litter and soil are assumed to be in steady state. This steady state assumption is unproblematic for the simulations of Δ^{14}C values and ^{14}C inventories in litter and soils.

Appendix B: Analytical solution for the soil organic carbon response to a step change in litter input

An analytical solution for the transient response of soil carbon stocks (Eq. (4) and Eq. (5)) can be given if a step change in litter input is assumed. For litter input rates of the form $u(t) = u_0$ for $t \leq t_0$ and $u(t) = u_0(1+\delta)$ for $t > t_0$ one obtains:

$$c_{fH}(t) = \begin{cases} \frac{u_0}{a} & \text{if } t \leq t_0 \\ \frac{u_0}{a}\left(1+\delta(1-e^{-a\cdot t})\right) & \text{otherwise,} \end{cases} \tag{9}$$

$$c_{sH}(t) = \begin{cases} \frac{u_0\cdot\nu}{as} & \text{if } t \leq t_0 \\ \frac{u_0\cdot\nu}{as(1-s)}\left(1+\delta(1-e^{-a\cdot s\cdot t}) - s(1+\delta(1-e^{-a\cdot t}))\right) & \text{otherwise,} \end{cases} \tag{10}$$

The relative increase of SOC stocks resulting from this increase in SOC input can be expressed as:

$$\frac{c_{SOC}(t)}{c_{SOC}(0)} = \frac{\rho - s}{1-s}\left(1+\delta\cdot(1-e^{-a\cdot t})\right) + \frac{1-\rho}{1-s}\left(1+\delta\cdot(1-e^{-a\cdot s\cdot t})\right), \tag{11}$$

where ρ is the relative contribution of fast humus to total SOC stocks for $t \leq t_0$:

$$\rho := \frac{c_{fH}}{c_{SOC}} = \frac{c_{fH}}{c_{fH}+c_{sH}} = (1+\frac{\nu}{s})^{-1}. \tag{12}$$

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Figure captions

Figure 1 : Structure of the decomposition model ForClim–D (Perruchoud, 1996).

Shaded boxes denote state variables, solid arrows indicate C mass fluxes and dashed arrows describe CO_2 losses associated with heterotrophic respiration. The parameters k_i are litter decomposition rates. ξ_i indicates the fraction of the litter efflux that is entering the fast humus compartment; the rest is respired to the atmosphere. a and $a \cdot s$ are the turnover rates of fast and slow humus C. ν is the fraction of the flux out of the fast humus compartment which enters the slow humus compartment. The absence of needle litter (*Conif. 1* and *Conif. 2*) and coarse woody debris (*Wood*) for the calibration of ForClim–D at Meathop Wood is indicated by open rectangles.

Figure 2 : Comparison of simulated and observed ^{14}C in the 0–5 cm and 0–15 cm mineral soil layer at Meathop Wood.

Simulations of ^{14}C in SOC were driven by the atmospheric ^{14}C history for the Northern Hemisphere of Enting et al. (1994). ^{14}C data were taken from Harkness et al. (1986) and Harrison and Harkness (1993) (see Tab. 2). Two data points (collection dates 1977 and 1982 denoted by "Topsoil (outlier)") were not used for parameter identification (see text).

All ^{14}C data are expressed in ‰ Δ^{14}C. The analytical measurement error is 10 ‰ Δ^{14}C, natural soil ^{14}C variability lies around 50–100 ‰ (Harrison et al., 1990).

Figure 3 : Effect of litter mineralization and humification on simulated ^{14}C in soil and heterotrophically respired CO_2 at Meathop Wood for a soil depth of 5 cm.

Atmospheric ^{14}C data for the Northern Hemisphere are from Enting et al. (1994). Note the higher levels of simulated ^{14}C if plant litter is bypassed ("bypass" simulation) compared to the data of Harkness et al. (1986) for the early 1970's and the convergence of the two simulations in the late 1980's.

All ^{14}C data are expressed in $^o/_{oo}$ $\Delta^{14}C$.

Figure 4 : Comparison of simulated and measured ^{14}C for the deciduous forest stand Haumättli investigated by Kuhn (1990) and Burri (1996).

Defaults for litter input and decomposition rates are given in Tab. 1 and resulted in an underestimation of measured ^{14}C values ("Topsoil (0–5 cm)"). A readjustment of litter inputs by +10% (1.32 t C ha^{-1} yr^{-1}) and SOC by 10% (11.9 t C/ha instead of 13.2 t C/ha) considerably improved the agreement between simulation and data ("Topsoil (0–5 cm, corr.)").

All ^{14}C enrichment data (Tab. 3) are expressed in $^o/_{oo}$ $\Delta^{14}C$. Measurement errors from accelerated mass spectroscopy ($\pm$7–9 $^o/_{oo}$ $\Delta^{14}C$) are indicated by error bars.

Figure 5 : Simulated $\Delta^{14}C$ air-biota disequilibrium and ^{14}C soil inventory at Meathop Wood for litter and soil (0–20 cm).

The $\Delta^{14}C$ air-biota disequilibrium was calculated as difference of observed atmospheric $\Delta^{14}C$ (Enting et al., 1994) minus simulated $\Delta^{14}C$ in heterotrophically respired CO_2. The ^{14}C soil inventory was determined from Eq. (8).

