Soils as sources and sinks of greenhouse gases

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Abstract: Soils annually emit between 6.8 and 7.9 Gt CO_2 equivalents, mainly as CH_4 from intact peatlands and from rice agriculture; as N_2O from unmanaged and managed soils; and as CO_2 from land-use change. Methane emissions attributable to other wetlands add another 1.6–3.8 Gt CO_2 equivalents. From a global standpoint, N_2O from unmanaged soils and CH_4 from peatlands and other wetlands make soils naturally net greenhouse gas emitters. In addition, the storage of carbon in soils and the fluxes of CH_4 and N_2O have been changed by anthropogenic effects towards emission rates 52 to 72% above those under natural conditions before the dawn of intensive agriculture and land-use change. Land-use changes on mineral soils induced most of the recorded losses of soil organic matter (SOM), but there is evidence that proper agricultural management of soil resources is able to recover some of these losses and to maintain soil functions. However, the discrepancy between so-called 'sequestration potentials' and the measures already adopted is amazingly large. Globally, only about 5% of the cropped areas is managed according to practices such as no tillage or organic farming. The contribution of soil loss by erosion, desertification and sealing to global oxidative SOM losses is uncertain; however, in the case of soil erosion, it is considered to be a major factor in global SOM decline. Mitigation options calculated for SOM restoration, reduced CH_4 and N_2O emissions are able to alleviate mean annual emissions by 1.2 to 2.9 Gt CO_2 equivalents, mainly as a result of carbon sequestration, which is the most efficient measure for the next few decades. In the longer term, however, the large potential for reducing CH_4 and N_2O emissions outweigh the finite capacity of soils to recover C. Integrated assessment of net greenhouse-gas fluxes is key for evaluating management practices aimed at reducing overall emissions. From the viewpoint of climate change and taking into consideration the mean fluxes of CO_2, CH_4 and N_2O, peatland protection is more favourable than peatland cultivation in the long term. The most important gaps in our understanding appear to be with regard to estimating fluxes along with soil erosion and desertification processes, in the extent of peatland cultivation; the role of black carbon formation, natural 'background' sequestration rates of undisturbed soils; and the net response of soils, particularly in cold regions, to global warming. With regard to the societal perception of soil contributing to the global cycling of greenhouse gases, it is important to emphasize that significant proportions of the emissions are inevitably linked to intensive agriculture.

Atmospheric concentrations of the greenhouse gases (GHGs) carbon dioxide (CO_2), nitrous oxide (N_2O), and methane (CH_4) are increasing significantly due to human activity, accounting for much of the hypothesized additional anthropogenic greenhouse effect (IPCC 2001). Most of this increase is assigned to CO_2, and the relative importance of CH_4 and N_2O in terms of global warming is determined by the mean residence time of these gases in the atmosphere and their absorptivity for infrared radiation, resulting in calculated global-warming potentials of CH_4 and N_2O relative to CO_2 of 23 and 296 for a 100-year time horizon (kg^{-1}, IPCC 2001). From the radiative forcing of 2.43 W m^{-2} between 1850 and 1990, 1.46 W m^{-2}, 0.48 W m^{-2} and 0.15 W m^{-2} stem from the atmospheric increase in CO_2, CH_4 and N_2O,

respectively (IPCC 2001). The rest is attributed to halocarbons.

Soils are key elements in the global transformation, processing, and turnover of CO_2, N_2O and CH_4. They not only harbour micro-organisms as the main actors of various transformation processes, they also store organic and inorganic carbon and are regarded as a potential 'sink' for CO_2. Soils also provide the space for agricultural and other land-use activities as important factors driving the atmospheric increase in GHGs. Soils serve as both sources and sinks of the major GHGs – both naturally and under the influence of anthropogenic activities.

The goal of this paper is to review, how and to what extent the soil-related sources and sinks of GHGs have been perturbed historically by direct human-induced activities and what the

current challenges are. This is achieved by analysing (1) the most important changes in land-use and management with respect to soils and GHGs; (2) the extent of global soil loss and its impact on the cycles of CO_2; and (3) the potentials of emission reductions and sequestration. A brief introduction to the processes of GHG formation is given in the following section.

Production, consumption, storage and emission of greenhouse gases in soil

The processing and transformation of greenhouse gases are mostly microbially mediated, and the soil architecture is thus the living space in which these processes occur. Organic carbon can be regarded as a repository for carbon dioxide fixed via photosynthesis, and is part of the solid and liquid phase in the soil. Together with inorganic carbon derived from parent material or precipitated via desiccation of pore-water rich in Ca^{2+} and HCO_3^- during capillary rise in arid and semi-arid regions, carbon stored in soil is among the most important C pools on the global scale, exceeding that in vegetation and the atmosphere by a factor of two (Bolin & Sukumar 2000). In contrast, production, consumption and emission are the main processes related to nitrous oxide and methane, and storage of these gases in the soil pore space is limited.

Measurements of the soil's C balance are carried out mainly by (1) direct measurement of the soil organic carbon (SOC) content, bulk density, and stone content for a particular depth and at different points in time; (2) integrated net CO_2 exchange over larger areas (e.g. at the field level) by using Eddy-covariance techniques; and (3) various chamber techniques to quantify the CO_2 efflux from soil. The first technique gives a direct estimate of potential gains and losses; however, it requires large sample numbers, owing to the heterogeneity of SOC in the field and the comparable small rates of change (e.g. P. Smith 2004a). Eddy-covariance quantifies the bidirectional CO_2 flux over a field with high resolution in time, but lacks differentiation between autotrophic and heterotrophic processes, and hampers an assignment of fluxes to soil. It thus needs to be complemented by additional methods (e.g. Ehman et al. 2002). For measuring soil respiration as well as soil CH_4 and N_2O fluxes, chamber techniques are most frequently used. As for SOC inventories, CO_2 chamber measurements are plagued by high spatial – and in addition – also temporal vari-

ability of fluxes (e.g. Davidson et al. 2002). This variability is the cause of much of the error documented for chamber-based N_2O and CH_4 fluxes.

Soil organic matter (SOM)

This is the product of residue decomposition and consists of a variety of constituents, some of which have not yet been identified chemically. In undisturbed ecosystems, the soil carbon stock, just like the vegetation carbon stock, is considered to be almost in a steady state between inputs (e.g. as plant residues) and outputs (e.g. as CO_2 from respiration) of carbon. This does include episodic fluctuations (e.g. flooding, storms, vegetation fires), whose effect on the carbon stock can still be described by a steady-state model if the corresponding time span is allowed to be long enough to capture those fluctuations (e.g. Körner 2003). With respect to soil C, a steady state can also be reached in managed or otherwise altered ecosystems if the strength of the disturbing force (e.g. frequency and intensity of tillage) remains constant. Under such conditions, the system adapts to the new situation (this often takes decades to centuries) until inputs again equal outputs. Under steady-state conditions, the relationship between SOC stock, input, output, and turnover is:

$$\text{C stock (t ha}^{-1}) = \text{input rate (t ha}^{-1}\text{ a}^{-1})$$
$$\times \text{ turnover time (a) and}$$
$$\text{Input rate (t ha}^{-1}\text{ a}^{-1}) = \text{output rate (t ha}^{-1}\text{ a}^{-1})$$

Evidently, a change in the C stock is attributed to (1) a change in the input or output rate (e.g. by changing the rate of residue input in agriculture or changing the output by erosion) or (2) a change in the turnover time (e.g. by changing tillage practices, which affects aeration and moisture as important abiotic forces influencing microbial activity). A significant increase in the C stock over time, namely carbon sequestration, is always associated with a non-steady-state system, as is the significant decline of the C stock. The reason that increased residue inputs do not infinitely increase the C stock is given by the soil's limited potential to store C. The state of maximum C storage is termed 'saturation', and the difference between the steady state and saturation with respect to soil C storage has been addressed by Six et al. (2002). The main reasons why SOM is accumulating in soil until it reaches a steady state can be explained by the ability of a mineral soil to protect organic materials against biochemical attack by several

physico-chemical mechanisms, together with the selective and relative enrichment of substrates that have a limited availability for micro-organisms (Sollins *et al.* 1996). In combination, these mechanisms probably cause an energetically unfavourable situation, where the costs for the production of enzymes and locally low substrate concentrations offset the benefit for the organisms. An additional stabilization pathway involves vegetation fires and the accompanying formation of charred products (black carbon, BC), which probably contribute to the accumulation of more resistant organic materials. Charcoal has been shown to be much more resistant to microbial degradation compared to non-charred precursors (Baldock & Smernik 2002).

In soils exposed to periodic or permanent waterlogging or water saturation, and in the interior of soil aggregates, oxygen deficiency and the accompanying decelerated decomposition due to thermodynamically more unfavourable reaction mechanisms, may cause accumulation of SOM, which, in the case of organic soils, may equal thousands of tons per hectare. The agricultural use of peatlands is almost unavoidably associated with a loss in soil C, because oxygen deficiency caused by water saturation is the major mechanism leading to the accumulation of peat, and its direction is reversed by drainage and subsequent peat oxidation. Hence, rates of oxidative peat decomposition can be affected mainly by regulation of the water table (Augustin *et al.* 1996; Kasimir-Klemedtsson *et al.* 1997).

