Radiocarbon, Vol 59, Nr 3, 2017, p 1021–1034 DOI:10.1017/RDC.2016.81
Selected Papers from the 2015 Radiocarbon Conference, Dakar, Senegal, 16–20 November 2015

SUBSOIL ^{14}C DYNAMICS IN DIFFERENT TYPES OF TROPICAL AND SUBTROPICAL SOILS UNDER DIFFERENT CROP MANAGEMENT

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ABSTRACT. Radiocarbon has been applied as a tracer to study carbon dynamics in different types of tropical soils, under paddy and non-paddy management on Java, Indonesia. The ^{14}C concentrations were measured in samples of total organic carbon as well as in alkali-soluble humic acids, insoluble humin, and plant remains (roots, seeds, leaves) obtained from three sites with Andosols, Alisols, and Vertisols at different altitudes. In addition, the abundance and distribution of plant macrofossils in the soil column and organic δ^{13}C values were determined. The results obtained so far are compared with those from a chronosequence of Cambisols in China. They indicate the input of fresh plant materials into the subsoil directly via roots and/or by soil cracks and bioturbation of aboveground litter. The total organic C and ^{14}C concentrations show the usual decrease with increasing depth in paddy and non-paddy soils, reflecting the influence of direct input as well as the downward redistribution of organic material as particulates or dissolved organic carbon (DOC) depending on soil type (pedogenesis, plough-pan formation, drying cracks), crop type and management, and climatic factors. A disturbance of the Andosol around 0.4 m depth and a change in profile properties around 0.63 m in the Alisol limit the general conclusions.

KEYWORDS: radiocarbon, paddy soils, carbon dynamics, Java.

INTRODUCTION

The signal of bomb radiocarbon in the deep subsoil (Baisden and Parfitt 2007) may imply that deep carbon (C) is not completely passive. Young and fresh organic matter, introduced from the top through the direct input of fresh plant remains (Bräuer et al. 2013a), stabilized-dissolved organic matter (Kalbitz et al. 2005; Hanke et al. 2014), and shoot roots and root exudates (Kuzyakov and Domanski 2000), may add fresh photosynthate to older subsoil organic matter and result in younger average ^{14}C concentrations. This supports other findings indicating that fresh organic C inputs tend to reactivate, microbially stable organic C in the deep soil layers (Fontaine et al. 2004, 2007). Given the indication of active organic C in the subsoil, the storage of atmospheric CO_2 as soil carbon, a key component of the fast reacting part of the global carbon cycle deserves attention. C inputs and storage in soils influence the global C dynamics and respond in different ways under different soil properties (i.e. mineralogy), climate, and crop management (Harrison et al. 1993; Trumbore et al. 1996; Torn et al. 1997). The influence of these factors on subsoil carbon turnover needs to be quantified.

Tropical soils store substantial amounts of organic C (Chiti et al. 2010), while worldwide in top soils an estimated 1500–2000 PgC (Pg: petagram = 10^{15} g) are present as organic matter (Houghton 2007), which is 2–3 times the amount in the atmosphere. Various types of soils exist under different climatic regimes and soil-related factors that constrain C dynamics are, partly, connected to organic carbon (OC) stabilization by clay minerals (Mikutta et al. 2005, 2006; Eusterhues et al. 2007; Lorenz et al. 2009) or the soil structure (Chevallier et al. 2010). Among climate aspects, moisture and temperature are two keys factors to control soil properties (Birkeland 1999). These factors, however, not only affect the decomposition rate of organic matter under a broad range of environmental conditions, i.e. temperate soils (Leirós et al. 1999), temperate monsoon soils (Dan et al. 2016), tropical rainforest (Knorr et al. 2005), Arctic soils

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(Moni et al. 2015), and permafrost region soils (Bracho et al. 2016), but also vegetation types (Adams 2010) and chemical composition of tissues (Cotrufo et al. 2009). Biomass inputs as derived from plants/vegetation are the main source of soil OC (Lal 1997), and their dynamics, in the top and subsoil layers, are influenced by crop rotations (Witt et al. 2000); environmental conditions, i.e. parent material and climate (Solly et al. 2015); type of crops; and crop management methods (Nieder and Benbi 2008).

The irrigated rice system is one of most important food production systems (Dobermann and Witt 2000). The submerged condition during the rice season creates specific characteristics related to C dynamics, such as redox cycling, dynamics of dissolved organic C, and association with mobilized Fe and clay minerals (Kögel-Knabner et al. 2010).

As part of Research Unit 995, "Biogeochemistry of Paddy Soil Evolution," we research paddy soils in both subtropical (phase 1) and tropical environments (phase 2). In phase 1, work on a soil chronosequence, derived from marine sediments in a subtropical area, demonstrated slower refreshment of old OC with fresh, plant-derived C in rice paddies than in non-paddies ("upland soils"), and a fast increase of OC in the top soil, already after 50 yr of cultivation (Bräuer et al. 2013a). Now, in the second phase the studies focus on three different soil types in a tropical climate.

Following the simple approach of previous studies (Rethemeyer et al. 2004; Bräuer et al. 2013a,b) and radiocarbon dating, we selected the insoluble and thus supposedly stationary humin fraction and the alkali-soluble, mobile humic acid fraction to highlight the age heterogeneity of the soil total organic carbon (TOC), commonly analyzed in soil science. ^{14}C analyses are still ongoing. The insolubility of the humin fraction is in part attributed to strong interaction with clay and Fe minerals and may thus be soil type dependent and of special interest. These bound humins are a complex mixture, characterized by labile compounds—i.e. carbohydrates, glucose, cellulose, peptides, peptidoglycan—and more stable compounds—i.e. aliphatic and lignin contents (Fabbri et al. 1996; Simpson et al. 2007). We also observed plant remains as a source of C input to study subsoil C dynamics. We compare rice paddies with neighboring non-paddy fields to investigate the effect of paddy management on soil carbon dynamics. We use the ^{14}C concentration of different soil carbon fractions to determine the influence of recently photosynthesized OC in these fractions.

MATERIAL AND METHODS

Study Sites and Sampling

Research was carried out on three different types of tropical soils of Indonesia, i.e. Andosol, Alisol, and Vertisol (FAO 2006). Andosols at Sukabumi-Perbawati, West Java, were derived from a colluvial fan deposit (lahar) of Mount Gede-Pangrango at an altitude from 850 to 1000 m above sea leavel (asl) on terraced upper slopes (see Figure 1). The mean annual temperature in 2001–2010 ranged from 20.4 to 21.8°C, with annual rainfall in the range of 2200–6200 mm/yr. Alisols at Bogor-Jasinga, West Java, developed from andesitic volcanic materials at an altitude of 240 m asl on terraced midslopes with a mean annual temperature range of 25.5 to 26°C and annual precipitation of 1900 to 4700 mm/yr. Vertisols at Ngawi-Ploso Lor, East Java, formed from basaltic materials and marine sediments, with ~70 m asl elevation on a stagnant alluvial plain. In Ngawi, mean annual temperature during the period 1973–1992 is estimated at 26.6°C (Amien et al. 1996), and the annual precipitation data from 2006–2014 show values from 980 to 4900 mm/yr. More details about soil characteristics have been discussed by Winkler et al. (2016).

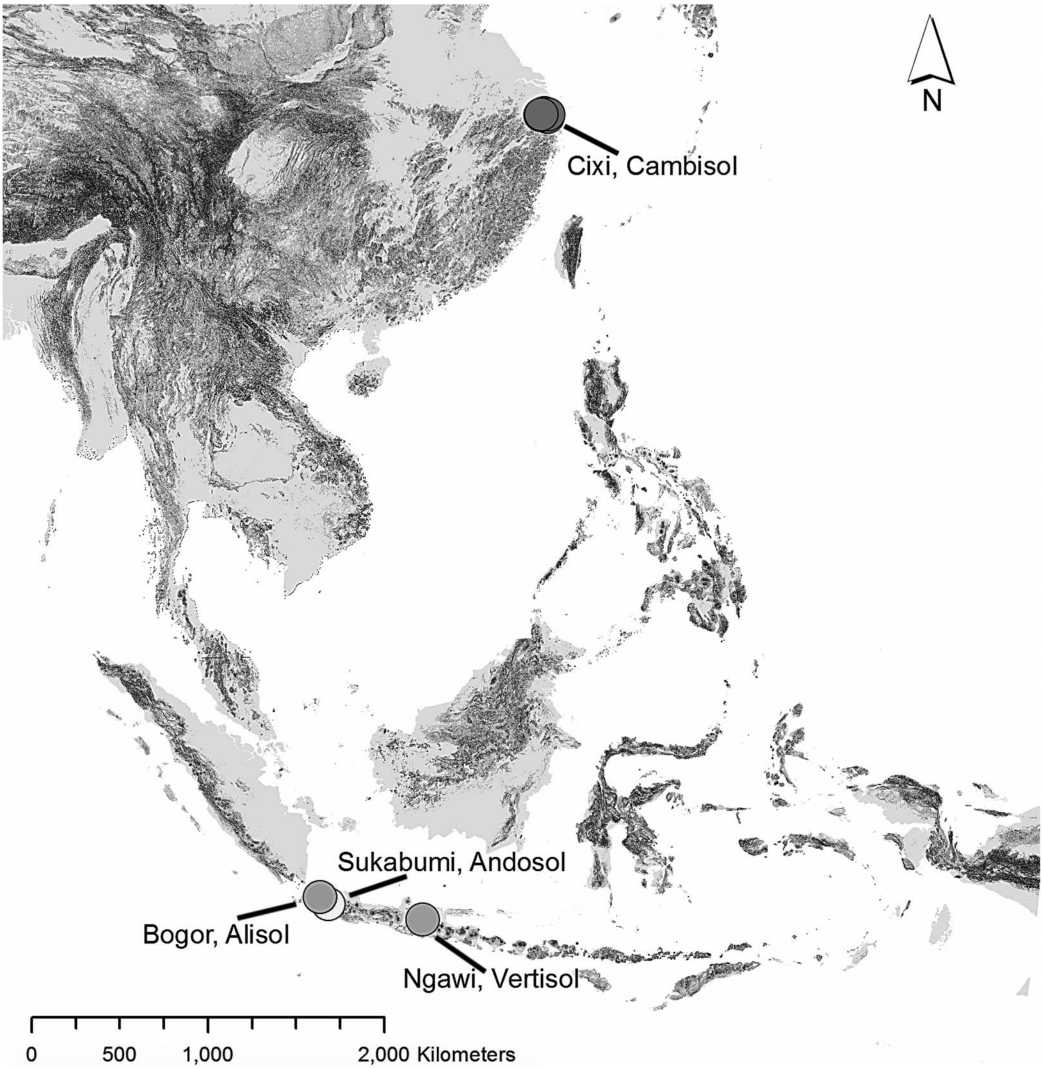

Figure 1 Map of research area in Java, Indonesia (bottom) and Cixi, China (top) (source of data: DIVA-GIS)

Soils were investigated from two neighboring sites under paddy and non-paddy cultivation. The paddies have a rotation of rice (C_3, Calvin cycle) with during the dry season mostly C_3 plants [e.g. cassava, talas, chili, banana, Chinese cabbage (bok-choi), and tobacco] except in Bogor-Jasinga where maize (C_4, Hatch-Slack cycle) was planted. The non-paddy crops in Ngawi-Ploso Lor and Sukabumi-Perbawati were sugarcane and maize (both C_4 plants), while in Bogor-Jasinga cassava, chili, and talas were planted (C_3). Photosynthesis favors CO_2 molecules containing the light isotope ^{12}C, which leads in C_3 plants to a relative deviation of δ^{13}C from the composition of the international standard PDB of –22 to –30‰. The more efficient CO_2 uptake of the C_4 cycle leads to reduced fractionation and δ^{13}C values of about –9 to –19‰ (Balesdent et al. 1987). Different crops will lead to the input of plant material with different δ^{13}C signatures into the soil and the TOC. Profiles were observed from all sites down to ~1 m depth, in paddies slightly below the groundwater level. Samples were taken from each layer of each profile over depth intervals of 1 to 5 cm as samples of ~5 kg to collect plant remains and as separate point samples over 1–2 cm depth, ~100–500 g, to analyze C and ^{14}C concentration.

Plant remains (roots, seeds, leaves, stems) were collected on a 2-mm and a 0.37-mm sieve by wet-sieving. Root material was separated by color, light versus dark, to test whether color correlated with ^{14}C content ("young versus old"). Some of the collected plant remains are depicted in Figure 4. A standard acid-alkali-acid treatment (Grootes et al. 2004) was used to remove carbonates and potentially mobile, contaminating fulvic and humic acids from the acid/alkali-insoluble remains. Humic acids were precipitated and collected separately by acidifying the alkali extract. Visible plant remains were first picked by hand and removed from the point samples of soil, which were the passed through a $\phi < 2$-mm sieve and oven-dried at 60°C. The acid-alkali-acid extraction procedure was used to obtain a humic acid and a humin fraction of soil organic carbon. Prepared samples were combusted to CO_2 at 900°C in quartz ampoules with CuO and silver wool, graphitized, and analyzed for ^{13}C and ^{14}C with accelerator mass spectrometry (AMS) in Trondheim (Nadeau et al. 2015). ^{14}C concentrations are expressed in percent modern carbon (pMC) according to Stuiver and Polach (1977). The total organic carbon (TOC) was determined via elemental analyzer (EuroEA 3000). A pretreatment with concentrated HCl without washing was used to remove carbonates in particular for the Vertisol samples with inorganic carbon (IC) values ranging from 0.1 to 0.6%. IC was assumed to be negligible in the Andosols and Alisols as implied by the $pH_{(H2O)}$ values 4 to 6, so total carbon (TC) would be equal to OC content. The grain-size distribution of particles in the point samples from the different paddy and non-paddy soil profiles was quantified via a Mastersizer (Hydro 2000G), using a standard protocol with HCl and H_2O_2 extraction in a water bath at 60°C to remove carbonates and soil organic matter and soil dispersion by shaking with $Na_4P_2O_7$ overnight.

RESULTS AND DISCUSSION

Here we present the distribution of grain sizes, organic carbon, and plant residues in the different paddy and non-paddy profiles and discuss the significance of the ^{14}C concentrations measured in different OC fractions in Java in comparison with earlier results obtained in China.

Grain-Size Distribution

The grain-size distributions of the different paddy and non-paddy soil profiles are shown in Figure 2. Fine-sized particles dominate in the subsoil in all profiles and the sand fractions amount to only ~10%, except in the Andosol. In the Alisol and Vertisol, clay makes up 40 to 50% of the paddy and 25 to 40% of the non-paddy subsoil. In the Cambisol, clay contributes ~30% and 25%, respectively, while medium silt dominates with 40 to 45%. The grain-size distribution is relatively constant with depth for the Alisol, the Cambisol, and the non-paddy Andosol and Vertisol. The strong scatter of the paddy Vertisol data may be the result of the alternating wetting and drying of rice cultivation and the accompanying, smectite[1]-dominated swelling and shrinking characteristics of the Vertisol, which result in vertical soil mixing through drying cracks.

The grain-size distribution of the paddy Andosol top section, down to ~0.4 m depth, is relatively coarse with 25 to 30% sand like the non-paddy. It grades to finer material, more similar to the other paddy soils, below that depth, implying a possible disturbance of the layers above. The Sukabumi-Perbawati Andosol fields are on colluvial fan deposits on the upper slopes of Mount Gede-Pangrango with the paddy field downslope, ~1.5 km

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[1] Name of 2:1 phyllosilicate group mineral species, which are favored in dry climates, impeded drainage, and basic parent material, a typical environment in which Vertisol is formed.

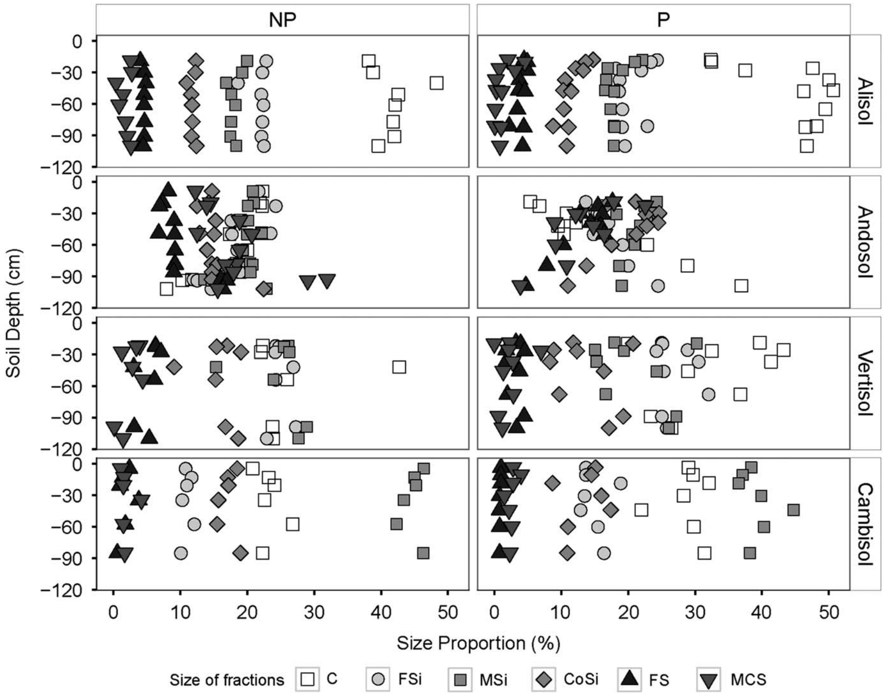

Figure 2 Distribution of size fractions throughout the three profiles of paddy (P) and non-paddy (NP) soils from Java and the 50-yr Chinese Cambisols (C = clay, FSi = fine silt, MSi = medium silt, CoSi = coarse silt, FS = fine silt, MCS = total of medium and coarse fractions).

away and ~110 m lower. We suppose the soils may have generally been affected by frequent deposition and reworking of volcanic ashes, while the paddy fields have, additionally, been cut and shaped into flat surfaces. Grain-size distribution and ^{14}C results (Figure 6), as well as the slight difference in Ti/Zr ratio (Winkler et al. 2016), all indicate this disturbance.

Total Organic Carbon (TOC) of Soil and Organic Matter Fractions

TOC values, representing a local time average of the soil carbon cycle that depends on past farming management and soil development-related processes, are in general relatively constant in the subsoil, around 0.5%. The values increase towards the topsoil and are a bit higher in non-paddy subsoil. C/N values appear to be decreasing with increasing depth (Figure 3).

Layers immediately below the plough-pan in the paddy fields in Java, generally, contained less TOC than non-paddy fields at similar depth (Figure 3). Yet, in the top ~0.4 m of the Sukabumi-Perbawati Andosol, the reverse appears to be true. Field observations indicate that these anomalous TOC values may be influenced by recent human disturbance, taking different soil materials from elsewhere to the site, e.g. in field terracing, as we found a differently compacted layer below the plough-pan indicating a past-tillage effect. ^{14}C results that are close to and above "recent" atmospheric values in the upper 0.4 m, and "old" below this depth (Figure 6) are consistent with this.

In the Cixi (China) area, in a chronosequence of polders with originally estuarine sediment[2] under paddy management, the total organic C content in the top soil of the paddy was higher

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[2] Originates from mixing and depositing materials of the Yangtze River in Hangzhou Bay.

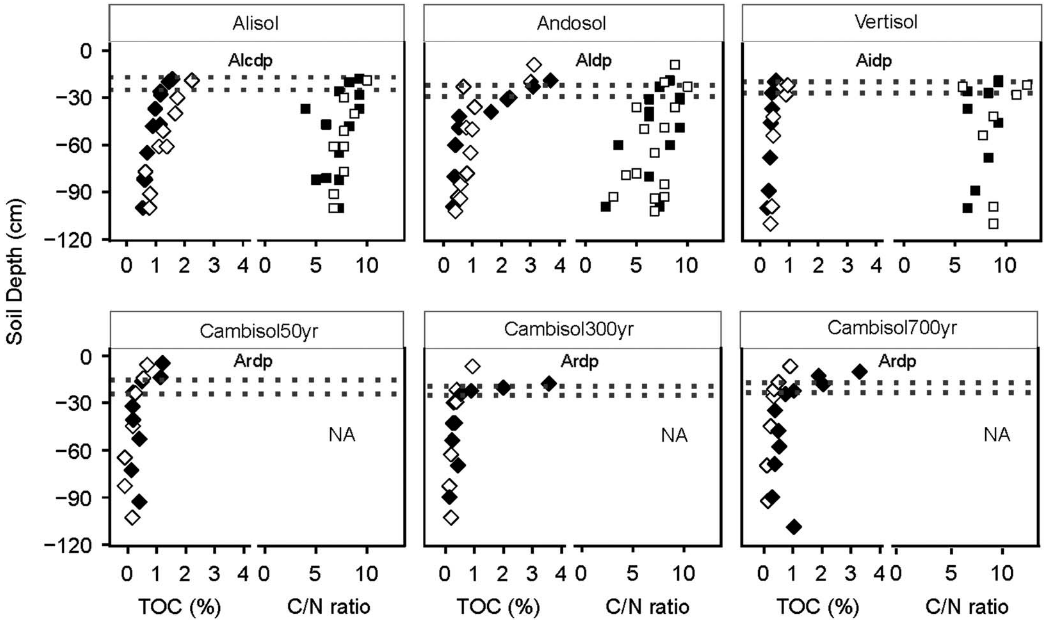

Figure 3 Total organic C in percent of dry soil weight and C/N ratio of soil in paddy (P) and non-paddy (NP) in Alisol, Andosol, and Vertisol from Java, Indonesia, and Cambisol 50, 300, and 700 yr from Cixi, China (Bräuer et al. 2013b). Filled and open symbols represent P and NP, respectively. Horizontal dotted lines show the compacted plough-pan. NA = data not available.

Table 1 Relative proportion of isolated SOC fractions to TOC from soils of Java under paddy (P) and non-paddy (NP) management.

			Fractions (% soil weight)		Proportion to TOC (%)	
Soil type	Depth (cm)	Land use	C-humic	C-humin	C-humic	C-humin
Alisol	18	P	0.18	1.16	12	73
Alisol	28	P	0.03	0.53	2	45
Alisol	30	NP	0.06	0.63	4	36
Alisol	61	NP	0.03	0.40	3	36
Alisol	91	NP	0.02	0.41	2	45
Andosol	9	NP	0.26	0.79	8	25
Andosol	50	NP	0.04	0.18	4	18
Andosol	65	NP	0.08	0.17	9	18

than in that of the non-paddy and increased by about a factor of 3 from 50 to 700 yr cultivation (Bräuer et al. 2013b). This has been shown to be associated with the increase in silt and clay-sized particles (Wissing et al. 2011).

The humin fraction accounts for most of the isolated soil OC (Table 1), presumably due to the high clay and fine silt mineral content of the soils (Figure 2). As OC is strongly bound by clay, the higher C-humin in the Alisol than in the Andosol soils is obviously due to the higher clay and fine particle content in the former (Figure 2).

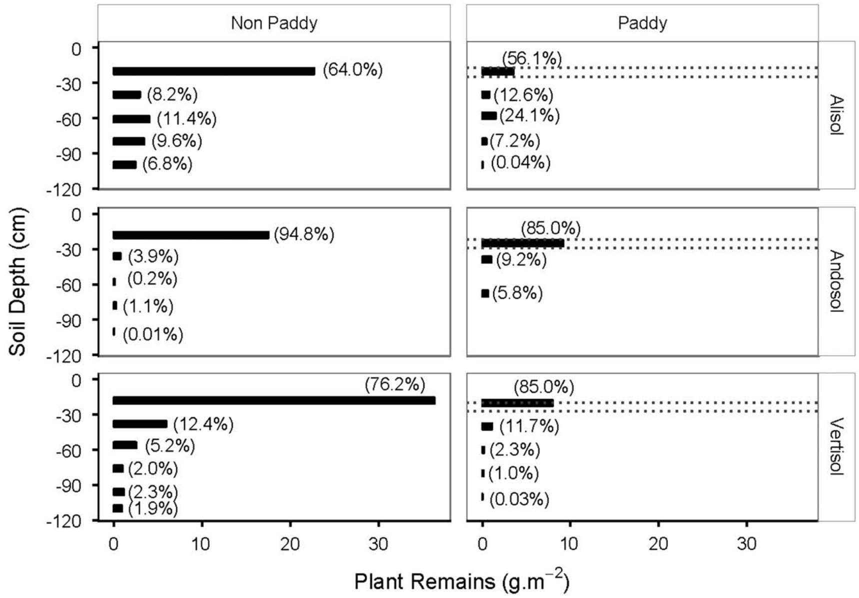

Figure 4 Abundance of collected plant remains in the Alisol, Andosol, and Vertisol. The numbers in parentheses represent the proportional values. The dotted horizontal line in the paddy figures represents the plough-pan.

Subsoil Macrofossils

Collected plant remains were more abundant in the non-paddy than in the paddy topsoil for all three soil types, which is similar to TOC trends. This continued in the Alisol and Vertisol subsoils, while for the Andosol subsoil the non-paddy contained fewer plant remains than the paddy (Figure 4). Below 0.30 m depth, the size of plant remains decreased, except in the Ngawi-Ploso Lor Vertisol, where shrinking and swelling (a characteristic of smectites under dry and wet conditions) allow deep input of aboveground material through large cracks. The higher abundance of plant remains in the non-paddy soils may be described as a consequence of multiple, inseparable factors. Differences in annual crop rotations and in crop varieties affect the input of plant material, both aboveground and as (deep) roots. For paddy fields, with one seasonal dry-land non-rice crop, rotations applied regularly are rice-rice-bok choi (Chinese cabbage) in Andosol, rice-rice-maize in Alisol, and rice-rice-tobacco in Vertisol. In the non-paddy, we have maize (C_4) in Sukabumi-Perbawati Andosol; cassava, chili, and talas (C_3) in Bogor-Jasinga Alisol; and sugarcane (C_4) in the Ngawi-Ploso Lor Vertisol (Dieterich 2012). In addition, hydrological management and soil factors such as accessibility of plough-pan and soil/clay mineral properties influence the distribution size of subsoil plant remains.

In the Alisol, an increase in small plant remains at around 0.63 m depth (Figure 4), for both paddy and non-paddy fields, coincides with a change in TOC from 0.8% below to ~1.3% above (Figure 3). This similarity in the pattern of distribution with depth of organic remains and TOC (Figure 3) suggests a close link between the two on a local scale (same soil type and climate).

Plant remains found in the subsoil were generally dark in color. Yet, we also found some light-colored, fresh-looking leaves and roots at depth. We measured the ^{14}C concentration of

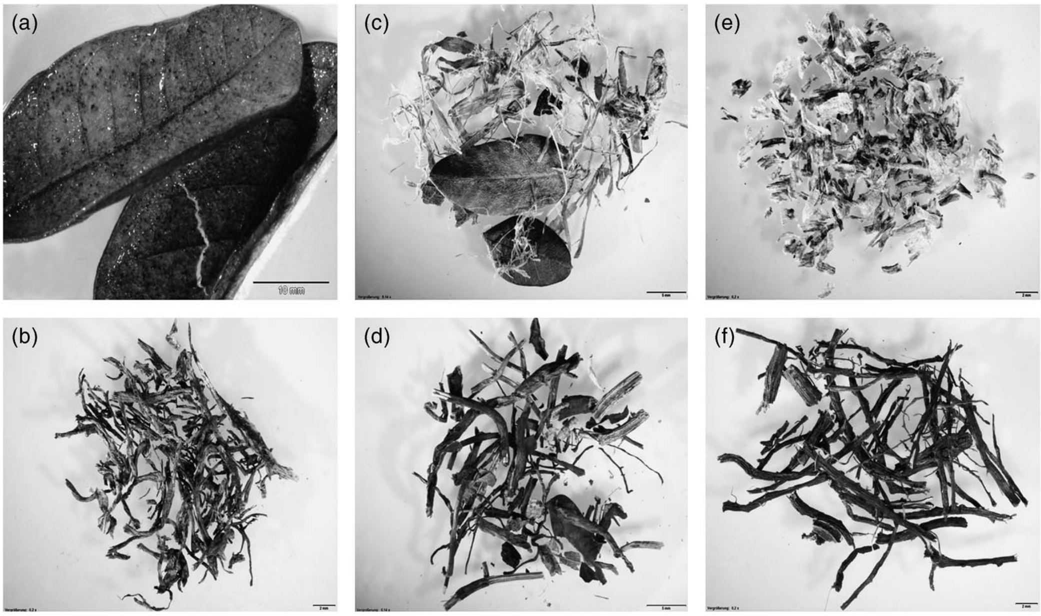

Figure 5 Some collected subsoil plant remains for ^{14}C dating: (a) and (b) dark roots and leaves from the Andosol profile; (c) and (d) leaves and dark-light roots from the Alisol profile; (e) and (f) light leaf fragments and dark roots from the Vertisol profile.

Table 2 Organic carbon content and isotopic composition of collected macrofossils at different depths.

Soil type	Crop type	Plant remains	Depth (cm)	TOC (%)	^{14}C (pMC)	δ^{13}C (‰)
Andosol	Maize	Dark roots	78–83	44	105.8 ± 0.2	–27.54 ± 0.46
Andosol	Paddy	Light leaf	98–102	46	103.8 ± 0.1	–28.35 ± 0.66
Alisol	Paddy	Light roots	100	53	105.3 ± 0.2	–32.35 ± 0.28
Alisol	Paddy	Light leaf fragments	100	64	60.2 ± 0.1	–28.94 ± 0.43
Alisol	Cassava	Dark roots	100–104	46	105.0 ± 0.2	–29.72 ± 1.70
Alisol	Cassava	Light roots	100–104	47	105.1 ± 0.1	–26.62 ± 1.59
Alisol	Cassava	Charcoal	80–84	64	73.7 ± 0.2	–28.15 ± 0.56
Vertisol	Paddy	Light leaf fragments	80–85	44	104.4 ± 0.1	–23.61 ± 1.84
Vertisol	Sugarcane	Dark roots	110–115	47	106.2 ± 0.2	–15.85 ± 1.73
Vertisol	Sugarcane	Dark roots	110–115	48	104.8 ± 0.2	–11.12 ± 1.40

several dark and light leaf and root fragments from the bottom of the profiles to check for a correlation between color, depth, and ^{14}C concentration. In the Alisol paddy at Bogor-Jasinga (Figures 5c, d; Table 2), we found light-colored roots with a ^{14}C content of ~105 pMC at 1 m depth but also light-colored leaves, containing 60.2 pMC, at the same depth. In a neighboring non-paddy field, we observed both dark and light-colored roots having a ^{14}C content of ~105 pMC at 1.00–1.04 m depth and charcoal of 73.7 pMC at 0.80–0.84 m depth. In the Sukabumi-Perbawati Andosol, we collected dark roots with ~105.80 pMC ^{14}C at 0.78–0.83 m depth in the non-paddy, and a "fresh" leaf with ~103.80 pMC at 0.98–1.02 m depth of the paddy field. Furthermore, dark roots with ^{14}C contents of ~104.8 and 106.2 pMC, observed at 1.10–1.15 m depth in the non-paddy field of the Ngawi-Ploso Lor Vertisol, are identified as sugarcane by their heavier C_4 $\delta^{13}C$ signal of ~–11.12‰ and –15.85‰, respectively. In the Ngawi-Ploso Lor paddy field, a light-colored leaf fragment, found at ~0.80–0.85 m depth, has a ^{14}C content of ~104.4 pMC and, most likely, a C_3 origin ($\delta^{13}C$ content ~–23.61‰). These results indicate that not only young root material may be found in the deeper subsoil, but that also young plant litter may be transported into the subsoil by soil mixing processes.

It is also clear that the color differences of organic remains (Figure 4) do not necessarily indicate ^{14}C age differences. Plant remains with a ^{14}C concentration close to that of atmospheric CO_2 below 0.70 m depth (Table 2) are an indication of direct input of macrofossils from top layers. The present results confirm the previous findings of Bräuer et al. (2013b) who obtained plant material and roots with a ^{14}C concentration higher than 105 pMC at ~1 m depth with one sample identified as rice roots by DNA analysis.

^{14}C Distribution as a Function of Depth

The changes in the amount of organic carbon and its ^{14}C concentration are determined by a complex interplay of several factors, such as crop type, crop management, and soil composition—including soil history, hydrology, and climate—that together determine input, remineralization, export, and stabilization of organic carbon in the soil. The three sites in Java provide data on three different volcanic soils in a tropical climate at different elevations to be compared with the earlier results from a fluviomarine soil at sea level in subtropical China (Bräuer et al. 2013a).

The ^{14}C concentrations of the humic acid and the humin fractions show, as expected, a gradual decrease with increasing depth (Figure 6). The TOC ^{14}C concentrations agree with these in general, but cannot be compared in detail, since they were obtained on samples averaged over a full soil layer (vertical bar) instead of on a 1–2 cm point sample. The humins, which make up ~20 to 50% of the TOC (Table 1), contain less ^{14}C than the mobile humic acids in the Alisol and in the Chinese Cambisol, a pattern expected both from downward transport as DOM and from microbial recycling. Most of the humin OC is strongly associated with clay minerals, which form the largest of the particle size fractions (Figure 2; Morra et al. 1991; Amelung et al. 1998; Schmidt et al. 1999). It may thus also be suggested that the lower ^{14}C concentration in the humin fraction is due to the dominance of older, mineral-stabilized OC. Lower ^{14}C concentrations in the mineral-associated OC than in TOC (“bulk soil”) have also been observed by Eusterhues et al. (2007). The deep Alisol samples show ^{14}C concentrations around 60 pMC, similar to those of the deep Andosol.

In the Sukabumi-Perbawati Andosol, the increase in clay- and silt-sized particles and Fe-mineral crystallinities below 0.4 m depth (Winkler et al. 2016) coincides with a break in the TOC (Figure 3, Andosol) and ^{14}C profile (Figure 6b). Above ~0.35 m depth, the ^{14}C results obtained so far give ^{14}C concentrations of humic acids above 100, up to ~111 pMC, corresponding to that of the atmosphere since ~1996 (Hua et al. 2013), while below 0.4 m it is less than 70 pMC. The high ^{14}C concentrations indicate carbon from the period with bomb ^{14}C since 1954, most likely preserved by the stabilization of SOM in organomineral and Al/Fe complexes (Takahashi and Dahlgren 2016). This is supported by the abundance of short-range-ordered Fe oxides and allophane and imogolite[3]-type phases in this soil (Winkler et al. 2016). Additional analyses are in progress to confirm this. This large drop in ^{14}C concentration probably reflects a change from a disturbed, terraced top soil with up to ~3% young TOC, to an

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[3] Amorphous to poorly crystalline hydrous aluminium silicate clay mineral; it exists in volcanic ash-derived soils.

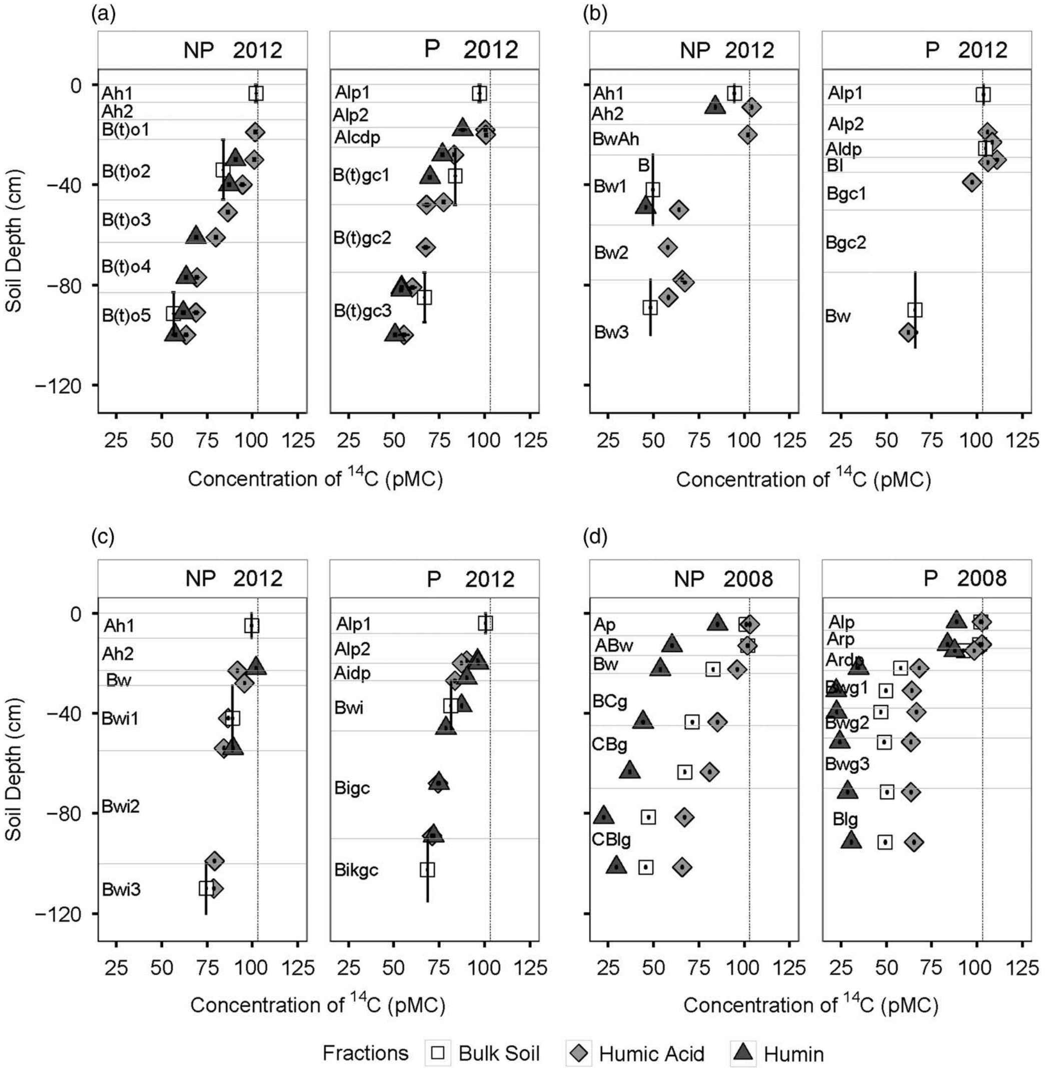

Figure 6 ^{14}C distribution profiles (in pMC) in different types of soils (a) Alisol, (b) Andosol, (c) Vertisol, and (d) Cambisol in paddy (P) and non-paddy (NP). Short vertical bars represent the layer thickness sampled, while dotted vertical lines indicate 2008 and 2012 atmospheric $^{14}CO_2$ concentrations, which were 105 and 103 pMC, respectively (Levin et al. 2010, 2013). Java belongs to the Southern Hemisphere atmospheric ^{14}C zone 3 (SH-3), and Cixi, China, is in NH zone 3 (Hua et al. 2013).

old subsoil with less than 1% TOC (Figure 3). The 111 pMC ^{14}C concentration found for humic acids in a differently compacted layer at 0.3 m depth below the plough-pan indicates a significant presence of organic material from the 1990s in this layer. This must have been the time at which the profile was disturbed.

In Ngawi-Ploso Lor (Figure 6, Vertisol), the ^{14}C concentration of the humic acids is lower than that of the humins in the upper 0.46 m, but at greater depth they are similar. These results are in contrast with earlier findings (Rethemeyer et al. 2004; Bräuer et al. 2013b) and with the results for the Java Alisols and Andosols. This could indicate a strong signal of the presence of young, acid-alkali-insoluble fine organic remains (e.g. cellulose) from aboveground litter or roots in the upper layers, for which the presence in the humin fraction has been reported (Fabbri et al. 1996; Simpson et al. 2007). Smectite-dominated “swelling and shrinking” of the Vertisol generates cracks during the dry seasons through which material can fall down into the subsoil (Birkeland 1999), and thus causes pedoturbation. Roots are dominantly placed in fine shrinkage cracks and between slickensides[4] (Nicoullaud et al. 1994), so they too may penetrate more easily to deeper layers. Together with the dense root system of the mature sugarcane at time of sampling and its decrease with increasing soil depth, this could explain the relatively large amount of young plant fragments found in the upper part of the sugarcane and rice field (Figure 4). The observed younger humin ^{14}C signal may then result from the presence of fine root and plant material incompletely removed by sieving and acid-alkali-acid treatment. The higher ^{14}C concentrations around 75 pMC, observed at depth (Figure 6c), are consistent with a larger input of young OC into the subsoil.

The TOC (Figure 3) and ^{14}C-humic (Figure 6) values in the top layers of the non-paddy Vertisol are lower than in the Alisol despite the high abundance of collected plant remains (Figure 4). ^{14}C concentrations deeper in the subsoil are, however, above those in the Alisol. This low ^{14}C concentration-depth gradient may be explained by pedoturbational mixing due to the shrinking/swelling of the Vertisol under alternating dry/wet conditions. The inversion in humic acid ^{14}C concentrations between 0.20 and 0.28 m depth and a relatively higher scatter of ^{14}C concentrations in the Ngawi-Ploso Lor subsoil are consistent with this explanation.

In the Chinese Cambisol, the difference in ^{14}C concentrations between the humin and the humic acid fractions is much larger due to much lower humin ^{14}C values. This may be attributable to the origin of the Cambisol, which was deposited as estuarine sediment with a significant fraction of eroded old organic carbon, resulting in a TOC with an original ^{14}C concentration of only about 50 pMC. Microbial recycling with DOM priming in the subsoil quickly lowered this already low initial humin ^{14}C concentration from ~50 pMC to ~30 pMC (Bräuer et al. 2013a). The volcanic ash falls forming the parent material of the Java soils and thus lack such old OC. The Cambisol TOC shows ^{14}C concentrations clearly intermediate between the humin and the humic acid fractions.

In the 50-yr Cambisol (Bräuer et al. 2013a), the effect of the plough-pan, decoupling the paddy topsoil from the subsoil C profile, can be clearly seen. The presence of the compacted layer at ~0.20 m depth reduced the root penetration as well as the infiltration of young organic remains and DOC with irrigation water and created a steep OC ^{14}C gradient across this layer. In the 50-yr non-paddy field, the absence of the compacted layer made the subsoil more accessible for young organic carbon. Consequently, infiltrating young OC raised the original ^{14}C concentration by increasing the proportion of young to old OC without significantly changing the OC concentration, and created a gentle ^{14}C concentration gradient down to groundwater at ~0.80 m depth. The contribution of root material (cf. Table 2) and root exudates to the subsoil will be spatially heterogeneous. DOC may, in part, simply pass through the soil via preferential flow channels (Gandois et al. 2014), and partly be transformed and mineralized by microbial carbon cycling and stabilized on mineral surfaces (Hagedorn et al. 2015). After about 300 yr, the delaying effect of the plough-pan on profile development is no longer clear (Bräuer et al. 2013a). In Java, the paddy and non-paddy profiles are similar. This probably reflects the fact that the plough-pan in these profiles was much less distinct than in the Cixi Cambisols as well as the higher age of the profiles. The discontinuity in the Andosol paddy at ~0.4 m depth then would indicate a disequilibrium, like in the 50-yr Cambisol.

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[4] A surface of cracks produced in soils containing high proportion of swelling clays.

In the context of atmospheric variations of $^{14}CO_2$ between the Northern and Southern Hemisphere, particularly during pre-bomb testing (Hua et al. 2013), the difference of ^{14}C concentration in topsoils of China and Indonesia is not significant, implying a wide distribution of bomb-derived ^{14}C.

On Java, climate differences are mostly related to altitude. Thus, our three sites varied in mean annual temperature from ~21°C for the Andosol at 850 to 1000 m elevation to ~26.5°C for the Vertisol at 70 m asl with rainfall ranging from ~2 to 6 m/yr and 1 to 5 m/yr, respectively. In the Cixi area, at sea level the annual temperature is 16.3°C with a mean precipitation of 1325 mm per year (Cheng et al. 2009). A comparison of the TOC values in Figure 3 and the ^{14}C profiles of Figure 6 gives no clear indication of a correlation with these climatic differences.

CONCLUSIONS

^{14}C is a valuable tracer of the influence of recent photosynthate on organic carbon in the subsoil of three Java paddy/non-paddy sites of different soil types. The results suggest the following conclusions:

- Plant material with a ^{14}C concentration close to that of the atmosphere, found down to ~1 m, the deepest level sampled, proves roots and/or bioturbation/pedoturbation can introduce organic carbon directly into the deeper subsoil.
- The influence of recent photosynthate in the humin and humic acid fractions of soil organic carbon decreases regularly with increasing depth in the subsoil. This decrease is not linked to soil horizons, but occurs continuously across soil layers.
- The ^{14}C concentration in the mobile humic acid fraction is generally higher than in the humin, which is consistent with a downward transport of "young" organic carbon as DOM and with the ^{14}C concentration gradients observed.
- A crop-dependent large input of litter and root material may overwhelm the effect of young DOM input and result in higher humin ^{14}C concentrations in the upper layers of the subsoil.
- Pedoturbation, related to the shrinking and swelling properties of the Vertisol, opening and closing cracks in the soil, not only facilitates introduction of plant material into the subsoil, but also leads to mixing of OC in the subsoil and to a lower ^{14}C-depth gradient.
- The origin of the sediment and its inherited organic carbon influence the ^{14}C concentration of the humin fraction.
- The ^{14}C concentration of TOC or any OC fraction thus does not primarily reflect ^{14}C decay of a uniform carbon pool, but rather the mixing of carbon from different sources with different ^{14}C concentrations.

ACKNOWLEDGEMENTS

We thank the German Research Foundation (DFG) for the funding of Research Unit FOR 995 "Biogeochemistry of Paddy Soil Evolution," and our many colleagues of FOR 995 for their support, friendly collaboration, and scientific exchange. We thank the team of the National Dating Laboratory in Trondheim for high-quality isotopic measurements and Sophia and Manfred of the Institute for Ecosystem Research for grain-size and C/N analyses. The constructive and critical comments of two reviewers contributed to this paper.

REFERENCES

Adams J. 2010. *Vegetation-Climate Interaction. How Plants Make the Global Environment*. 2nd edition. Chichester: Wiley.

Amelung W, Zech W, Zhang X, Follett RF, Tiessen H, Knox E, Flach KW. 1998. Carbon, nitrogen, and sulfur pools in particle-size fractions as influenced by climate. *Soil Science Society of America Journal* 62(1):172–81.

Amien I, Rejekiningrum P, Pramudia A, Susanti E. 1996. Effects of interannual climate variability and climate change on rice yield in Java, Indonesia. In: Erda L, Bolhofer WC, Huq S, Lenhart S, Mukherjee SK, Smith JB, Wisniewski J, editors. *Climate Change Vulnerability and Adaptation in Asia and the Pacific*. Dordrecht: Kluwer Academic. p 29–39.

Baisden WT, Parfitt RL. 2007. Bomb ^{14}C enrichment indicates decadal C pool in deep soil? *Biogeochemistry* 85(1):59–68.

Balesdent J, Mariotti A, Guillet B. 1987. Natural ^{13}C abundance as a tracer for studies of soil organic matter dynamics. *Soil Biology and Biochemistry* 19(1):25–30.

Birkeland PW. 1999. *Soils and Geomorphology*. New York: Oxford University Press.

Bracho R, Natali S, Pegoraro E, Crummer KG, Schädel C, Celis G, Hale L, Wu L, Yin H, Tiedje JM, Konstantinidis KT, Luo Y, Zhou J, Schuur EAG. 2016. Temperature sensitivity of organic matter decomposition of permafrost-region soils during laboratory incubations. *Soil Biology and Biochemistry* 97:1–14.

Bräuer T, Grootes PM, Nadeau M-J. 2013a. Origin of subsoil carbon in a Chinese paddy soil chronosequence. *Radiocarbon* 55(2):1058–70.

Bräuer T, Grootes PM, Nadeau M-J, Andersen N. 2013b. Downward carbon transport in a 2000-year rice paddy soil chronosequence traced by radiocarbon measurements. *Nuclear Instruments and Methods in Physics Research B* 294:584–7.

Cheng YQ, Yang LZ, Cao ZH, Ci E, Yin S. 2009. Chronosequential changes of selected pedogenic properties in paddy soils as compared with non-paddy soils. *Geoderma* 151:31–41.

Chevallier T, Woignier T, Toucet J, Blanchart E. 2010. Organic carbon stabilization in the fractal pore structure of Andosols. *Geoderma* 159(1–2): 182–8.

Chiti T, Certini G, Grieco E. 2010. The role of soil in storing carbon in tropical rainforests: the case of Ankasa Park, Ghana. *Plant and Soil* 331(1): 453–61.

Cotrufo MF, Gado ID, Piermatteo D. 2009. In: Kutsch WL, Bahn M, Heinemeyer A, editors. *Soil Carbon Dynamics. An Integrated Methodology*. New York: Cambridge University Press. p 76–90.

Dan W, Nianpeng H, Qing W, Yuliang L, Qiufeng W, Zhiwei X, Jianxing Z. 2016. Effects of temperature and moisture on soil organic matter decomposition along elevation gradients on the Changbai Mountains, Northeast China. *Pedosphere* 26(3):399–407.

Dieterich U. 2012. Description of 18 profiles of Andosols, Ferralsols, and Vertisols under paddy and non-paddy use in Java (Indonesia) [Bachelor's thesis]. Munich: Technische Universität München.

Dobermann A, Witt C. 2000. In: Kirk GJD, Olk DC, editors. *Carbon and Nitrogen Dynamics in Flooded Soils*. Los Banos: IRRI. p 1–27.

Eusterhues K, Rumpel C, Kögel-Knabner I. 2007. Composition and radiocarbon age of HF-resistant soil organic matter in a Podzol and a Cambisol. *Organic Geochemistry* 38(8):1356–72.

Fabbri D, Chiavari G, Galletti GC. 1996. Characterization of soil humin by pyrolysis(/methylation)-gas chromatography/mass spectrometry: structural relationships with humic acids. *Journal of Analytical and Applied Pyrolysis* 37(2):161–72.

FAO. 2006. *Guidelines for Soil Description*. 4th edition. Rome: Food and Agricultural Organization (FAO).

Fontaine S, Bardoux G, Abbadie L, Mariotti A. 2004. Carbon input to soil may decrease soil carbon content. *Ecollogy Letters* 7(4):314–20.

Fontaine S, Barot S, Barré P, Bdioui N, Mary B, Rumpel C. 2007. Stability of organic carbon in deep soil layers controlled by fresh carbon supply. *Nature* 450(7167):277–80.

Gandois L, Teisserenc R, Cobb AR, Chieng HI, Lim LBL, Kamariah AS, Hoyt A, Harvey CF. 2014. Origin, composition, and transformation of dissolved organic matter in tropical peatlands. *Geochimica et Cosmochimica Acta* 137:35–47.

Grootes PM, Nadeau MJ, Rieck A. 2004. ^{14}C –AMS at the Leibniz-Labor: radiometric dating and isotope research. *Nuclear Instruments and Methods in Physics Research B* 223–224:55–61.

Hagedorn F, Bruderhofer N, Ferrari A, Niklaus PA. 2015. Tracking litter-derived dissolved organic matter along a soil chronosequence using ^{14}C imaging: biodegradation, physico-chemical retention or preferential flow? *Soil Biology and Biochemistry* 88:333–43.

Hanke A, Sauerwein M, Kaiser K, Kalbitz K. 2014. Does anoxic processing of dissolved organic matter affect organic–mineral interactions in paddy soils? *Geoderma* 228–229:62–66.

Harrison KG, Broecker WS, Bonani G. 1993. The effect of changing land use on soil radiocarbon. *Science* 262(5134):725–6.

Houghton RA. 2007. Balancing the global carbon budget. *Annual Review of Earth and Planetary Sciences* 35:313–47.

Hua Q, Barbetti M, Rakowski Z. 2013. Atmospheric radiocarbon for the period 1950–2010. *Radiocarbon* 55(4):2059–72.

Kalbitz K, Schwesig D, Rethemeyer J, Matzner E. 2005. Stabilization of dissolved organic matter by sorption to the mineral soil. *Soil Biology and Biochemistry* 37(7):1319–31.

Knorr W, Prentice IC, House JI, Holland EA. 2005. Long-term sensitivity of soil carbon turnover to warming. *Nature* 433(7023):2003–6.

Kögel-Knabner I, Amelung W, Cao Z, Fiedler S, Frenzel P, Jahn R, Kalbitz K, Kölbl A, Schloter M. 2010. Biogeochemistry of paddy soils. *Geoderma* 157(1–2):1–14.

Kuzyakov Y, Domanski G. 2000. Carbon input by plants into the soil. Review. *Journal of Plant Nutrition and Soil Science* 163(4):421–31.

Lal R. 1997. Land use and soil management effects on soil organic carbon dynamics on Alfisols in western Nigeria. In: Lal R, Kimble JM, Follet RF, Stewart BA, editors. *Soil Processes and the Carbon Cycle*. Boca Raton: CRC Press. p 109–26.

Leirós MC, Trasar-Cepeda C, Seoane S, Gil-Sotres F. 1999. Dependence of mineralization of soil organic matter on temperature and moisture. *Soil Biology and Biochemistry* 31(3):327–35.

Levin I, Naegler T, Kromer B, Diehl M, Francey RJ, Gomez-Pelaez AJ, Steele LP, Wagenbach D, Weller R, Worthy DE. 2010. Observations and modelling of the global distribution and long-term trend of atmospheric $^{14}CO_2$. *Tellus B* 62: 26–46.

Levin I, Kromer B, Hammer S. 2013. Atmospheric $\Delta^{14}CO_2$ trend in Western European background air from 2000 to 2012. *Tellus B* 65:20092.

Lorenz K, Lal R, Jiménez JJ. 2009. Soil organic carbon stabilization in dry tropical forests of Costa Rica. *Geoderma* 152(1–2):95–103.

Mikutta R, Kleber M, Jahn R. 2005. Poorly crystalline minerals protect organic carbon in clay subfractions from acid subsoil horizons. *Geoderma* 128(1–2):106–15.

Mikutta R, Kleber M, Torn MS, Jahn R. 2006. Stabilization of soil organic matter: association with minerals or chemical recalcitrance? *Biogeochemistry* 77(1):25–56.

Moni C, Lerch TZ, Knoth de Zarruk K, Strand LT, Forte C, Certini G, Rasse DP. 2015. Temperature response of soil organic matter mineralisation in arctic soil profiles. *Soil Biology and Biochemistry* 88:236–46.

Morra MJ, Blank RR, Freeborn LL, Shafii B. 1991. Size fractionation of soil organo-mineral complexes using ultrasonic dispersion. *Soil Science* 152(4):294–303.

Nadeau MJ, Vaernes E, Svarva HL, Larsen E, Gulliksen S, Klein M, Mous DJW. 2015. Status of the "new" AMS facility in Trondheim. *Nuclear Instruments and Methods in Physics Research B* 361:149–55.

Nicoullaud B, King D, Tardieu F. 1994. Vertical distribution of maize roots in relation to permanent soil characteristics. *Plant Soil* 159(2):245–54.

Nieder R, Benbi DK. 2008. *Carbon and Nitrogen in the Terrestrial Environment*. Dordrecht: Springer.

Rethemeyer J, Grootes PM, Bruhn F, Andersen N, Nadeau MJ, Kramer C, Gleixner G. 2004. Age heterogeneity of soil organic matter. *Nuclear Instruments and Methods in Physics Research B* 223–224:521–7.

Schmidt MWI, Rumpel C, Kögel-Knabner I. 1999. Evaluation of an ultrasonic dispersion procedure to isolate primary organomineral complexes from soils. *European Journal of Soil Science* 50(1): 87–94.

Simpson AJ, Song G, Smith E, Lam B, Novotny EH, Hayes MHB. 2007. Unraveling the structural components of soil humin by use of solution-state nuclear magnetic resonance spectroscopy. *Environmental Science and Technology* 41(3): 876–83.

Solly EF, Schöning I, Herold N, Trumbore SE, Schrumpf M. 2015. No depth-dependence of fine root litter decomposition in temperate beech forest soils. *Plant and Soil* 393(1):273–82.

Stuiver M, Polach HA. 1977. Discussion: reporting of ^{14}C data. *Radiocarbon* 19(3):355–63.

Takahashi T, Dahlgren RA. 2016. Nature, properties and function of aluminium-humus complexes in volcanic soils. *Geoderma* 263:110–21.

Torn MS, Trumbore SE, Chadwick OA, Vitousek PM, Hendricks DM. 1997. Mineral control of soil organic carbon storage and turnover. *Nature* 389(6647):3601–3.

Trumbore SE, Chadwick OA, Amundson R. 1996. Rapid exchange between soil carbon and atmospheric carbon dioxide driven by temperature change. *Science* 272(5260):393–6.

Winkler P, Kaiser K, Kölbl A, Kühn T, Schad P, Urbanski L, Fiedler S, Lehndorff E, Kalbitz K, Utami SR, Cao Z, Zhang G, Jahn R, Kögel-Knabner I. 2016. Response of Vertisols, Andosols, and Alisols to paddy management. *Geoderma* 261:23–35.

Wissing L, Kölbl A, Vogelsang V, Fu JR, Cao ZH, Kögel-Knabner I. 2011. Organic carbon accumulation in a 2000-year chronosequence of paddy soil evolution. *Catena* 87(3):376–85.

Witt C, Cassman K, Olk DC, Biker U, Liboon SP, Samson MI, Ottow JC. 2000. Crop rotation and residue management effects on carbon sequestration, nitrogen cycling and productivity of irrigated rice systems. *Plant and Soil* 225(1):263–78.