Compost Quality and Soil Improvement

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Soil Carbon Stocks and Global Change

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Abstract
Feedbacks between soil carbon fluxes to and from the atmosphere and climate change could strongly influence future climate developments. Not only does temperature influence plant derived carbon input into soil and decomposition rates, but droughts, fires, floods and thawing also could affect environmental constrains on decomposition rates. Global change processes, such as agricultural management practices, e.g. tilling technologies, fertilizers and land use, also have considerable influence on soil carbon stocks and their distribution and, through these, on climate change.

1 Introduction
Soil is a major player in climate change: through its carbon uptake and efflux it influences climate and through changes in temperature, moisture etc. soil in turn is influenced by climate change. Changes in agricultural management practices, as e.g. tillage technologies, fertilizers, etc. evolve and also have considerable influence on soil carbon stocks and their distribution and, through these, on climate change. On the other hand, climate change can call for changes in management practices.

2 Climate Change and its Influence on Carbon Fluxes from and to the Soil

2.1 The Carbon Cycle
By far the largest stock of carbon is found in the solid earth (10^9 Gt C), followed by that in the ocean (4.10^4 Gt C). The carbon stored in the world's soils (1500 Gt C) - including peatlands, wetlands and permafrost – exceeds that in the atmosphere (750 Gt C) by a factor of two and that of vegetation (550 Gt C) by a factor of almost three (Kromp-Kolb und Formayer 2005). In principle, soil organic carbon is derived from autotrophic organisms - higher plants and autotrophic microorganisms. Dead plant and microbial material is the most important energy source for soil microorganisms. The organic debris are either mineralized - i.e. completely

decomposed to CO_2 - or humified. The latter process leads to soil organic matter fractions of widely varying turnover half lives, from several decades to thousands of years. Carbon exchange processes between the atmosphere and vegetation are very rapid, those with the surface waters of the ocean are quick, while those with the soil are slow and with the deep ocean very slow. The flux from soil to the atmosphere is due to "soil respiration" (Davidson and Janssens 2006), comprising root respiration and microbial decomposition of organic matter, while the input into the soil is mainly an indirect one through plant derived inputs into the soil. These fluxes are of the order of 60 GT C/year for plant and soil respiration (Esser 1990, Schlesinger 1997) and 120 Gt C/year for biomass production.

The near equilibrium between carbon uptake by the atmosphere and carbon depletion into the ocean and uptake by the biosphere including the soil induced but slow changes in atmospheric concentration over geological time periods. In the last 650.000 years carbon dioxide concentrations in the atmosphere fluctuated between about 180 ppm and 280 ppm in eight cycles of the order of 100.000 years (IPCC 2007). Over the last 150 years, however, the carbon dioxide concentration rose to 380 ppm, exceeding the natural range of the last 650,000 years by far, and continue to rise by about 2 ppm per year. The bulk of this rapid increase is considered to be a consequence of the burning of fossil fuels by man. Annual fossil CO_2 emissions averaged 7.2 Gt C per year in 2000-2005. Although this is little compared to the natural fluxes, it is more than plants, soils and the oceans could take up: about half of the anthropogenic input remains in the atmosphere.

2.2 Climate Change

Through the greenhouse effect influencing the radiation budget of the earth and a number of feedback mechanisms in the climate system temperatures and carbon dioxide concentration are closely interrelated and temperatures rise and fall with the change in carbon dioxide concentration. CO_2 radiative forcing has increased significantly over the last decades (20% from 1995 to 2005, the largest in any decade in at least the last 200 years). It is now (IPCC 2007) calculated to be 1,66 W / m^2, significantly larger than radiative forcing of other greenhouse gases, such as methane (0,48 W / m^2) and of N_2O (0,16 W / m^2).

The increase of carbon dioxide and other green house gas concentrations in the atmosphere has led to considerable changes in climate. Global average air temperature rose by about 0.8°C within the last 150 years, with local increases partially significantly higher (e.g. Alpine area: 2°C). Cold days, cold nights and frosts occur less frequently, while the number of hot days, hot nights and heat waves has increased. In Austria, for instance, days with temperatures above 30°C have doubled within the last 50 years (Kromp-Kolb und Schwarzl 2005). At continental and regional scales numerous other long-term changes in climate have been observed. These include (IPCC 2007):

- Precipitation has significantly increased in eastern parts of North and South America, northern Europe and northern and central Asia. Regional varia-

tions in the change of precipitation patterns are large, however.