^{14}C data are expressed in $^o/_{oo}$ $\Delta^{14}C$. The ^{14}C soil inventory is given in units of 10^{12}atoms$\cdot m^{-2}$.

Figure 6 : Simulated relative response (expressed as [%/100]) of C stocks (fast plus slow humus) for step increases of C input at Meathop Wood (see Eq. (11)).

We assumed a turnover time of 25 years for fast humus with a contribution of 68% to the SOC inventory of the 0–20 cm mineral soil layer. Relative increases of input fluxes considered were 10, 20 and 35% which compare to the response of SOC to an instantaneous doubling of atmospheric CO_2 for growth enhancement factors β of 0.1, 0.2 and 0.35 respectively.
The horizontal lines at 8 and 12% emphasize the differences for distinct β factors and overall slow response of SOC.

Figure 7 : Relative changes in NLOC–stocks (left axis) as induced by CO_2 fertilization. Here, productivity is assumed to increase logarithmically ($\beta = 0.35$) with atmospheric CO_2 (right axis) as prescribed according to the IPCC IS92a scenario and S550 pathway. The S550 profile approaches stabilization at roughly twice the preindustrial CO_2 level (i.e. 550 ppmV) by 2150. The concentration for IS92a was calculated from prescribed anthropogenic carbon emissions with the Bern Model (Schimel et al., 1996b; Joos et al., 1996). Simulated NLOC changes represent changes of carbon stocks sequestered by litter and soil organic matter in the 0–20 cm mineral soil.

Table captions

Table 1 : Estimates of Litter Input and Decomposition Rates at Meathop Wood and Haumättli.

Litter inputs were taken from Sykes and Bunce (1970) and Harrison and Harkness (1993) for Meathop Wood. For Haumättli, foliar and twig litterfall was estimated based on the litterfall model of Meentemeyer et al. (1982). Fine root litter input was assumed to equal fine root production which was determined after Nadelhoffer and Raich (1992). All fine root litter inputs refer to the 0–5 cm soil layer. Decomposition rates for fine root and foliar litter were determined by matching simulated and observed levels of ^{14}C at Meathop Wood. Decomposition of twigs was adopted from Pastor and Post (1985) and Bugmann (1994). Estimates of fine root and twig litter decomposition for Meathop Wood were also applied for the Haumättli site.

Table 2 : Observed $\Delta^{14}C$ and SOC Inventory in Soil at Meathop Wood.

$\Delta^{14}C$ data (in ‰) for the 0–5, 5–10, and 10–15 cm layers correspond to measurements in Spring (Sp.) or Autumn (Au.) (Harkness et al., 1986). Values for the 0–10 cm and 0–15 cm layers are SOC inventory-weighted averages (see Harrison and Harkness (1993)). ^{12}C denotes carbon inventories in t C/ha. For a soil depth of 0–20 cm we used 52 t C/ha.

Table 3 : Organic Carbon Content, Soil Bulk Density and Soil Isotopic Data at Haumättli.

All data are for the 0–5 cm forest soil layer sampled in 1969, 1988 (Kuhn, 1990), 1993 (Burri, 1996), and 1995 (this study). Soil organic matter was measured following aggregation of seven soil cores collected within two hectares (Burri, 1996). Bulk densities for all collection dates were assumed to equal the value measured in 1993 by Burri (1996). Isotopic ratios were measured by accelerated mass spectroscopy (AMS) at the AMS laboratory in Zürich (PSI/ETH).

Table 4 : Comparison of Published Turnover Times and Fractions of Fast and Slow Overturning Humus in the Topsoil Layers.

The estimates of Harrison et al. (1993a), Harrison et al. (1993b), Harrison et al. (1995), and Harrison (1996) are based on a world-wide compilation of forest and grassland soil samples. Those of Townsend et al. (1995) and Trumbore et al. (1996) are from an altitudinal transect in Hawaii and in the Sierra Nevada. The range of this data reflects turnover time estimates derived for different sites and not an uncertainty as the range indicated for this study. The latter resulted from parameter intervals obtained for $RMSE = 0.03$ (cf. Tab. 5) via Eq. (7). Abbreviation: nd = not determined.

Table 5: FORCLIM–D Model Parameter Values and Uncertainty Ranges at Meathop Wood.

The table gives defaults and uncertainty ranges for FORCLIM–D model parameters (cf. Fig 1) which control the dynamics of litter and topsoil C (0–20 cm). The root mean square error ($RMSE$, see p. 7) determined from observed and simulated ^{14}C data was on the order of the measured soil ^{14}C variability (Harrison et al., 1990) and within 15% of the final minimal value ($RMSE = 0.0262$). ξ_{foliar}, $\xi_{f.roots}$, and ν are dimensionless transfer ratios which control the fraction of decomposing litter C being transferred to the fast humus compartment (ξ_i) and the fractional C flux between the fast and the slow humus compartment (ν). c_{fH}, c_{sH} denote the percentage of fast and slow humus in 20 cm topsoil, and τ_{fH}, τ_{sH} the respective turnover times in years.

