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Review

Carbon losses from soil and its consequences for land-use management

Julian J.C. Dawson *, Pete Smith

School of Biological Sciences, University of Aberdeen, Cruickshank Building, St Machar Drive, Aberdeen, AB24 3UU, UK

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Abstract

This paper reviews our current knowledge and understanding of carbon processes in the terrestrial ecosystem with a view to reducing soil carbon losses by optimising land-use and land management. Processes that influence the fate of carbon (in both terms of quantity and quality) are important in determining soil fertility, quality and health as well as consequences for future environmental change scenarios. We need to understand the processes that determine soil carbon losses and the fate of the carbon once lost from the soil in order to provide sustainable solutions for mitigating these carbon losses as part of land management "best practice" and balancing national carbon budgets. Here we review the amount of carbon within the UK terrestrial pool, the processes involved and factors influencing carbon transport to and from soils, the fate of the carbon once it has been lost from the soil environment and land-use scenarios that affect carbon losses. We conclude with possible management options to reduce soil carbon loss and identify gaps in knowledge in order to better understand carbon processes in the terrestrial environment.

Keywords: Carbon; Soil; Land-use; Management; Climate

Contents

* Corresponding author. Tel.: +44 1224 272259; fax: +44 1224 272703.
E-mail addresses: j.j.dawson@abdn.ac.uk (J.J.C. Dawson), pete.smith@abdn.ac.uk (P. Smith).

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1. Introduction

A recent paper by Bellamy et al. (2005) has suggested losses of carbon (C) from soils in England and Wales has occurred at a mean rate of 0.6% yr^{-1} over the period 1978–2003. This stimulated much interest in the possible mechanisms for that loss. Moreover, this work described data from the upper 15 cm of soil, ignoring C stocks and processes at depth. Although climate change may be partly responsible (Bellamy et al., 2005), the UK has undergone substantial land-use/management changes within the same time period. Arable soils that tend to be more depleted in soil organic C, have shown increases in area with concomitant reductions in permanent and temporary grasslands (King et al., 2005). To provide sustainable solutions for managing and mitigating these C losses, this paper, with particular reference to the UK situation, describes our current understanding of processes (climatic, edaphic and land-use and management) determining C losses from soils, and the fate of that C. Here, we review estimates of the amount of C within the UK terrestrial pool, the processes involved and factors influencing C transport to and from soils and the rate of C loss. We conclude our review by suggesting possible management options to reduce soil C loss and identifying gaps in our understanding of soil C processes.

2. Terrestrial carbon pools

The C cycle can be conceptualised as a three compartment system comprising terrestrial (soil and vegetation), atmospheric and oceanic pools (IPCC, 2001; Janzen, 2004). Climate, geology and land management practices are the main factors determining soil and vegetation type, and thus are initial controls on the amount of C within the terrestrial pool (Hope et al., 1994; Yoo et al., 2006). Previously, extrapolating data from vegetation or soil groups, life zone classes and land cover have been used to estimate terrestrial C stores at a set time point (Dixon and Turner, 1991; Milne and Brown, 1997; Bradley et al., 2005). Moreover, decadal changes in soil C have been determined at sites in the UK, such as arable soils under different long-term management regimes (Blair et al., 2006; Kemmitt et al., 2006) and forest soils (Billett et al., 1990).

2.1. Carbon in UK vegetation

Despite accounting for <5% of the UK terrestrial C pool, the changes in vegetation type, distribution and productivity are integral to whether terrestrial systems act as C sinks or sources (Milne and Brown, 1997). Vegetation directly affects atmospheric CO_2 concentrations (and other greenhouse gases); influences heat and moisture exchange between land and the atmosphere and prevents soil erosion (Krinner et al., 2005). By combining established databases, biomass partitioning, forest inventories, ecological surveys, land classification and remote-sensing techniques, Milne and Brown (1997) derived spatial databases describing C in dominant vegetation types (agriculture, forestry, semi-natural and permanent grasslands). Other land categories such as rock, bare ground and urban areas were assumed to have no C. However, recently, the C content of gardens has also been estimated (Bradley et al., 2005). The vegetation C pool of

Great Britain (GB — England, Wales and Scotland) was estimated at $113.8 \pm 25.6 \times 10^{12}$ g (Milne and Brown, 1997) with Northern Ireland representing an additional $3.8–4.4 \times 10^{12}$ g (Cruickshank et al., 1998, 2000).

Milne and Brown (1997) have detailed the distribution of vegetative C and the different C densities of vegetation cover types in GB. Non-forested cover types, e.g. arable crops, pasture, horticulture and unimproved grasslands contained *ca.* 1×10^3 kg C ha^{-1} and shrub, heath and bogs had *ca.* 2×10^3 kg C ha^{-1} in their vegetation. The densities of C in forests, totalling 80.1% of the vegetative C in GB (55% in Northern Ireland), were calculated by converting timber volumes into biomass C. The average net C density, weighted by the area of species distribution with age, for 15 main forest species varied between from 4.3 to 34.9×10^3 kg C ha^{-1} for conifers and 36.4 to 90.6×10^3 kg C ha^{-1} for broadleaf trees Milne and Brown (1997). Conifers have a lower net C density than equivalent aged broadleaved species as well as a higher incidence of younger forest stands in the UK (Cannell, 1999).

2.2. Carbon in UK soils

The majority of C in the UK terrestrial pool is stored belowground in soils. By integrating soil series databases and land cover, the total amount of C in the soils of GB was estimated at $9838 \pm 2463 \times 10^{12}$ g (6948×10^{12} g in Scotland and 2890×10^{12} g in England and Wales). The error term was mainly due to uncertainties in bulk density and depth of Scottish peats. Peatlands (upland blanket, lowland raised bogs and fen peats) consisted of 5162×10^{12} g of C; the majority (4523×10^{12} g of C) in Scotland. In England and Wales, stagnogleys, brown earth and raw peat soils contained the majority of C. A detailed study of C contents and distribution of soils of GB has been described by Milne and Brown (1997). In a separate study, the total soil C of Northern Ireland was estimated at 386×10^{12} g of C (Cruickshank et al., 1998), with 42% of the C present in peat soils.

However, a work on a typical British moorland soils in the North Pennines (Moor House NNR), comparing true surface areas, with the broader based national inventory data for the same area, estimated soil C stores were 3 times lower in the more detailed smaller-scale study (Garnett et al., 2001). Another cause of variability in estimates has been partly due to sampling to different soil depths (Soussana et al., 2004). Bradley et al. (2005) recently assembled a new soil C and land-use database for the whole of the UK with weighted soil parameters (bulk density, organic C and texture) at 2 separate depths of 0–30 cm and 30–100 cm. These recently revised estimates

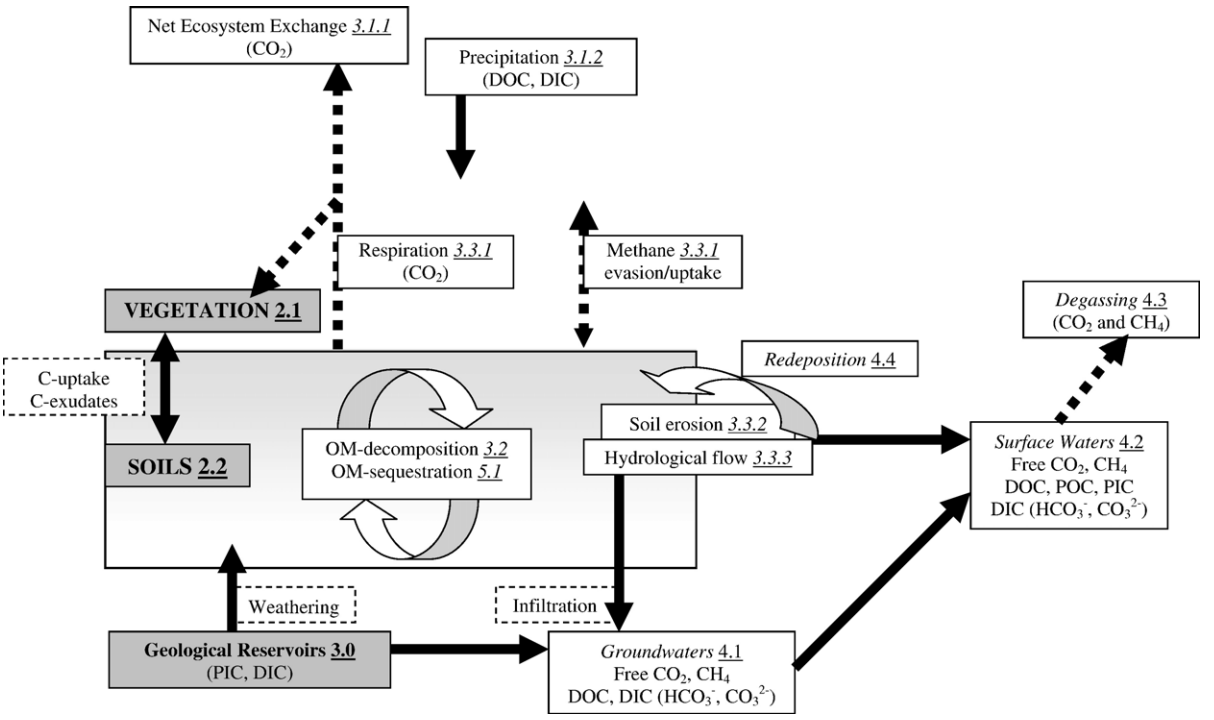

Fig. 1. Terrestrial Ecosystem Pathways and processes to determine a total carbon budget (Worrall et al., 2003a; Billett et al., 2004). The numbers relate to the section of the review where each process is described.

suggested that soil C in GB was 5434×10^{12} g in Scotland, including 3248×10^{12} g in peats below 1 m (Milne et al., 2001), and 2079×10^{12} g in England and Wales, totalling 7513×10^{12} g; a reduction by *ca*. 24% on the estimate by Milne and Brown (1997).

3. Processes affecting the turnover of C in soil

Carbon is sequestered in soils from the atmosphere, in organic form via vegetation deposition and the accumulation of recalcitrant organic matter (OM), and inorganically from parent materials by bicarbonate weathering of silicate minerals (Chadwick et al., 1994). Terrestrial C is subject to recurrent dynamic processes that determine its subsequent fate in the wider environment (Fig. 1). The main form that C enters the terrestrial biosphere is as carbon dioxide (CO_2). This is part of a dynamic equilibrium between CO_2 inputs as gross primary production (GPP) and losses of CO_2 via respiration from both vegetation and soil biomass (Cao and Woodward, 1998). On average, *ca*. 40–50% of C inputs from GPP by photosynthesis is used to support autotrophic respiration and returned back to the atmosphere as CO_2; the remainder is fixed as net primary production (NPP) in vegetative biomass (Schlesinger and Andrews, 2000; Janzen, 2004), which varies with climate, plant productivity and season (Falge et al., 2002). The NPP enters the soil as organic C in the short-term (hours–days) and medium-term (weeks–months) by rhizodeposition or in the longer-term by decomposition of litter and fragmented plant structures (Ostle et al., 2000; Jones and Donnelly, 2004). Once in the soil, it is converted back to CO_2 and CH_4 via soil (heterotrophic) respiration processes.

Net ecosystem production (NEP) has been defined as the difference between GPP and ecosystem respiration (Lovett et al., 2006) and "represents the total amount of organic carbon in an ecosystem available for storage, export as organic carbon, or non-biological oxidation processes". Loss mechanisms such as harvest, fire and insect damage remove C and the potential for plants to acquire C for a period of time, which combined with riverine export and the respiratory processes, counterbalances the terrestrial CO_2 input from GPP. The equilibrium between these processes determines whether terrestrial ecosystems act as a C source or sink and termed net biome production (Cao and Woodward, 1998; IGBP, 1998; Schlesinger and Andrews, 2000; Janzen, 2004). Recently, it has been suggested that the rate of accumulation in an ecosystem is equivalent to its NEP, contrary to the historical definition, which links more coherently with NPP and respiration processes enabling non-respiratory ecosystem fate mechanisms to be determined separately (Lovett et al., 2006).

Processes are inter-related and temporally diverse with terrestrial C turnover time varying from days to thousands of years (Jenkinson, 1990; Post and Kwon, 2000). Belowground soil C tends to turnover more slowly than vegetative C and is better protected from disturbance (IGBP, 1998).

3.1. Inputs of carbon to soil

3.1.1. Inputs of carbon from the atmosphere via vegetation

Net ecosystem exchange (NEE) of CO_2 shows major diurnal fluctuations (Beverland et al., 1996; Valentini et al., 2000) as photosynthetic processes dominate during daylight hours (Hargreaves et al., 2003). Moreover, NEE values also vary seasonally and by land-use and its management. It has been estimated that at the UK level, forestry ($-0.11 \pm 0.04 \times 10^3$ kg C ha^{-1} yr^{-1}) and grasslands ($-0.24 \pm 0.20 \times 10^3$ kg C ha^{-1} yr^{-1}) have negative NEE — ecosystem C gain; croplands ($0.14 \pm 0.10 \times 10^3$ kg C ha^{-1} yr^{-1}) generally have positive NEE — ecosystem C loss. Peatlands ($0.28 \pm 0.13 \times 10^3$ kg C ha^{-1} yr^{-1}) across the UK also show C losses (Janssens et al., 2005), although this is not always the case, particularly from the undrained peatlands alone (-0.2 to -0.5×10^3 kg C ha^{-1} yr^{-1}, Cannell et al., 1999) and individual studies (-0.25 to $-0.28 \pm 0.03 \times 10^3$ kg C ha^{-1} yr^{-1}, Hargreaves et al., 2003; Billett et al., 2004). In a pan-European study, the NEE for the UK suggested an overall loss of terrestrial C ($0.06 \pm 0.26 \times 10^3$ kg C ha^{-1} yr^{-1}, Janssens et al., 2005), although Milne and Mobbs (2005) estimated a slight yearly increase in C since 1998 for the UK (215×10^9 g C yr^{-1} in 2000), mainly due to C influxes from forestry, but there is a high degree of uncertainty in both studies.

3.1.2. Inputs of carbon from the atmosphere via deposition

External sources of C enter the terrestrial biosphere via wet, occult and dry deposition (Cerný et al., 1994; Vautz et al., 2003). On a mean annual basis, rainfall in the temperate zone contains dissolved organic carbon (DOC) concentrations in the range of 0.82–2.00 mg C L^{-1} (Meyer and Tate, 1983; Koprivnjak and Moore, 1992; Willey et al., 2000); dissolved inorganic carbon (DIC) is generally lower than DOC (Likens et al., 1977; Willey et al., 2000). Occult deposition on vegetation can comprise 30% of the total input of precipitation (Burgess and Dawson, 2004) but is highly dependent on vegetation type (height and surface area) and local climate (temporal variability in mist, fog and cloud production). Dry deposition is the retention of particulate matter by vegetative, soil or water surfaces (Cresser and Edwards, 1987).

As precipitation falls onto vegetation, its organic C content generally increases before reaching the soil surface (Moore, 2003). DOC concentrations can be 3 and 10 times higher in throughfall and stemflow, respectively, than in the original precipitation from leaching of plant-derived organic acids but this will vary with vegetation type (Koprivnjak and Moore, 1992; Edmonds et al., 1995; Inagaki et al., 1995).

As precipitation infiltrates the soil profile, biogeochemical C cycling increases its DOC and gaseous C concentration. Hydrological variability in terms of spatial and temporal changes in flow path characteristics within soil horizons can be as important as biological activity in determining the actual C concentrations in soil waters (Kalbitz et al., 2000). Lower down the soil profile, removal (adsorption) of DOC from the soil water occurs via organo-mineral interactions on the surfaces of Fe and Al oxides and hydroxides in the mineral horizons (Hope et al., 1994; Kalbitz et al., 2000; Moore, 2003). Other factors governing DOC retention in the mineral horizons include pH and texture, particularly silt and clay particles and formation of soil aggregates (Hagedorn et al., 2000; Jones and Donnelly, 2004). In highly organic soils, only small reductions in soil water DOC concentrations occur with depth due to a lack of available mineral binding capacity (Fiebig et al., 1990; Grieve, 1990a).

3.2. Decomposition of carbon in terrestrial ecosystems

Temperature, plant residue composition, nutrient and substrate (OM) availability determine the rate of biologically driven OM decomposition and mineralisation to DOC, CH_4 and CO_2 (Moore et al., 1998; Martens, 2000; Blanco-Canqui and Lal, 2004). These parameters interact with soil physico-chemical factors including moisture, pH, redox potential, inhibitors (e.g. trace metal ions, phenolics) and activators (e.g. cations) that influence the activity of the soil decomposer community (micro-organisms, fungi, invertebrates) and OM degradative extracellular enzymes (Freeman et al., 2001a; Bonnett et al., 2006). The rate of CO_2 production from OM is regulated through aerobic and fermentative respiration. Under anaerobic conditions, DOC and CH_4 (by methanogenesis) rather than CO_2 may be the foremost end product of decomposition, although proportions change as aerobic conditions prevail (Freeman et al., 2004a).

Soil temperature, in particular, is a major factor in the regulation of decomposition rate, which influences observed seasonal patterns of higher soil DOC and CO_2 concentrations in the summer compared to the colder winter months (Castelle and Galloway, 1990; Jones and Mulholland, 1998a; Hope et al., 2004; Bonnett et al., 2006). However, the temperature sensitivity of soil OM decomposition remains a topic for debate (Davidson and Janssens, 2006) with some studies suggesting that recalcitrant OM is not sensitive to temperature variation (Giardina and Ryan, 2000), others suggesting that non-labile OM is more sensitive to temperature than labile pools (Knorr et al., 2005), or that recalcitrant and labile pools have a similar temperature sensitivity (Fang et al., 2005).

Although plants tend to be composed of similar C compounds, e.g. celluloses, starches, proteins, lipids and phenolic compounds, the proportions of each depends on species type and age. The C:N ratio can be used to explain residue quality influencing turnover rates and the type of C strongly impacts its ability to be decomposed, with lignin a particularly recalcitrant form (Cheshire and Chapman, 1996; Martens, 2000). Physical and chemical protection of soil OM depends on the ease with which it can be encapsulated within stable aggregates. The quality and quantity of plant residues determine the amount of soil C in these aggregates. However, the physico-chemical interactions between soil OM and its mineral constituents and "the mechanisms responsible for the dynamics of aggregate formation and stability," with respect to C decomposition in soils are not well understood or quantified (Blanco-Canqui and Lal, 2004; Basile-Doelsch et al., 2005).

Three main methods of organic C stabilisation have been proposed: micro-aggregation (53–250 μm) formation within macro-aggregates; physically binding with clay and silt particles and biochemically by formation of recalcitrant soil OM compounds (Six et al., 2002; Bronick and Lal, 2005; Sleutel et al., 2006). Light fractions and particulate OM that do not bind within aggregates (unprotected OM) generally remain more susceptible to microbial decomposition (Six et al., 2002; Sleutel et al., 2006). This labile OM decomposes more rapidly but tends to only constitute 3–5% of total OM (Jones and Donnelly, 2004). The majority of soil OM occurs in micro and macro-aggregates that form around the light fractions through binding of C to mineral particles to form heavy fractions of organic C; these have residence times of decades (Post and Kwon, 2000). Recalcitrant material that is physically or biochemically protected may have turnover times of hundreds to thousands of years (Jenkinson, 1990; Post and Kwon, 2000).

Disturbance increases breakdown of macro-aggregates, which "diminishes C stabilisation in newly formed micro-aggregates within macro-aggregates" (Six et al., 2004). However, C can be protected from decomposition by placement in the intra-aggregate or organo-mineral complexes in deeper mineral horizons (Jones and Donnelly, 2004). Alkaline soils also decrease decomposition as precipitation of hydroxides, oxides, phosphates and

inorganic C enhance aggregation by the formation of secondary carbonates (Bronick and Lal, 2005). The rate of decomposition varies considerably but it is ultimately determined by the accessibility of the microbial community to the soil OM. Thus, decomposition/mineralisation of OM results in losses of soil C as CO_2 and CH_4 to the atmosphere; as POC and DOC by erosion processes and as losses of gaseous, dissolved and particulate C to surface waters.

3.3. Losses of carbon from soils

3.3.1. Losses to the atmosphere

The main losses of C from soil is in the form of CO_2 from OM mineralisation although fires cause direct C emissions to the atmosphere and changes to vegetation and species composition (Harden et al., 2000; Mollicone et al., 2002) that affect the dynamics of terrestrial C stores for subsequent decades (Lal, 2004a). The gaseous C efflux from soils is initially dependent on the rate of production of CO_2 (or CH_4) within the soil–plant root system, and subsequently on the rate of gaseous diffusion and mass flow from soil waters to the atmosphere; a function of soil moisture and textural properties (Pauss et al., 1990; Skiba and Cresser, 1991; Whalen and Reeburgh, 2000). The rates of soil CO_2 efflux in UK ecosystems can range from 0.16–1.0 g C m^{-2} d^{-1} in peatland and wetland systems (Fowler et al., 1995a; Worrall et al., 2005; Lloyd, 2006); 3.4–3.8 g C m^{-2} d^{-1} in forested areas (Valentini et al., 2000) although this varies further with soil type, depth and species (Fang and Moncrieff, 2005) and 1.2–7.8 g C m^{-2} d^{-1} in shrub and heathland soils (Jensen et al., 2003; Emmett et al., 2004). The efflux of CO_2 from non-organic managed grassland and cropland systems varies substantially due to the variety of land practices undertaken but tends to be greater with disturbance (Fortin et al., 1996; Vleeshouwers and Verhagen, 2002; Bol et al., 2003; Lohila et al., 2004; Kemmitt et al., 2006). Moreover, high soil CO_2 effluxes from agricultural organic soils in Finland have been measured at 1.8–25.9 g C m^{-2} d^{-1} (Maljanen et al., 2001).

Methane is lost to the atmosphere by diffusion, ebullition or via vascular plants (Moore et al., 1998; Whalen, 2005) but generally is confined to relatively localised wetland areas. The estimates vary both temporally and spatially (0.43–236 mg C m^{-2} d^{-1}) as landscape changes from vegetation covered fenlands (lower rates of CH_4 losses), via bogs to open water ponds (higher rates of CH_4 losses) with corresponding differences in water table height (Fowler et al., 1995b; Hargreaves and Fowler, 1998; MacDonald et al., 1998; Cao et al., 1998; Heikkinen et al., 2002). However, CO_2 losses in the same wetland tend to be *ca.*100 times higher than CH_4 effluxes (Fowler et al., 1995a) although relative greenhouse gas warming potentials of CH_4 compared to CO_2 is significantly (30 times) higher (Ashmore, 1990). In addition, *ca.* 20–55% of CH_4 is oxidized in aerobic surface horizons prior to diffusion to the atmosphere (Cao et al., 1998; Whalen and Reeburgh, 2000; Whalen, 2005) and methanotrophic CH_4 consumption can also occur (Whalen and Reeburgh, 2000; Whalen, 2005), particularly in organic agricultural soils (CH_4 uptake of 0.27 mg C m^{-2} d^{-1}, Maljanen et al., 2004). Ebullition can comprise 17–52% of the total CH_4 efflux in some Swedish wetlands (Christensen et al., 2003).

3.3.2. Losses due to soil erosion

Erosion studies have ranged from site-specific plots to regional scales (McHugh et al., 2002). The main parameters that are assessed include erosion rate and the distribution and degree of erosion within the area under investigation (Evans and Brazier, 2005). The erosion on hillslopes is highly variable; vegetation that maintains stability can be sensitive to disturbance and land-use changes, potentially enhancing mobilisation of surface soils by water (Stott and Mount, 2004). A study of *ca.* 400 sites in the uplands of England and Wales showed that there was extensive degraded land (a total of *ca.* 2.5% of the area surveyed), of which 50% lacked vegetative cover and were likely to erode further at a faster rate. However, there are large variations in percentage erosion for areas under investigation (Table 1), mainly due to the spatially heterogeneous nature of the process. Moreover, its episodic nature makes erosion processes over time difficult to assess (Evans and Warburton, 2005). Altitude and slope have been correlated with extent of erosion, although impacts by humans (recreational and land management) and animals via overstocking were also important contributors to the erosional process (McHugh et al., 2002). On many farmland hillslopes, erosion rates from tillage by mechanised agriculture were similar to erosion rates caused by water (Govers et al., 1999).

There are 2 types of erosion by water; sheet erosion as flow over vegetated surfaces and channel erosion, which is limited to where soils lack vegetative cover. Overland flow occurs, removing topsoil and hence substantial OM translocation, when runoff is greater than the infiltration capacity of the soil (Evans and Brazier, 2005). Carbon and nutrients from water eroded soil is relocated downslope from one area to another or transported to surface waters (Stallard, 1998; Smith et al., 2001c; Liu et al., 2003). For soils close to the river network, higher stream discharges from increased stream size and energy enhance erosion processes (Lawler et al., 1997). Again, river bank erosion and subsequent suspended sediment production (Table 1) can vary considerably within and between sites (Stott and

Table 1
Soil erosion rates in UK terrestrial ecosystems

Details of study	Land erosion (% of site area)	Reference
[a]Blanket peat, Moor House, N. Pennines	17%	Warburton (2003)
399 field sites (NSI), upland England and Wales	2.46%	McHugh et al. (2002)
Modelled uplands, using soil associations and erosion risk	1.84%	Evans (2002)
250 field sites (NSI), England and Wales	45% (of field sites)	Harrod and Fraser (1998)
Bare peat and soil, Bakewell, Peak District	1.6%	Phillips et al. (1981)
Catchment, Peak District	2%.	Evans (1974)
Valley slopes, Plynlimon	19%	Thomas (1965)
[b]*Riverbank erosion (mm yr^{-1})*		
R. Arrow, Warwickshire	32.6	Couper and Maddock (2001)
R. Arrow, Warwickshire, post flood events	59.8	Couper and Maddock (2001)
R. Swale — moorland, Yorkshire	70–260	Grove (2000)
Cyff — moorland, Plynlimon, Wales	31–66	Stott (1999)
Tanllwyth — forest, Plynlimon, Wales	35 (forest) 89 (harvesting)	Stott (1999)
Monachyle — moorland, Trossachs, Central Scotland	59	Stott (1997)
Kirkton — forest, Trossachs, Central Scotland	47	Stott (1997)
Tanllwyth, Hafren and Hore — Plynlimon forest catchments, Wales	12.9–64.0	Bull (1996)
River Ilston, West Glamorgan, Wales	27	Lawler (1993)
Narrator Brook — moorland, Devon	0.7	Murgatroyd and Ternan (1983)
Narrator Brook — forest, Devon	5.2	Murgatroyd and Ternan (1983)
Suspended sediment (×10^3 kg ha^{-1} yr^{-1})		
15 sites, mid-Wales, Northern England and Central Scotland uplands		Stott and Mount (2004)
Undisturbed moorland	0.15 (0.01–0.41)	Stott and Mount (2004)
Ploughing and ditching	0.42 (0.03–1.22)	Stott and Mount (2004)
Mature forest	0.30 (0.15–1.18)	Stott and Mount (2004)
Harvesting forest	1.63 (0.44–4.63)	Stott and Mount (2004)
Smisby, Derbyshire — arable lowland	0.95–5.56	Walling et al. (2002)
Smisby, Derbyshire — pasture lowland	0.02–3.54	Walling et al. (2002)
Rosemund, Herefordshire — arable lowland	0.33–5.30	Walling et al. (2002)
13 rivers, NE England	0.13–0.35	Robson and Neal (1997)
13 rivers, SW England	0.04–0.58	Walling and Webb (1987)

[a]Data mainly from McHugh et al. (2002).
[b]Data mainly from Stott and Mount (2004).

