

# Radiocarbon based assessment of soil organic matter contribution to soil respiration in a pine stand of the Campine region, Belgium

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**Abstract** The contribution of decomposing soil organic carbon (SOC) to total annual soil respiration (SR) was evaluated by radiocarbon measurements at a Scots pine stand growing on a plaggen soil in the Belgian Campine region. Two approaches were used to estimate the contribution of different C pools to SR. In the first approach, the variations in  $^{14}\text{C}$  content of soil  $\text{CO}_2$  efflux were monitored during one year (2003) and compared to the atmospheric and SOC  $^{14}\text{C}$  signatures to determine the contribution of “fast” (root respiration and fast decomposing SOC) and “slow” cycling C pools to total SR. In the second approach an estimate of the total heterotrophic soil respiration (Rh), comprising the slow cycling C and

the heterotrophic part of the fast-cycling C pools, was derived applying a box model based on the amount of the bulk SOC pool and its  $^{14}\text{C}$ -derived mean residence time (MRT). The quantification of the Rh and the decomposition rate of the slow-cycling SOC allows to indirectly determining the contribution of the heterotrophic C that decompose within a year. Measurements of total SR performed in the field allowed assessing the contribution of the different C pools to total soil C efflux. On an annual basis, the fast-cycling C was the main contributor to SR, about 85%, while the contribution of the slow-cycling C (with MRT >1 yr) to total SR was 15%. Total annual Rh was 36% of total SR, which is in the lower range

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reported for temperate coniferous forests. The comparison of Rh with other estimates for the same site (47–50% of total SR) suggest a possible underestimation of the C flux from the mineral soil. In fact, the “very old” C contained in the plaggen horizon strongly affects the signature of the mostly young C leaving the soil. In conclusion, our results indicate that the contribution of SOC decomposition to total soil CO<sub>2</sub> flux in this forest is less than 40%, and at least half of it comes from organic compounds less than 1 year old.

**Keywords** Forest soil · Flux partitioning · Radiocarbon (<sup>14</sup>C) · Soil CO<sub>2</sub> efflux · Soil organic carbon (SOC)

## Introduction

Soil is the largest reservoir of organic carbon (C) in terrestrial ecosystems, storing about 1550 Pg of carbon in dead organic matter (Lal 2004). Under increasing atmospheric CO<sub>2</sub> concentrations and subsequently changing climate, this large organic carbon stock is likely to change and interact with the climate system (Jenkinson et al. 1991; Bond-Lamberty and Thomson 2010). In fact, soil organic carbon (SOC) is an active pool that via soil respiration (SR) contributes to the annual flux of CO<sub>2</sub> towards the atmosphere. Such a SOC contribution in the 1990s globally exceeded eighteen-fold the rate of fossil fuel release (Prentice et al. 2001). In temperate forests, soil respiration accounts for 60–90% of total ecosystem respiration, therefore representing a major flux in the terrestrial carbon cycle (Luysaert et al. 2007). Soil CO<sub>2</sub> is produced by SOC decomposition and root respiration, which are the heterotrophic and autotrophic components of soil CO<sub>2</sub> flux, respectively. An increase in root respiration may indicate increased carbon inputs to the soil through greater photosynthesis, specific root activity, or root biomass (Hungate et al. 1997); hence it is not an evidence of enhanced soil carbon losses. In contrast, an increase in the contribution from SOC decomposition to SR reduces the potential for soil carbon storage (Borken et al. 2006), if not balanced by increased C inputs. It is therefore important to recognize the relative contributions of autotrophic vs. heterotrophic components of SR to provide more insight in the interactions between global climate change and soil processes.