Nitrous oxide

This is an intermediate product of two main opposing processes: denitrification and nitrification. It occurs naturally in almost all ecosystems. In the enzymatic reduction chain of nitrate to nitrogen, nitrous oxide is produced together with NO and N_2 as the two other products of denitrification by various micro-organisms. A reaction of oxygen as an electron acceptor with an organic or inorganic electron source has a more negative ΔG, and is thus thermodynamically more favourable than a reaction with nitrate. Therefore, denitrification particularly occurs under anoxic conditions. However, nitrous oxide emissions have also been measured in aerated soils (Müller *et al.* 2004). These emissions are probably caused by anoxic conditions in the micropores that have been caused by limited oxygen diffusion or high oxygen consumption rates. Nitrification is the oxidation of ammonia to nitrate along a redox

gradient that converts nitrogen from the N(–III) to the N(+V) state, and is favoured under oxic conditions. Like denitrification, nitrification is associated with both chemo-autotrophic and heterotrophic organisms (Paul & Clark 1996). Because the reaction rate is proportional to the concentration of the reactants, the availability of either nitrate or ammonia together with organic matter for heterotrophic or CO_2 and oxygen for autotrophic transformations mainly determines the production rate of nitrous oxide. The efflux from soil is further modified by the diffusivity and the diffusive path length, because uptake of nitrous oxide in the upper few centimetres derived from deeper horizons can be quantitatively important (Neftel *et al.* 2000). Not surprisingly, nitrous oxide emission rates from soils typically peak after fertilization events (i.e. when there is a surplus of mineral N), and after rainfall (increasing oxygen deficiency and increasing microbial activity; Ball *et al.* 1999; Simek *et al.* 2004).

Methane

Methane in soils is a product of biochemical reduction processes and can typically be found in anoxic waterlogged ecosystems having redox potentials below –200 mV, e.g. floodplains, peatlands, marshes, and in lowland rice-cropping areas. Because of their oxygen limitation and the accompanying accumulation of organic debris, wetland soils producing methane are often rich in organic matter. Methane is produced by methanogenic bacteria along two different pathways: acetoclastic methanogenesis, and H_2/CO_2 methanogenesis. During acidogenesis, anaerobic decomposition of organic matter produces intermediates such as low-molecular-weight fatty acids, acetate, H_2 and CO_2. In sulphate-rich sediments (e.g. marshes), sulphate serves as the final electron acceptor, and the organic material is finally mineralized to CO_2 and H_2S. If sulphate is absent (either consumed or naturally in low abundance, e.g. in freshwater systems), the intermediates of anaerobic decomposition are processed further by methanogenesis. During acetoclastic methanogenesis, acetate is converted to CO_2 and CH_4, and during H_2/CO_2 methanogenesis, H_2 and CO_2 react to form CH_4 and H_2O. The latter pathway is also responsible for CH_4 emissions from ruminants.

If this methane enters oxic soil horizons with higher redox potentials, it can be oxidized to CO_2 and water by a number of bacteria, so its warming impact is reduced to a large extent. Because of this important transformation, it is

worth taking a closer look at the transport mechanism of methane in soils and plants. Methane from soil reaches the atmosphere by three different pathways: ebullition, diffusion, and transport in the aerenchyma of vascular plants. Only in the case of diffusion are the transport rates slow enough to allow a significant proportion of the methane to be oxidized before it enters the atmosphere, while, during convection in the case of gas bubbles or gas transport in the plant aerenchyma, the rate of decrease relative to that of transportation is considered to be small. Oxidation is also an important process of methane consumption in aerobic soils. Its rate is highly dependent on climatic conditions and soil management and is greatest under undisturbed forest, for still-unknown reasons (Robertson & Grace 2004).

Land-use change and agricultural intensification, and their impact on carbon dioxide fluxes

Land-use changes

Land-use changes are important driving forces for changes in the soil's state and they lead to many of the important threats to global soil resources, such as loss of organic matter,

erosion, sealing or physical deterioration (e.g. Van-Camp et al. 2004).

The conversion from forest to agricultural land began with the onset of the Neolithic age around 7000 years BP and became far more noticeable after the arrival of Bronze-age ploughs between 6000 and 5000 years BP (Ruddiman 2003, and references cited therein). Yet by 2000 years BP, agricultural land-use had spread throughout large areas of Eurasia, North Africa, Eastern China, India, Central- and South America (Roberts 1998). Along with the rapid increase in the human population during the nineteeth and twentieth centuries, the agricultural use of soil has increased steadily. FAO estimates for the changes in the area of land under different uses, during the 1960 to 1990s are given in Figure 1. The area occupied by agricultural land has expanded from 4.51 Gha (1 Gha = 10^9 ha) in 1961 to 5.01 in 2000, and the forest area has been reduced by approx. 0.2 Gha from 4.37 to 4.17 Gha between 1961 and 1994 (FAOSTAT data 2004). Most pronounced is the increase in permanent pastures by 0.33 Gha between 1961 and 2000. At the same time, the arable area has expanded by c. 0.13 Gha. The expansion of arable land in developing countries from 2006 to 2030 is estimated to increase by 25% to 28% for rain-fed and irrigated crop production, respectively (FAO 2003),
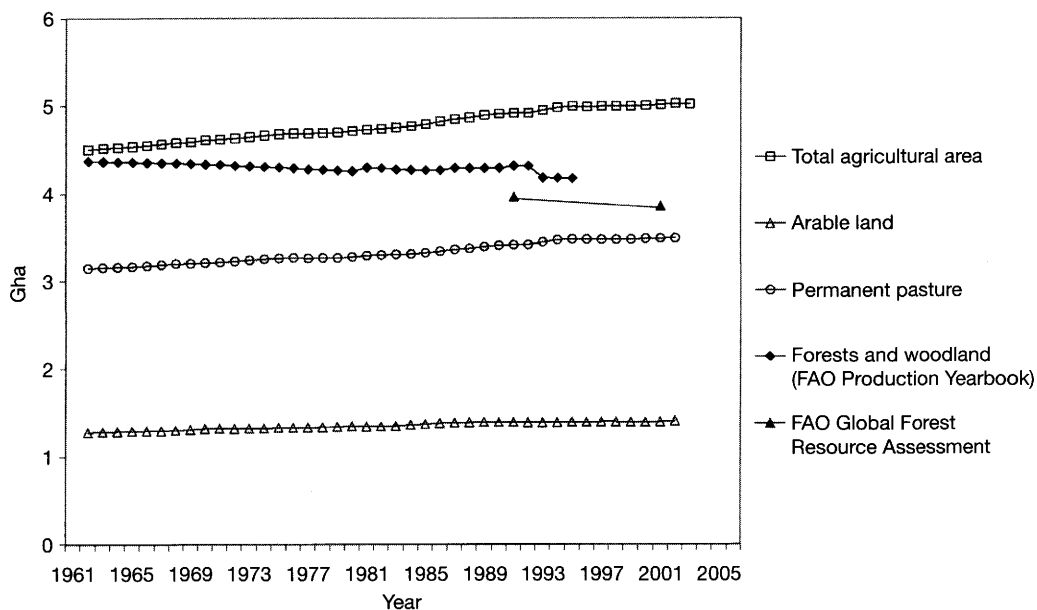
Fig. 1. Major world land-use categories from 1961 to 2004.
Source: FAOSTAT data (2004).

with maize being the most important single contributor to this land-use change. According to the FAO Global Forest Resource Assessment (FRA 2000), the forest area decreased from 3.96 to 3.87 Gha between 1990 and 2000. Differences in forest cover estimates are caused by the methodology used (see FRA 2000); however, the trend is consistent. Recent deforestation during the 1990s occurred in the tropical regions (net annual loss 12.3 Mha), while the temperate regions gained forest area (+2.9 Mha) by natural expansion, afforestation and reforestation. This represents a general pattern during the twentieth century, where the geographical distribution of land-use change shifted towards the subhumid and humid tropics, accompanied by the still ongoing clearance of large areas of tropical forests and a concomitant increase in tropical pastures (Houghton 2003). Agroforestry, a management system that integrates trees into farmland, is included in the FAO categories 'arable land' and 'forest land', and accounts for approximately 0.4 Gha, mostly allocated in the tropics.

Conversion of native grasslands to cropland, managed pastures or managed forest plantations is another important element of global land-use change. Grasslands (tropical savannas and temperate grasslands, both natural and anthropogenic, without semi-deserts) cover approx. 3.5 Gha, or 23% of the global land area (Bolin & Sukumar 2000). For example, grassland conversion has been documented for current land-uses in comparison to the former, potential natural prairie vegetation for large parts of the Central and Western US (Sobecki *et al*. 2001); for the semi-arid grasslands in China (Zhao *et al*. 2005); or for South American grasslands (Farley *et al*. 2004).

Cultivation of wetland soils, and of peatland soils in particular, is of major importance for the terrestrial C budget, although its contribution by area is rather small. The estimated total area of peatland soils is 254–480 Mha (Gorham 1991; Paustian *et al*. 1997; WBGU 1998; Spiers 1999; Moore 2002). According to Spiers (1999) and Moore (2002), the vast majority of peatlands are

allocated in the boreal and sub-arctic regions of Canada, Alaska and Russia, in particular. Estimates of the area of tropical peatland range from 30.6 to 45.9 Mha, of which the majority is located in Indonesia (17–27 Mha), Malaysia (1.5 Mha) and China (1–3 Mha) (van Dam *et al*. 2002). More recent estimates on peatland extension for West Siberia (Sheng *et al*. 2004) and Indonesia (Page *et al*. 2002) indicate that the total area of peatlands might even be higher. Thus, peatlands are the most important single category of freshwater wetlands, the latter covering approximately 530 to 570 Mha (Spiers 1999), including swamps as a category similar to peatlands. The share of peatlands drained or used for peat-cutting has been estimated to be 11.5 Mha and 4.4 Mha, respectively (Gorham 1991), and 39 Mha for both drained and cut peatlands by Paustian *et al*. (1997). OECD (1996) figures on the drainage of wetlands in 1985 are 26% worldwide (56–65% of the available wetlands had been drained for intensive agriculture in Europe and North America; the figures for tropical and subtropical regions were 27% for Asia, 6% for South America and 2% for Africa). With respect to an estimated total area of 860 Mha of wetlands in the OECD study and the figures on peatland areas amounting to approximately 400 Mha, it is possible that the drainage and/or cutting of peatlands exceeds the values of 15.9–39 Mha cited above.