The parameters ξ_{foliar}, $\xi_{f.roots}$, s, and s/ν were identified with SUFI (Abbaspour et al., 1997) and a constrained by Eq. (7). The ratio s/ν representing the ratio of ^{12}C masses for the fast and the slow humus compartment was expressed by c_{fH} and c_{sH} and rate parameters $(a,\ 1/(a \cdot s))$ converted to figures for C turnover via $\tau_{fH} = 1/a$ and $\tau_{sH} = 1/(a \cdot s)$. Due to homogeneity of the topsoil at Meathop Wood ξ_{foliar}, $\xi_{f.roots}$, s, and ν were derived from ^{14}C data for the 0–5 cm soil layer and extrapolated to 0–20 cm soil depth.

Table 6 : Change in Soil and Litter Carbon Storage Induced by a Postulated CO_2 Fertilization of Vegetation.

Simulated carbon storage in nonliving organic carbon (NLOC) as induced by a postulated plant growth enhancement due to increasing atmospheric CO_2 levels. The same effect was assumed for vegetation growth and litter production for the sake of simplicity. Atmospheric CO_2 levels were prescribed according the IPCC scenario IS92a and S550. Two different estimates of the preindustrial NLOC stock were considered as calculated by assuming an area of $13.2 \cdot 10^{12} \cdot \mathrm{m}^{-2}$ for cold deciduous forests (Dai and Fung, 1993; King et al., 1997) and initial C densities of litter and mineral soil (0–20 cm soil depth) of 50 and 70 t C ha^{-1}, respectively (Perruchoud, 1996). Numbers in parentheses indicate the change relative to the preindustrial carbon stock.

Litter Type	Input rate [t C ha^{-1} yr^{-1}]	Decomposition rate k [yr^{-1}]
	Meathop Wood	
Acer pseudoplatanus L.	0.048	0.470
Betula spp.	0.168	0.430
Corylus avellana L.	0.212	0.470
Fraxinus excelsior L.	0.365	0.510
Quercus petraea Liebelein	0.625	0.430
Fine roots	0.400	0.690
Twigs	0.640	0.220
	Haumättli	
Foliage	1.200	0.540
Fine roots	0.600	0.690
Twigs	0.700	0.220

Table 1: Estimates of Litter Input and Decomposition Rates at Meathop Wood and Haumättli.

Year	0–5 cm		5–10 cm		10–15 cm		0–10 cm		0–15 cm	
(Season)	Δ^{14}C	^{12}C	Δ^{14}C	^{12}C	Δ^{14}C	^{12}C	Δ^{14}C	^{12}C	Δ^{14}C	^{12}C
1973 (Au.)	160		10		-50		94		55	
1975 (Au.)	220		15		—		130		—	
1977 (Au.)	310	18	15	14	-40	12	180	32	120	44
1984 (Sp.)	170		90		50		135		112	
1986 (Sp.)	130		22		-50		82		47	

Table 2: Observed Δ^{14}C and SOC Inventory in Soil at Meathop Wood.

Collection date	C–content [%]	Bulk density [g/cm^3]	Δ^{14}C [‰]	δ^{13}C [‰]	Laboratory number
1969	2.38	—	158	–25.7	ETH–15004
1988	2.28	—	174	–21.9	ETH–15006
1993	2.40	1.12	164	–26.3	ETH–15007
1995	—	—	125	–28.5	ETH–15329

Table 3: Organic Carbon Content, Soil Bulk Density and Soil Isotopic Data at Haumättli.

Humus turnover time		Fast humus			
Fast [yr]	Slow [yr]	fraction [%]	Soil type	Depth	Reference
25	3700	75	Native forest and grassland	A horizon	Harrison et al. (1993b)
25	3700	50	Cultivated	A horizon	Harrison et al. (1993a)
25	4700	82	Native forest and grassland	A horizon	Harrison et al. (1995)
25	4700	83	Native forest and grassland	A horizon	Harrison (1996)
12	4700	70	Recovering forest soil	A horizon	Harrison et al. (1995)
15–30	nd	70–75	Forest and pasture	20 cm	Townsend et al. (1995)
35–62	nd	50–90	Temperate forest	12–23 cm	Trumbore et al. (1996)
9-50	*155-10018*	*38-74*	*Native temperate forest*	*20 cm*	*This study*

Table 4: Comparison of Published Turnover Times and Fractions of Fast and Slow Overturning Humus in the Topsoil Layers.

Parameter	Default	*RMSE*		
		0.03	0.028	0.027
ξ_{foliar}	0.51	0.06–0.50	0.06–0.50	0.39–0.50
$\xi_{f.roots}$	0.27	0.27–0.90	0.27–0.90	0.32–0.84
ν	0.00329	0.0018–0.0887	0.0018–0.0187	0.0018–0.0082
c_{fH} [%]	68	38–74	57–74	65–74
c_{sH} [%]	32	26–62	26–43	26–35
τ_{fH} [yr]	25	9–50	13–50	15–29
τ_{sH} [yr]	3570	155–10018	511–10018	993–5838

Table 5: ForClim–D Model Parameter Values and Uncertainty Ranges at Meathop Wood.