Flux measurements alone cannot distinguish the CO<sub>2</sub> produced by autotrophic vs. heterotrophic sources, despite different approaches having been used for this purpose (Hanson et al. 2000; Kuzyakov 2006; Subke et al. 2006). Against this background, the tracing of “bomb” radiocarbon in terrestrial ecosystems has emerged as a powerful tool for quantifying the two components of soil respiration (Dörr and Münnich 1986; Borken et al. 2006; Hahn et al. 2006; Schuur and Trumbore 2006; Trumbore et al. 2006; Trumbore 2006). “Bomb” <sup>14</sup>C was released by atmospheric nuclear weapons tests mainly in the 1950s and early 1960s, when the amount of radiocarbon in atmospheric CO<sub>2</sub> was nearly doubled. Some countries continued atomic tests also afterwards, the last one being performed by China in 1980 ([www.ctbto.org](http://www.ctbto.org)). After the nuclear tests period (1950–1960), the amount of <sup>14</sup>C in atmospheric CO<sub>2</sub> declined mainly because bomb <sup>14</sup>C was mixed into ocean and in minor part because it was mixed in terrestrial C reservoir and diluted by the combustion of radiocarbon-free fossil fuel (Levin and Hesshaimer 2000). Depending on the residence time of organic matter in soil and the relative contribution of the two components of soil respiration (autotrophic and heterotrophic), the <sup>14</sup>C concentration of soil CO<sub>2</sub> differs from that of atmospheric CO<sub>2</sub>. It is usually assumed that the CO<sub>2</sub> produced by root respiration has the same <sup>14</sup>C concentration of the atmosphere (Wang et al. 2000), the CO<sub>2</sub> derived from the decomposition of organics synthesised before the 1950s is depleted in <sup>14</sup>C by radioactive decay, and the CO<sub>2</sub> derived from the decomposition of organics synthesised after the 1950s is enriched in CO<sub>2</sub> by the contribution of bomb radiocarbon. On these bases, monitoring the <sup>14</sup>C concentrations in SR and identifying the signature of the pools contributing to SR allow partitioning soil CO<sub>2</sub> flux into recent and older components.

The aim of this study is to estimate the contribution of different C pools to total soil respiration using radiocarbon measurements. In particular, we determined the contribution of: i) a “fast” cycling C pool, which includes the root-respired C and the fast decomposing SOC with short ecosystem residence time, ii) a “slow” cycling C pool, the one from the decomposition of SOC with residence time longer than a few years, and iii) the total heterotrophic respired C, comprising the entire slow cycling pool and the non-autotrophic fraction of the “fast” one.

## Materials and methods

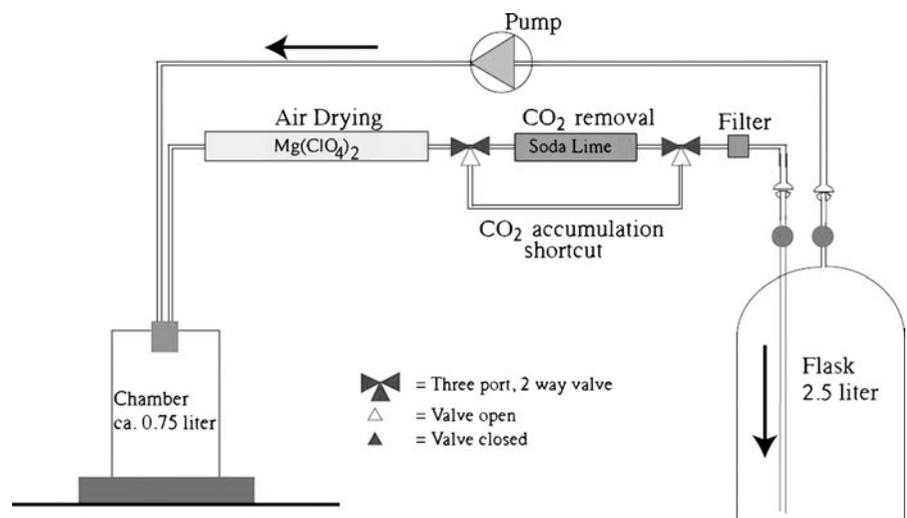
The study site is an even-aged, 74-year-old Scots pine (*Pinus sylvestris* L.) stand, representing a portion of a 150 ha mixed coniferous/deciduous forest—De Inslag—in Brasschaat (51° 18' 33" N, 4° 31' 14" E), in the Campine region, Northern Belgium. The mean annual temperature (MAT) is 9.8°C and the mean annual precipitation (MAP) is 750 mm (Carboeurope IP database 2011). During our study, in 2003, Europe experienced anomalous climatic conditions, with July temperatures up to 6°C above the long-term mean (1900–2003), and annual precipitation deficit up to 300 mm, 50% below the average as reported by Ciais et al. (2005). At our study site, July temperature was only 1.4°C above the average of the previous five years, but annual precipitation was more than 200 mm less than the average. The site was a low productive heathland in the 18th century (Gemeentekrediet 1965), up to the beginning of the 20th century when the stand was planted. The relatively high SOC stock of the stand, 172.6 Mg ha<sup>-1</sup>, is due to substantial human-induced additions of organic material, often several hundred years old, a common amendment practice in this sandy region that resulted in plaggen formation (Chiti et al. 2009). The soil, which is classified as Endogleyc Regosol (Dystric, Arenic, Transportic) according to the World Reference Base for Soil Resources (IUSS Working Group 2006), is covered by an organic layer about 8 cm thick, which is stable

throughout the year, and classifiable as Dysmoder type of humus. More detailed information on the site characteristics is available in Curiel Yuste et al. (2005a) and Chiti et al. (2009).