Land-use changes and impacts on soil-derived CO_2 fluxes

Losses and gains of SOC are closely linked to the bulk CO_2 flux attributable to land-use change. For the period 1850–1980, the annual net CO_2 flux from land-use change was almost identical for the tropics (tropical Asia, America, Africa) and the non-tropics (Canada, US, Europe, former Soviet Union, China Pacific, North Africa and the Middle East); however, the ratio of CO_2 tropics to CO_2 non-tropics has changed considerably during the 1980s and 1990s (Houghton 2003; Table 1). For the 1980s

Table 1. *Estimates of the main annual sources (+) and sinks (−) of carbon resulting from different types of land-use change and management, since 1850*

Region	1850–1980 (Gt C)	1980–1989 (Gt C a^{-1})	1990–1999 (Gt C a^{-1})
Tropical areas	+56.7	+1.93	+2.20
Other areas (non-tropics)	+57.6	+0.06	−0.02
Source contribution from the tropics (%)	50	97	100

Based on Houghton (2003).

and 1990s, annual emissions from land-use change have been estimated to be 7.3 ± 2.9 and 8.1 ± 2.9 Gt CO_2, respectively. During the period 1850–2000, global CO_2 emissions from land-use change were estimated to be 572 ± 202 Gt CO_2 (Houghton 2003) or 3.8 ± 1.4 Gt CO_2 annually – 87% of which is derived from forest areas and approximately 13% from cultivation of mid-latitude grasslands (Bolin & Sukumar 2000, and references cited therein). In addition to these more recent developments, Ruddiman (2003) also estimated the C lost by the early forest clearances since the Neolithic age. This estimate of 822–914 Gt CO_2 even exceeds the loss induced by land-use changes since 1850, but its magnitude has been questioned (Joos et al. 2004).

What are the implications of these land-use changes for the soil's function as a source and sink of CO_2? The global amount of SOC (from 0 to 100 cm depth) has been estimated to be 1200–1600 Gt (Batjes 1996; Paustian et al. 1997; Jobbagy & Jackson 2000). Based on the difference between precultivated and current SOC stocks in cultivated soils, Paustian et al. (1997) estimate a global loss of 40–60 Gt due to cultivation, from which the major part (approximately 80%) is derived from aerated mineral soils and the remainder (approximately 11 Gt) is from the drainage of 'wetland soils', that is, Histosols and Gleysols (FAO). While the SOC loss estimated so far only accounts for a small fraction of the global SOC stock (2.5 to 6%), it contributes one-third to the estimated C loss of 156 ± 55 Gt caused by land-use change since 1850. Two-thirds are attributable mainly to deforestation.

In many parts of the world, forest fires (whether ignited by humans or caused naturally) are taken as the cue for land-use change to agricultural or pastoral uses. While the CO_2 released by vegetation burning has been considered for global estimates (Houghton 2003), the net effects on the soil's C balance are hardly ever addressed. Potentially, the formation of black carbon (BC) is another quantitatively important budget component accompanying vegetation fires. Its net effect on the global C budget is still poorly understood. The annual rate of BC formation has been estimated to account for 0.05–0.2 Gt C, 80% of which has been assigned to vegetation fires (Kuhlbusch 1998). Most of these fires are human-induced and can also contribute to large CO_2 emissions from burning peatlands (e.g. Page et al. 2002). Black carbon is assumed to represent between 1 and 6% of the total SOC, with much higher concentrations in regions affected by frequent vegetation fires (Gonzalez-Perez et al. 2004).

Estimates of the amount of soil organic carbon stored in peatlands vary widely between 225 Gt C (643 t ha^{-1}; IPCC 2000), 43.6 Gt C (1120 t ha^{-1}; Paustian et al. 1997, only for artificially drained peatlands), 455 Gt C (1300 t ha^{-1}; recalculated from Moore 2002, only for temperate peatlands), and 541 Gt C (Bergkamp & Orlando 1999; tropical peatlands 1700–2880 t ha^{-1}, boreal and temperate peatlands 1314–1315 t ha^{-1}). Data from K. A. Smith et al. (2004) for the West Siberian Lowland indicate mean C stocks of approximately 1200 t ha^{-1}. For Swiss temperate fens, Leifeld et al. (2005) estimated a mean C stock of 1600 t ha^{-1} for intact peatlands, and 710–1050 t ha^{-1} for peatlands drained several decades ago. On the basis of an assumed global C stock in peatlands of between 225 and 541 Gt, the loss of 11 Gt is equivalent to 2–4.9% of this reservoir. In some regions, the actual loss induced by peatland cultivation and peat excavation might significantly exceed this percentage. Considering that the previous area of intact peatlands in Central Europe of 5.9 Mha has been reduced to a current value of 0.29 Mha (Succow & Joosten 2001), and assuming C stocks of between 1300 and 1600 t ha^{-1} for intact – and of between 710 and 1120 t ha^{-1} for cultivated – peatlands, the historical peatland cultivation in Central Europe alone accounts for a C loss of 0.68–3.9 Gt C, or 7–51% of the former peatland C stock.

To attribute the SOC loss discussed above to particular regions or land-use types is challenging, chiefly because of the regional imbalance in available soil data. In temperate regions, C losses resulting from the conversion to agriculture have been documented frequently (e.g. Cihacek & Ulmer 1997; W. N. Smith et al. 1997; Lal et al. 1998; Mikhailova et al. 2000). As pointed out above, tropical forest clearance and conversion to pasture is currently the quantitatively most important land-use change globally. This land-use change is accompanied by a strong reduction in biomass C, but not necessarily in soil C. In a review of 115 studies from 300 datapoints worldwide, Conant et al. (2001) found significant evidence that, during the conversion from tropical forest to pasture, the soil carbon stock increases more often than it decreases. A similar result has been reported by Guo & Gifford (2002), who reported significant increases in the SOC stock after conversion from forest to pasture for many paired plots, particularly in regions with high precipitation. For the Ecuadorian Andes, Farley et al. (2004) reported a significant decline in SOC in the A

horizon of pine plantations for all age classes (five to 25 years) after conversion of the native tussock grasses, which were not cleared or burned before first planting. Results on increasing SOC after tropical forest clearance and pasture establishment, or decreasing SOC after the introduction of plantations do not exclude a possible decline in SOM after forest clearance (e.g. Tiessen *et al.* 2003), but indicate that, under proper management, SOC stocks can at least be preserved in many cases. Agroforestry is a suitable alternative that can reduce the negative effects on SOC caused by slash-and-burn agriculture followed by the introduction of arable rotations, particularly in tropical regions. Relative to undisturbed forests, agroforestry systems are able to maintain 80 to 100% of the former C stock and may also improve the productivity of degraded tropical pastures (IPCC 2000).

Precipitation is a major driver for both plant productivity and SOC (Jobbagy & Jackson 2000). It also significantly affects the rate and direction of SOC change when land use is altered. In a comparison of six adjacent plant communities used either as grassland or where shrub/woodland encroachment was allowed, along a precipitation gradient in the South-Western USA, Jackson *et al.* (2002) showed that there was a significant negative linear relationship between the change in soil C and the annual precipitation. For sites with annual precipitation above 300 mm, woody encroachment reduced SOC contents (up to 50% of the initial value at the Engeling site with 1070 mm annual rainfall), while, for the dry sites, SOC slightly increased. These data emphasize the need to account for co-variables, when effects of land-use change on SOC stocks are evaluated.

Soil degradation and sealing and its impact on the global soil C storage

Severe soil degradation by desertification, erosion or sealing has been recognized as an important threat to the soil reserve (Van-Camp *et al.* 2004); however, it has been less intensively discussed in terms of the soil's role as a source and a sink of GHGs. In contrast to the land-use issues addressed above, soil loss has an almost irrevocable negative effect on SOC storage. Oldeman *et al.* (1991) estimated the extent of human-induced soil degradation for the world. Out of four categories of soil degradation, strong degradation (i.e. terrain not reclaimable at farm level), and extreme degradation (i.e. irreclaimable terrain destruction beyond

restoration) as topsoil loss induced by water and wind erosion, together account for 170.6 Mha. The three main causative factors for soil degradation, in decreasing order of magnitude, are overgrazing, deforestation and agricultural mismanagement. More recent global data are missing, however, UNEP (2002) states that, despite trends towards soil conservation and resource management, there is no clear indication that the rate of land degradation has decreased. Desertification refers to land degradation in arid, semi-arid and subhumid areas due to anthropogenic activities (UNEP 1993). It can be exacerbated by long, continuous dry periods (Nicholson 2001). Eswaran *et al.* (2001) estimated the total area vulnerable to desertification to be 4.3 Gha, of which 0.79 Gha are classified as very highly vulnerable. Desertification not only leads to severe reductions in land productivity of up to 50% (Eswaran *et al.* 2001), but, like erosion in humid regions, it leads to a decline in the soil-profile thickness and thus to a decline in its potential to store C.

The impact of soil erosion on the global carbon budget has recently been discussed by Lal (2003). He estimated that 4.0–4.6 Gt C are eroded annually by water erosion, 0.8–1.2 Gt of which is emitted as CO_2 to the atmosphere. The major fraction is redistributed over the landscape and, to a minor extent, buried in depositional sites of terrestrial and aquatic ecosystems. The corresponding area prone to erosion is 1.09 Gha for water erosion – 0.75 Gha of which are severely affected, and 0.55 Gha for wind erosion – 0.30 Gha of which are severely affected (Lal 2003). However, an estimate of the C lost by wind erosion is not provided. According to these figures, of the approximately 13.6 Gha total land area, 12% is affected by erosion in general, including 8% by severe erosion. A simulation on global soil dust emissions estimated the total dust loading to be 1.92 Gt, of which only a minor fraction of 0.12 Gt was attributed to agricultural areas, mostly in semi-arid regions of Africa, continental Asia and Australia (Tegen *et al.* 2004). On the basis of mean SOC concentrations of roughly 1% for soils in semi-arid regions (Batjes 1996, 1999), the total mass of SOC transported by wind erosion becomes 19.2 Mt, with an anthropogenic contribution of approximately 1.2 Mt C. This indicates the very small share of wind to global SOC erosion. For an area prone to severe wind erosion of 0.55 Gha, a global soil dust emission of approximately 1.92 Gt yields an annual soil loss by wind erosion of 3.5 t ha^{-1}, corresponding to 0.035 t SOC ha^{-1} a^{-1}. As for water erosion, the fate of the eroded material in

terms of its global contribution to the CO_2 budget is largely unknown. In addition to the mass loss by erosion and the accompanying reduction in topsoil thickness, the remaining soil is often relatively enriched in sand (Zhao *et al.* 2005), which reduces further plant growth and its future capacity to protect plant residues entering the soil.

Soil sealing by urbanization and road construction can also be regarded as affecting the SOC storage capabilities of soils. Although sealed areas are considered to increase world-wide, owing to changes in population, habits and economic structures, reliable figures on their extent are sparse. Recent global estimates on the extent of urban and built-up areas range from 25.6 to 39.5 Mha (WRI 2005), but probably underestimate the existing expansion. Germany and Switzerland are considered here as examples to compare the values of WRI (2005) with the corresponding data on urbanization as provided by the national authorities of these countries (BFS 2001; BMU 2002). While the WRI (2005) estimates that 0.027 and 0.84 Mha fall under the category 'urban and built-up' for Switzerland and Germany, respectively, the corresponding national data are 0.26 (including areas not sealed but assigned to urbanization, settlement and road construction) and 2.1 Mha (sealed areas).

A conservative estimate, taking only areas into account that are severely threatened or already lost for future C sequestration owing to strong and extreme erosion as well as to sealing, gives a total global value of 196–210 Mha. Soil loss and the concurrent decline in SOC caused by the processes discussed above has undoubted consequences for many soil functions at the affected sites, including the soil's function as a carbon reservoir and a CO_2 buffer. The situation is more complex when the global CO_2 budget is to be assessed. Lal (2003) points out that the fate of the eroded material with respect to its potential contribution to atmospheric CO_2 is largely unknown, and cites ranges from 0% to 100% for the oxidizable fraction. The same holds for SOC stored in soils that are excavated and displaced along with construction activities. Although this material is lost as a possible accumulator for additional C, its original C will most probably not be fully oxidized. After its displacement, soil may be deposited on to existing surface horizons, at depressional sites as colluvium; in river floodplains, estuaries and deltas; and, finally, on continental shelves. Many of these geomorphological landscapes provide ecological conditions that are less favourable for decomposition compared to the place of origin,

and may thus serve as a carbon sink in the medium term. Even under favourable conditions, only a fraction of the SOC is readily decomposable (Kiem *et al.* 2000). With respect to the global CO_2 balance, the amount of soil displaced and the fate of SOC as a part of it can be regarded as largely unknown and are thus a subject of high priority for research.

Measures to combat soil degradation and increase C sequestration

With increasing awareness of soil degradation, decline and, more recently, the role of soils in the global cycle of greenhouse gases, agricultural policy has begun to support measures for soil conservation and soil fertility improvement. These measures include several conservation tillage practices; conversion of erodible land from active crop production to permanent vegetative cover or to agroforestry; adapted crop rotations and adapted grassland management; and improved fertilization, as well as organic farming practices (Lal 2004*a*). They may already have led to a deceleration in SOC loss and a reduction in erosion, and a partial regeneration of SOC stocks in many parts of North America, South America and Europe, and are considered to contribute to soil carbon sequestration (e.g. West and Post 2002; Lal 2004*a*; Holland 2004; P. Smith 2004*b*). Global estimates on the potential to sequester C in soil by means of reclaiming degraded land, range from 16 to 78 Gt (Paustian *et al.* 1997; Batjes 1999; Lal 2004*b*).

Batjes (1999) distinguished various scenarios for C sequestration for degraded soils, based on the classification given by Oldeman *et al.* (1991). He found sequestration potentials between 14 ± 7 Gt C (restoration of degraded agricultural land and improved management of arable land) and 20 ± 10 Gt C (additional restoration of degraded extensive grasslands, forest regrowth on degraded land) to be realistic. It is worth noting that these sequestration measures have the potential to recover substantial amounts of the 40–60 Gt estimated as human-induced losses in SOC as a result of cultivation and land-use changes. Although the importance of soil C sequestration as a tool to mitigate global warming should not be neglected, the attainable capacity of the soil as a C sink is most likely to be smaller than the historical losses induced by human activity (Lal 2004*b*). The effectiveness of no tillage and other soil conservation practices also depends on their continuous application, because even a single ploughing event may lead to the loss of all additionally

sequestered carbon (Stockfisch *et al.* 1999), and probably more importantly, to the mineralization of the sequestered organic N.

Conservation agriculture includes any practice which reduces, changes or eliminates soil tillage and avoids burning of residues, thus maintaining sufficient surface residue throughout the year (ECAF 2005). Zero or minimum tillage and direct seeding are important elements of conservation agriculture. The global extent of no tillage has been estimated to be 58 Mha (FAO 2001). The area under other soil-conserving practices for arable land is probably much larger. A comparison of the European figures (ECAF 2005) for no tillage (0.96 Mha which correspond well with the data in FAO 2001) with the European figures on total conservation tillage (10 Mha) indicates that, on a global level, the area under conservation tillage is also likely to exceed the 58 Mha given by FAO (2001). Organic farming has also been assumed to contribute to soil conservation (e.g. Shepherd *et al.* 2002). Studies comparing management effects of conventional v. organic farming often reveal a positive effect of organic farming on SOM storage (Condron *et al.* 2000; Pulleman *et al.* 2000; Pulleman *et al.* 2003), mainly because of a higher share of leys and grain legumes in arable rotations, and higher rates of manure application. These management practices seem to balance out the higher mechanical disturbance that is often needed in organic systems for the purpose of mechanical weeding. Differences in SOM might be smaller when compared to integrated farming systems with similar rotations and less mechanical destruction of the soil's structure. The area under organic farming was estimated to be 24.1 Mha in 2003, and 26.5 Mha in 2005 (Yussefi & Willer 2004; IFOAM 2005). This area adds to the area under conservation tillage practices, because these two systems typically are mutually exclusive. Conversion of marginal land to permanent vegetative cover in the US is estimated to be

c. 14.8 Mha (Lal *et al.* 1998). The underlying Conservation Reserve Program (CRP) has not only led to a decline in soil erosion, but also to increases in SOC (Gebhart *et al.* 1994). Agroforestry has been practised traditionally in the humid tropics, and is increasingly being adopted to prevent soil erosion and improve soil fertility (Oelbermann *et al.* 2004). While those benefits, together with higher biomass C stocks, have been reported frequently, the effects on SOC are not consistent. The net effect will depend not only on crop productivity, and thus residue return, but on competition for water and nutrients between arable crops and trees, and on the particular residue management (Montagnini & Nair 2004).

Table 2 is a compilation of the most important data discussed above. Assuming that soil loss mainly occurs in agriculture, strong and extreme erosion concerns 3.5% of that area. In addition, 0.51–0.78% of agricultural soils have been destroyed by urbanization, which adds up to a total of *c.* 4% of the world's agricultural soils that have been lost or which are severely threatened by human interference. These areas are no longer available for storing biogenic C, thus burdening the remaining area with respect to C storage potentials. In contrast, 5.5% of the arable land (1.7% of the agricultural area) is affected by measures that actively contribute to the protection of soil and soil C. Adding the area under CRP in the US reveals that 2.0% of the agricultural area is given active protection. These data, although still a fragmentary approximation, show the gap between potentials for SOC sequestration and protection, and the current situation.

Agricultural intensification and nitrous oxide emissions

The key to understanding global N_2O emissions from soil is the close relationship between the

Table 2. *Some key figures on the global use of soils (Mha)*

| Agricultural areas		Areas degraded by erosion		Urban and built-up areas	Soil-conservation areas in cropping systems	
Arable crops*	Pastures	Strong[†]	Extreme[†]		No tillage	Organic farming
1530	3490	170.6	4.7	25.6–39.5	58	26.5

*Includes 130 Mha permanent crops. [†]For definitions, see the text.
Sources: Land use: FAOSTAT data (2004); erosion: Oldeman *et al.* (1991); urban and built-up areas: WRI (2005); no tillage: FAO (2001); organic farming: Yussefi & Willer (2004) and IFOAM (2005).

amount of N cycled in terrestrial ecosystems and the amount of N_2O released from them. The factors that affect nitrous oxide emission from the soil are partially reflected in the IPCC guidelines for the calculation of GHG emissions from soil (IPCC 1996), whereby input rates of N into the soil by organic and mineral fertilizers and by fixation and plant residues are multiplied by the corresponding emission factors, in order to calculate the nitrous oxide emissions. Important emission factors for N_2O, according to the IPCC (1996) are $1.25 \pm 1\%$ for fertilization, 0.5–3% for the storage of manure, 0–1% for the storage of slurry, and 0.2–12% for leached nitrate-N (the percentage of the applied mineral or organic N for fertilization, percentage of stored manure or slurry-N, and the percentage of leached N, respectively).

Under natural or semi-natural conditions, many terrestrial ecosystems, particularly in temperate regions, are N-limited, and losses to adjacent compartments, such as to groundwater or the atmosphere, are small (e.g. Asner *et al.* 1997). Nitrogen in natural or semi-natural terrestrial systems mainly originates from N fixation and from lightning. Before the disseminating agricultural cultivation of legumes, terrestrial organisms fixed *c.* 90–140 Mt N per year, and an additional 5–10 Mt originated from lightning (Vitousek *et al.* 1997). With the advent of commercial ammonia production using the Haber–Bosch process and the selective breeding of N-fixing crops, global amounts of N entering the soil have increased significantly. Vitousek *et al.* (1997) estimated the annual fixation by agricultural crops to be 32–53 Mt. The production of N fertilizers peaked in 1996 (90 Mt N); 87 Mt were produced in 2002 (FAOSTAT 2004). Figure 2 illustrates the chronological progression of the global N production and the yields of major N-fixing crops. Between 1961 and 2002, fertilizer-N production increased by a factor of 6.7, soybean production by a factor of 7.7, and the production of other pulses by a factor of 1.6. Dividing the annual fertilizer-N production by the arable cropping area (Fig. 1) yields a calculated N application of 10 kg N ha^{-1} a^{-1} in 1961, and 62 kg N ha^{-1} a^{-1} in 2002. These basic figures indicate a dramatic increase in N_2O emissions from managed soils during the twentieth century. Global estimates of fertilizer-induced N_2O emissions from agricultural soils are highly uncertain and range from 0.6–14.8 Mt N_2O–N a^{-1} (IPCC 2001). Emissions from natural soils, including background emissions from agricultural soils, were estimated to be 6.8 Mt N_2O–N a^{-1} (range 3.3–9.9; IPCC 2001; based on Kreileman & Bouwman, 1994). Some of the N_2O from 'natural' soils may be already derived from anthropogenic atmospheric deposition.
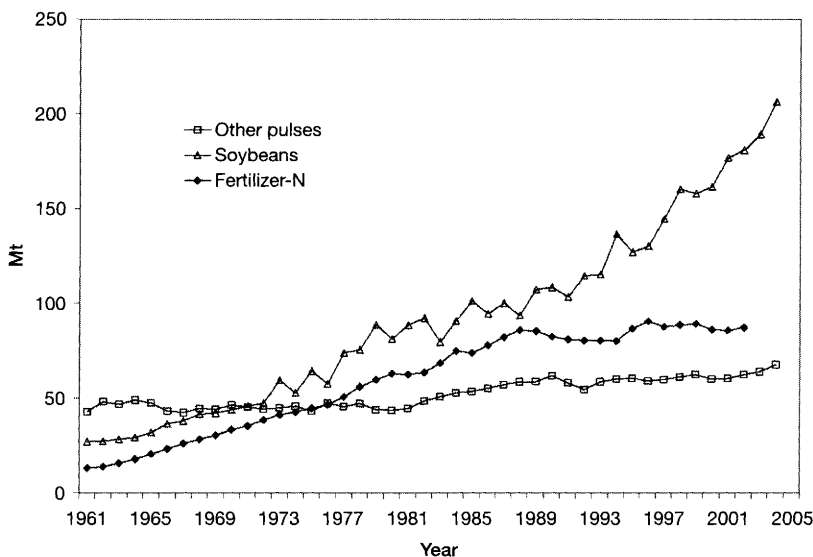
Fig. 2. World production of mineral N fertilizer and selected N-fixing crops from 1961 to 2004. Other pulses include beans, peas, chickpeas, cowpeas, pigeon-peas, lentils, bambara beans, vetches and lupins. *Source*: FAOSTAT data (2004).

The frequently applied emission factors for N fertilization (IPCC 1996; based on Bouwman 1994) for the calculation of national GHG emissions, imply a linear relationship between the amount of N added to the soil and the amount of N_2O released from it. In contrast to the IPCC (1996) data, more recent studies suggest non-linear responses between N_2O emission and N application and include additional factors to estimate the release of N_2O (Bouwman *et al.* 2002; Flynn *et al.* 2005), which may lead to higher or lower emission estimates, depending on management and environmental conditions. By using an empirical modelling approach, which included 846 N_2O measurements from all over the world, Bouwman *et al.* (2002) estimated that the annual emission from fertilized agricultural fields (including those where manure was applied) to be 2.8 Mt N_2O–N. Kaiser & Ruser (2000) evaluated six long-term plots in Germany and confirmed the range of emission factors as proposed by the IPCC (1996); however, they did not find any significant correlation between N application and N_2O emission, and 40% of their observed variability in N_2O emissions remained unexplained by Bouwman's model (1994). Kaiser & Ruser assigned the residual variability to crop- and site-specific effects on N_2O emission. In this study, approximately 50% of the measured emission occurred during winter, thus emphasizing the need for long-term measurements. Bouwman *et al.* (2002) showed that N_2O emissions, irrespective of climate, soil type and fertilizer type, were not linearly, but disproportionately related to the amount of applied N. Significant factors for the prediction of N_2O emissions were the N application rate, as well as fertilizer type, climate, SOC content, soil texture, drainage status, soil pH, and crop type. For global assessments, three findings from their study are of particular importance: (1) mean fertilizer-induced emissions were lower than the values proposed by IPCC (1996), for both mineral N and manure N; (2) grasslands had lower emissions than croplands, apart from poorly drained soils where emissions may exceed those from crop cultivation on well-drained soils; and (3) N_2O emissions from tropical and subtropical soils exceeded those from temperate soils. Hall & Matson (1999) discussed N_2O emissions after N additions in tropical forests on the Hawaiian Islands. They found higher N_2O emissions for P-limited than for N-limited forests, and they cautiously assigned this response to the inability of the P-limited system to retain much of the anthropogenic N. Phosphorus limitation is also a common property of many agricultural tropical soils, because of the prevalence of Fe and Al oxides/hydroxides and allophane with a high sorption capacity for phosphate, particularly at low pH values. This chemical property, in conjunction with higher temperatures in the subtropical and tropical regions, might contribute to the higher N_2O emissions per unit applied N, as found by Bouwman *et al.* (2002).

Draining organic soils typically causes a tremendous increase in N_2O emissions. Zeitz & Velty (2002) reported emissions of 0.0–0.8 and 0.3–26.9 kg N_2O–N ha^{-1} a^{-1} for wet and drained fens in Germany, respectively. Augustin *et al.* (1996) compiled data from Finland and the US, showing N_2O emissions from undrained fens to be 0.4–0.5 kg N_2O–N ha^{-1} a^{-1}, and 1.2–13.1 N_2O–N ha^{-1} a^{-1} after drainage. Similarly, Kasimir-Klemedtsson *et al.* (1997) reviewed studies from Finland, Sweden and The Netherlands that provided evidence for an increase in N_2O emissions from zero to 5.7 ± 3.2 (drained, grassland) and 9.6 ± 7 kg N_2O–N ha^{-1} a^{-1} (drained, cereals), respectively.

In view of the increasing rate of additional N input via fertilizers and biological N fixation, global N_2O emissions from soils have probably increased sharply during the last five decades. Projections for the use of fertilizers are based on the assumption of an increase in fertilizer consumption, but at a lower annual rate of approximately 1% than in the past, due to improvements in nutrient use efficiency and a projected slowdown of increases in crop production (FAO 2003). From 2006 to 2030, increases in fertilizer use in the industrial countries, especially in Western Europe, are expected to lag significantly behind increases in other regions of the world, because of changes in agricultural policies. Thus, more detailed knowledge on the controlling factors for N_2O emission in subtropical and tropical agricultural ecosystems is needed in order to identify appropriate mitigation options. The FAO report (2003) did not distinguish between different chemical elements; however, N fertilizers, as the single most important nutrient, will probably be used in proportion to this projection. In contrast to FAO (2003), an extrapolation of past trends in fertilizer use projects a much higher consumption. Tilman *et al.* (2001) predicted a fertilizer-N application of 135 Mt for 2020 and 236 Mt for 2050, corresponding to annual increases of 2.8% and 3.4% relative to 2000, respectively. Their estimate, based on linear regressions between fertilizer consumption and time, global population, and global gross domestic product (GDP) as independent variables, is higher than

that of the FAO (2003), mainly because of high projections when using GDP as a predictor for N consumption. Extrapolations using population growth roughly halve the projected increase by 2050, to 1.8% annually.

Mitigation strategies should be geared to site- and management-specific demands and cannot be generalized easily. Dinitrogen monoxide is an unwanted excess product of N fertilization and, like ammonia and nitrate, can be decreased by appropriate mitigation. Since a further global increase in N fertilization and biological N fixation is probable, an increase in the N utilization efficiency is a key factor not only to reduce N_2O emission, but environmentally harmful N emissions in general. The global N utilization efficiency (i.e. N output from harvest and live-stock, divided by N inputs) was approximately 50% in 1996 (Mosier 2002). For German agri-culture, Schweigert & van der Ploeg (2002) calculated the annual N surplus from the differ-ence between the N input by commercial fertil-izers and imported fodder, and the N output by animal and crop products for the period from 1951 to 2000. This surplus increased from c. 10 kg ha^{-1} in the 1950s to 120 kg ha^{-1} to 70–80 kg ha^{-1} in the 1990s. For farms in England, Leach et al. (2004) reported N surpluses of up to 250 kg ha^{-1} a^{-1} for pig/arable and dairy farms, while N surpluses on arable farms without animal husbandry were 100 kg ha^{-1} a^{-1}. The corresponding N utilization efficiency was 50% for arable farms, while that for animal husbandry (beef/sheep or dairy) was below or close to 20%.

A number of management practices have been identified that can reduce the amounts of fertilizer used and improve the efficiency rate. Among these, timing multiple applications rather than single ones during a growing season, avoiding application of manure/slurry during bare fallow seasons, a balanced fertiliza-tion (i.e. considering possible growth limi-tations induced by shortages in other nutrients), dissolution of N fertilizer in irriga-tion water, application below the soil surface, and measures subsumed as precision farming (i.e. N application depending on actual crop demands as measured by on-site fluorescence, and on soil characteristics and yield potentials) are important ones (Vitousek et al. 1997; Khosla et al. 2002; Mosier 2002). Like recom-mendations for the increase in N utilization efficiency, management practices to reduce N_2O emissions have been assessed (Dalal et al. 2003). These measures include matching the supply of applied N to the spatial and temporal

requirements of the crops and pastures, split-ting fertilizer applications; balancing N with other nutrient supplies; and using cover crops to take up residual mineral N from applications to the preceding crop or from mineralization of N fixed during legume or ley phases. Sehy et al. (2003) showed that site-specific N fertilization within a single field resulted in a 34% decrease of N_2O emissions, while maintaining crop yields. In the area on their field where site-specific differences in soil water constrained yields, higher N inputs were not able to increase crop yields.

The adoption of nitrification inhibitors is a chemical tool used to reduce N_2O emissions and to improve the efficiency of fertilizer utilization. Nitrification inhibitors are chemicals that reduce the rate at which ammonium is converted to nitrate, by interfering with the metabolism of nitrifying bacteria. The loss of N from the root zone can be reduced by maintain-ing applied N in the ammonium form during periods of excess rainfall prior to N uptake by crops. Nitrification inhibitors were shown to reduce N_2O emissions by 60% and 42% when applied together with cattle slurry and calcium ammonium nitrate, respectively (Merino et al. 2002).

Carbon and nitrogen sequestered by reducing the soil tillage or by land-use changes require long-term conservation measures to avoid the release and reflux of the elements to the atmos-phere and other ecosystem compartments. Ploughing of grassland soil and cultivation of bare soils leads to temporarily increased N_2O emissions (Baggs et al. 2000). Increased nitrogen mineralization induced by tillage is typically more difficult to manage because microbes are also active during phases when bare soil is exposed or when there is or reduced uptake of nitrogen by plants.

Methane: implications of peatland dynamics and rice agriculture

Natural wetlands and rice agriculture are among the most important single sources of the global CH_4 flux. From the c. 500–600 Mt CH_4 emitted annually worldwide, between 203 and 337 Mt are derived from wetland and rice soils (IPCC 2001). Estimates on the amount of CH_4 emitted from wetlands alone range from 115 to 237 Mt CH_4 a^{-1} (IPCC 2001, and references therein). On the other hand, aerobic soils oxidize c. 29 Mt CH_4 annually (Smith et al. 2000). The major quantitative process of atmospheric CH_4

consumption is its reaction with OH radicals in the troposphere (IPCC 2001).

Methane emissions from peatlands which are not drained, mined or otherwise used agriculturally vary widely, depending on the water level and oxygen supply, nutrient availability and temperature. Bergkamp & Orlando (1999) estimated that natural wetlands in boreal and temperate regions emit $0.11–0.20$ t CH_4 ha^{-1} a^{-1}, and tropical wetlands $0.35–0.37$ t CH_4 ha^{-1} a^{-1}. Similar CH_4 emission rates of $0.1–0.28$ t CH_4 ha^{-1} a^{-1} were reported by Kasimir-Klemedtsson et al. (1997) for Finland and Sweden. Augustin et al. (1996) reviewed studies from bogs and fens in Sweden, Finland, Germany and the US (temperate zone) and reported CH_4 emissions to range from 0.01 to 2.4 t CH_4 ha^{-1} a^{-1}. In their study, CH_4 emissions from bogs (i.e. ombrotrophic peatlands) were always higher than those from fens (minerotrophic peatlands). In Finland, Martikainen et al. (1996) measured fluxes of $0.15–0.45$ t CH_4 ha^{-1} a^{-1} for a series of natural bogs and fens, and Nykänen et al. (1996) reported fluxes of $0.15–0.26$ t CH_4 ha^{-1} a^{-1} for two different years at yet another site. Values for wetlands, including other types than peatlands, were modelled by Cao et al. (1998). They estimated the mean CH_4 emissions to be 0.15, 0.55 and 0.71 t CH_4 ha^{-1} a^{-1} for northern, temperate and tropical wetlands, respectively, considering a total wetland area of 504 Mha. The value for this area lies below the range for freshwater wetlands given by Spiers (1999), so it is assumed that freshwater wetlands, including peatlands, comprise the area modelled by Cao et al. (1998). The total emission was estimated to be 92.3 Mt CH_4 a^{-1}. If the emission from peatlands is assumed to be proportional to that of all wetlands considered by Cao et al. (1998), the global contribution by peatlands would become $46.5–73.3$ Mt CH_4 a^{-1} for the above-mentioned peatland area of 254–400 Mha. Smith et al. (2004) suggest a CH_4 emission attributable to the northern hemisphere peatlands of c. 60 Mt CH_4 a^{-1}, based on interpolar (i.e. Greenland/Antarctic) methane gradient data from ice cores and assumptions on global distributions of CH_4 sources and sinks. This value lies well within the estimate given above, keeping in mind that more than 90% of the world's peatland area is located in the northern hemisphere. Based on the data of Cao et al. (1998), a mean calculated emission rate of 0.18 t CH_4 ha^{-1} a^{-1} can be derived, which is within the range of the studies from field trials cited above. It must be noted, however, that

measured ranges of CH_4 emissions from peatlands, even at the same site but under different conditions, can vary by two orders of magnitude. Similar to aerobic soils, natural peatlands may also oxidize some atmospheric CH_4 during drier periods; however, this flux is small relative to the annual release (Nykänen et al. 1996).

Along with the drainage and cultivation of natural peatlands for agricultural purposes or peat mining, CH_4 fluxes, in general, are drastically declining, while CO_2 fluxes change from a small sink to a strong source. From those studies cited above that use paired fields to distinguish between fluxes of natural v. drained conditions, it can be concluded that lowering the water table by only a few decimetres reduces CH_4 emissions in most cases by more than 90%, or even converts the peatland from a net source to a net sink of CH_4. On the other hand, CO_2 release resulting from peat oxidation can reach levels exceeding those from mineral soils by one to two orders of magnitude. An emission range of $3.7–70$ t CO_2 ha^{-1} a^{-1} is given by Bergkamp & Orlando (1999) for the agricultural use of wetlands in boreal/temperate regions. Freibauer & Kaltschmitt (2001) estimated CO_2 fluxes from drained organic soils to be 10 ± 5 t CO_2 ha^{-1} a^{-1} for leys and grasslands, and 15 ± 5 t for arable crops. The authors stress that their estimate is applicable to farming on organic soils, and not for the conversion of pristine peat soils to farmed soils. For temperate regions in Switzerland, Leifeld et al. (2005) estimated an annual loss of 35 ± 8 t CO_2 ha^{-1} by peat oxidation from organic soils used for agriculture. Compared to the CO_2 emission from drained peatlands, C accumulation by peat growth in intact peatlands is small. Gorham (1991) attributed a mean annual accumulation of c. 0.23 t C ha^{-1} to northern peatlands, or 0.076 Gt C globally for an approximated peatland area of 346 Mha.

Global CH_4 emissions from rice agriculture amount to approximately 60 Mt CH_4 (FAO 2003). This value lies within the range of estimates compiled by IPCC (2001), but is higher than the range of 25–54 Mt CH_4 given by Sass et al. (1999). Rice agriculture increased almost steadily from 115 Mha in 1961 to 153 Mha in 2004 (it was 86 Mha in 1935) and increasingly contributed to the world's total agricultural area (2.56% in 1961, 2.94% in 2002). It is expected to increase to 164 Mha by 2030 (FAO 2003). Most of this area is wetland rice (around 85%), which is considered to be the main source of CH_4 from rice agriculture (IPCC 1996). In contrast to the pronounced uncertainty with regard to global

CH_4 emission from rice agriculture, various case studies address the drivers of CH_4 emission from wetland rice, and possible reduction strategies. Among the manageable factors known to affect CH_4 emission from rice fields, two are frequently cited to be of major importance:

(1) *Water regime/irrigation status.* For continuously flooded fields, IPCC (1996) gives a seasonally integrated emission factor (i.e. cumulative emission over the growing season) of 12–28 g CH_4 m^{-2}. Because CH_4 production is strongly related to the redox conditions (Yu & Patrick 2004), the maximum default emission factor under continuous flooding is reduced by intermittently flooding, or when water depths are below 0.5 m (IPCC 1996). In a study evaluating the effect of rice field management on the composite effect of N_2O and CH_4 emissions, Yu et al. (2004) noted the highest CH_4 emissions reductions, of more than 70%, for non-flooded but wet soil conditions; however, they observed that some of this reduction was offset by increased N_2O emissions. Park & Yun (2002) discussed measured CH_4 emissions at various sites in Korea. They showed that the mean emission factors were 0.24 g CH_4 m^{-2} d^{-1} for continuous flooding and 0.15 g CH_4 m^{-2} d^{-1} for intermittent flooding in situations where no organic matter was added. Emissions under direct seeding on dry paddies were 40–50% lower than those in transplanted rice. Integrated CH_4 emissions factors in transplanting cultures were 27 g CH_4 m^{-2} for early-maturing rice, but 35 g CH_4 m^{-2} for late varieties. During the Methane Asia Campaign, Mitra et al. (2002) revealed higher integrated emission factors of 5–29 g CH_4 m^{-2} and 22–57 g CH_4 m^{-2} for fields with continuous flooding, but with different SOC contents, while the corresponding ranges for intermittently flooded fields were 0.06–3 g CH_4 m^{-2} and 0.6–24 g CH_4 m^{-2}, respectively. Emission reductions of >70% for non-flooding but wet conditions relative to flooding were reported by Yu et al. (2004). In a modelling study for India, Pathak et al. (2005) derived annual net emissions of 1.07–1.10 Mt CH_4–C for continuous flooding of 42.25 Mha rice fields, and 0.12–0.13 Mt for the same area but under intermittent flooding. Under conditions of continuous irrigation, the depth of the water table also has a pronounced effect on CH_4 emissions (Liu & Wu 2004).

