Chapter 7
Accounting More Precisely for Peat and Other Soil Carbon Resources

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Abstract In the context of "recarbonization", it is important to know where the soil C stocks are located and how much of these are prone to emission to the atmosphere. While it may appear to be a trivial question considering available global estimates and maps, yet there is a strong need to emphasize that erroneous estimates are made in assessing the global soil C stocks. Without doubt, peatlands hold the single most important soil C stock at the global scale, and these soils are mostly located in the northern latitudes between 50°N and 70°N. However, there are additional wetlands or other ecosystems which also hold potentially relevant amounts of soil C stocks. From the soil science perspective, it implies that there

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are other hydromorphic soils, besides Histosols and potentially other soil types, also containing relevant amounts of soil C stock. Differences in scientific approaches, which include terminology, definitions, depth to which soil C is considered, and bulk density, etc., lead to different estimates of soil C stocks. Recent estimates indicate that peatlands cover only 3% of the global land surface but contain 40% of the soil C stocks to 1-m depth. Consequently, only small differences in the estimate of the land coverage lead to great differences in the soil C stock estimates. Typically peatlands, wetlands and other ecosystems rich in soil C, cover only small parts of the landscapes, and yet are not easily accounted for by any inventory or mapping attempts. With estimates presented in this chapter, hydromorphic soils, aside Histosols, add 10% soil C stock to the estimates of peatland's Histosols. Additionally, non hydromorphic Podzols add another 10% to the soil C stock. Above all, soils from the steppe biome must also be considered. The soil C stock of Cryosols (frozen soil C not separated from peatlands) contain as much as 1,500 Pg C, which is as much C as the total stock estimated in world soils to 1-m depth. Thus, coordinated and substantial efforts are needed to improve the mapping of ecosystems, particularly of those which are rich in soil C stocks. One option is to improve remote sensing techniques for wetlands. These efforts must be undertaken quickly because soil C stocks are being depleted not only by the positive feedback with the climate system but also directly by land use change. The conversion of peatlands to agricultural and forestry uses is not sustainable because of the depletion C stocks, and especially not for conversion of peatlands for "biofuels" production.

Keywords Hydromorphic soils • Histosols • Fluvisols • Gleysols • Planosols • Chernozems • Phaeozem • Kastanozem • Greyzem • Podzols • Podzols • Aerobic decomposition • Anaerobic decomposition • Turnover rate • Wetlands • Methane • Soil carbon budget • Seasonally inundated • Mire • Marsh • Swamp • Fen • Bog • Temperate peatland • Tropical peatland • Biofuel • Oil palm • Biodiesel • Sphagnum • Peatland conversion • Ethanol • Prairie • Methane • Water table • Permafrost • CO_2-equivalent • Peatland distribution • Land cover • Land use change • Abiotic • Anoxic sites • Carbon sequestration

Abbreviations

C	carbon
DOC	dissolved organic C
GCC	global carbon cycle
GLS	Global land cover
GLCC	Global land cover characteristics
GHGs	greenhouse gases
LCCS	land cover classification system
Mha	million ha

OM	organic matter
SOC	soil organic carbon
SOM	soil organic matter
GWP	global warming potential

7.1 Introduction

There are myriads of articles and book chapters on the importance of peat for global and local ecosystem carbon (C) budgeting (see review by Limpens et al. 2008). Both large C pools and biogeochemical processes in peatland are of utmost importance for the climate system. Feedbacks of releasing or storing atmospheric C are possible (Davidson and Janssens 2006). The rationale for writing this chapter lies in the fact that there are some important aspects that still need to be highlighted more intensively. The focus on solely peat and may be even the permafrost does not account for all important soil C recourses. There may be other organic rich soils which have not yet been studied more thoroughly e.g. other hydromorphic (or wetland) soils, Podzols and Chernozems or similar soils from the steppe. Last but not the least a book on “Recarbonization of the Biosphere – Ecosystems and the Global Carbon Cycle” would be incomplete without a chapter on peat.

Peat is an organic material formed *in situ* from the remains of plants (and animals) under anaerobic conditions, and with a minimal thickness of 30 cm (Joosten and Clarke 2002) to 40 cm (National Wetlands Working Group 1997) above the mineral horizon. Thus, peatland refers to a peat covered landscape (Rydin and Jeglum 2006). All peats are formed in wetlands, but not all wetlands are associated with peat. Wetlands are characterized by saturated soil conditions or standing water (inundation), hydromorphic soils of a high organic matter (OM) content, and hydrophytic (water loving) vegetation. A water body without hydrophytic vegetation growth is a pond or a lake, but not a wetland, because not all wetlands have the conditions to form peat (Rydin and Jeglum 2006).

There are many **wetland definitions**, which can differ largely from one to another even though it appears to be a rather straight forward expression. A wetland, as suggested by the nature of the name, consists of two interacting natural media: water and soil (Richardson et al. 2001). Generally, wetlands experience water saturation for a sufficient duration of time during the year and usually are defined from “botanical” aspects. Wetlands are permanently or seasonally inundated or water saturated areas with a vegetation which is adapted to water saturated soil conditions (Joosten and Clarke 2002). Aselmann and Crutzen (1989) divided wetlands into six categories: bogs, fens, swamps, marshes, floodplains and shallow lakes. According to the Ramsar Convention, wetlands are areas of marsh, fen, peatland, or water, whether natural or artificial, permanent or temporary, with water that is static or flowing, fresh, brackish, or salt (Ramsar Information Bureau 1971). Mitsch and Gosselink (2007) divided wetlands into two main groups, coastal and inland wetlands.

Table 7.1 Types of wetlands related to peat (Adapted from Tiner 1998, 1999; Rydin and Jeglum 2006; Mitsch and Gosselink 2007)

Type	Description
Mire	A wetland or wet spongy earth, and dominated by living peat-forming plants
Marsh	A wetland subject to continuous or frequent floods, with standing or flowing water, and with or without peat
Swamp	A wetland flooded under a shallow depth of water, with swamp-woody forests and shrub forests (transitional), and comprising fresh, brackish or sea water
Fen	A wetland with either shallow depth of standing water or water table just beneath the soil surface. Fens are associated with peat depth often >40 cm, and are two types basin fen (topogenous) and sloping fen (soligenous).
Bog	A wetland (also called moor or quagmire) associated with acidic peat derived from mosses and (lichens under arctic climates), and either receiving natural precipitation (rain-fed or ombrotrophic) or acidic groundwater.

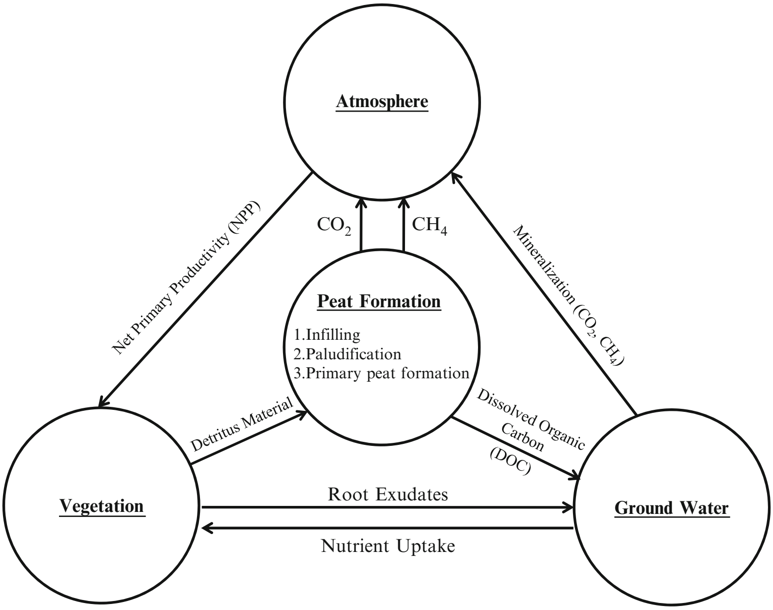

Fig. 7.1 Schematic of carbon cycling in peatland

In this context, peatlands belong to the group of inland wetlands and represent the largest area (3.5×10^6 km^2) of wetlands worldwide (5.8×10^6 km^2) (Mitsch et al. 2009). Apparently wet forest ecosystems are not particularly mentioned, but at least from mangroves, floodplain forests and forested peatlands it is clear that forests and wetlands can coincide. The most important consequence of water saturation is the oxygen depletion leading to conditions of low redox potential to which plants have to adapt. Therefore, more soil oriented definitions could be useful but not easy to map either with or without means of remote sensing. Different types of wetlands are described in Table 7.1. A schematic of C cycling in peat land indicates the C flow between peat on the one hand and atmosphere, vegetation and the groundwater on the other. Thus, world's peatlands are a major reservoir of soil C pool, and play an important role in the global carbon cycle (GCC) through interaction with vegetation or the biosphere, the hydrosphere and the atmosphere (Fig. 7.1).