- Spring snow cover in the Northern Hemisphere shows a 5% drop during 1980s.
- Temperatures at the top of permafrost layer have generally increased by up to 3°C since the 1980s.
- The maximum area covered by seasonally frozen ground has decreased by about 7% in the Northern Hemisphere since 1900, in spring up to 15%.
- Glaciers have increasingly retreated since the early 1990s, with a special boost in Europe in the hot summer of 2003.
- The frequency of heavy precipitation events has increased over most land areas - consistent with warming and increases of atmospheric water vapour. However, regional differences are large: Thus there is an increase of 50% in western Austria, while values remained constant in the Vienna area.
- Drying has occurred in the Sahel, the Mediterranean, southern Africa and parts of southern Asia.
- The frequencies and intensities of extreme weather including droughts, heavy precipitation and heat waves has changed.
- More intense and longer droughts have been observed since the 1970s, particularly in the tropics and subtropics.

Global Climate Models show that continued greenhouse gas emissions at or above current rates would cause further warming and induce many changes in the global climate system during the 21st century that would very likely be larger than those observed during the 20th century. According to the latest IPCC report (2007) the best estimate of global warming for a low emission scenario (B1) is 1.8°C (likely range is 1.1°C to 2.9°C), and for high emission scenario (A1FI) is 4.0°C (likely range is 2.4°C to 6.4°C). Near term projections are rather insensitive to the choice of scenario, as the climate system has considerable inertia, and even if greenhouse gas emissions had been kept constant at the level of the year 2000, a temperature rise of 0,1°C per decade would have to be expected over the next decades.

Projected warming in 21st century is expected to be larger over land, especially in northern latitudes and least over the Southern Ocean and parts of the North Atlantic Ocean. Precipitation increases are very likely in high latitudes, while decreases are likely in most subtropical land regions. Snow cover is projected to contract further and widespread increases in thaw depth in most permafrost regions is expected. It is very likely that hot extremes, heat waves, and heavy precipitation events will continue to become more frequent. Extra-tropical storm tracks are projected to move polewards with consequent changes in wind, precipitation, and temperature patterns over Europe.

In the Alpine area significantly hotter and dryer summers and warmer and somewhat wetter winters are expected. The number of days with temperatures above 30°C are expected to at least double again within the next 50 years throughout Austria (Formayer et al. 2004, Frank et al 2005). The increased precipitation

in winter will to a growing percentage be rain, rather than snow, thus further decreasing snow cover. This implies dryer soils, due to less and earlier input of moisture from thawing snow and increased absorption of solar radiation by the snow-free soil. Higher snow lines associated with warmer temperatures, increased thunderstorm activity in some areas and possibly more intense precipitation from Mediterranean cyclones (Vb-cyclones) will enhance the risk of flooding in Austria (Formayer et al. 2007 in preparation).

2.1 Climate Sensitivity of Carbon Fluxes from and to the Soil

Carbon fluxes between soil and atmosphere are climate dependant: obviously biomass production is influenced by temperature, precipitation, snow cover, extreme events, etc., both directly and indirectly through pests and diseases whose occurrence is climate dependant. In addition, increase in carbon dioxide concentrations enhance production. At present an increase in production is expected with warming and rising CO_2 concentrations on the global scale, in spite of considerable agricultural areas experiencing a loss of production due to diminishing precipitation, increasing water stress or acceleration of phenological phases. The primary input to the soil, leaf and root detrius, is therefore also expected to increase globally – extracting increasing amounts of carbon from the atmospheric reservoir, constituting a negative feedback.

However, temperature also influences the rate of decomposition. A well known rule-of-thumb claims that the rate of decomposition of soil organic matter (SOM) doubles every 10°C of temperature increase. This would imply increasing efflux of carbon dioxide from the soil into the atmosphere with increasing temperature, and thus a positive feedback. These counteracting processes make a quantitative understanding of temperature and climate sensitivity of soil-vegetation-atmosphere carbon fluxes essential.