Period	NLOC change [Gt C] (% change)	
	IS92a	S550
	NLOC(1770 AD) = 66 Gt C	
1770–2000	3.4 (5.1)	3.4 (5.1)
1770–2050	7.7 (11.7)	6.6 (10.0)
1770–2100	13.1 (19.9)	9.4 (14.2)
1980–1989	0.4 (0.7)	0.4 (0.7)
2090–2099	1.0 (1.5)	0.4 (0.7)
	NLOC(1770 AD) = 92 Gt C	
1770–2000	4.7 (5.1)	4.7 (5.1)
1770–2050	10.8 (11.7)	9.2 (10.0)
1770–2100	18.4 (19.9)	13.1 (14.2)
1980–1989	0.6 (0.7)	0.6 (0.7)
2090–2099	1.4 (1.5)	0.6 (0.7)

Table 6: Change in Soil and Litter Carbon Storage Induced by a Postulated CO_2 Fertilization of Vegetation.

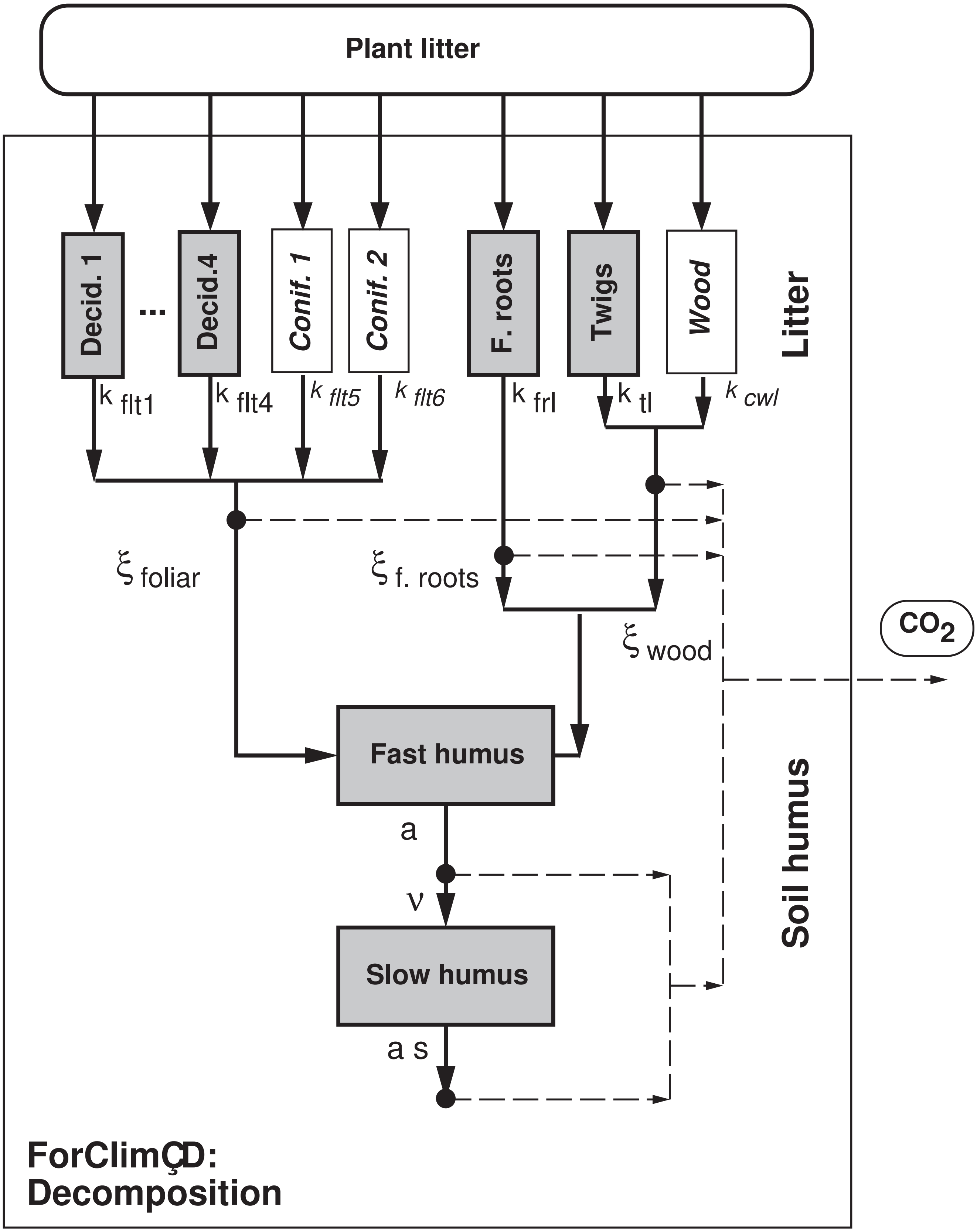
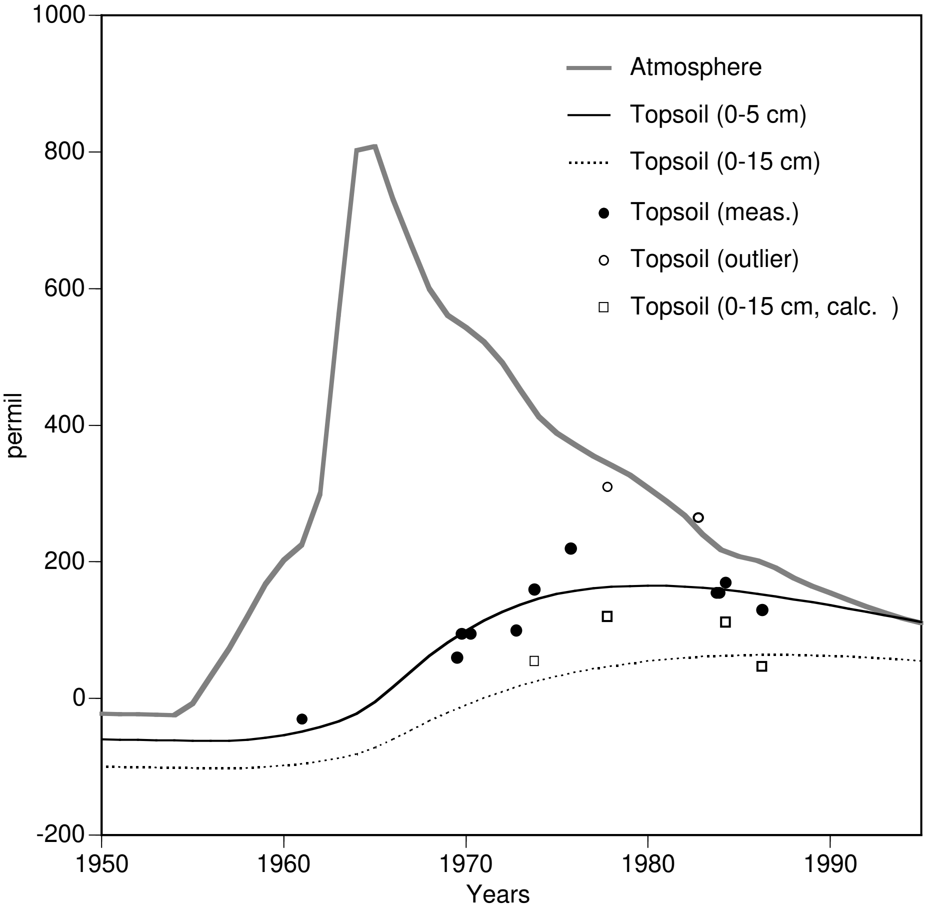