7.2 Peat Formation

Undisturbed peatlands are a net sink for atmospheric CO_2. The C sink in peatlands is attributed to a high productivity. The magnitude of net C sequestration in natural peatlands ranges from 20 to 1,120 kg C ha^{-1} $year^{-1}$ (Strack 2008). Sequestration of C in peatlands happens through photosynthesis by hydrophytic vegetation and some of it being retained as the net gain in the ecosystem. The latter is estimated in C ha^{-1} $year^{-1}$ at 160–2,440 for herbs and graminoids, 500–1,800 for shrubs, and 1,220–1,880 for bryophytes (Rydin and Jeglum 2006). Because of the low rate of decomposition under anaerobic/inundated conditions, OM accumulates as partially decomposed peat. An anaerobic environment created by waterlogging/inundation reduces the rate of decomposition in peatlands, and is the principal determinant of the rate of net C sequestration. In addition to anaerobiosis, the rate of decomposition is also determined by the composition of the biomass, pH, redox potential, temperature, and can be aerobic and/or anaerobic. Composition of the biomass varies among plant species. The aerobic decomposition increases with drop in the water table, and the anaerobic with the rise in the water table and inundation. It is the anaerobic decomposition which leads to methanogenesis. Hence, natural peatlands are also net sources of CH_4. The anaerobic conditions in peatlands enhance methanogenesis with CH_4 flux of 20–330 kg C ha^{-1} $year^{-1}$ (Strack 2008). With global warming potential (GWP) of ~21, CH_4 emission is an important factor affecting the magnitude of radiative forcing by trace gases. In addition, the loss of C from wetlands also occurs in the form of dissolved organic C (DOC). The global efflux of DOC is estimated at 30–210 kg C ha^{-1} $year^{-1}$ (Strack 2008). Despite the emission of CH_4 and loss of DOC, natural peatlands are net C sink.

Under complete anaerobic conditions, decomposition is slower than the rate of C input as biomass. Thus, peat formation by accumulation of biomass through net ecosystem C gain occurs by three principle processes (Rydin and Jeglum 2006; Wieder and Vitt 2006): (i) **Infilling or terrestrialization** is the process by which new peat formation occurs on the margins of the wetland and in regions with shallow waters of a pond or a slow-flowing streams/rivers, (ii) **Paludifucation** is the process of peat formation over the ground which was previously not flooded or was less wet, and thus does not have any prior sediment deposition. Thus, these types of peat often have woody mass and stump underneath. Paludification can either happen by the rise of water table and new waterlogging of an adjacent upland, or by changes in pedogenesis. The latter is set-in-motion by alterations in hydrologic processes within the soil profile (e.g., decrease in permeability, runon, pan formation). Paludification is promoted by natural or anthropogenic factors like deforestation, fire, etc., (iii) **Primary** peat formation happens when new peat is deposited on top of the mineral soil such as on land exposed by sea level change, fresh volcanic deposit, glacial moraine, mined lands etc. The basic conditions of inundation and growth of hydrophytic vegetation are important pre-requisites.

Other processes of peat formation include **dischargelspring** and **ombrotrophication**. Dischargelspring implies creation of continuous wetland along a hillside due to the lateral flow or interflow (spring). In comparison, the term ombrotrophication implies transformation of mineral soil to peat formation by an upward development of peat surface (Rydin and Jeglum 2006). This transition can be triggered by any climatic shift.

On the basis of formation, there are two principal types of peat. **Ombrotropic bogs** involve peatlands domed above the surrounding landscape and the input of water and nutrients occurs through the atmosphere. Thus, the peat principally involves the remains of *Sphagnum* mosses (Gorham 1991). In contrast, **minerotrophic fens** are formed where water enriched in bases and nutrients percolates into the peat from the surrounding soil. In case of poor, generally sandy substrate, minerotrophic fens may be poor in nutrients, favoring vegetation generally found in ombrotrophic bogs. Minerotrophic fens primarily contain reed, cattail and sedges, and in case of poor fens, *Sphagnum* mosses and shrubs.

Humification (chemical and structural alteration of OM) is another on-going and a continuous process in peat formation. The degree of humification can be assessed by visual examination or by laboratory-based analyses (Blackford and Chambers 1993). Laboratory analysis is based on different extractability of OM with strong acid or bases and classification and fulvo acid, humic acid and humin (Waksman 1936). However, the methodological procedure can create a strong artifact in assessment of the humification (Caseldine et al. 2000), and the concept of different extractibility is currently questioned (Schmidt et al. 2011).

Boreal and sub-arctic peatlands, comprising a total C pool of 455 Pg accumulated since the post-glacial period, have on average accumulated C at the rate of 0.096 Pg $year^{-1}$ (Gorham 1991). The present rate of C sequestered in peatlands is ~0.076 Pg $year^{-1}$ (Clymo 1984; Gorham 1991). The long-term drainage, on the other hand, can exacerbate oxidation and emission of 0.0085 Pg $year^{-1}$, to which should be added the emission from peat fuel at 0.026 Pg $year^{-1}$ (Gorham 1991). In addition CH_4 emission from world peatlands is estimated at 0.046 Pg $year^{-1}$ (Gorham 1991).

Peatlands are also located in the tropics. However, tropical peat lands differ from those in temperate regions in vegetation and other physiographic characteristics (Table 7.2). With ~90% of the global area of peatlands located in northern latitudes (MacDonald et al. 2006), those in Europe and North America are among the most widely studied.

With a large C pool (~550 Pg), peatlands are an important component of the GCC (Yu et al. 2010). There is a strong and growing interest in peatlands because of the projected risks of positive feedback to climate change (Friedlingstein et al. 2006), by which peatlands may become a major source (rather than being a sink) of atmospheric CO_2 and other greenhouse gases (GHGs). Large areas of peatlands

Table 7.2 Differences in temperate vs. tropical peatlands (adapted from Rydin and Jeglum 2006)

Characteristics	Tropics	Temperate
1. Vegetation	Tropical rainforest	Sphagnum and herbaceous
2. Material	Partly decomposed, woody, and covered with leaf litter, with a high hydraulic conductivity	Non-woody and relatively more decomposed, with a distinct *catotelm* layer of low hydraulic conductivity
3. Gradient	Gentle (<0.5 m/km)	Steeper gradient
4. Depth	>25 m	<10 m
5. Duration	Since the last 5,000 years, but some as old as 40,000 years	Glacial melting
6. Uses	Source of timber	Source of peat

have been drained for conversion to agricultural and forestry lands uses. Drained peat decompose rapidly (Fargione et al. 2008), and drained peatlands are prone to fire (Strack 2008). Together with other wetlands, they also form the largest natural source of atmospheric CH_4.

Thus, the objective of this chapter is to describe the difficulties of accounting for soil C resources, offer management strategies for restoring the wetlands, and minimize the risks of emissions of GHGs from peatlands.

7.3 Ecological Characteristics of Peatlands and Other Ecosystems Rich in Soil C

Peatlands are wetland with accumulated (partially) decomposed OM (peat, i.e., material with high fibre content). According to most definitions peat has at least 30% (dry mass) of OM. The definition of peatland varies between the different classifications from 20 up to 70 cm thickness of accumulated peat layer (Joosten and Clarke 2002). Histosols, the typical peat soils, are defined by WRB (2007) by having a histic horizon (>30% OM and >10 cm thickness) of 40 cm thickness within the upper 80 cm. National definitions differ from this. For example in Germany peat soils are defined as soils with >30% dry mass of OM and >30 cm thickness (AG-Boden 2005). One of the most wide-ranging studies of northern peatlands was conducted by Gorham (1991), who used 30 cm of peat as minimum to distinguish between peat and non-peat.

Different definition of peatland or wetland affects directly estimates of organic C stored in the soils of these ecosystems. For example, 'organic soils' and peatlands are commonly found in both North Russia and North America. According to the Russian definition these contain ≥30 cm thick organic horizons (e.g., histic horizon). In contrast, in North America 'organic soils' and peatlands, by definition, have to be thicker than 40 cm (Tarnocai et al. 2009). Consequently,

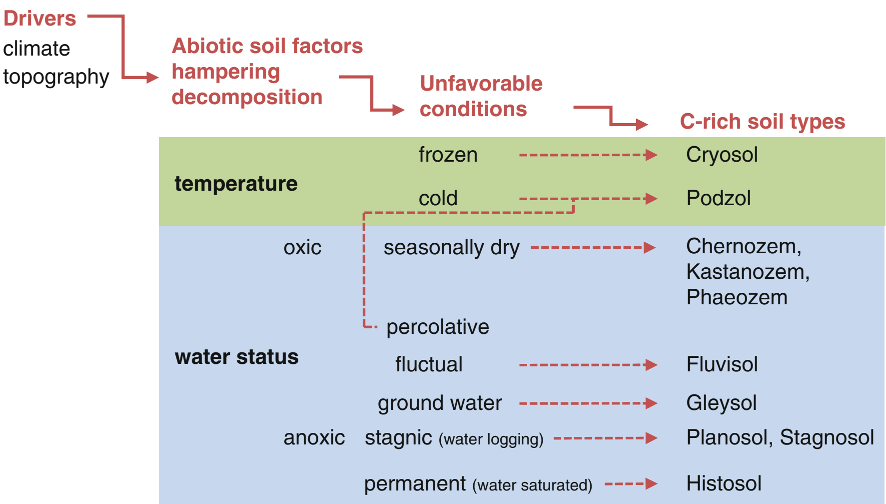

Fig. 7.2 Main pathways of organic matter preservation in soils

areas (1,162–2,730 × 10^3 km^2) and C stocks (94–215 Pg) of peatlands in published estimates vary widely (Tarnocai et al. 2009) (see also Fig. 7.2).