In fact temperature sensitivity of decomposition is no simple matter. The "intrinsic temperature sensitivity" of decomposition, due to the inherent kinetic properties based on molecular structure and ambient temperature is an idealised concept that needs to be complemented by the "apparent temperature sensitivity", the observed response to temperature taking account of environmental constraints (Davidson and Janssens 2006). The multiple processes of environmental constraints that govern availability of substrates to enzymes are especially sensitive to specific aspects climate, such as drought, fire, flooding and freezing, but are not yet well understood. As shown above, considerable changes in these climate components must be expected especially affecting peat and wetlands, as well as permafrost areas. In these cases the uncertainties regarding to the degree to which the currently constrained decomposition rates are temperature sensitive are of lesser importance than those regarding the area of permafrost that is likely to melt or how much of the peatland area is likely to dry significantly. Assuming that 25% of the estimated stocks of carbon in peatlands and permafrost is subject to loss due to global warming in the twenty-first century, this potential loss would be two to three times larger than C losses from mineral soils using current soil

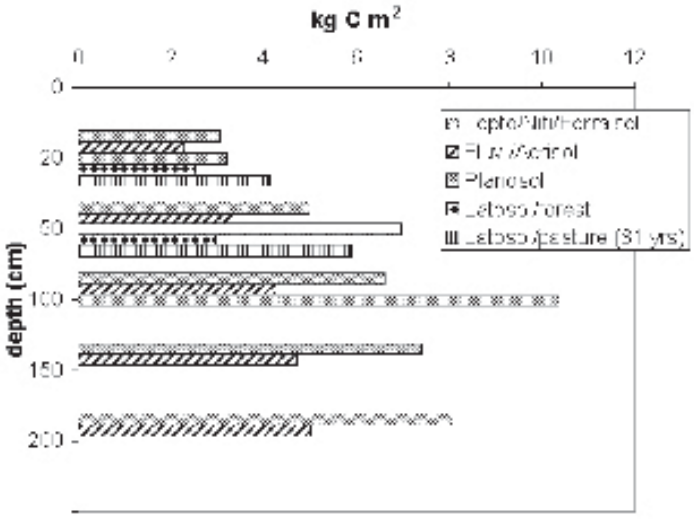
C models (Davidson and Janssens 2006). This, as many other processes not yet properly quantified or included in climate models, indicates that climate change might accelerate beyond present expectations.

Some global change processes, such as improved soil management (for example, conservation tillage) and land-use change however, could help to enhance soil carbon stocks.

3 Soils Carbon in Agroecosystems

3.1 Carbon Distribution in Soils[1]

As organic matter input occurs both onto the soil surface (litter and manures) and within the soil profile (root exudates and dead root material), the 3-dimensional spatial distribution of organic carbon sources at the micro- , meso- and macroscales varies widely between soils of different origin, and according to landuse and management. The depth distribution of organic carbon in soils is influenced by various factors, especially by (i) tillage, which leads to more or less unified C_{org} concentrations within the tillage depth and increases soil aeration; (ii) the depth of the rooting zone, as plants transfer 20 - 30 % of the assimilates into the soils (Kuzyakov and Domanski 2000); (iii) the weathering status of the soil and



thus the depth of penetrable soil horizons; (iv) the activity of the soil meso- and macrofauna; (v) the chemical soil properties, as high clay contents favor the formation of organo-mineral complexes; (vi) soil erosion and colluvial enrichment; and (vii) soil management including historical land-use (e.g. plaggen-manuring

1 The present chapter contains parts of a previously published review paper (Kandeler et al., 2003).

(North Germany), terra preta (Amazonia) caused unusually high organic carbon contents (Springob and Kirchmann 2002).

Fig. 1: Depth distribution of organic carbon stocks in a range of different soil profiles (from Kandeler et al. 2003): Lepto/Niti/Ferralsol and Fluvi/Acrisol: Thailand; Planosol: mean values of four soil profiles under pasture at 2300 m a.s.l. in the Grossglockner massif, calculated from Gerzabek et al. (2004); Latosol: data from Neill et al. (1996, modified)

Figure 1 shows the high variability of C_{org} depth distributions within different soil types. Tropical soils (Lepto-/Niti-/Ferralsol and Fluvi-/Acrisol) show increasing organic C contents down to 2 and more meters. Especially the deeply weathered soils store considerable amounts of C_{org} in deeper layers. Plant cover may change the C-stocks drastically. Fig. 1 gives an example from a chronosequence investigated in the Amazon region. A latosol that had already been under grassland for 81 years had significantly more Corg than under forest. A planosol from a footslope in the Alps contains the highest amount of Corg of all the presented soils within the first 100 cm of its profile due to the unfavorable mineralization conditions (wet, periodically frozen subsoil). The high Corg stocks in deeper layers are due to colluvial materials on which this latter soil developed.

3.2 Dynamics of Organic Matter Decomposition in Agroecosystems

Soil management and agricultural practices significantly influence microbial decomposition of organic carbon in soil. In Austria, Gerzabek et al. (2005) evaluated C-inventories of different agricultural soils (Table 1). Soil use classes can be ranked according to their average C_{org} stocks (0-50 cm) as follows: vineyards ~ cropland < orchards/gardenland < intensive grassland < extensive grassland. The obtained ranking clearly reflects the balance between organic matter input by plant residues and possibly organic manures, and mineralization. Generally, most soil carbon is stored in topsoils (0-20 cm) in all land-use classes. The lowest C_{org} contents in topsoil layers are observed in intensively tilled systems (cropland and vineyards), the highest in extensive grasslands, which in most cases are characterized by slow mineralization of organic matter.

Comparing the subsoil C-inventories under different landuse reveals smaller differences than for the topsoil layers. This indicates that subsoil-C is less influenced by the present landuse than the topsoil. Franzlübbers (2002) recently suggested to use the C-stratification, which is also a common phenomenon in natural ecosystems, as an indicator for soil ecosystem functioning and especially for detecting management-induced changes. Calculating the stratification ratio (quotient of C_{org} inventories in the 0-20 cm and 20-50 cm layers) yielded the following ranking: vineyards (2.15) < cropland (2.27) < orchards/gardenland (2.71) < intensive grassland (2.95) < extensive grassland (3.75). Land-use classes exhibiting the lowest ratios seem to have the largest potential for SOM build up. Of course, this approach requires to take into account the site-specific potential of a soil to store

Tab. 1: Weighted median values of organic carbon contents (t C / ha) and stored Corg-amounts (Mt C) in Austria's agriculturally used soils (Gerzabek et al. 2005).

Depth (cm)	0-20	20-50	0-20	20-50	area
	t C ha^{-1}		Mt C		10^3 ha
Cropland	41.3 ± 5.0*	18.2	57.8	25.5	1397
Grassland	60.5 ± 9.1	20.5	56.7	19.3	938
Extensive grassland/Alpine Meadows	91.8 ± 15.1	27.2	92.3	27.3	1005
Vineyards	39.3 ± 5.5	18.3	2.1	1.0	52.5
Orchards/Domestic Gardens	57.0 ± 14.6	21.0	1.5	0.5	27.1
Sum			210.4	73.6	3420

* ± 0.5 x AAD (%) : absolute average deviation from median

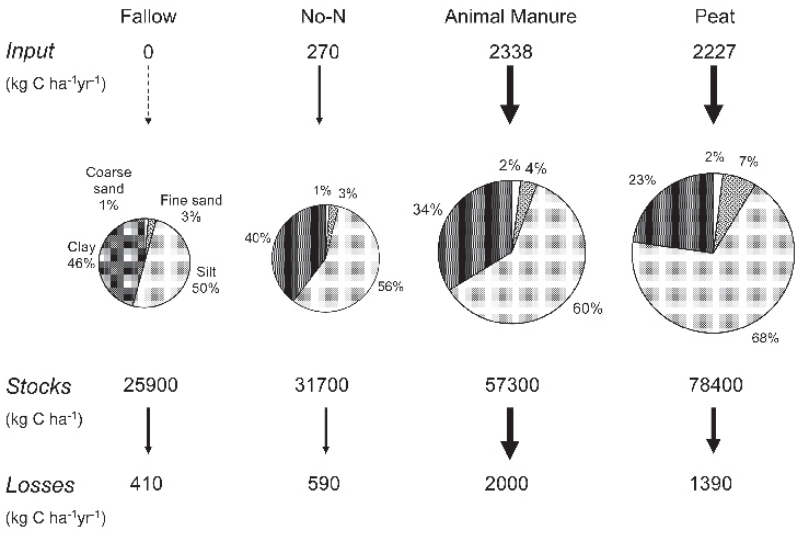


Fig. 2: Response of carbon pools to four selected treatments in the Ultuna experiment: Carbon input, C-partitioning in particle size fractions, C stocks and C-losses as observed until 1998 (Kirchmann et al. 2004); the aerea of the circles reflects the size of the carbon stocks in the 0-20 cm layer.

organic carbon, governed e. g. by the clay content, which influences the size of the physically protected organic matter.