Mount, 2004) impacting on C cycling processes downstream or in lakes (Walling et al., 2002; Stott and Mount, 2004).

The amount of C mobilized by erosion processes for England and Wales has been estimated at 0.20–0.76×10^{12} g C yr^{-1}, of which 0.08–0.29×10^{12} g C yr^{-1} was re-deposited and 0.12–0.46×10^{12} g C yr^{-1} was transported to surface waters (Quinton et al., 2006). Certain soils are more prone to erosion; erodible soil textures tend to be lighter fine sand or coarse silt (Brazier et al., 2001; Evans and Brazier, 2005). The maximum soil erosion rate indicates its "full erosive potential"; 3.30×10^3 kg ha^{-1} yr^{-1} for clays and medium loams, sands and light loams were 108.3×10^3 kg ha^{-1} yr^{-1} (Brazier, 2004). In studies across Britain, mean soil erosion varied between 0.22×10^3 kg ha^{-1} yr^{-1} to 11.3×10^3 kg ha^{-1} yr^{-1} (Quine and Walling, 1991; Walling and Quine, 1991; Brazier et al., 2001). Moreover, where localized erosion has been measured, rates $>100\times10^3$ kg ha^{-1} yr^{-1} have occurred (Brazier, 2004).

Wind erosion can cause a substantial loss of soil C in exposed areas and can be transported thousands of km before redeposition, most of it on land (Smith et al., 2001c; Yan et al., 2005). Factors, such as distribution and speed of wind and soil erodibility parameters including dryness, texture and land-use establish surface roughness of non-vegetated soils to determine the temporal and spatial differences in wind erosion (Böhner et al., 2003; Yan et al., 2005). In terms of the UK agricultural land, East Anglia, Midlands and Yorkshire are most affected by wind erosion. At UK sites in East Anglia, the highest rates of wind derived soil erosion were estimated at $>3.00\times10^3$ kg ha^{-1} in March, September and November (Warburton, 2003; Böhner et al., 2003). Moreover, the prevalence of high winds in upland ecosystems and organic bare soils with low bulk density, where overgrazing or recreational

Table 2
Potential changes in soil C storage in terms of conversion of land-uses

Land-use change	Net C rate[a] and *uncertainty* ($\times 10^3$ kg C ha^{-1} yr^{-1})	Reference
Arable to ley:arable rotation	1.6	Smith et al. (1997)
Arable to grassland (50 yr)	0.3–0.8	IPCC (2000)
Arable to grassland (35 yr)	0.63	Jenkinson et al. (1987)
Arable to grassland (15–25 yr)	0.3–1.9±0.6, *110%*	(Vleeshouwers and Verhagen, 2002; Guo and Gifford, 2002; Murty et al., 2002)
Arable to grassland short leys (20 yr)	0.35	Soussana et al. (2004)
Arable to permanent pasture	0.27	Post and Kwon (2000)
Arable to forestry (115 yr)	0.52+1.53(C in veg.)	Hooker and Compton (2003)
Arable to forestry	0.62+2.8(C in veg.)	(Smith et al., 2000; Falloon et al., 2004)
Arable to forestry (25 yr)	0.3–0.6, *>50%*	(Guo and Gifford, 2002; Murty et al., 2002)
Arable to forestry	0.5–1.4, *>50%*	Maljanen et al. (2001)
Permanent crops to arable	−0.6 and 1.0–1.7, *>50%*	(Smith et al., 1996; Guo and Gifford, 2002; Murty et al., 2002)
Grassland — arable (20 yr)	−0.95±0.3, *95% CI*	Soussana et al. (2004)
Grassland — arable	−1.0 to −1.7, *>50%*	(Smith et al., 1996; Guo and Gifford, 2002; Murty et al., 2002)
Grassland — afforestation (general, 90 yr)	0.1±0.02, *95% CI*	Soussana et al. (2004)
Moorland — grassland	−0.9 to −1.1	Soussana et al. (2004)
Forestry — arable	−0.6	(Guo and Gifford, 2002; Murty et al., 2002)
Forestry — grassland	−0.1±0.1, *95% CI*	Soussana et al. (2004)
Native vegetation — grassland	0.35	Conant et al. (2001)
Peatland — cultivation	−2.2 to −5.4	Freibauer et al. (2004)
Wetland — arable (temperate and boreal)	−1.0 to −19	Watson et al. (2000)
Wetland restoration	0.1–1.0	Watson et al. (2000)
Revegetation on abandoned arable	0.3–0.6, *>50%*	Poulton (1996b)
Revegetation on wetlands from arable	2.2–4.6, *>50%*	Kamp et al. (2001)
Revegetation on wetlands from grassland	0.8–3.9,*>50%*	Kamp et al. (2001)
Conservation	>2.2, *>50%*	Freibauer et al. (2004)