The story of feedbacks between soils and the climate system (Fig. 7.1) is predominately a C story. Plants assimilate C by performing photosynthesis and building up the OM. The term soil organic matter (SOM) refers to the total sum of all substances that occur in soils and contain elements in organic form (Collins and Kuehl 2001). The SOM is a complex mixture of organic residues of plant and animal origin (Kögel-Knabner 2002) that accumulate in the soil and undergo continuous transformation.

Compared to “regular” oxic sites a larger part of this C is stored below-ground at anoxic sites. The reason is that decomposition is hampered by anoxic or hypoxic conditions. Other conditions hampering decomposition are acidic, frozen or cold conditions (Fig. 7.2).

The SOM transformation predominately depends on **temperature and water status** which are the main controls on soil microbial activity. Transformation consists of a series of steps which can be associated by gaseous by-products such as nitrous oxide (N_2O) and end-products CO_2 (aerobic and anaerobic decomposition) and CH_4 (anaerobic decomposition). These gaseous C and nitrogen (N) products of SOM decomposition are the main feedbacks to the climate system. All individual steps are governed by a complex interplay of (a) soil organisms, (b) the quantity and quality of OM (Paul and Clark 1996), and (c) prevailing environmental conditions (e.g., temperature, water supply, Fig. 7.2).

In general, aerobic decomposition of OM is more efficient and more rapid than anaerobic decomposition. D’Angelo and Reddy (1999) postulated that aerobic C

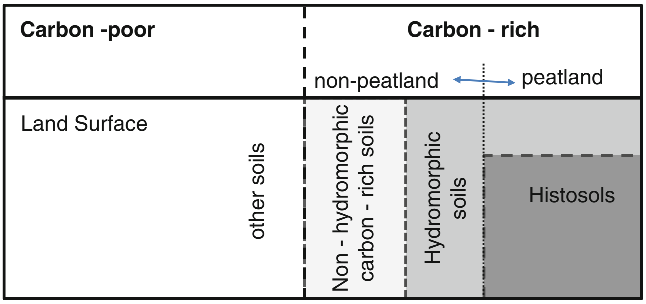

Fig. 7.3 Soil C characteristics of peatlands, wetlands and other ecosystems

transformation is about 3 times faster than an anaerobic transformation. Therefore, in **hydromorphic soils**, the accumulation rate is much higher and hence overrides the rate of decomposition. For that reason, the quantity of OM tends to increase in hydromorphic soils (e.g., undecomposed plant tissues) to a greater extent than in aerated soils. However, both thickness and darkness of H and Ah horizons are positively correlated with the exposure time of the soils to reducing conditions (Thompson and Bell 1996).

The OM undergoes a much slower transformation in acidic soils in cold and humid climates. In strongly acid soils, the growing conditions for microorganisms are poor, resulting in low levels of biological oxidation of OM (Primavesi 1984). As a result, Podzols usually have much thicker organic horizons than their less acidic counterparts (like Cambisols). The pH-values affect humus formation in two ways: hampered decomposition, and biomass production. However, decomposition is usually more affected than biomass production, resulting in accumulation of OM.

Extremes in climate trigger the turnover rates of SOM. A prominent example is the formation of soils under steppe vegetation (Chernozems, Phaeozems and Kastanozems). These biomes are dominated by cold winters and dry, hot summers. Plant productivity by these conditions is less reduced than decomposition.

Besides the lack of monitoring networks with adequate temporal and spatial resolution (Prechtel et al. 2009), one of the major obstacles to correctly account for soil C stocks and identifying the spatial hot spots are differing definitions from diverse scientific disciplines and cultures (countries, administration). The latter are the cause why maps on soil C, peatland etc. may show abrupt changes at administrative borders which are not explainable by the natural causes. This is among numerous reasons for very different regional estimates that have to rely on mapping which applies to any attempt to identify soil C hotspots and account for soil C stocks. One illustrating example is the fact that forests are frequently and solely considered to be upland and therefore non-wetland ecosystems. This simplification leads to serious bias for landscape biogeochemical budgets, e.g., methane (Fiedler et al. 2005) and soil C budgets.

7.4 Predominant Soils of Peatlands and Other Ecosystems Rich in Soil C

At the pedological perspective, wetlands build **hydromorphic soils**. These include biologically active organic as well as mineral soils in which the water contents are permanently or temporally large enough to inhibit oxygen diffusion into soil and stimulate anaerobic processes. In turn, these are characterised by accumulation of SOM. In total, hydromorphic soils cover 17.4×10^6 km^2 (based on FAO 1995) or 12.9% of the land surface, which is more than other wetland estimates. Diagnostic horizons of soils are outlined in Table 7.3.

The most prominent hydromorphic soils are **Histosols** (2.5×10^6 km^2 and 1.8%) which are formed in response to permanent water saturation. FAO (1995) estimates the global area of Histosols at about 3.3×10^6 km$^{2.}$ In theory, Histosols and peatlands should be approximately the same, however, these estimate are lower than most peatland estimates.

Histosols, by definition according to both WRB (2007) and the US soil taxonomy (2006), are soils consisting primarily of organic materials. Soil type is extremely poorly drained because the OM has a high water retention capacity.

Estimated average C contents of Histosols range between 720 and 1,250 Mg C ha^{-1} (0–100 cm depth) and 1,230–2,640 Mg C ha^{-1} (0–200 cm) (Batjes 1996). The majority of Histosols are located in the boreal, subarctic and low arctic regions of the Northern hemisphere. During the revision of soil classifications (US Soil Taxonomy 2006; WRB 2007), a new soil type defined was 'Cryosols' (WRB 2007)/ Gelisols (US Soil Taxonomy 2006) which includes some soils formerly classified as Histosols. Therefore, it is difficult to precisely differentiate between Histosols and Cryosols. By definition, **Cryosols** are soils at or below the freezing point of water for two or more years. The extent of permafrost (should be Crysols) is estimated by

Table 7.3 Some typical diagnostic horizons (WRB 2007) of carbon-rich soils with some properties important for SOC storage

Cause for OM accumulation	Diagnostic horizons (WRB 2007)	Properties
Too wet	Folic	10 cm, organic, <30 days water saturation
	Histic	10 cm, organic, >30 days water saturation
Too dry	Mollic, umbric	20 cm, >0.6[a]/2.5% SOC
	Voronic	35 cm, >1.5[a]/6% SOC
Too acidic	Sombric	Illuvial humus
	Spodic	Illuvial humus, >0.5% SOC, pH <5.9
	(Fulvic, melanic)	30 cm, >4% SOC, andic soils
Anthropogenic	Anthric	Wide range
	Hortic	20 cm, >1% SOC,
	Plaggic	20 cm, >0.6% SOC, commonly acidic

[a]If Munsell chroma <3

Tarnocai et al. (2009) to be 18.8×10^6 km^2 (equal to 16% of the total soil area) but these estimates vary and the actual extension will change with the projected climate changes.

Cryosols can store extremely high amounts of C. In a recent study of Tarnocai et al. (2009) the C storage in Cryosols is estimated between 1,400–1,700 Pg worldwide. This estimate include: (a) C stock in circumpolar soils at 496 Pg to 1-m depth, (b) 1,024 Pg to 3-m depth, and (c) the estimated C stock of 1,672 Pg in all Cryosols and peatland including thus far unconsidered stocks of 648 Pg contained in the deltaic deposits, and the Sibirian yedoma sediments below 3-m depth. Nonetheless, these estimates are large because only 1,500 Pg is the estimate C stock to 1-m depth in all world soils (Batjes 1996).

OM decomposition is generally faster under periodic than continuous water saturation. Therefore, Histosols contain the highest SOM content among all other wetland soils. All other hydromorphic soils (i.e. Stagnosols, Planosols, Gleysols, and Fluvisols) (WRB 2007), are mineral soils and formed in response to temporary water saturation. The diffusion of oxygen into hydromorphic soils is drastically curtailed, at least episodically, forcing biological and chemical processes to change synchronously (or slightly delayed) with water saturation. The water regimes transform soils from an aerobic to an anaerobic environment and vice versa leading to predominantly redoximorphic features (e.g., mottled soil matrix). These changing redox conditions are associated with highly dynamic biogeochemical cycling.

Predominant mineral hydromorphic soils are **Gleysols** (8.3×10^6 km^2, 6%) which are saturated by ground water near the surface (for long time). These soils occur at foot slopes or at landscape depressions. The prevalent water table creates a typical redox gradient in soil profiles. Soil horizons above ground water table are mostly oxic and characterized by Fe accumulation which explains their brown colour. Grey coloured horizons below the water table are predominately anoxic. The highest C accumulation (up to 150 Mg C ha^{-1} in 0–30 cm) is observed in the topsoil which increases with increase in the mean of annual water table.

Fluvisols (3.2×10^6 km^2, 2.4%) are young soils in alluvial deposits, and are widely distributed in floodplains or terraces including alluvial plains, river fans, valleys and tidal marshes which are inundated or flooded frequently. The hydrology of Fluvisols is dictated by the periodically presence of ground water and flood water. The accumulation of C in Fluvisols is influenced by import as well as export of fresh OM due to alluvial transport and wet conditions. The lateral and horizontal changes in sediment deposits results in a large variability in the C content of Fluvisols. Additionally, Fluvisols are commonly associated with Gleysols in great river deltas and Histosols in the meandering river areas. Such a physiographic combination makes soil survey maps of floodplains difficult to interpret regarding the C stocks, because they cannot adequately delineate all characteristics required to fully characterize these soils.

The most prominent mineral hydromophic soils formed on concave or flat topography and plateaus are **Planosol** and **Stagnosols.** These soils are formed by the influence of stagnant water caused by dense soil layers, which impede water percolation. These soil types (3.4×10^6 km^2 equal to 2.5% of the total land) are distributed among all continents and climatic zones with maxima in the zone between 10° and 50°N covering 35% of the total area. Global estimation of C stocks (1-m depth) of these soils range from 57 to 138 Mg C ha^{-1}.

The largest anthropogenic wetlands are paddy soils, classified as hydromorphic **Anthrosols** (WRB 2007). These soils occupy 1.55×10^6 km^2 of the total land surface (http://beta.irri.org/index.php/). Paddy soils can originate from different soils (e.g., Andosols, Cambisols, Vertisols), however, specific management leads to alternating redox conditions and therefore to soil properties and morphology independent of the antecedent soil unit. The formation of these Anthrosols is induced by tilling the wet soil (puddling), and the flooding and drainage regime associated with the development of a plough pan (Kögel-Knabner et al. 2010). Anthropogenic activities (large inputs of plant residues, organic fertilizers, temporal flooding) enhance accumulation of SOC in these soils (Lal 2002, 2004; Sahrawat 2004).

Podzols occur mainly in humid areas in the boreal (3.2×10^6 km^2) and temperate zones (1.3×10^6 km^2), and locally in the tropics (1×10^5 km^2). On the global scale, Podzols store a significant amount of C (275 Pg in 0–200 cm depths), especially in the 'spodic horizon'. Humus accumulation is enhanced due to strongly acidic soil conditions, but the mechanisms are not well understood (Sauer et al. 2007).

Chernozems (equivalent to Mollisols of the US Soil Taxonomy) and the associated soils (**Kastanozem, Phaeozem and Greyzem)** are mineral soils which usually develop ubiquitously under steppe vegetation. Chernozems constitute 4–6% of the total global land area. By definition, Chernozems must have a mollic epipedon which is a thick (≥25 cm), dark coloured surface horizon containing C content ≥6 g kg^{-1} and a base saturation of ≥50% (WRB 2007). Factors which trigger genesis of Chernozems include semiarid to subhumid climates, grassland vegetation, and a calcium-rich parent material. The combination of these factors control the amount of SOM by favouring increased below-ground biomass production, deposition of lignin-rich residues and development of stabilizing bonds with Ca that slow the rates of OM decomposition (Thompson and Bell 2001). Chernozems are highly productive soils, but have a vulnerability to decomposition.

7.5 Distribution of Peatland and Hydromorphic Soils

Peatlands cover only approximately 3% of the land surface (Yu et al. 2011), or approximately 4×10^6 km^2, but contain 40% of the SOC to 1-m depth (612 Pg out of 1,550 Pg). Regarding the huge C stocks in peatlands it is important to realize that (a) the area covered by peatland is not the only factor for assessing the importance of these ecosystems to global biogeochemical cycle, and (b) erroneous estimates are made by different assumptions about the area under peatlands. Estimating the mean or actual depth and bulk density of peatland are also very challenging. Hence global and regional (continental) estimates of C stored in peatlands vary with approach and the time when the estimate was made (Fig. 7.4).

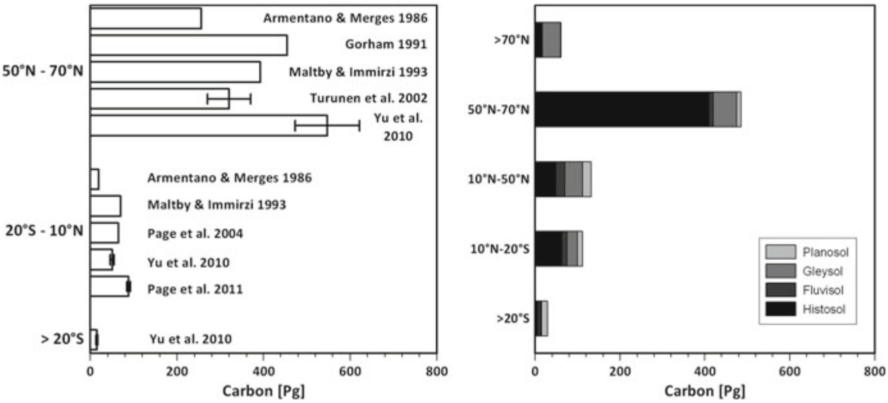

Fig. 7.4 *Left*: Zonal distribution of soil carbon of peatlands by various authors. *Right*: Zonal distribution of soil carbon of hydromorphic soils (2 m depth) (FAO 1995; Batjes 1996)

Therefore, studies are inherently imprecise and coordinated efforts are needed to assess the C stocks of peatland and hydromorphic soils. The qualities of those estimates depend on the transparency of the chosen procedure, and providing the degree of uncertainty. One uncertainty that is particularly difficult to capture is the spatial extent, because wetlands and peatlands are not clearly defined. It is also important to know the location and area of other C-rich but nonpeat and partially wet soils which are even more difficult to delineate. The comparison between the low land area and high importance as C stock provides an important reason to improve the scientific understanding of peat, peat-like and wetland systems. Credible data are needed about the land area and associated processes for more precisely predicting the future soil to atmosphere C feedbacks.

A large proportion of the global peatland area is located in the northern latitudes (50–70° north), especially in boreal and subarctic regions of Canada and Russia (Aselmann and Crutzen 1989). The humid and cold climatic conditions in these regions are suitable for a high rate of peat accumulation. In addition to the largest area of peatlands at these latitudes, there soils are also characterized by the highest C densities (Yu et al. 2011). Northern peatlands accumulated about 550 Pg of C since the last glacial maximum (Yu et al. 2010) with a mean net rate of about 0.1 Pg C year^{-1} (Gorham 1991). The formation of peatlands in the high northern latitudes started mainly after the last glacial maximum with the peak accumulation in the early Holocene (Yu et al. 2010). These peatlands store about 15–30% of the world's soil C stock (Limpens et al. 2008) which is lower than the estimate stated above. All hydromorphic soils, which include Histosols, store approximately 41% of the global soil C stock. Podzols store an additional 10%.

The second largest land area covered by peatland is found between 20° south and 10° north (Aselmann and Crutzen 1989) with approximately 11% of the global peatland area. Moreover the second largest concentration of C with about 85 Pg or

15–19% of the global peatland C stock is estimated in tropical peatlands. Most of the tropical peatlands occur in Southeast Asia (with about 5% of the global peatland area), and with the largest C stock in Indonesia (Page et al. 2011). Tropical peatlands have the lowest C accumulation rates during the Holocene, but these peatlands started to form much earlier than northern peatlands (Yu et al. 2010).

Peatlands in the southern hemisphere, mainly in Patagonia, South America, have accumulated approximately 15 Pg of C since the last glacial maximum (Yu et al. 2010). These peatlands are characterized by the highest accumulation rate of C during the Holocene with 220 kg C ha^{-1} $year^{-1}$ compared to northern and tropical peatlands (Yu et al. 2010). Understanding the mechanisms of these high accumulation rates is of utmost importance particularly in the context of the need for "recarbonization" of the biosphere. Additionally, the spatial distribution of peat and peat-like ecosystems may be underestimated for Africa (a case study is presented later in this chapter). Besides evapotranspiration and precipitation, topography also plays an important role in peat formation. Peatland formation can also occur in even in wetlands of the semiarid regions, which are abundant in Africa (Tooth and McCarthy 2007). In the tropics, countries with known and mapped peat lands include Indonesia (27 Mha), Brazil (5.5 Mha), Peru (5.0 Mha), Papua New Guinea (2.0 Mha), Malaysia (2.5 Mha), Congo (1.5 Mha), Uganda (1.4 Mha), Colombia (1.0 Mha), Venezuela (1.0 Mha) and Zambia (1.0 Mha).

The huge amount of C stored in peatlands has mainly been accumulated since the last glacial maximum with a net sink rate of more than 5 Pg of C per century on average (Yu et al. 2010). Mainly due to water-saturated (hypoxic = oxygen low) soil conditions, build-up of SOM is faster than decay and these wet and mostly peaty soils act as prolonged atmospheric C sinks. The fate of this locked C in a changing world (See Chap. 18) under changing conditions is not understood, but is important to predicting any feedback mechanisms with climate system (Davidson and Janssens 2006).

7.6 Differences Between Wetland and Non Wetland Soils

7.6.1 A Case Study South Africa

Only very preliminary work has been done in African and South American peatlands, and at this stage even their areal extent is not well quantified, much less their C content, vegetation composition, or C cycle characteristics (Frolking et al. 2011). Therefore, soil C content of four different soils was determined in an investigation in the Drakensberg, Republic of South Africa. Two wetlands and two non wetlands were chosen for the study in this region. The differences between wetland and dryland were obvious in the vegetation and in the geomorphology, but only slight differences were measured in the top soil C content (see Fig. 7.5). Wetland soils have a depth of >0.5 m and up to 1.5 m, whereas the dryland soils were only about 20–30 cm thick. However, a thickness of 30 cm would qualify these soils as peatland

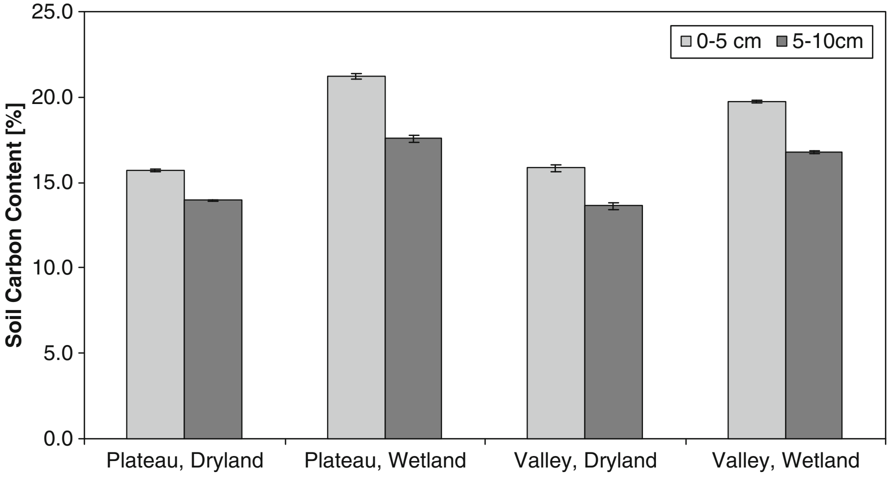

Fig. 7.5 Mean soil organic carbon content [%] with standard deviation of four different locations [n = 3] with depths of 0–5 cm and 5–10 cm, Drakensberg region, Republic of South Africa

(for example Gorham 1991). This case study shows that even non wetland soils may have high C contents, and C content and stocks are underestimated. Therefore, the simple differentiation of soils into high C-rich wetlands and C-poor drylands is inadequate. Based on the large areal extent of the dryland compared to the wetland soils, credible assessment of the C content is an important component in estimating regional and global C stocks.

7.7 Global Soil Carbon Hot Spots: Potential Sources for Atmospheric CO_2

In the context of "recarbonization", it is important to know where the soil C hot spots are. While it may appear to be a trivial question considering available global estimates and maps (FAO 1995), yet there is a strong need to stress that erroneous estimates are made in precisely accounting for the global soil C resources. The difficulty lies not only in insufficient means to measure soil C resources at the global scale – because of the insufficient data – but also due to other factors such as consideration of different depths, including peat and other wetland soils, and different definitions or perceptions of soil types, all of which hamper obtaining the reliable estimates of global and regional soil C stocks. Therefore, there exists a large uncertainty not only about the magnitude of soil C stocks but also which part is prone to being emitted into the atmosphere. World soils most likely store more than the "official" estimate of 1,500 Pg C (Fig. 7.6), which is already twice the amount contained in the atmosphere. The available estimates of soil C stocks, for example, do not include the latest estimates for peatland C stock of Yu et al. (2010). Furthermore,

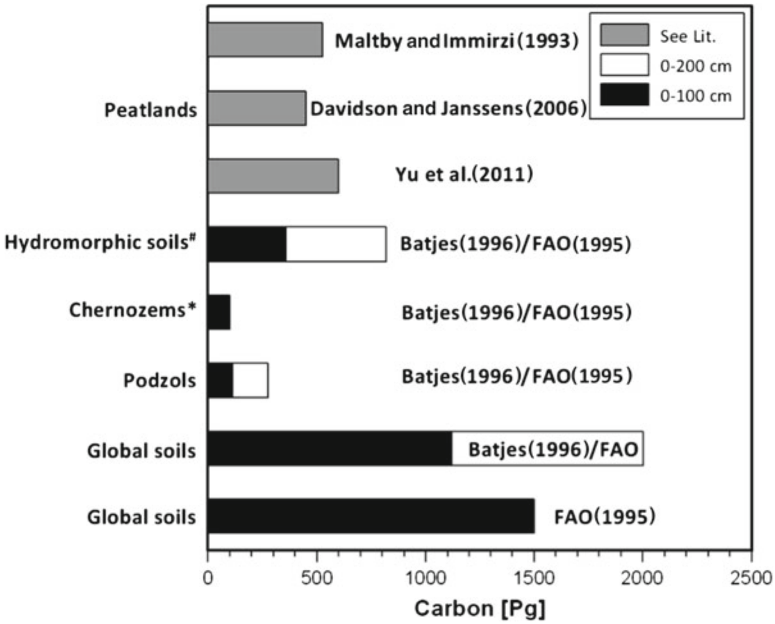

Fig. 7.6 Global soil carbon estimations. Hydromorphic soils# (Histosols, Fluvisols, Gleysols, Planosols), Chernozems* (Chernozem, Phaeozem, Kastanozem, Greyzem), (FAO 1995; Batjes 1996)

the present estimates also underestimate the deep soil C storage. Nonetheless, the deep soil C is not prone to being emitted into the atmosphere in the near future. Therefore, a credible inventory of vulnerable and less vulnerable soil C stocks would be advantageous for predicting feedbacks to the climate system. In this context, the first simplest step is to differentiate between wetland and upland ecosystems. Hydromorphic soils store a greater part of soil C (Fig. 7.6) which is, at the same time, undeniably highly vulnerable to changes in both temperature and precipitation.

Grouping the world into wet- and dryland or low and high C-rich soils is not as simple as it may seem. Addressing the exchange processes between the atmosphere and the biosphere by disregarding the complexities can lead to erroneous estimates. Soils characterized by transient water logging conditions (semi-wet) are very common and react, at least partially like wetlands. Therefore, all wetlands must be included, as has been done for the present study (Fig. 7.6). All hydromorphic soils (0–200 cm) potentially store 818 Pg C which is more than the latest estimates for peatland alone, even though the former conservative estimates by Batjes (1996) were used and multiplied by the land area given by the FAO (1995).

There are more conditions that hamper OM decomposition leading to higher soil C resources. When summers are too dry for decomposition, typically C-rich Chernozems are formed in the steppe. These Chernozems (includes Chernozem, Phaeozem, Kastanozem, Greyzem) store at least another 100 Pg C (Fig. 7.6). Potentially these steppe soils store about another ~25% in 1–2 m depth as compared to the top 1 m (Mikhailova and Post 2006). Podzols have both C rich topsoils due to hampered decomposition and high subsoil C due to percolation transport. The rough estimate of the mean values given by Batjes (1996) and multiplied by the area derived from the FAO soil type maps is 275 Pg C (for 0–2 m, Fig. 7.6). All C-rich soils and particularly any in-between states need to be included in biogeochemical feedbacks estimates between soils and the climate system. And even less C-rich soils are important because they cover larger part of the terrestrial surface. The global estimates derived from Batjes (1996) and area given by the FAO is 1,120 Pg C (0–1 m) and 2,003 Pg C (0–2 m). Using the soil C data provided by the FAO (1995), the mean value is 1,501 Pg C (also reported by Schlesinger 1997; Glatzel 2011), with a range of 963–2,057 Pg C.

7.8 Peatland Conversion to Agricultural Use

Drainage, deforestation and tillage lead to rapid decomposition. In the temperate zone, rate of oxidation of drained peatlands, used for seasonal/annual crops, can be 1–2 cm $year^{-1}$. The rate (Mg C ha^{-1} $year^{-1}$) of CO_2-C emission from drained and cultivated peatlands have been measured at 2.8–6.7 for Germany, 4.9–11.2 for Poland, 3.5–17 for Sweden, 2.2–8.2 for the Netherlands, 5.5 for Russia, 1.9–2.3 for Canada, 3.3–8.3 for Finland (Strack 2008). The site-specific rate of C emission depends on local conditions related to temperature and moisture regimes, pH and the soil and crop management practices.

Area of non-tropical peatlands converted to managed ecosystems are estimated at 25 Mha for agriculture and 15 Mha for forestry (Joosten and Clarke 2002). Total area of peatlands drained for agricultural and forestry uses in Europe is estimated at 10–30 Mha (Joosten and Clarke 2002; Rydin and Jeglum 2006), with large land use conversion in Russia, Finland, Sweden, U.K., Netherlands, etc. As much as 14% of the peatland area in Europe has been converted to managed ecosystems. Large areas of peatlands have also been converted to agricultural and forestry land uses in Thailand, Malaysia and Indonesia, and additional areas are being deforested and drained in southeast Asia for establishment of biofuel plantations. More than 50% of the original peatlands in south-east Asia have already been converted. Deforestation for timber is a serious threat. Drainage ditches for log transport alter the hydrology, and exacerbate the risks of fire.

Rather than emission, drained and cultivated peatlands may be a small sink for CH_4. The sink capacity also depends on site-specific conditions and varies widely among locations. Further, there is a large spatial and temporal variability within each site. The CH_4 uptake on cultivated peatlands has been measured at rate (kg CH_4 ha^{-1} $year^{-1}$) of −0.2 to −4.9 (Strack 2008). Depending on the moisture regime, cultivated peatlands can also emit CH_4 at rates of 0.2–3.6 kg CH_4 ha^{-1} $year^{-1}$ (Strack 2008).