(2) *Organic matter status/organic matter addition.* In most of the above-mentioned studies, SOM content and organic matter application were included as additional factors. Yu et al. (2004) reported an emission reduction of 57% for flooded fields without amendment, relative to the application of 30 t ha^{-1} a^{-1} organic manure. Liu & Wu (2004) proposed the removal of the rice straw (i.e. reducing the addition of organic matter) from the first harvest, before planting the second crop, thus reducing the CH_4-emission by one order of magnitude. Significantly higher emissions were reported by Kimura et al. (2004), when senescent rice leaves at a quantity naturally occurring in the field were left in the field, in comparison to their removal. Mitra et al. (2002) showed that soils with lower SOC contents (<0.7%) emitted 5–29 g CH_4 m^{-2} and 0.06–3 g CH_4 m^{-2} under continuous and intermittent flooding, respectively, and soils with a higher SOC content (>0.7%) emitted 22–57 g CH_4 m^{-2} and 0.6–24 g CH_4 m^{-2} under continuous and intermittent flooding during the growing season, respectively. In the study of Park & Yun (2002), the amendment of 5 t rice straw ha^{-1} in autumn increased the emission from 0.24 to 0.33 g CH_4 m^{-2} d^{-1} (continuous flooding), and from 0.15 to 0.24 g CH_4 m^{-2} d^{-1} (intermittent flooding).

## Integrated assessment of greenhouse gas fluxes from soil

System analysis

An integrated assessment of any anthropogenic use of soil should account for all of the soil functions, and this equally applies to soils as sources and sinks of GHGs. Measures implemented to sequester carbon in soil often have benefits for other soil functions as well; however, their gross effect may be counterbalanced by changes in the exchange rate of N_2O and CH_4. Integrated assessments of GHG balances have been discussed by Robertson et al. (2000) and Robertson & Grace (2004), for example, for a maize/soybean/wheat cropping system in the US Midwest, and for a rice/wheat/cowpea cropping system in India. For the US system, annual crop rotations with three practices (conventional tillage, no tillage, organic without mineral N but with legumes in the rotation) were compared together with perennial systems (alfalfa and

poplar) and forest successional communities (on a time-scale of 10–50 years). Net global-warming potentials (GWP) were positive (i.e. net GHG source) for any of the annual systems, and neutral-to-highly negative for perennial crops and successional communities. They showed C sequestration to be of major importance for no tillage, perennial crops and early successions, and N_2O emission to be highest in the annual crops and for alfalfa. Pronounced CH_4 oxidation occurred mainly in late successional forests. For the Indian rice system, the net GWP was one order of magnitude greater, mainly due to CH_4 emissions from rice. Both studies indicated marked differences in net GWP, which must be known before management strategies aimed at reducing GHG emissions are implemented. Six *et al.* (2004) compiled data from soil-derived GHG emission comparisons between conventional and no-tillage strategies for humid- and dry-temperate climates, mainly from the northern hemisphere. They found the net GWP to be highly time-dependent and net GHG mitigation to occur under no tillage in humid regions only when practised for more than 10 years. During the first years, no tillage led to high N_2O emissions, and to a net loss of soil C in dry climates. A net C sequestration could only be ascertained in the long term (20 years). Cumulative soil-related GHG effects over 20 years indicated that no tillage outperformed conventional tillage only

in humid climates. They concluded that N_2O emissions drive much of the trend in net GWP, suggesting that improved nitrogen management is essential for realizing the full benefit of C sequestration. Further illuminative comments on the net CO_2 balance of some mitigation measures can be found in Schlesinger (1999).

Net GWP has been compared only rarely for peatlands under natural v. cultivated conditions. It has been shown that GHG fluxes in peatlands are typically much higher than those of mineral soils; that pristine peatlands are important emitters of CH_4; and that they respond dynamically to drainage and cultivation measures. To assess the atmospheric impact of natural v. drained peatlands, a rough estimate of the integrated GHG fluxes of natural v. drained peatlands is provided. This calculation, based on the data discussed above, illustrates the directional flux under two different conditions: natural and managed. For CO_2 fluxes, I assume a sink for intact peatlands of 0.84 t CO_2 ha^{-1} a^{-1}, and a mean source of 37 t CO_2 ha^{-1} a^{-1} under drainage, with an exponential rise of the emission to maximum over time. Methane emissions from the intact peatland are estimated to be 0.18 t CH_4 ha^{-1} a^{-1}, and are estimated to be zero after cultivation. Nitrous oxide emissions are assumed to be 0.5 kg N_2O–N ha^{-1} a^{-1} for intact peatland, and 5 kg N_2O–N ha^{-1} a^{-1} for cultivated peatland at the beginning of drainage. Dinitrogen monoxide emission during drainage
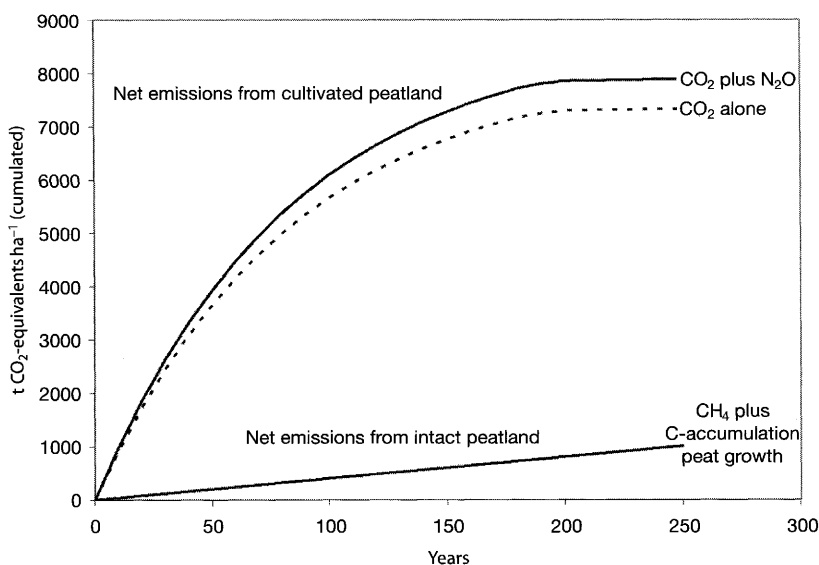
Fig. 3. Cumulative greenhouse gas emissions of natural and cultivated peatlands (CO_2-equivalents ha^{-1} a^{-1}), estimated by means of typical flux rates. For details, see the text.

J. LEIFELD

is considered to behave proportionally to CO_2 (i.e. it declines along with the decline in annual CO_2 emission rates) because C and N mineralization are tightly coupled. It is also assumed that the intact peatland stores 2000 t C ha^{-1} (peat thickness 2 m; Batjes 1996), so that oxidative peat consumption ceases after 200 years. The results of this calculation are given in Figure 3, adopting global-warming potentials for a 100-year horizon of 23 for CH_4 and 296 for N_2O (IPCC 2001). The curves clearly indicate that, over the selected time horizon, which is determined by the duration of peat decomposition, cultivation is highly unfavourable compared to leaving the peatland in its natural state. Methane emissions would need to accumulate for 1700 years to cause similar levels of global warming to those of the drained peatland over 200 years. Within the selected time horizon, annual CH_4 emission rates of approximately 1.59 t ha^{-1} would have a similar global-warming effect to that of the drained peatland. Such rates have been reported sporadically (Augustin et al. 1996), but are considered to lie at the upper limit.

Global estimates

Global estimates of the annual fluxes of CO_2, CH_4 and N_2O from soils can be addressed tentatively with the data discussed in the previous paragraphs. This summary adopts the 40–60 Gt CO_2–C lost by cultivation, 11 Gt of which are delivered from peatland drainage (Paustian et al. 1997), assuming that these changes occurred exclusively and steadily since 1850 (i.e. over a period of 150 years). Estimates of CH_4 emis-

sions from rice agriculture (currently 60 Mt), wetlands (115–237 Mt), from peatlands as a fraction of wetlands (46.5–73.3 Mt CH_4), of N_2O emissions from agricultural soils (2.8 Mt N_2O–N) and from natural soils (6.8 Mt N_2O–N) are taken from the data compiled above. Emission savings by peatland cultivation (39 Mha cultivated $\times$ 0.18 t CH_4 ha^{-1} a^{-1}) are calculated based on the data published by Paustian et al. (1997) and on the emission rate used to calculate the net flux of peatlands above. A value of 29 Mt CH_4 for global soil CH_4 oxidation is taken from Smith et al. (2000), and C sequestration by peat growth in intact peatlands from Gorham (1991). The annual fluxes are summarized in Table 3. Methane emissions from intact peatlands and other wetlands and N_2O from natural soils are the most important single sources, followed by CH_4 from rice agriculture; N_2O from agricultural soils; and CO_2 from land-use change. The latter category is an estimated mean for the period since 1850, and fluxes may have been distinctly above or below the calculated range at any one point in time. If, for example, the soil CO_2-flux is considered to be proportional to the total CO_2-flux from land-use changes and land management, annual rates may have been higher since the 1950s than before (see Houghton 2003 for comparison). The data imply that, on a global scale, soils are net GHG sources, and that this source function has been accelerated by human perturbation. Under natural conditions, soils emitted 4.86 to 7.67 Gt CO_2-equivalents (2.65 to 5.45 Gt from wetlands minus 0.67 Gt from CH_4 oxidation minus 0.28 Gt by peat growth, plus 3.16 Gt