Soil respiration was measured twice a month from early January 2003 until the end of January 2004, using a closed dynamic chamber system (Infrared Gas Analyzer: CIRAS-1, soil chamber: SRC-1, both PP-Systems, Hitchin, UK) and ten soil collars randomly distributed within the stand. To mitigate potential bias due to high spatial variability, the surface area sampled by the chamber was enlarged from 78.5 cm<sup>2</sup> to 314 cm<sup>2</sup> by attaching a PVC-rim to the base of the chamber (Fig. 1). The bottom side of the PVC rim has a slot in which a rubber joint provided an airtight seal for the soil collars that were permanently installed in the soil of the stand (Janssens et al. 2000; Curiel Yuste et al. 2003). The modification of the chamber was assessed to do not alter the CO<sub>2</sub> fluxes (Janssens et al. 2000). Each of the ten measurements was made in duplicate and the mean was used in subsequent calculations, as reported in Curiel Yuste et al. (2005b).

In the same period (January 2003–January 2004), soil CO<sub>2</sub> was collected for isotope analysis in triplicate once per season, except in summer, when it was collected twice, at the beginning and at the end of the season. For this purpose, we used a modified dynamic flow chamber system, analogous to the one used for soil CO<sub>2</sub> flux measurements (Norman et al. 1997; Gaudinski et al. 2000). The soil chamber

**Fig. 1** Modified dynamic flow chamber system used for collecting the CO<sub>2</sub> from soil

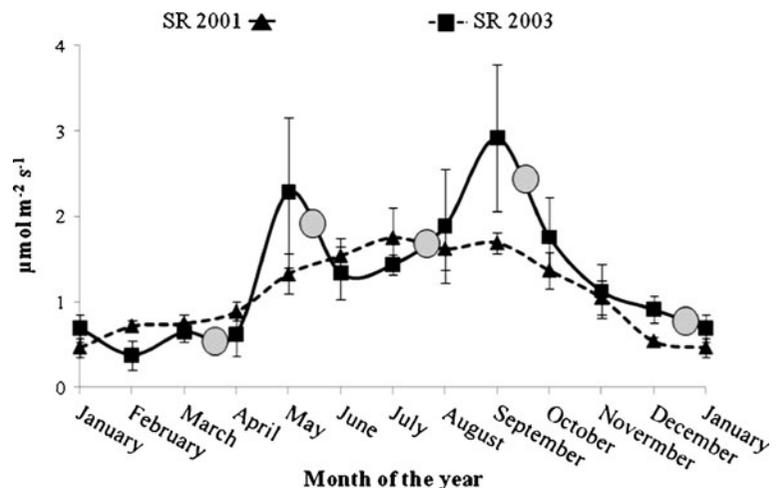


volume of approximately 0.75 L was extended with the 2.5 L of a glass flask (Fig. 1), the soil atmosphere being flushed in series through its two vacuum stop-cocks (Louwers, Hapert, NL), one of them connected to the bottom of the flask by internal glass tubing. The components were connected by 6.35 mm o.d. polyethylene-coated aluminium tubing (Dekoron Synflex 1300). After leaving the soil chamber, the air flow first passed through a  $\text{Mg}(\text{ClO}_4)_2$  filled water trap. As a second step in conditioning the air in the sampling system, a soda lime  $\text{CO}_2$  trap was used to remove all atmospheric air  $\text{CO}_2$  from the system after installation, i.e. under the soil chamber, in the tubing, drying cartridge and sample flask. Only the soda lime short cut tubing stayed with its air  $\text{CO}_2$  content. At the given length of 15 cm, the volume is 2.2 mL and its contribution negligible. The air was pumped into the soil chamber by a membrane pump (KNF Neuberger, Freiburg, Germany) at more than  $2 \text{ L min}^{-1}$ . As it was a closed system, the same amount of air has been sucked out of the soil chamber as had been pumped into it, without the need of a significant pressure gradient at the chamber. By this, the pressure in the chamber remained equal to atmospheric pressure. After 15 min of air circulation, as the third step in sampling, the soda lime cartridge was removed from the cycle by switching in the short-cut tubing using the three-port-two way-valves, so that the air between the soil chamber and the flask was dried, while  $\text{CO}_2$  started to accumulate in the system. The  $\text{CO}_2$  evolving from soil into the chamber was mixed with the  $\text{CO}_2$ -free air with which the flask

had been pre-filled. For the first sampling, a minimum flushing time of 15 min had been applied. The flask was closed, hence the  $\text{CO}_2$  contained in it was of pure and isotopically undiluted soil efflux origin. With better knowledge of the system and the soil fluxes, the minimum flushing time has been extended to 25 min for the following samplings. Aimed-at  $\text{CO}_2$ -concentrations were around atmospheric, in order to keep the sample preparation comparable to the atmospheric routine samples. Actual  $\text{CO}_2$ -concentrations in the flasks varied between 200 ppm in some samples from the first day, up to 460 ppm later on. Dates of sampling are sketched in Fig. 2.

For  $^{14}\text{C}$  measurements, soil  $\text{CO}_2$  was cryogenically extracted and converted into graphite targets according to Aerts-Bijma et al. (1997). The graphite targets were analysed for their  $^{14}\text{C}$  concentration by Accelerator Mass Spectrometry (AMS) at the Centre for Isotope Research of the University of Groningen, The Netherlands. The AMS measures directly the isotopic ratios  $^{14}\text{C}/^{12}\text{C}$  and  $^{13}\text{C}/^{12}\text{C}$  (van der Plicht et al. 2000), with typical measurement uncertainties of 4‰ and 0.1‰, respectively (Meijer et al. 2006). Radiocarbon data are expressed as  $\Delta^{14}\text{C}$  according to Stuiver and Polach (1977). For the smaller samples, the measurement uncertainties are larger than for the samples at standard size. I.e., at atmospheric  $\text{CO}_2$  concentrations (about 0.37 mg C) we do find uncertainties of 4‰ to 5‰ in  $\Delta^{14}\text{C}$ , whereas the smallest samples (about 0.2 mg C) have typical uncertainties of 7‰ (5‰ to 10‰), however, without a clear size dependency. The  $^{14}\text{C}$  signature of the

**Fig. 2** Monthly soil  $\text{CO}_2$  efflux in 2003 and 2001 (the latter is a year without climate anomalies). Grey dots indicate the sampling campaign for  $^{14}\text{C}$  measurements during 2003. Error bars represent standard deviation of the mean ( $n=10$ )



respired C is the sum of the contributions of substrates that are producing CO<sub>2</sub> at different rates and that have different radiocarbon concentrations. Positive  $\Delta^{14}\text{C}$  values reveal enrichment in  $^{14}\text{C}$  from nuclear weapons testing, mostly in the 1960s, while negative  $\Delta^{14}\text{C}$  values indicate that the organic matter has resided in soil long enough for significant  $^{14}\text{C}$  decay.

#### Contribution of fast and slow-cycling C to annual soil respiration

Following the annual  $^{14}\text{C}$  variations in the atmosphere and in the soil atmosphere allows partitioning the soil CO<sub>2</sub> flux into: a) “fast-cycling” C, the one from root respiration and decomposition of organic material with mean residence time (MRT) in soil shorter than 1 year, and b) “slow-cycling” C from the decomposition of organic material with higher residence time. The relative contribution of these two CO<sub>2</sub> sources to SR can be calculated by a simple two component-mixing model as described by Dörr and Münnich (1986), assuming the  $^{14}\text{C}$  signature of the fast-cycling C to be the same of the current atmospheric CO<sub>2</sub> (Wang et al. 2000) and the  $^{14}\text{C}$  signature of the slow-cycling C the same of the bulk SOC. In the former case we referred to  $^{14}\text{C}$  atmospheric values recorded in 2003 from the nearest  $^{14}\text{C}$  station, located at Smilde, in The Netherlands, about 300 km north-east of the study site (Palstra et al. 2008), while in the latter case we used the weighed means of  $^{14}\text{C}$  and SOC concentrations measured in place by Chiti et al. (2009), in the organic layer and the top two mineral soil horizons (Table 1). Accordingly, the relative contribution of the fast and slow-cycling C to total soil CO<sub>2</sub> efflux was calculated following Eq. 1:

$$^{14}\text{C}_{\text{SR}} = ^{14}\text{C}_{\text{atm}}C_1 + ^{14}\text{C}_{\text{SOC}}C_2 \quad (1)$$

$^{14}\text{C}_{\text{SR}}$  =  $^{14}\text{C}$  signature of respired CO<sub>2</sub>;

$^{14}\text{C}_{\text{atm}}$  =  $^{14}\text{C}$  signature of atmospheric CO<sub>2</sub> from Smilde station, The Netherlands;

$^{14}\text{C}_{\text{SOC}}$  =  $^{14}\text{C}$  signature of SOC, weighed mean of the  $^{14}\text{C}$  in the organic layer and the top 45 cm of mineral soil (A1 and A2 mineral horizons);  
C<sub>1</sub>, C<sub>2</sub> = relative contribution of the fast and slow cycling C pools (C<sub>1</sub> + C<sub>2</sub> = 1).

#### Heterotrophic contribution to annual soil respiration

Accounting for the SOC stock and its MRT allows calculating the heterotrophic output C flux from a specific compartment (*e.g.* soil horizons, depth, physical or chemical fractions), as described by the box model proposed by Harrison et al. (2000). The box model assumes that the annual input and output of C of any SOC compartment are equivalent. The  $^{14}\text{C}$  content of the C input to the plant box corresponds to that of the atmospheric CO<sub>2</sub>, but in the model one has to take into account the residence time of C in the plant before it can be incorporated into the soil. The  $^{14}\text{C}$  content of the output equals that of any given box in the respective year. The C fluxes from the litter and the mineral soil, down to 45 cm depth, were hence calculated according to Eq. 2:

$$\text{C flux (Mg ha}^{-1}\text{ y}^{-1}) = \text{C pool (Mg ha}^{-1})/\text{C MRT (years)} \quad (2)$$

Data on quantity and MRT of SOC for 2003 were taken from Chiti et al. (2009). To obtain an estimate of total soil heterotrophic respiration (Rh), the model-derived fluxes from all soil horizons were summed. With our data it was theoretically possible to apply the two pools model proposed by Hahn et al. (2006), which derives soil Rh fluxes by  $^{14}\text{C}$  measurements of SOC. However, we realized that the application of such a model is not possible at this site, because it assumes as responsible of soil respiration only the portion of soil where bomb C prevails by far. However, in the studied plaggén soil the presence of bomb C is strongly diluted by the presence of ancient organic C, hence preventing any discrimination of the influence of bomb C already in the top cm of soil.

**Table 1** Carbon stocks, mean residence time of SOC and heterotrophic respiration (Rh) in the investigated soil horizons. Numbers on the right side are the uncertainties

	Horizon	Organic C [Mg ha <sup>-1</sup> ]	$\Delta^{14}\text{C}$ (‰)	MRT (years)	Rh [Mg C ha <sup>-1</sup> yr <sup>-1</sup> ]
Pine	Litter	29±0.7	158.0±4.5	17±1	1.705±0.125
	A1	12±0.3	-142.7±5.1	1480±40	0.008±0.001
	A2	95±0.5	-170.5±4.3	1789±45	0.053±0.001

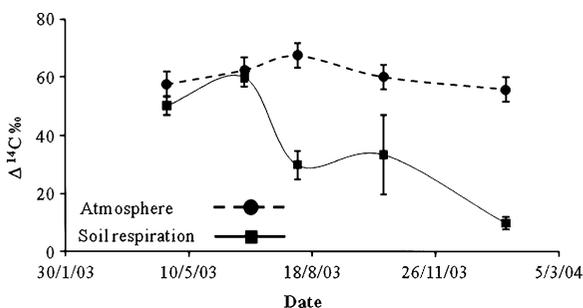
To test the reliability of the estimated Rh fluxes we compared them with other Rh estimates available for this site, obtained using different methodologies such as the mass balance approach (Janssens et al. 1998), the SECRETS (Stand to Ecosystem CaRbon and EvapoTranspiration Simulator) model (Sampson et al. 2001), and a mix of eddy covariance and chamber-based CO<sub>2</sub>-flux measurements (Nagy et al. 2006).

## Results

The annual trend of temperatures in 2003 does not show substantial differences with the average trend of the five previous years, while the annual precipitation decrease is about 25% compared to the average of the same five years (Fig. 4).

The annual soil C efflux in 2003 amounted to  $5.0 \pm 0.5 \text{ Mg ha}^{-1} \text{ yr}^{-1}$ . This efflux is not significantly different from that of 2001 (t test,  $p < 0.05$ ), which amounted to  $4.1 \pm 1.6 \text{ Mg C ha}^{-1} \text{ yr}^{-1}$  (Curiel Yuste et al. 2005b) and did not experience anomalous climatic conditions. Looking at the monthly data in Fig. 2, the slightly higher annual SR in 2003 could be attributable to higher fluxes during two periods, April to May and August to September, the latter period being significantly different from the same period in 2001 (t-test,  $p < 0.05$ ).

The <sup>14</sup>C concentrations in the atmosphere and the soil atmosphere are shown in Fig. 3. The <sup>14</sup>C signature of the soil CO<sub>2</sub> does not increase significantly (t-test,  $p < 0.05$ ) from spring (50.2‰) to early summer (60.1‰), while it decreases significantly at



**Fig. 3** General pattern of  $\Delta^{14}\text{C}$  in 2003 for soil atmosphere and outer atmosphere [(data of the latter are from Smilde station, The Netherlands, by Palstra et al. (2008)]. Error bars represent standard deviation of the mean ( $n=3$ )

the end of the same season (29.9‰). Later it remains rather constant until winter, when it drops to 9.8‰.

The calculated contributions of the fast and slow-cycling C to total SR are reported in Table 2. The fast-cycling C resulted the main contributor of SR all over the year, showing two peaks: in spring (when it represents  $95 \pm 5\%$  of total SR) and in early summer ( $98 \pm 7\%$  of total SR). On the contrary, the relative contribution of the slow-cycling C increased through the year, with a maximum of  $31 \pm 4\%$  of total SR in winter.

The estimated Rh in 2003, obtained using the box model, was  $1.766 \pm 0.125 \text{ Mg C ha}^{-1} \text{ yr}^{-1}$ . The litter layer was the main contributor to this flux,  $1.705 \pm 0.125 \text{ Mg C ha}^{-1} \text{ yr}^{-1}$ , while the underlying mineral A1 and A2 horizons contributed much less to Rh,  $0.008 \pm 0.001$  and  $0.053 \pm 0.001 \text{ Mg C ha}^{-1} \text{ yr}^{-1}$ , respectively (Table 1).

## Discussion

In spite of the differences in annual mean and seasonal pattern of precipitation, total SR in the hot and dry 2003 was not significantly different (t test,  $p < 0.05$ ) from SR in the milder and moister 2001 (Figs. 2 and 4). In fact, in 2003 the low SR rates during the summer drought were compensated by high SR rates during two periods of high precipitation: April to May and August to September, which accounted for almost two third of annual precipitation in 2003. Additionally, spring 2003 was warm and although precipitation was well below the average, there was no drought stress (shown in Konôpka et al. 2005, Table 1). During the drought in July and early August 2003, decomposition virtually stopped and high root mortality occurred, as demonstrated for the same stand by Konôpka et al. (2005). With the first rains, there was thus an unexpectedly higher than normal substrate availability, especially in the upper 30 cm, where root biomass is concentrated (Janssens et al. 1999; Xiao et al. 2003). If dead roots were decomposed and became a significant source of soil-respired CO<sub>2</sub> in September, the <sup>14</sup>C signature of soil-respired CO<sub>2</sub> should have a value that is equal or higher than the atmospheric value (Gaudinski et al. 2000; Vargas et al. 2009). However, the signature of soil-respired CO<sub>2</sub> is always quite far from the atmospheric signature. A convincing explanation for

**Table 2** Contribution of fast- and slow-cycling C to total soil respiration (SR) in the seasons of 2003, expressed as a percentage of total SR and absolute value ( $\text{Mg C ha}^{-1}$ ). Numbers on the right side are the uncertainties

	Fast-cycling C		Slow-cycling C	
	% of total SR	$\text{Mg C ha}^{-1} \text{ yr}^{-1}$	% of total SR	$\text{Mg C ha}^{-1} \text{ yr}^{-1}$
Spring	95±5	1.05±0.06	5±5	0.06±0.06
Early summer	98±7	0.63±0.03	2±7	0.01±0.03
Late summer	77±8	0.64±0.06	23±8	0.19±0.06
Autumn	83±10	1.51±0.18	17±10	0.30±0.18
Winter	69±4	0.43±0.03	31±4	0.18±0.03

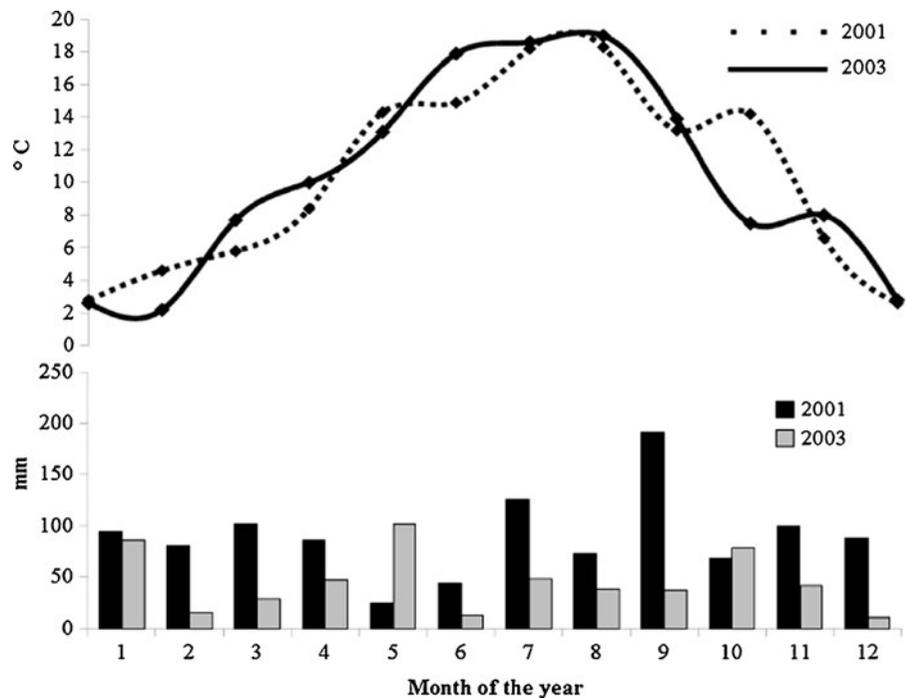
this discrepancy could be that also a minor contribution of the plaggen horizon to total SR has a great impact in term of  $^{14}\text{C}$  signature of the  $\text{CO}_2$  fluxes, because of the very old age of the SOC in plaggen (Chiti et al. 2009).

The  $^{14}\text{C}$  activity of the  $\text{CO}_2$  evolving from the soil was similar to the one of atmosphere in spring and early summer, while in the other seasons it was depleted in  $^{14}\text{C}$ . Such a depletion could be the result of a decreased influence of root respiration and/or a reduced decomposition of recently synthesized C substrates and/or an increased contribution of microbial decomposition of older SOC. Actually, a lower  $^{14}\text{C}$  concentration in soil  $\text{CO}_2$  between summer and winter suggests a reduced contribution of root respiration due to physiological changes when trees

approach dormancy. The difference in  $\Delta^{14}\text{C}$  between soil and atmospheric  $\text{CO}_2$  is larger at the beginning of summer ( $50.3\pm 2.0\%$ ) and smaller at the end of the season ( $20.1\pm 5.3\%$ ), probably due to an higher input of fresh SOC in the first part of the summer period, as discussed above. Roots are very dynamic in this stand, likely because Scots pines continue to photosynthesize in the winter in such a mild climate, and thus C is allocated to the rhizosphere throughout the year. This fact would partly explain the marked contribution of the fast cycling C to total SR all year long, with a minimum (69% of SR) in winter, when tree activity is actually reduced.