In addition to gaseous emissions, other environmental impacts of drainage and cultivation of peat include loss of dissolved organic carbon (DOC), emission of N_2O

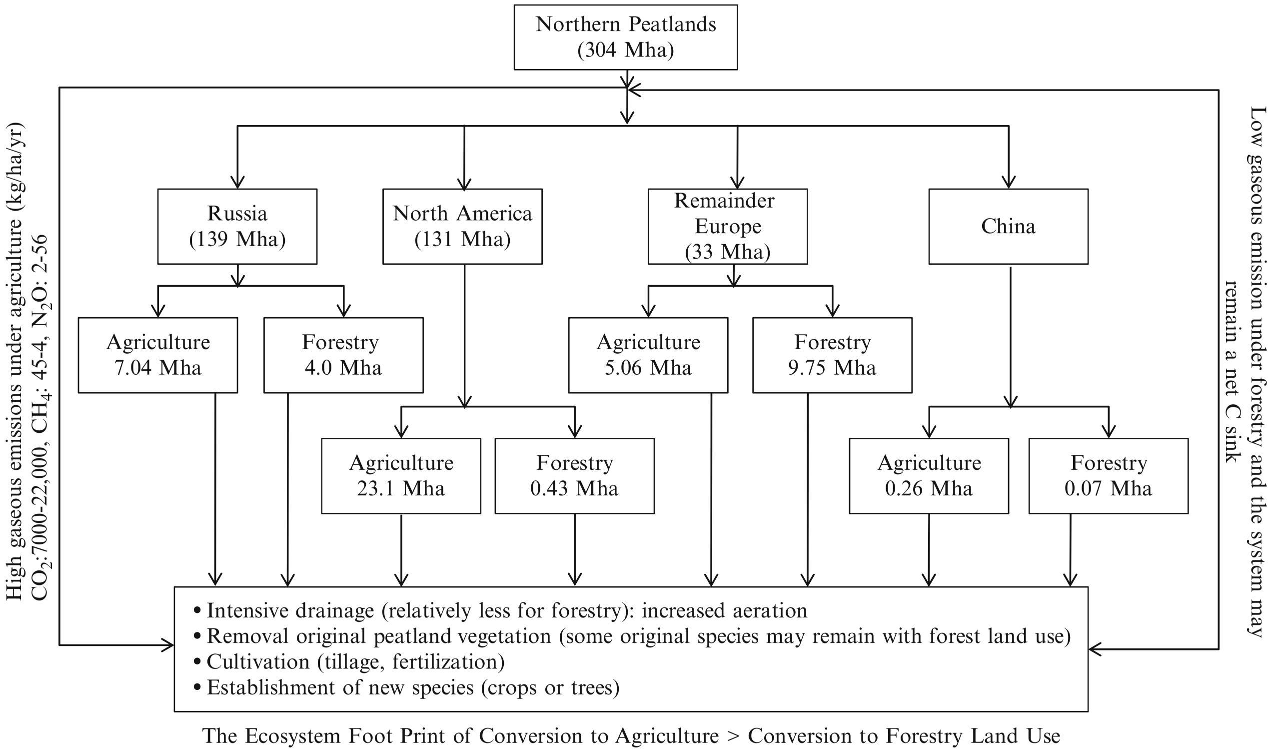

Fig. 7.7 Anthropogenic perturbations of northern peatlands for agriculture and forestry (Redrawn from the data by Lane et al. 2009). C density agriculture, forestry and other uses

Table 7.4 Land clearing and biofuel debt (Recalculated from Fargione et al. 2008)

Biofuel	Former land use	Location	C debt ($Mg\ CO_2/ha$)	Time to repay biofuel C debt (year)
Palm biodiesel	Tropical forest	Indonesia/Malaysia	702 (175)	86
Palm biodiesel	Peatland forest	Indonesia/Malaysia	3,452 (2,952)	423
Soybean biodiesel	Tropical forest	Brazil	737 (200)	319
Sugarcane ethanol	Cerrado (woods)	Brazil	165 (100)	17
Soybean biodiesel	Cerrado (grasslands)	Brazil	85 (85)	37
Corn ethanol	Central grassland	USA	134 (10)	93
Corn ethanol	Abandoned cropland	USA	69 (5)	48
Prairie biomass ethanol	Abandoned cropland	USA	6	1
Prairie biomass ethanol	Marginal cropland	USA	0	No debt

The number in parenthesis is soil carbon as a part of total carbon debt

through the nitrification and denitrification processes, and leaching of nitrates. Drainage and lowering of water table can lead to emission of N_2O at 2–56 kg N_2O ha^{-1} $year^{-1}$ (Strack 2008).

Increase in demand for production of biofuel has accentuated the rate of deforestation of peatlands in Southeast Asia. Deforestation, drainage and cultivation create a large ecosystem C debt (Fargione et al. 2008). Deforestation and drainage adversely impact both the biomass and soil C stocks. In addition, burning exacerbates the ecosystem C loss. Fargione et al. (2008) estimated the C debt computed as CO_2 released during the first 50 years of land use conversion. The data in Table 7.4 show CO_2 debt and the time needed to repay the debt through the production of biofuel. The highest C debts (Mg CO_2 ha^{-1}) for biomass, and soil, respectively, was 3,452 and 2,952 with 423 years to repay it were estimated for conversion of tropical peatlands (Indonesia, Malaysia) to oil palm plantation for the biodiesel production. In comparison, establishment of soybean-based biodiesel plantation in Brazil created CO_2 debt of 737 and 200 with 319 years to repay it for the tropical rainforest ecosystem, compared with 165 and 100 for 17 years to repay it for soybean-based biodiesel from the cerrado/savanna ecosystem. Thus, conversion of peatlands to biodiesel production is not sustainable. Conversion of virgin peat swamp forest to oil palm (including biomass, fire), created a total C-loss of 17 Mg ha^{-1} $year^{-1}$ over 25 years (Hergoualc'h and Verchot 2011).

7.9 Interaction with the Climate System

Increasing temperatures lead to increasing microbiological activity and, therefore, enhanced decomposition (Arrhenius equation, Vicca et al. 2009). Thus, a direct feedback of climate warming and more CO_2 emission from soils may appear convincing. However, plant growth might also be enhanced and sensitivity of decomposition

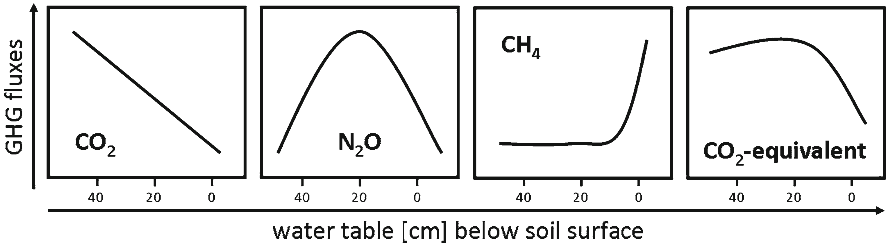

Fig. 7.8 Schematic relationship between greenhouse gas fluxes and water table in wetland soils (According to Jungkunst et al. 2008)

processes is not completely understood (Davidson and Janssens 2006). A more comprehensive discussion can be found in Chap. 18. On the other hand, higher temperatures lead to higher evapotranspiration which eventually leads to dryer conditions that may hamper decomposition despite the better temperature conditions. From a wet and peatland perspective, drying usually leads to a shift from anoxic conditions (low redox values) that hamper decompositions to more favorable oxygen richer conditions (higher redox values). Therefore, increasing temperature and higher redox values will both enhance soil C decomposition from hydromorphic soils and increase the feedback loop. In some cases predicted increase in precipitation can promote low redox condition associated with soil C lockage. Increasing variability of precipitations is the most difficult feature of predicted climate change conditions because (a) both agriculture and ecosystem cannot easily adapt to ever changing and "unreliable" conditions, and (b) feedback reactions are most difficult to predict because every wetness state is associated to a different composition of GHGs releases from soils (Fig. 7.8).

For example the latest observations that atmospheric CH_4 is increasing at lower rates is explained by: (a) drying of northern wetlands, or (b) a change in the agriculture practice in rice fields mainly in Southeast Asia (Heimann 2011). The change in the rice agriculture in Asia in the last three decades has contributed to the reduction in CH_4 emission (Kai et al. 2011). A greater application of fertilizer and shorter inundation period can reduce the CH_4 emissions from these soils (Heimann 2011; Kai et al. 2011). However, both less anoxic conditions and more available fertilizer (mainly N) promotes nitrous oxide (N_2O) emissions. New agriculture practices likely reduce CH_4 emissions, but potentially increases N_2O emissions (Frolking et al. 2004). For a complete atmospheric – soil feedback assessment, all three important GHGs (i.e. CO_2, CH_4 and N_2O) must be considered (Jungkunst and Fiedler 2007).

The net release of CO_2, as the common "end-product" of OM decomposition, is substantially lower under anoxic than oxic conditions (Fig. 7.8). However, under strictly anoxic conditions the release of CH_4 is high. The debate, what is worse for the climate system: (a) high CO_2 release from dried peat, or (b) low CO_2 and associated high CH_4 under wet peat conditions, is common. Details are even more complicated. Both N and C cycles are closely coupled and cause the conjoined feedbacks to the climate systems. In this context, N_2O is highly important, which is preferably released under intermediate wet conditions that fall in the hypoxic category (0–5% O_2) (Fig.7.8). Further, N_2O is particularly important for managed soils and vegetation free peat e.g. in the artic (Repo et al. 2009). Generally, N_2O has to be considered for estimation of the impact of climate change in the northern peatlands. A warming of this region would probably increase the N_2O emissions from these ecosystems and which will be a positive feedback to climate change (Elberling et al. 2010; Jungkunst 2010). Besides being large CH_4 sources, several northern peatlands (drained, low C:N ratio and the absence of vegetation) are considered as hot spots for N_2O emissions (Marushchack et al. 2011). Therefore, despite the available knowledge, there is a need for more process-based understanding on peat, wetland and soil biogeochemistry. Theses alarming findings from higher latitudes are unlikely to be restricted to boreal and arctic hydromorphic soils.

7.10 Climate Change and the C Cycle in Peatlands

The climate-induced uncertainties in the fate of C in peatlands are due to the changes in water table and the temperature regimes. These uncertainties may be especially large in regions where melting of the permafrost, thermokarst erosion, and formation of melt/thaw lakes may change the surface and sub-surface hydrology. Yet, another uncertainty with a strong impact may be caused by the frequency and intensity of fire. The climate change and other anthropogenic factors have pertubated the natural balance.

The projected climate change may also affect the C cycle in peatlands through changes in temperature and moisture regimes, and the attendant alterations in the water table (Strack 2008). For example, Roulet et al. (1992) predicted the increase in temperature by 3°C and rainfall by 36.5 cm year^{-1} may lower the water table in peatlands by 14–22 cm. These factors may lead to alterations in species compositions, and thus affect the quantity and quality of biomass addition (Weltzin et al. 2000). An increase in atmospheric CO_2 concentration may have no net effect on ecosystem C budget, because any gains by the CO_2 fertilization effect on the biomass production may be negated by increase in respiration at elevated temperature and the attendant decomposition (Strack 2008). Increase in the length of the growing season may alter the total biomass production in northern latitude but not so in the tropical or equatorial regions. Therefore, the SOC pool in the northern peatlands may be vulnerable to decomposition and to creating a positive feedback. Bridgeham et al. (2008) studied the effects of changing water and temperature regimes over 8 years in 54 peat monoliths from a bog and fen in northern Minnesota, USA. Increase in water availability increased C sequestration in bogs but not in fens. These trends were primarily due to water table effects on Sphagnum moss production in bogs, and to decomposition in fens. The controlled study indicated that peatlands can gain or lose large amounts of soil C until the new equilibrium with the water table depth is attained. The projected climate change may also lead to other C cycle-hydrological feedbacks at a larger/global scale. The projected climate change may lead to new peat accumulation in sub-arctic climate where the prevailing cool climate led to the development of permafrost (Zoltai and Tarnocai 1975) due to increased biomass production and water levels.

7.11 Distribution of Soil Carbon Resources

The distribution of soil carbon (0–1 m) is similar to the distribution of area covered with soils (Fig. 7.9). The exceptions are the “boreal” latitudes (50–70°N) that have proportionally more C than soil area. For the other for zones area and soil carbon stocks correlete well ($r^2 = 0.926$) and the average million square kilometer hold 12 Pg soil C (0–1 m). This number for the “boreal” latitudes is 20 Pg soil C (0–1 m) per 10^6 km^2. This elucidates the relative importance of the “boreal” latitudes for soil C resources, but the absolute importance of the “mid-latitudes” (10–50°N) is at least as important (Fig. 7.9). If the soil type specific mean values given by Batjes (1996) are used and multiplied with the area of the individual soil types by the FAO (1995), the mid-latitudes appear even more important in absolute numbers than the boreal regions. For many estimates, these common mid-latitudes are not considered, whereas tropical zones are. In the context of these estimates, the “tropical” latitudes (20°S–10°N) appear less important. Nonetheless, all zones must be appropriately considered.

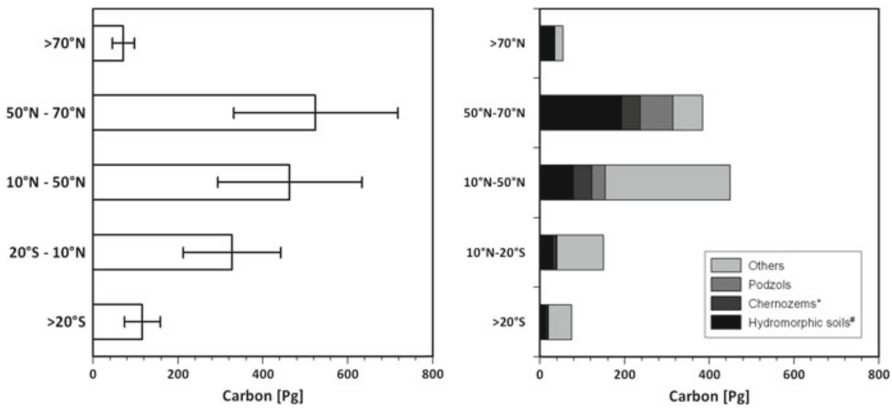

Fig. 7.9 Soil carbon resources referring to 1 m soil depth. *Left*: Zonal distribution of soil carbon (FAO 1995). *Right*: Zonal distribution of soil carbon of different soil types (by FAO 1995; Batjes 1996), Hydromorphic soils (Histosols, Fluvisols, Gleysols, Planosols), Chernozem* (Chernozem, Phaeozem, Kastanozem, Greyzem)

7.12 Peat Extraction

In regions rich in peat, peat extraction has been a part of traditional land use for millennia (Charman 2002). Peat has been used, among others, as surgical dressing, diaper absorbent and in traditional Chinese medicine (Harris 2008), as a prerequisite for manufacturing of whisky, for horticultural purposes and last but not the least as energy source. Presently, 8,105 km^2 of peat area is under active peat extraction for energy purposes using industrial techniques. Most of the affected area (6,590 km^2) is located in the boreal zone, and an additional 1,609 km^2 are in the temperate zone, but only 6 km^2 in the tropics (IPS 2011). These estimates do not include peat extraction for other purposes, such as that for horticultural use, which is widespread in Canada and Baltic countries, and for many small-scale operations in Latin America, Africa and elsewhere.

This quick glance at the diversity of peat use shows that, during the last century, only the scale of peat extraction but not the land use type itself has changed. On the global scale, considering the large peat resources, peat use for energy purposes is small, amounting to 17 Gg $year^{-1}$ (World Energy Council 2007). Since, 80% of the peatlands, most of them located in high latitudes, are still pristine, the question of the sustainability of peat extraction depends on the scale: Globally, more peat is formed in the world's intact peatlands than is extracted for energy purposes. However, considering the smaller scales, it becomes clear that peat extraction contributes to species loss as well as to jeopardizing the last remaining peatlands in many countries in the temperate zone.

7.13 Peat Restoration

As has been already discussed, using peatlands for forestry, agriculture, or extracting peat is a C-consuming process, leading to a net rise in GHG emissions. Recognizing this, numerous measures for restoring peatlands have been adopted since 1980s.

The aim of peatland restoration is to regenerate a self-sustaining naturally functioning mire ecosystem that accumulates carbon and retains nutrients from through-flowing waters (Wheeler and Shaw 1995). There are numerous strategies of peatland restoration ranging from just raising the water table while continuing agricultural use less intensively to flooding the peatland and converting it, for some time, into a shallow lake. The choice of the best restoration method depends on environmental conditions, the specific goals, and the available infrastructure. Until recently, the major goal of peatland restoration has mostly been motivated by nature conservancy. More recently, following the concept of "ecosystem services" (SCEP 1970), the "sponge" function of peatlands (e.g., protection from flooding and, especially their ability to store C) have become the main motivation of peatland restoration.

As restoration of peatlands focusing on the mitigation of GHG emissions is a very recent concept, there is no universally accepted "best practice" suited to all situations. In general, the local situation determines the best practice. However, as a general guideline, it is well known that a permanent water table at −10 to 0 cm below ground minimizes the emissions of GHGs (Couwenberg et al. 2011). Nonetheless, it is difficult to maintain these conditions throughout the year.

Flooding is much easier and also economical in most locations. In Central Europe, some studies have shown extremely high CH_4 release following flooding, which results in elevated net GHG emission compared to the prior drained situation (Augustin and Chojnicki 2008; Glatzel et al. 2011). It is not yet clear, however, whether these emission peaks are representative of the flooding situations, and they are probably transient. Regardless of this, duration of the transient phase is not yet known, and is related to the trophic status of the site. In any case, a transient period of increased CH_4 release may contribute to GHG mitigation when followed by an extended period of C storage and low emissions. Thus, it is necessary to ensure decades of high water table in order to maximize GHG sequestration following flooding.

7.14 Feedbacks to Climate Change

The ability of peatlands to maintain the C-sink function under changing environmental conditions will depend on the balance between C-inputs and outputs (i.e., the net ecosystem C balance, NECB). Furthermore, fluxes of CH_4 and N_2O must also be taken into account (Schulze et al. 2009), and numerous uncertainties exist regarding how the NECB of terrestrial soils will be influenced by increasing concentrations of atmospheric CO_2, warming, and increases in drying and rewetting or freezing and thawing cycles (see Chap. 18). For example, warming and CO_2 fertilization increase biomass production, but also the rate of mineralization. As a result, the NECB may be unaffected if effects of increased biomass production are negated by increase in the rate of SOC-mineralization. The situation is even more complex in hydromorphic soils. Despite processes relevant to terrestrial soils, water level and hydraulic conductivity are dominant controls on biomass production, species composition, decomposition, gas fluxes and production of dissolved organic carbon(DOC) (Limpens et al. 2008).