A major factor influencing the C_{org} contents of agricultural soils is the application of organic residues that differ in their stability against decomposition. Figure 2 presents an example from the Ultuna long-term agricultural field experiment in

Sweden (Uppsala). This experiment was established in 1956 on a clay loam (Eutric Cambisol) and is based on application of equal amounts (app. 2000 kg C_{org} ha^{-1} yr^{-1}) yet different compositions of organic carbon. Therefore, changes in organic carbon contents in topsoil (0-20 cm) can directly be related to the stability of the applied organic matter. Figure 2 shows the development of C_{org} stocks in four selected treatments during the last 42 years. The bare fallow plot lost approximately one third of its initial C_{org} content due to mineralization. Root and stubble input (cereals, rape and fodder beet) in the non-nitrogen treated plot resulted in slightly higher C_{org} contents as compared to bare fallow. The third and fourth treatments involved adding organic materials of quite high stability, such as animal manure and peat that are less available for microbial degradation and mineralization and increased the C_{org} level considerably. Note that changes in humus contents are still ongoing and no equilibrium between C_{org} input and mineralization has been observed in the Ultuna experiment until now. According to Figure 2, most of the C_{org} introduced into the Ap-horizon was stored in the silt-sized fraction, and only little reached the clay fraction during the decomposition process. Relative contributions of C_{org} in silt increased from 52% in fallow plots to nearly 70% in peat plots, while C_{org} in clay-sized particles (2 – 0.1µm) decreased from 36% (fallow) to 19% (peat). The proportions of C_{org} derived from organic amendments decreased from sand to silt and clay fractions. Organic carbon in the latter fraction consisted of approximately one third amendment–C (animal manure, peat), whereas in sand-sized particles, most C_{org} originated from organic amendments (Gerzabek et al. 2001). The silt-sized fraction can therefore be regarded as a medium-term C-sink due to the build-up of stable microaggregates. The clay fraction contains the most stable and oldest C_{org} fraction (Buyanovsky et al. 1994). Other important stabilization mechanisms are the recalcitrance of the humified material itself and organo-mineral complexes between humified material and inorganic soil constituents like clay minerals and pedogenic oxides (Gerzabek et al. 2006).

Besides the input of more or less stabilized organic materials from plant residues and fertilizer input, soil tillage influences soil carbon stocks considerably. There is a significant variety in different tillage options, which are widely used. These include the traditional plowing, minimum tillage options and in between e.g. the use of cultivators. These measures have a distinct impact on the aeration of the topsoil and the water regime. Both factors influence significantly the mineralisation dynamics. In most cases, minimum tillage methods lead to an increase of soil organic matter stocks. From a long-term field experiment in Lower Austria (Fuchsenbigl) it can be concluded that already after 11 years of different tillage management a significant increase of organic carbon stocks were observed in the minimum tillage plots as compared with the plowing treatment (Palli et al. 1999). Revisiting the long-term experiment after another 5 years supporting the previously measured trends (Tatzber, pers. communication).

4 Conclusions

- Soil organic carbon is the largest pool within the terrestrial carbon cycle and about twice as large as the atmospheric reservoir; the fluxes between these pools exceed the annual fossil CO_2 emissions by about an order of magnitude. However, the imbalance produced by this additional input has led to a rapid increase of global carbon dioxide concentration in the atmosphere and thus to climate change.
- The feedback process to climate change induced temperature rise tends to be negative through increases in plant-derived carbon input to soils and positive due to warming-induced acceleration of decomposition. It is expected that positive feedbacks will dominate in the long term. However, quantification of temperature and climate sensitivity of currently constrained decomposition rates is still highly uncertain, as is the extent of the affected areas (e.g. permafrost and peatlands).
- Humus has manifold functions in soil and is one of the important soil parameters, which can be drastically influenced by human activity.
- Especially soil functions like the production function, the filter, buffer and transformation function and the gene reserve function is heavily influenced by humus contents and quality.
- Active management of humus levels include land use in general, tillage, crop rotation, fertilization and treatment with organic manures/residues. The major driving factors in this respect are the amount and quality of input of organic material (plant residues, organic fertilizers), soil aeration and soil moisture.

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