Table 3. *Estimated global annual greenhouse gas fluxes for important soil-related sources (Mt CO_2 equivalents)*

CO_2 from land-use change		CO_2 uptake by peat growth	CH_4 emission savings[1]	CH_4 rice agriculture	CH_4 intact peatlands (all wetlands)	CH_4 oxidation[4]	N_2O from soils	
Mineral soils	Peatland drainage						N_2O agricultural soils[5]	Natural soils[6]
710–1200	269	−279	−162	1380	1070–1690[2] (2645–5451)[3]	−667	1302	3160

[1] Originating from peatland drainage.
[2] Range only for intact peatlands.
[3] Range for all wetlands.
[4] By aerobic soils.
[5] Fertilizer-induced.
[6] Including background emission from agricultural soils.
Based on Gorham (1991); Paustian et al. (1997); Cao et al. (1998); WBGU (1998); Spiers (1999); L. C. Smith et al. (2000); IPCC (2001); Bouwman et al. (2002); Moore (2002); FAO (2003). For the method of calculation, see the text. Negative values indicate sinks.

as N_2O. Human-induced emissions are 0.71–1.20 Gt CO_2 from land-use change, plus 0.27 Gt CO_2 from peatland drainage, plus 1.38 Gt as methane from rice agriculture, plus 1.30 Gt as nitrous oxide from agricultural fertilization, minus 0.16 Gt methane emission savings deriving from peatland drainage. According to this rough estimate, emissions, including wetlands other than peatlands, have increased from 4.86–7.67 Gt to 8.36–11.65 Gt CO_2-equivalents, namely by 51 to 84%. Total anthropogenic GHG emissions are approximately 33 Gt CO_2-equivalents per year (including CH_4 and N_2O from agriculture; WRI 2005). Annual soil fluxes due to land-use change and agriculture account for 3.50–3.99 Gt CO_2 equivalents of the total, and are thus a significant contribution to the anthropogenic source.

Not included in this examination are the naturally low sequestration rates that occur in soil along with profile development. Schlesinger (1990) indicated long-term rates of carbon accumulation of c. 2.4 g C m^{-2} a^{-1}, resulting in a global annual accumulation of 0.32 Gt C, or 0.40 Gt C if peat growth is also included. He points out that these rates probably overestimate actual ones, because the data include initial phases of soil developments after glacier retreat or volcanic eruptions, which typically have higher rates than older soils. He also stresses that such a global accumulation is consistent with global estimates on organic carbon transported by rivers. Hedges et al. (1997) give similar estimates on the global amount of organic carbon discharge to the sea (0.25 Gt dissolved C and 0.15 Gt particulate C). They conclude that much of this material is from forests and is soil-borne, and that there is growing evidence for rapid and remarkable extensive mineralization of terrestrial organic matter in the ocean. Together, these findings imply that, once the soil is in a steady state, much of the organic matter entering it will be discharged and subsequently mineralized in the ocean unless not oxidized in $situ$.

Climate change is expected to affect the element balance of soils. Decreasing C stocks of temperate soils have been discussed already in the context of climate change (Bellamy et al. 2005), referring to potential CO_2 sources and atmosphere–plant–soil interactions beyond the scope of this chapter. Under conditions of warming, arctic and boreal regions are vulnerable to thawing. While thawing of permafrost soils seems possible (Waelbroeck et al. 1997), there is quite a lot of uncertainty about the direction and magnitude of potential warming impacts and feedback loops because of the

many interacting forces (Stokstad 2004). For example, thawing the permafrost potentially increases decomposition of stored C and thus may turn a system from a net sink to a net source, while on the other hand it may increase net primary production and thus enhance C inputs. Effects on both, CO_2 and CH_4 emissions are also difficult to predict because they depend strongly on the hydrological situation.

Mitigation potentials

A tentative attempt is made to discuss potentials for the reduction of global greenhouse-gas emissions from soils on the basis of three of the major strategies cited above. This approach does not claim to be complete, but is intended to stimulate discussions in terms of quantity and longevity of abatement. The three strategies are:

(1) Carbon sequestration in soil of 10–30 Gt CO_2–C, according to Batjes (1999), for a timeframe of 50 years until a new steady state is reached (i.e. 0.73 to 2.2 Gt CO_2 a^{-1}), with an exponential rise to that point over time;

(2) Annual reductions of N_2O emissions from managed soils by 10% and 20% relative to the current level of c. 2.8 Mt N_2O–N (Bouwman et al. 2002) taking into consideration an underlying annual increase in N fertilization of 1%, according to the FAO (2003), and assuming that global N_2O emissions are proportional to N fertilizer application. An annual reduction of 10–20% seems feasible in view of the much higher reduction reported in the literature;

(3) Annual reduction of CH_4 emissions from rice agriculture by 20 and 30% from current values of 60 Mt CH_4 and considering an increase in area to 164 Mha by 2030 (FAO 2003). The attainable annual reduction is small relative to that reported in most of the studies for measures of water and organic matter management, but it was chosen to account for the higher N_2O emissions reported in certain cases.

The results of these strategies are summarized in Figure 4. Over the next 50 years, C sequestration will be the most effective measure, fixing between 37 and 110 Gt CO_2. Cumulative emission savings are 8.4–16.8 Gt CO_2-equivalents for improved management of N_2O, and 13–20 Gt CO_2-equivalents for the reduction of CH_4 from rice agriculture. For a scenario with small C sequestration rates, combined N_2O plus CH_4 savings are in the same order of magnitude as C sequestration after 50 years. Mean annual
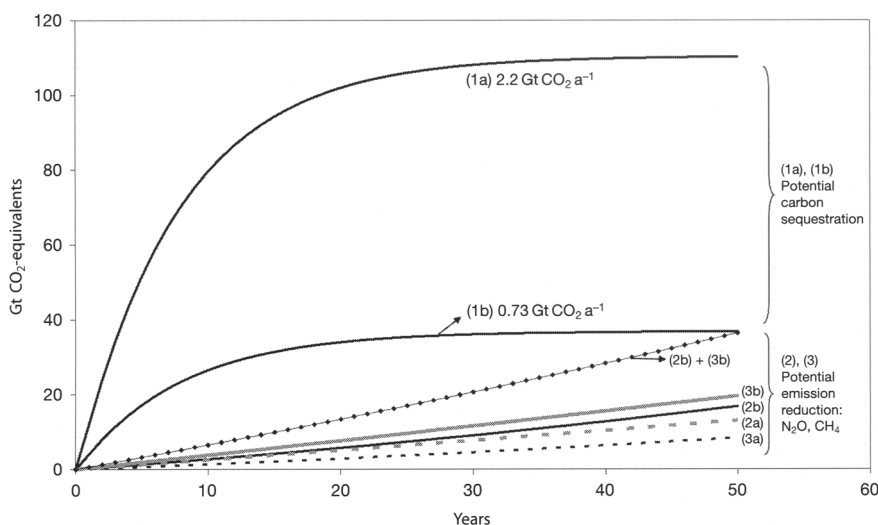
Fig. 4. Cumulative global CO_2 mitigation potentials of soils. (**1a**), (**2b**) C sequestration; (**2a**) reduction in N_2O emissions by 10%; (**2b**) reduction in N_2O emissions by 20%; (**3a**) reduction in CH_4 emissions 10%; (**3b**) reduction in CH_4 emissions by 30% (Gt CO_2 equivalents). For further explanations, see the text.

emission savings (i.e. CO_2 plus N_2O plus CH_4) amount to 1.2–2.9 Gt CO_2-equivalents, which is a significant share of the anthropogenic emission of 3.5–4.0 Gt CO_2-equivalents attributed to soils. Considering that conservation tillage is one of the measures contributing to the C sequestration capacity, it must be noted that the potential for N_2O emission reduction will probably be smaller because of higher N_2O emissions under no tillage (Six *et al.* 2004). While the pattern for these three strategies seems obvious for the next few decades, the relative contribution of emission savings from N_2O and CH_4 will increase over time, because of the finite potential of soils to store C. Long-term investigations should thus specify the interaction between C-sequestration and fluxes of the other GHGs, and should focus on emission savings of the non-CO_2-GHGs. Compared to soil C sequestration, their cumulative potential is infinite.

Peatland management has not been included in the discussions of mitigation potentials, because data on the area of cultivated peatlands and their future projection are highly uncertain. However, the integrated assessment of GHG fluxes in Figure 3 has shown that it would be reasonable to protect intact peatlands, not only in view of hydrological and ecological functions, but also with respect to global-warming mitigation. Cultivation of 1% of the actual peatland area would imply additional mean annual emissions (emissions at the beginning of cultivation are supposed to be higher) of 96–147 Mt CO_2,

and a reduction in CH_4 emissions of only 11–17 Mt CO_2-equivalents.

Conclusions

On the global scale, soils are net sources of GHGs under natural conditions as well as under conditions of land-use change and agricultural intensification, particularly with respect to the large amounts of N_2O and CH_4 that they emit. Greenhouse-gas fluxes from soils are important budget components in the terrestrial elemental cycles, and are also significant if the overall anthropogenic emissions are considered. Human activity has increased soil-derived fluxes, and it is likely that, despite mitigation potentials yet to be implemented, fluxes will remain above the pre-agricultural level. This is mainly induced by the elevated CH_4 and N_2O emissions, while proper agricultural land management is qualified to maintain and restore SOM levels in many cases. There are enormous uncertainties in estimates related to (1) fluxes, particularly for CH_4 from intact and for CO_2 from drained peatlands, and (2) areas, particularly of cultivated peatlands. There are also gaps in our understanding of the fate of SOC transferred by erosion, sealing and desertification, as well as in the role of black carbon formation. Beyond what has been addressed in this paper, the responses of soil-related GHG fluxes to climatic changes are crucial for the future role of soils in the global element cycles.

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