In pristine peatlands, vegetation influence the C-cycle in several ways: (a) species composition influences the C-input into the ecosystem; (b) litter chemistry determines the rate of decomposition (Limpens and Berendse 2003); (c) diffusion, hydraulic conductivity and redox conditions are influenced by the structure of the peat, which is determined by the peat-forming genus (Limpens et al. 2008); and (d) an increase in graminoid species can increase CH_4 emissions by transport through aerenchymatic tissue (Nilsson et al. 2001). Shifts in vegetation are driven by water level, temperature and N-deposition. Decreasing water level as well as increasing temperature and N-availability stimulate growth of vascular plants at the expense of *Sphagnum* species. This in turn increases the rate of decomposition and, thus, diminishes

C-storage in peatlands. Warming is predicted in all regions with peatlands. However, increasing precipitation in temperate regions may counteract the effects of warming if water level in peatlands can be maintained (Limpens et al. 2008). For some peatland regions in North America and Southeast Asia, significant decreases in precipitation are predicted (Li et al. 2007), and droughts and heat waves are likely to occur more often almost everywhere on the globe (Christensen et al. 2007). Exceptional droughts can increase decomposition, and also increase the risk of fire. Frequency and severity of fire is important for the NECB of several ecosystems. Often, fires affect vegetation and only a few centimeters of the topsoil. However, this may be different in the case of organic soils. Globally, around 2 Pg C year^{-1} (1997–2009) were emitted by fires, 3% of this was ascribed to tropical peat fires and 15% to extra-tropical forest fires (van der Werf et al. 2010). The latter might also include forest with peaty soil. Even in tropical montane cloud forests, organic soils are present due to wet and relatively cool conditions (Schawe et al. 2007) and forests fires are prominent (Roman-Cuesta et al. 2011). While vegetation recovers soon after burning, loss of C from organic soil has been reported to be as large as 0.8–1.8 Tg C year^{-1} in the Andes for the period 2000–2008 (Roman-Cuesta et al. 2011). Not only in the tropics but also in mid-latitude (30–50°N, Poulter et al. 2006) and in northern peatland ecosystems (Turquety et al. 2007; Wieder et al. 2009), fire is an important factor in affecting the NECB. Poulter et al. (2006) estimated that the C-emission by fire was as high as 110 Mg C ha^{-1} for a single fire event in North Carolina. With a fire frequency of less than 20 years, the peat ecosystem would turn into a net-source of CO_2. For a Canadian bog, a fire return interval of <60 years was estimated to turn that ecosystem into a net-source of CO_2 (Wieder et al. 2009). Neither of these studies (Poulter et al. 2006; Wieder et al. 2009) took CH_4 emissions into account.

The examples described above show that an increase in fire frequency may turn peatland ecosystems into sources of CO_2. However, single catastrophic events in small regions can have even more severe impact on emissions. During an abnormally long, dry period caused by El Nino in Indonesia in the year 1997, the peat swamp forest of those regions caught fire. Page et al. (2002) estimated that on average 50 cm of the peat layer was destroyed and between 0.8 and 2.8 Pg C was released into the atmosphere from peat and the burnt vegetation. This emission was equivalent to between 13% and 40% of mean global C-emission from fossil fuel combustion in 1997. Further, the drained peatland was much more affected than that under undisturbed locations. This example shows that peat fires are a severe threat to climate, and emphasize the need for appropriate and careful management of peatland ecosystems.

7.15 Remote Sensing Possibilities to Capture Peat- and Wetland More Precisely

What are the options to improve scientific knowledge on the spatial distribution of peat- and wetlands without having to auger the entire globe?

As already stated in the previous sections, the knowledge about the global spatial coverage and latitudinal distribution of wetlands and in particular of peatlands is of utmost importance for a reliable estimate of the global C resources in soils. Remote sensing is a valuable means to map and monitor the status and changes of land surface properties at scales ranging from regional to global. Coarse to medium resolution remote sensing data deliver a globally consistent and objective source of information for a spatially explicit mapping of the global distribution of potential C stocks in terms of land cover type maps. However, there is still considerable uncertainty in estimates of the area and distribution of the relevant land cover types (e.g. peatlands, wetlands, wet forests) and hence of the stored C globally (Herold et al. 2008; Krankina et al. 2008).

These uncertainties are attributed to a number of limitations that are either determined by the technical specification of the sensor (wavelength, spectral and spatial resolution) or the derived data products (e.g. land cover maps). The following is a brief overview about some of these challenges. Despite its distinct canopy structure (a mix of tree canopy, a shrub layer, and a continuous layer of herbaceous vegetation with presence of mosses), mapping peatlands as a distinct type of land cover from remote sensing data is difficult. Most classifications of land cover focus on a dominant life form and are not well suited to mixtures of several life forms (e.g., Land Cover Classification System (LCCS), Di Gregorio 2005).

This is the reason why peatlands do not occur in most global land cover maps. Instead, they are often associated with water saturated conditions. Therefore, they are usually included in wetlands when defining thematic classes for land cover type mapping. However, it is well known that significant portions of peatlands, especially in the boreal zone do not experience prolonged inundation periods and furthermore, large parts of peatlands in some regions are drained. Considering wetlands, those areas are mapped as a separate class in most global land-cover maps. Because of the global availability of such maps, wetlands often dealt as a proxi to define the occurrence of peatlands in the global C cycling models (Krankina et al. 2008). An overview of the current global land cover products and the thematic representation of peatlands or wetlands respectively is given in Table 7.5.

Besides the occurrence of water, another proxy for peatlands is the vegetation cover. As stated above, peatlands have a characteristic canopy structure but on the other hand, this is also the main limitation when using the remote sensing data to map the spatial extent of peatlands. Although, the spectral signatures of peatland herbaceous vegetation and mosses distinctively differ from other vegetation types (e.g. lower spectral reflectance in the NIR and SWIR region) and they can be mapped from high to very high resolution satellite data, the spatial resolution of globally operating sensor systems and existing land cover maps is not adequate to detect single vegetation types and sometimes even omits the existence of small sized peatlands (Pflugmacher et al. 2007).

An approach to overcome the limitations in pixel size and categorical mapping is the modeling of "continuous fields" of quantitative surface parameters, e.g. the percentage of vegetation or tree cover within one single pixel (sub-pixel analysis, spectral unmixing Hansen et al. 2005). Another option is the synergistic use of passive and active remote sensing systems (e.g., radar or lidar Rosenqvist et al. 2007; Li and

Table 7.5 Global remote sensing based land cover products and representation of wetlands within the data

Product	Sensor	Year	Spatial resolution	Relevant class for wetlands	Scientific reference
GLCC	AVHRR	1992	1 km	Herbaceous wetland/wooded wetland	Loveland et al. (2000)
GLC2000	SPOT-VGT	2000	1 km	Regularly flooded shrub and/or herbaceous cover	Bartholomé and Belward (2005)
MODIS land-cover	MODIS	2000	1 km	Permanent wetlands	Friedl et al. (2002)
GlobCover	MERIS	2005	300 m	Closed to open (>15%) grassland or woody vegetation on regularly flooded or waterlogged soil	Arino et al. (2008)

Chen 2005). The advantage of radar sensors is that they can penetrate clouds and that they are sensitive to variations in top soil properties (e.g., soil moisture, grain size). However, global mapping based on active sensor systems is cost-intensive and time consuming and at present, there are no global products available with regard to wetland and/or peatland mapping.

Summarizing, the representation of peatlands in global land cover mapping as well as the definition and mapping accuracy of other relevant land cover types for the spatially explicit estimation of global C resources is challenging and should be an important objective for future global mapping campaigns. An effective and low-cost solution could be to build on existing global estimates of wetlands and other classes and use the crowdsourcing approach (Fritz et al. 2009) to validate and adjust global land cover maps for specific use in biogeochemical modeling.

7.16 Conclusions

Will peatland and other soil C-rich ecosystems maintain their C sink function remains to be the major climate change related question to be answered by soil scientists and other biogeochemists. Practical advice how to manage land use on soil rich in C cannot be given with simple answers which are applicable around the world. Rather than altogether ignoring these questions, it is important to address them systematically and objectively. It is difficult to exactly know how much vulnerable soil C recourses there are. For realistic bottom-up approaches to estimate potential feedbacks between soils and the climate systems, it is important to develop more precise inventories of soil C resources because net GHG exchanges are not only process-driven but also determined by area of OM (C and N) hotspots. This article has provided not only some rough global estimates on soil C estimates based on simple assumptions but also differentiated these into different zones (latitudes) and soil types. This information may trigger more research on the spatial distribution of soil C and attempts to divide these estimates into vulnerable and less vulnerable soil C stocks. In this context, the scientific communities of peatland research should join forces with all groups focusing on other C- rich soils. Furthermore a focus mainly on the northern (50–70° N) and the tropical latitudes (20°S–10°N) may not be sufficient for a global perspective.

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