According to the method of Harrison et al. (2000), the soil C flux attributable to SOC decomposition in 2003 was  $1.8\pm 0.1 \text{ Mg ha}^{-1} \text{ yr}^{-1}$ . As a

**Fig. 4** Monthly mean temperature (above) and monthly mean precipitation (below) during the years 2001 and 2003 for the Brasschaat site (Carboeurope IP database 2011)



consequence, the annual heterotrophic respiration represented  $36 \pm 2\%$  of total soil respiration, a value that matches almost perfectly the 32% contribution to total SR estimated by the SECRETS model for the years 1997–1998 (Sampson et al. 2001). On the contrary, the comparison with other estimates indicates a possible underestimation in the Rh contribution we found. For example, the heterotrophic respiration estimate for 1997 done by Janssens et al. (1998) for the same stand using a mass balance approach amounts to 47% of total SR, while a more recent one obtained by subtracting autotrophic respiration and slash decomposition from total ecosystem respiration provides an Rh of 3.9–4.3 Mg C ha<sup>-1</sup> yr<sup>-1</sup> (Nagy et al. 2006), much higher than the 1.8 Mg C ha<sup>-1</sup> yr<sup>-1</sup> found in this study. Furthermore, although the relative contribution of Rh to SR at this site is well within the range reported in the review of Subke et al. (2006), annual Rh as absolute value is smaller than the Rh typically observed in temperate coniferous forests. Harrison et al. (2000) reported a range in Rh from 1.4 to 5.1 Mg C ha<sup>-1</sup> yr<sup>-1</sup> for European coniferous forests, while Luyssaert et al. (2007) reported an average Rh of 4.2 Mg C ha<sup>-1</sup> yr<sup>-1</sup> for the same forests. The very low Rh and total soil respiration at our study site are probably related to the low site fertility (Xiao et al. 2003; Curiel Yuste et al. 2005a) and the unfavourable conditions for microbial activity imposed by the extreme soil acidity: the pH in CaCl<sub>2</sub> is in fact 2.8 in the litter layer and 3.6 in the mineral soil (Curiel Yuste et al. 2005a). Despite it could be expected a higher contribution of the mineral soil compared to the 4% we found, our results are in agreement with the estimates of Rh, based on the <sup>14</sup>C content of SOC, performed by Harrison et al. (2000) in several coniferous forests throughout Europe, who report fluxes from the litter ranging from 0.66 Mg ha<sup>-1</sup> yr<sup>-1</sup> to 2.39 Mg ha<sup>-1</sup> yr<sup>-1</sup>, and a contribution of the mineral soil to the total Rh ranging from 2% to 60%. The few other papers that distinguish the Rh contribution of the litter vs. the one from mineral soil, show contrasting results. For example, Sulzman et al. (2005) found a contribution to total SR of 19% for litter and 58% for mineral soil in an old growth coniferous forest, while Nadelhoffer and Raich (1992) report contributions up to 70–80% of total SR for litter plus root components in some forest ecosystems. It is very likely that the used <sup>14</sup>C based approach tends to underestimate the

fluxes from mineral horizons. In fact, these are mixtures of SOC with a range of turnover times from a few years to several centuries (Gaudinski et al. 2000; Trumbore 2000; Koarashi et al. 2009). Looking at the <sup>14</sup>C signature of the humic acid and the unextractable SOC from the A1 mineral horizon (0–8 cm) of the same stand obtained by Chiti et al. (2009), both fractions in the 0–4 cm have a positive  $\Delta^{14}\text{C}$  of 125.6‰ and 3.5‰, respectively, while in the 4–8 cm depth both of them have a highly negative  $\Delta^{14}\text{C}$ , -90.2‰ and -137.0‰, respectively. If we look at the whole A1 horizon (0–8 cm), the presence of bomb C in the 0–4 cm is completely diluted by the ancient C in the 4–8 cm, resulting in a final  $\Delta^{14}\text{C}$  for the bulk horizon of -142.7‰. Such a superimposition actually leads to an underestimation of the C flux from the A1 horizon. This is the reason why it is important to apply the radiocarbon based method on single fractions rather than on bulk soil in the purpose of reducing heterogeneity and providing a more precise value of Rh flux. However, the small contribution of the mineral soil to total SR could also be an indication of substantial loss of SOC as dissolved organic carbon (DOC), as observed by Gielen et al. (2009).

One criticism to our approach may arise from the consideration that the turnover time of SOC and, consequently, the heterotrophic C flux from soil are hard to estimate based on radiocarbon analysis of the bulk soil alone, as demonstrated by Torn et al. (2009). However, we verified that the <sup>14</sup>C-based approach produces realistic results if caution is exercised. Additionally, also alternative methods have their own drawbacks and assumptions, as well reviewed by Hanson et al. (2000).

In conclusion, at this site, radiocarbon measurements showed the potential to effectively estimate the contribution of different C sources to total soil respiration.

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