Plant litter
Decid. 1
...
Decid.4
Conif. 1
Conif. 2
F. roots
Twigs
Wood
Litter
k flt1
k flt4
k flt5
k flt6
k frl
k tl
k cwl
ξ foliar
ξ f. roots
ξ wood
CO2
Fast humus
a
ν
Slow humus
a s
Soil humus
ForClim-D:
Decomposition

Figure 1:



Atmosphere
Topsoil (0-5 cm)
Topsoil (0-15 cm)
Topsoil (meas.)
Topsoil (outlier)
Topsoil (0-15 cm, calc.)
permil
1000
800
600
400
200
0
-200
1950
1960
1970
1980
1990
Years

Figure 2:

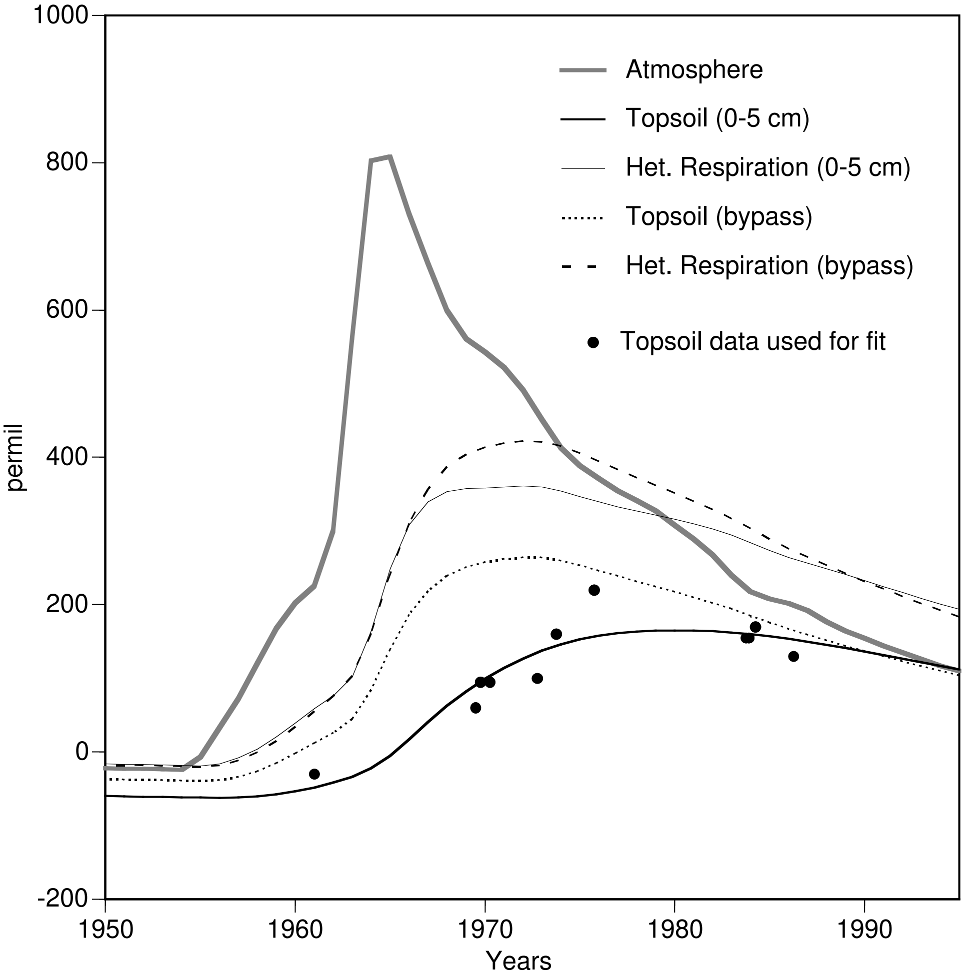
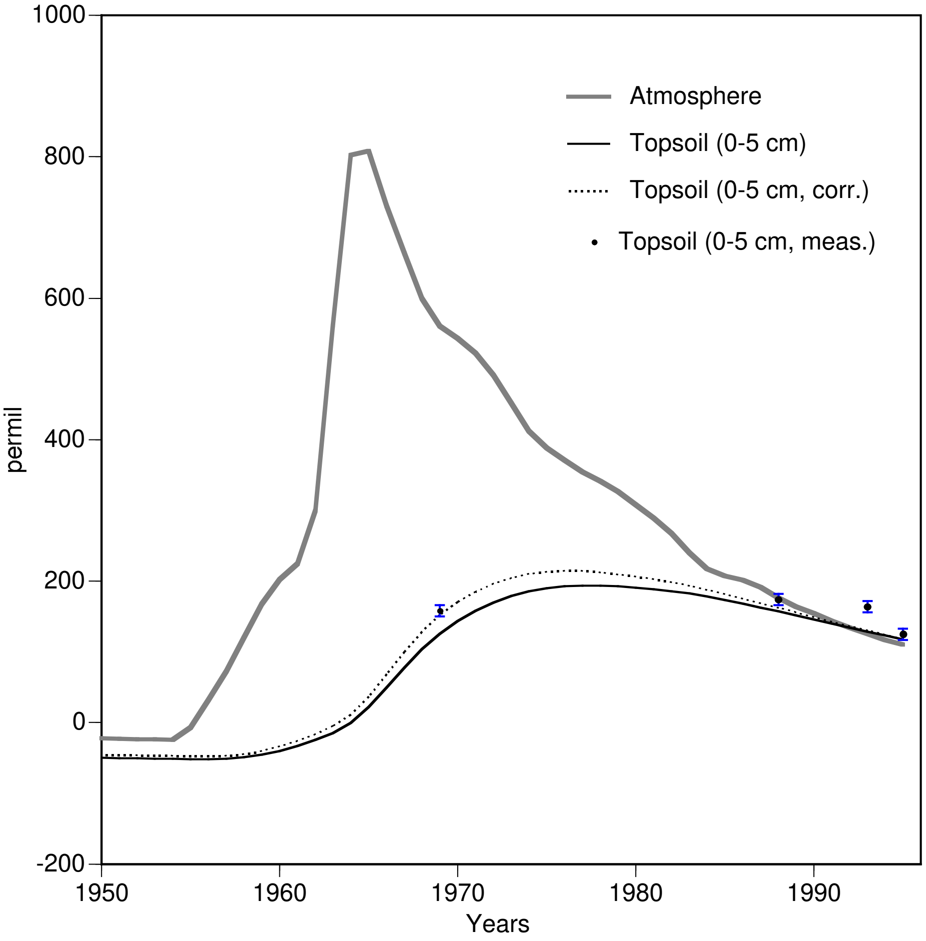

1000
800
600
400
200
0
-200
permil
1950
1960
1970
1980
1990
Years
Atmosphere
Topsoil (0-5 cm)
Het. Respiration (0-5 cm)
Topsoil (bypass)
Het. Respiration (bypass)
Topsoil data used for fit

Figure 3:



Atmosphere
Topsoil (0-5 cm)
Topsoil (0-5 cm, corr.)
Topsoil (0-5 cm, meas.)
1000
800
600
400
200
0
-200
permil
1950
1960
1970
1980
1990
Years

Figure 4:

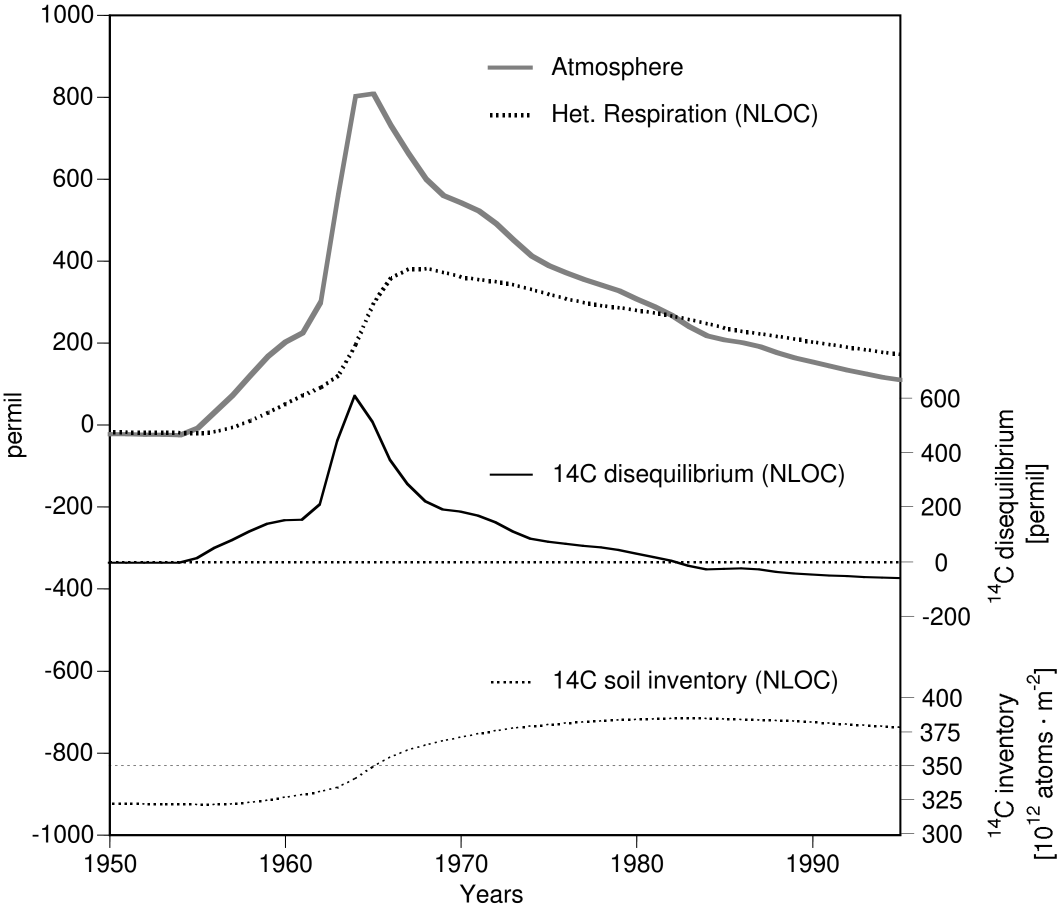
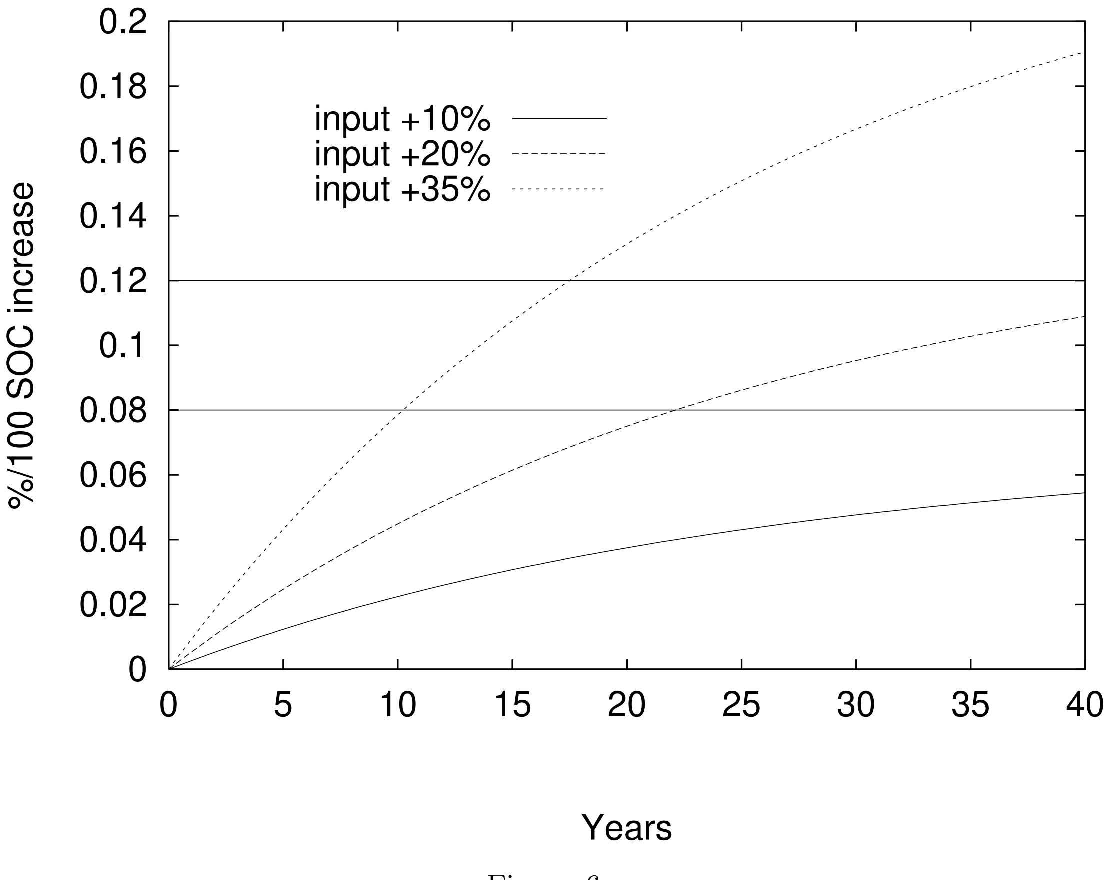
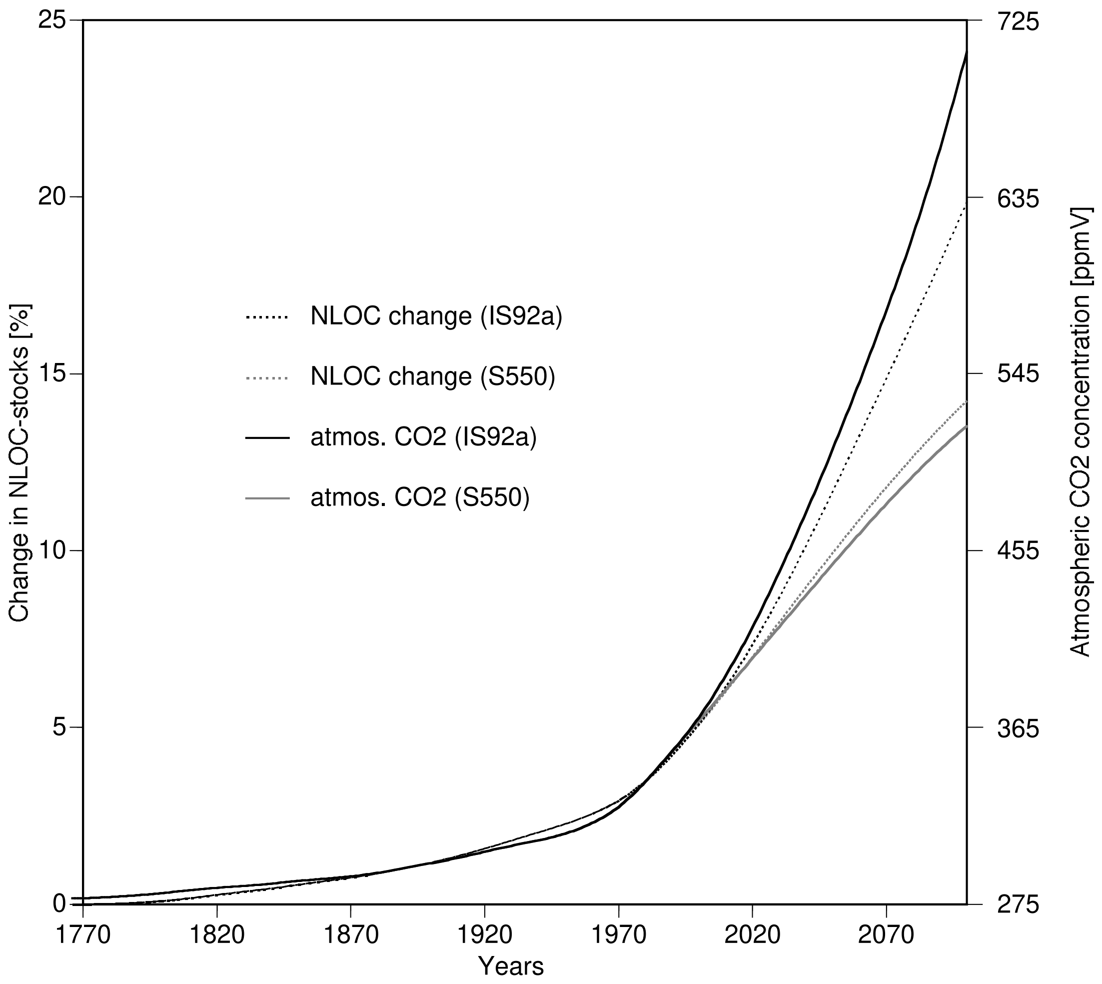

Atmosphere
Het. Respiration (NLOC)
14C disequilibrium (NLOC)
14C soil inventory (NLOC)
permil
1000
800
600
400
200
0
-200
-400
-600
-800
-1000
1950
1960
1970
1980
1990
Years
600
400
200
0
-200
^{14}C disequilibrium [permil]
400
375
350
325
300
^{14}C inventory [10^{12} atoms · m^{-2}]

Figure 5:



%/100 SOC increase
0.2
0.18
0.16
0.14
0.12
0.1
0.08
0.06
0.04
0.02
0
input +10%
input +20%
input +35%
0
5
10
15
20
25
30
35
40
Years

Figure 6:



NLOC change (IS92a)
NLOC change (S550)
atmos. CO2 (IS92a)
atmos. CO2 (S550)
Change in NLOC-stocks [%]
Atmospheric CO2 concentration [ppmV]
Years
0
5
10
15
20
25
275
365
455
545
635
725
1770
1820
1870
1920
1970
2020
2070

Figure 7: