Organic Carbon Turnover in Three Tropical Soils under Pasture after Deforestation

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ABSTRACT

Deforestation is one of the main reasons for the global net release of CO₂ from soil to atmosphere. Estimates of CO₂ emission from soils are highly variable, mainly due to limited data of C dynamics in soils after forest clearing. The objective of this study was to calculate the changes in soil organic carbon (SOC) storage after deforestation in three soil types in the Atlantic Zone of Costa Rica with help of the δ¹³C method. Changes in bulk density, which normally accompany land use changes, had a profound influence on the results of the calculations. Deforestation, followed by 25 yr of pasture, caused a net loss of 21.8 Mg ha⁻¹ in SOC for an Eutric Hapludand and 1.5 Mg ha⁻¹ for an Oxic Humitropept. The SOC changes in time were studied on a deforestation sequence on an Andic Humitropept. In the first years after forest clearing, decomposition of tree roots caused an extra input of SOC, which influenced the δ¹³C signal. Decomposition of forest C and increase of pasture C were mathematically described for several depths. A considerable influence of depth on decomposition rates was found. The strong stabilization of organic C by Al-organic matter complexes probably caused the relatively small net C loss from SOC since forest clearing.

Increasing atmospheric concentrations of CO₂ and other trace gases may affect the global climate in the near future. Soils play an important role in the production and consumption of CO₂. Soil-vegetation systems can act as a CO₂ sink or a CO₂ source, depending on decomposition rate and rate of SOC formation (Van Breemen and Feijtel, 1990). When forest is cleared, the soil turns into a CO₂ source.

The annual relative increase in global atmospheric CO₂ concentration is 0.5%, which corresponds with 3.6 Pg C yr⁻¹ (Bouwman, 1990). Estimates of global CO₂ release caused by deforestation are between 1.0 and 3.2 Pg C yr⁻¹. Forest clearing as a source of atmospheric CO₂ is second only to release from the burning of fossil fuels (Houghton et al., 1985; Tans et al., 1990). Data from Detwiler and Hall (1988) illustrate the significance of tropical forest clearing. They estimated the CO₂ release from nontropical forest regions in 1980 to be around 0.1 Pg C and for tropical forests to be 0.42 to 1.60 Pg C. Of this amount, 0.1 to 0.3 Pg C were attributed to decreases in SOC content.

The decrease in SOC content depends on the land use established after forest clearing. Detwiler (1986) estimated that cropping of tropical forest soils reduces their C content by 40%; the use of these soils for pastures reduces it by about 20%.

The capacity of soils to sequester C is considerable. Soils developed in volcanic ash (Andisols) have a very high organic matter storage capacity, due to the amorphous components that stabilize organic matter (Mizota and van Reeuwijk, 1989). Boudot et al. (1986) found a negative correlation between the C mineralization rate in Andosols and amorphous Al, Fe, and allophane. In spite of the stabilizing effect of amorphous components in Andosols, considerable SOC losses have been found in a secondary regrowth, compared with a mature tropical forest site (Raich, 1983).

Techniques, based on isotope fractionation of plant material and SOC have stimulated quantitative SOC turnover studies (e.g., Balesdent et al., 1987). Plant C contains distinctly less ¹³C than atmospheric CO₂ and, during decomposition of plant material in the soil, further fractionation takes place. To quantify changes in ¹³C, C isotope ratios defined as ¹³R = ¹³C/¹²C can be used. In practice, the δ¹³C value, which is the relative difference between isotope ratios of sample and standard, is used: δ¹³C‰ = (¹³Rsample / ¹³Rstandard) - 1000. The international standard is V-PDB, which has a ¹³R of 0.01124 (O'Leary, 1981). Atmospheric CO₂ has a δ¹³C of -7.5‰ (Mook, 1986), values of plant C range from -22 to -34‰ in C₄ plants, to around -15‰ in C₃ plants.

During decomposition of plant material, a small enrichment in ¹³C takes place. Generally, δ¹³C values of SOC increase with only 2 to 3‰ in depth, compared with the 15% caused by the change between C₃ and C₄ vegetation (Balesdent et al., 1987). Therefore, changes from C₃ to C₄ plants, provide an in situ labeling (Balesdent et al., 1987). Quantification requires detailed information on vegetation history.

In the Atlantic Zone of Costa Rica, large areas of tropical lowland forest have been cleared in the last 40 yr (Veldkamp et al., 1992). More than 50% of the cleared area has been put under pastures with low dry matter production, mainly consisting of C₄ grasses. Because forests are dominated by C₃ species, this shift in vegetation makes it possible to apply the δ¹³C method.

The objective of this study was to quantify the changes in SOC storage after deforestation under pastures with help of the δ¹³C method. Based on these changes in SOC, an estimate will be made of the total C losses caused by deforestation of three different soil types in the wet tropical lowlands of Costa Rica.

MATERIALS AND METHODS

Study Sites

The study was carried out in the Atlantic Zone of Costa Rica. The climate is humid tropical; mean annual temperature is 26 °C and mean annual rainfall is about 4000 mm. Precipitation exceeds evapotranspiration in all months.

The study area is situated on the foot slopes of the Turrialba volcano, and is covered with laharic, fluviolaharic, and fluvial geological information system; V-PDB, belemnite from the Pee Dee formation in South Carolina. **Significant at the 0.01 probability level.
Table 1. Bulk density, $\rho_b$ organic C ($C_{org}$), pH in water, and cation-exchange capacity (CEC) of selected samples of three different soil types under forest and pasture in the Atlantic Zone of Costa Rica.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Forest $\rho_b$</th>
<th>Forest $C_{org}$</th>
<th>Forest pH</th>
<th>Forest CEC</th>
<th>Pasture $\rho_b$</th>
<th>Pasture $C_{org}$</th>
<th>Pasture pH</th>
<th>Pasture CEC</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.00-0.05</td>
<td>0.58</td>
<td>7.4</td>
<td>5.6</td>
<td>38.2</td>
<td>0.82</td>
<td>3.95</td>
<td>5.5</td>
<td>48.4</td>
</tr>
<tr>
<td>0.30-0.35</td>
<td>0.83</td>
<td>5.1</td>
<td>6.0</td>
<td>28.0</td>
<td>0.92</td>
<td>2.17</td>
<td>6.1</td>
<td>30.2</td>
</tr>
<tr>
<td>0.60-0.65</td>
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<td>4.2</td>
<td>6.2</td>
<td>16.0</td>
<td>1.04</td>
<td>0.83</td>
<td>6.4</td>
<td>24.8</td>
</tr>
</tbody>
</table>

Deposits, mostly of andesitic composition. Tropical lowland rain forest is the natural vegetation. Tosi (1969) distinguished the life zones Premontane Wet Forest and Tropical Wet Forest in the area. After deforestation, one of the dominant land use types is pasture.

An important assumption of the $\delta^{13}C$ method is that the soils under pasture are directly comparable to those under forest. For that reason, particular attention was paid to site selection. A deforestation map was made and combined with an available soil map in a GIS (Veldkamp et al., 1992). Corresponding forest and pasture sites were selected within 1 km of each other. On all sites, clearing was done by small farmers without heavy machinery. Valuable tree trunks were removed along a few tracks and others were cut and left on the sites. Generally, stumps were not removed. If burning was done, it was restricted to piles. The location of former burning piles was easily identified by the charcoal left in the soil. Soil profiles were sampled only where there were no morphological indications of significant disturbance or burning during clearing. Pasture grasses are normally planted directly after clearing, without any further land preparation treatment.

Three well-drained soil types were sampled: an Eutric Hapludand, an Oxic Humitropept, and an Andic Humitropept (Soil Survey Staff, 1990). The studied soil profiles contained only negligible amounts on rock fragments. In Table 1, some important soil characteristics are summarized. Two sites of the Eutric Hapludand were selected: one under natural forest vegetation and the other site that had been manually cleared about 25 yr earlier and put under tropical carpetgrass pasture [Axonopus compressus (Sw.) P. Beauv.]. Two Oxic Humitropept sites were selected: one under forest vegetation, where some trees had been removed, and the other under 25-yr-old pasture [Andic Humitropept, five sites were selected: one site in natural forest, four others in pastures dominated by Ischaeum indicum (Houtt.) Merr. of 3, 5, 10, and 18 yr. On the younger pasture sites, decomposing tree trunks and stumps were still present.

Sampling and Sample Processing

Soil bulk density was sampled at five depths ($n \geq 8$, depending on variability) using $0.3 \times 10^{-3}$ m$^3$ stainless steel rings. The samples were oven dried for 24 h at 105 °C. Disturbed soil samples of about 1.5 kg were taken at 0.05-m intervals to 1.0 m. Soil samples were homogenized, air dried, and sieved to pass 2 mm. Litter and root material were separated manually from the soil. Air-dry soil, litter, and root samples were stored for $^{13}C$ analysis. Organic C and $^{13}C$ analyses were carried out in the Isotope Physics Laboratory of the University of Groningen, the Netherlands. To measure the $^{13}C$ composition, soil samples were first ground and treated with 1 M HCl to remove any CaCO$_3$. The samples were then oven dried at 80 °C. About 1 g of the samples was burned in the presence of CuO under pure O$_2$ at 900 °C. Water vapor was trapped on dry ice. The CO$_2$ and NO$_x$ were trapped with help of liquid air and analyzed on a SIRA 9 mass spectrometer (VG Isogas Ltd., Cheshire, England). The laboratory reference was calibrated against V-PDB, using the international standard NBS 19 (available from the National Bureau of Standards). Analytical precision determined as the standard deviation obtained on different combustions of the same homogenized sample is better than 0.05% $\delta^{13}C$.

Soil pH was measured by glass electrode in supernatant above a 1:2.5 soil/water suspension. Cation-exchange capacity was determined at pH 7 by NH$_4$OAc. Iron and Al were determined on the disturbed soil samples by atomic absorption spectrophotometry in pyrophosphate extracts and in acid oxalate extracts (Mizota and van Reeuwijk, 1989). Clay mineralogy was determined by x-ray diffraction on the clay fraction.

Calculations

Carbon contents of soil samples, $C_s$ (kg C kg$^{-1}$ soil), were converted to TOC (Mg ha$^{-1}$), based on soil layer thickness, $L$ (m), and bulk density, $\rho_b$ (Mg m$^{-3}$):

$$TOC = C_s L \rho_b \times 10^4 \quad \text{[1]}$$

Forest clearing and cultivation may cause compaction. On the Andic Humitropept, deforestation sequence bulk density gradually increased in time, which suggests that in this case not clearing practices but cattle trampling was the main reason for compaction (Keller et al., 1993). On all pasture sites bulk density may have an important effect on C-balance calculations, because samples in the forest at a certain depth are not directly comparable with samples in the pasture at the same depth. For example, if the bulk density of the first layer increases from 0.6 Mg m$^{-3}$ in the forest to 0.8 Mg m$^{-3}$ in the pasture, the 0.1-m top layer in the forest corresponds to a 0.075-m layer in the pasture. Thickness of layers in the pasture soils were adjusted assuming that the relationship between $\rho_b$ and depth in the pasture was originally the same as found in the corresponding forest profiles.

The amounts of SOC$_s$ and SOC$_p$ were estimated from the equation (Vitorello et al., 1989)

$$TOC \delta^{13}C_p = (SOC_s \delta^{13}C_s) + (SOC_p \delta^{13}C_p) \quad \text{[2]}$$

where:

$$\delta^{13}C_s = \delta^{13}C \text{ value of sample from pasture soil, corrected for compaction;}$$

$$\delta^{13}C_t = \delta^{13}C \text{ value of corresponding sample from forest soil;}$$

$$\delta^{13}C_p = \delta^{13}C \text{ value of pasture residues (litter and roots).}$$

In Eq. [2], variations in $\delta^{13}C$ due to mineralization after a change in land use are ignored.
RESULTS AND DISCUSSION

Organic Carbon Dynamics in Different Soil Types

The SOC profiles of forest and pasture in the Eutric Hapludand differed considerably (Table 2). The TOC decreased in the top 0.1 m from 48.1 to 34.3 Mg ha\(^{-1}\), assuming that the soil under forest can be considered as the precursor of the soil under pasture, and if changes in bulk density are ignored. Differences in deeper layers were small. For the Oxic Humitropept under pasture a small increase in SOC was found in the top 0.1 m (27.6 Mg ha\(^{-1}\)) compared with the forest (26.4 Mg ha\(^{-1}\)). When a correction for compaction is made, the TOC under pasture in the top 0.1 m of the Eutric Hapludand appears to have decreased to 24.6 Mg ha\(^{-1}\), almost 30% less than without correction. For the Oxic Humitropept, the corrected TOC in the topsoil is slightly decreased to 24.9 Mg ha\(^{-1}\). Without correction for compaction, the loss of organic C is thus systematically underestimated, especially in top horizons, where compaction is severe. Further calculations therefore all involved correction for compaction. In the forest soils \(\delta^{13}C\) increased from between \(-27.01\) and \(-27.44\)% in the top 0 to 0.05 m to about \(-25.00\)% between 0.5- and 0.7-m depths (Fig. 1). Similar trends were found by Desjardins et al. (1991) and Volkoff and Cerri (1987) for forest soils in the humid tropics of Brazil. Nadelhofer and Fry (1988) discussed that extra forest litter inputs to soil lower \(\delta^{13}C\) values of topsoils. They also linked increasing \(\delta^{13}C\) values with depth to soil age and organic matter decomposition. In the forest profiles of this study, \(\delta^{13}C\) values increased with depth to a value of about \(-25.0_{\text{oo}}\). This suggests that \(-25.0_{\text{oo}}\) is the \(\delta^{13}C\) value typical for the PSOC fraction. No differences occurred in \(\delta^{13}C\) values in different topsoils, suggesting that forest litter at the different sites had similar isotopic composition.

Decomposable and Passive Soil Organic Carbon

Soil organic C was divided in two fractions: DSOC, which consists of the SOC that decomposes within 25 yr (Parton et al., 1987), and PSOC, which consists of SOC that decomposes at slower rates. After 25 yr of pasture, the SOC fraction has been mineralized and the remaining SOC consists of PSOC. This passive soil C fraction is much larger for the Hapludand than for the Humitropept (Fig. 1). In the Hapludand, the passive C fraction remains about constant to a depth of 0.3 m, whereas for the Humitropept it decreases with depth from >7 Mg ha\(^{-1}\) in the top 0.0 to 0.05 m to <5 Mg ha\(^{-1}\) in the 0.25- to 0.30-m layer.

In the forest profiles, the increase in \(\delta^{13}C\) values during decomposition can be used to quantify the DSOC and PSOC pools. From the following equation both fractions can be calculated under the assumption that the PSOC has a typical \(\delta^{13}C\) value:

\[
TOC_t = (\text{DSOC} \delta^{13}C) + (\text{PSOC} \delta^{13}C_p) \quad [3]
\]

where:

\[
\delta^{13}C_t = \delta^{13}C\text{ value of forest soil}, \\
\delta^{13}C_D = \delta^{13}C\text{ value of DSOC}, \\
\delta^{13}C_P = \delta^{13}C\text{ value of PSOC}.
\]

The maximum \(\delta^{13}C\) value in each profile was used as \(\delta^{13}C_t\). The \(\delta^{13}C\) value for decomposable organic C must lie between that of fresh plant material (\(\delta^{13}C = -32.0\%o\)) and SOC in the soil surface (\(\delta^{13}C = -27.5\%o\)). The value of \(\delta^{13}C_p = -28.9\%o\) was used, which was measured for decomposing litter in the forest. In the two Humitropepts DSOC occurs only in the top 0.5 m (Table 3). In the Hapludand DSOC occurs to a depth of 0.7 m. The total SOC pool in the top 0.3 m of the Hapludand is about 53 Mg ha\(^{-1}\) in the Andic and Oxic Humitropepts 28 and 23 Mg ha\(^{-1}\), respectively. The PSOC pool is also larger in the Hapludand (45 Mg ha\(^{-1}\)) than in the Humitropepts (23 and 30 Mg ha\(^{-1}\), respectively).

Organic Carbon Dynamics after Forest Clearing

The \(\delta^{13}C\) profiles under pastures of different ages in an Andic Humitropept were studied to follow the changes in SOC with time after deforestation. In the forest profile \(\delta^{13}C\) increases from \(-27\%o\) at the top to about \(-24\%o\) at 1.0-m depth. The \(\delta^{13}C\) profiles under pasture differ considerably from the forest profile (Fig. 2). When a C\(_f\) forest vegetation is replaced by C\(_g\) grassland vegetation, normally an increase in \(\delta^{13}C\) is expected. At 3 and 5 yr after deforestation, however, \(\delta^{13}C\) at a depth between 0.2 and 0.4 m was distinctly lower than in the forest (Fig. 2). An extra input of forest-derived C, from decomposing tree roots, is probably responsible for this decrease. When calculations based on the \(\delta^{13}C\) values of young pastures are performed, this extra input of C derived from tree roots has to be taken into account. Pasture-derived organic C will be underestimated for the young pastures if the SOC profile under forest is taken as reference for calculations, as is usually done (Vitorello et al., 1989).

For root biomass and litter on the forest floor, I used data from Raich (1983), who studied the C cycle in a forest in the same area on a very similar soil. A \(\delta^{13}C\) value of \(-31.98\%o\) was used for forest roots, which was measured for fresh tree roots. For calculation purposes, litter on the forest floor was added to the top 0.05-m layer. Results of the calculations are summarized in Table 4.

For the 0- to 0.1-m layer, changes with time in SOC\(_t\) and SOC\(_p\) were calculated (Fig. 3). An exponential increase in SOC\(_p\) was assumed according to

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Eutric Hapludand</th>
<th>Oxic Humitropept</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Forest corrected</td>
<td>Forest corrected</td>
</tr>
<tr>
<td></td>
<td>uncorrected</td>
<td>uncorrected</td>
</tr>
<tr>
<td></td>
<td>Mg ha(^{-1})</td>
<td>Mg ha(^{-1})</td>
</tr>
<tr>
<td>----------</td>
<td>------------------</td>
<td>------------------</td>
</tr>
<tr>
<td>0.00-0.05</td>
<td>21.5</td>
<td>11.5</td>
</tr>
<tr>
<td>0.05-0.10</td>
<td>26.6</td>
<td>13.1</td>
</tr>
<tr>
<td>0.10-0.15</td>
<td>15.8</td>
<td>13.3</td>
</tr>
<tr>
<td>0.15-0.20</td>
<td>11.6</td>
<td>16.2</td>
</tr>
<tr>
<td>0.20-0.25</td>
<td>13.9</td>
<td>10.6</td>
</tr>
<tr>
<td>0.25-0.30</td>
<td>8.5</td>
<td>10.4</td>
</tr>
<tr>
<td>0.30-0.40</td>
<td>12.7</td>
<td>17.9</td>
</tr>
<tr>
<td>0.40-0.50</td>
<td>14.5</td>
<td>14.3</td>
</tr>
</tbody>
</table>
Fig. 1. The $\delta^{13}$C profiles and corresponding soil organic carbon derived from forest ($SOC_f$) and from pasture ($SOC_p$) profiles of an Eutric Hapludand and an Oxic Humitropept under 25 yr of pasture.

$$SOC_p = TP[1 - \exp(-r_{TP} t)]$$

where $t$ is time (yr), TP is the maximum (steady-state) level of $SOC_p$, and $r_{TP}$ is the relative growth rate (yr$^{-1}$). Best fit ($R^2 = 0.99^{**}$) was obtained with TP = 12.2 Mg ha$^{-1}$ and $r_{TP} = 0.19$ yr$^{-1}$. With these parameters, the maximum value of $SOC_p$ will be reached in about 30 yr.

At $T = 0$ (forest) the sum of organic C in the soil (0–0.1 m) plus the aboveground litter in the forest floor (24.7 Mg C ha$^{-1}$) and the root biomass (2.3 Mg C ha$^{-1}$) in the top 0.1-m layer was used. A double exponential function was fitted to the $SOC_f$ observations, assuming a decomposable (DF) and a passive (PF) fraction of a the soil organic C:

$$SOC_f = DF \exp(-k_{DF} t) + PF \exp(-k_{PF} t)$$

where $k_{DF}$ and $k_{PF}$ are the relative decomposition rate constants. Parameters with the best fit ($R^2 = 0.99^{**}$) were: $DF = 41.5$ Mg ha$^{-1}$; $k_{DF} = 0.46$ yr$^{-1}$; PF =

Table 3. Decomposable soil organic carbon (DSOC) and passive soil organic carbon (PSOC) pools in three soil types under forest.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Eutric Hapludand</th>
<th>Oxic Humitropept</th>
<th>Andic Humitropept</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DSOC</td>
<td>PSOC</td>
<td>DSOC</td>
</tr>
<tr>
<td>0.00–0.05</td>
<td>13.8</td>
<td>7.7</td>
<td>11.2</td>
</tr>
<tr>
<td>0.05–0.10</td>
<td>19.2</td>
<td>7.4</td>
<td>4.2</td>
</tr>
<tr>
<td>0.10–0.15</td>
<td>7.9</td>
<td>8.0</td>
<td>2.8</td>
</tr>
<tr>
<td>0.15–0.20</td>
<td>4.2</td>
<td>6.8</td>
<td>2.2</td>
</tr>
<tr>
<td>0.20–0.25</td>
<td>4.7</td>
<td>9.2</td>
<td>1.7</td>
</tr>
<tr>
<td>0.25–0.30</td>
<td>2.8</td>
<td>5.8</td>
<td>1.3</td>
</tr>
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<td>0.30–0.40</td>
<td>3.4</td>
<td>9.3</td>
<td>1.0</td>
</tr>
<tr>
<td>0.40–0.50</td>
<td>3.8</td>
<td>10.6</td>
<td>0.1</td>
</tr>
</tbody>
</table>
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12.2 Mg ha\(^{-1}\); and \(k_{DF} = 0.003\) yr\(^{-1}\). Gonzalez and Sauerbeck (1982), who estimated the decomposition rate constants for fresh maize (Zea mays L.) straw on a site in the Atlantic Zone of Costa Rica, found values of 3.4 and 0.15 yr\(^{-1}\). Their overall rate constant corresponds to the rate constant for the decomposable fraction \((k_{DF})\) in this study.

Extrapolation in time of the equations should be done with care, because only five observations in time were used. After the first years, the amount of SOC\(_f\) seems to stabilize at about 12 Mg ha\(^{-1}\) for the 0- to 0.1-m layer. Of the SOC originally present (=26.9 Mg ha\(^{-1}\) for the 0- to 0.1-m layer) about 45% consists of PSOC and the rest of DSOC. Total soil C (SOC\(_f\) + SOC\(_p\)) rapidly decreases after deforestation and stabilizes after about 5 yr (Fig. 3). This rapid initial decrease is mainly caused by decomposing litter and roots. If less litter would be left after clearing, the initial TOC value would be lower, but after about 5 yr TOC would not differ significantly. After 30 yr, TOC stabilizes at about 24 Mg ha\(^{-1}\) for the 0- to 0.1-m layer. Comparing this number with the TOC pool in the forest (26.8 Mg ha\(^{-1}\), without taking litter and roots into account) reveals that in the Andic Humitropept under low-productivity pasture the SOC content in the top layer tends to decrease but only very slowly.

### Organic Carbon Decomposition with Depth

Equation [5] was fitted to the observations at three depths to evaluate the relation between changes in SOC and depth in the profile. Parameters of the best fits are given in Table 5. The relative decomposition rate of the decomposable SOC\(_f\) pool \((k_{DF})\) is higher in the 0- to 0.1-m layer than in deeper layers. The high concentration of relatively easily decomposable litter in the top layer is probably responsible for this effect. In the deeper layers, the decomposable pool (DF) probably consists mainly of tree roots, left in the soil after forest clearing. The rate

<table>
<thead>
<tr>
<th>Depth</th>
<th>SOC(_f)</th>
<th>SOC(_p)</th>
<th>SOC(_f)</th>
<th>SOC(_p)</th>
<th>SOC(_f)</th>
<th>SOC(_p)</th>
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<th>SOC(_p)</th>
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<td>0.0</td>
<td>6.0</td>
<td>0.4</td>
<td>7.5</td>
<td>0.6</td>
<td>4.6</td>
<td>1.6</td>
<td>5.2</td>
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<td>0.0</td>
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<td>0.4</td>
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<td>3.5</td>
<td>0.2</td>
<td>3.0</td>
<td>0.6</td>
<td>3.8</td>
</tr>
</tbody>
</table>

† Including litter on the forest floor and roots in the profile.
Table 5. Parameters describing the exponential increase in soil organic carbon derived from pasture (SOC_P) and the exponential decrease in soil organic carbon derived from forest (SOC_F) at three depths in a deforestation sequence on an Andic Humitrope.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>TP (Mg ha⁻¹)</th>
<th>r_P (yr⁻¹)</th>
<th>DF (Mg ha⁻¹)</th>
<th>k_P (yr⁻¹)</th>
<th>PF (Mg ha⁻¹)</th>
<th>k_F (yr⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0-0.1</td>
<td>12.2</td>
<td>0.19</td>
<td>42.5</td>
<td>0.44</td>
<td>11.1</td>
<td>0.003</td>
</tr>
<tr>
<td>0.1-0.2</td>
<td>8.2</td>
<td>0.05</td>
<td>7.0</td>
<td>0.23</td>
<td>9.0</td>
<td>0.003</td>
</tr>
<tr>
<td>0.2-0.3</td>
<td>3.6</td>
<td>0.05</td>
<td>5.6</td>
<td>0.24</td>
<td>5.7</td>
<td>0.001</td>
</tr>
</tbody>
</table>

+ TP and r_P: steady-state level and relative growth rate of SOC_P.
+ DF and k_P: decomposable SOC_P and decomposition rate constant.
+ PF and k_F: passive SOC_P and decomposition rate constant.

Mechanisms Stabilizing Soil Organic Carbon

Extraction with acid oxalate dissolves Al in allophane and organically bound Al (Al₀). Pyrophosphate is supposed to dissolve only humus-Al (Alₚ). Subtraction of Al₀ from Alₚ results in the Al present in allophane (Alₐ) (Mizota and van Reeuwijk, 1989). Stepwise multiple regression was done to correlate the Al and Fe extractions in the PSOC fraction. Changes in PSOC could be explained by sesquioxide fractions (R² = 0.96**) in the Eutric Hapludand with the relationship:

\[ SOC_f = -2.95 + 14.98 \times Al_p + 3.24(Al_o - Al_p) \]  

Stable Al–organic matter complexes (Al₀) thus play a dominant role in the stabilization of SOC. The contribution of (Al₀ – Alₚ) indicates that allophane (Al₀ – Alₐ) also plays a role in stabilizing organic C.

For the Oxic Humitrope, the explained variance of the regressions using PSOC as dependent variable was also high (R² = 0.97**):  

\[ SOC_f = 0.36 + 6.9Al_p + 3.6Fe_0 \]  

So, in the Oxic Humitrope, Al–organic matter complexes are closely correlated to the magnitude of the stable SOC_f pool (Mizota and van Reeuwijk, 1989).

CONCLUSIONS

For a correct application of the δ¹³C method, detailed information on changes in bulk density accompanying changes in land use is critical. In the first years after clearing, decomposing tree roots cause a decrease in δ¹³C instead of an increase, which would be expected under pastures. Deforestation followed by 25 yr of pasture caused a net loss of 21.8 Mg C ha⁻¹ for the Eutric Hapludand and of 1.5 Mg C ha⁻¹ for the Oxic Humitrope. Due to the strong stabilization of SOC in Al–organic matter complexes in soils of volcanic origin and the continuous cover of grassland, this decline in SOC after forest clearing was less than usually reported in other studies.

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