[a]+ve value indicates soil C-gains, −ve value indicates soil C-losses.

activities occur, means that wind, as well as water, is a major cause of peat erosion. Mean net erosion rates from upland peaty sites at Moor House NNR were calculated at 0.47×10^3 kg ha^{-1} yr^{-1} (Warburton, 2003).

3.3.3. Losses via hydrological processes

The first stage in the loss of soil C to surface waters is the movement of OM from soil to interstitial pore waters. Flow paths are mainly determined by topography and soil type. Soil water, travels preferentially along flow paths and is modified as it moves through organic and mineral soils to the groundwater and the river network. Indeed, the distribution of the soils in relation to the drainage system can be a major influence on C inputs to surface waters (Billett and Cresser, 1992), in particular riparian soils in controlling C flow (Fiebig et al., 1990; Fiebig and Lock, 1991; Smart et al., 2001). However, the linkage of soil with the drainage system is temporally dynamic and liable to disturbance (Fraser et al., 2001; Billett et al., 2006).

The transport and temporal variability of DOC and POC losses from the bulk soil is linked to discharge in response to precipitation events and changing flow paths through different soil types containing contrasting amounts of OM (Grieve, 1990b; Hinton et al., 1998). Significant positive correlations between DOC, POC and discharge have been found in many UK catchments (Walling and Webb, 1981; Tipping et al., 1988; Grieve, 1994; Hope et al., 1997a; Dawson et al., 2002) as old soil water containing DOC and gaseous C is flushed out (Grieve, 1990b). Moreover, previously dry areas of the soil and river channel become immersed delivering further C to the stream (Pearce et al., 1986). The relationship between POC and discharge is more complex than that of DOC and discharge because POC requires a threshold energy within the stream before significant transport from the soil occurs (Thurman, 1985; Lawler et al., 1997).

In acid organic soils, DIC (HCO_3^- and CO_3^{2-}) concentrations in soil waters are lower than in alkaline mineral soils as the weathering and dissolution of alkaline parent material is substantial, forcing the carbonate equilibria (including free CO_2) towards dissolved anionic forms of C; the proportion being dependent on pH, temperature and ionic content of the water (Stumm and Morgan, 1981). The concentrations of DIC show an inverse relationship with discharge in UK catchments, as the contribution of deeper groundwater is reduced and the influence of soil water from the upper organic horizons

increases at higher discharges (Reid et al., 1981; Hill and Neal, 1997; Neal et al., 1997a,b; Dawson et al., 2002, 2004).

The porosity and moisture levels of soil controls the proportion of gaseous C that is transported via soil pore waters to streams (Skiba and Cresser, 1991; Dawson et al., 2001). In, well aerated, freely draining mineral soils, the rate of respired CO_2 lost from the soil surface can be high and losses to the streams only constitute a minor component of the CO_2 losses. However, deep wet organic peats store dissolved CO_2 and CH_4 in a solution more readily and gaseous evasion from the surface is reduced. Instead, CO_2 and CH_4 are transported laterally in the soil pore waters increasing their losses via hydrological processes (Clymo and Pearce, 1995; Fowler et al., 1995a; Billett et al., 2004).

During individual storms, the properties of soil pore waters and hence surface water chemistry can change substantially (Soulsby, 1995; Hagedorn et al., 2000). Changes will vary depending on soil type, the dominant influence on flow tends to be the top 15–20 cm surface layer (Bishop et al., 1990 in Hope et al., 1994; Worrall et al., 2003a), i.e. the organic horizon containing the majority of labile C. Peak concentrations of DOC coincide with or are close to maximum discharge (Chapman et al., 1993; Soulsby, 1995; Dawson et al., 2002, 2004). In relation to particulate material, 60% of the overall suspended sediment load was transported in only 2% of the time in the River Exe and a 90% flux in 10% time (Walling and Webb, 1985). Although many studies have shown strong relationships between C concentrations and discharge for a given soil-surface water system, catchment specific factors that affect the actual relationship include the soil and vegetation type and hence quantity of potentially available OM for export. Moreover, seasonal and hydrological (e.g. hysteresis and substrate exhaustion following consecutive storm events) variations in C concentrations for a particular discharge, also increase variability (Walling, 1977; Dawson et al., 2002).

3.4. Climate and land-use change effects on soil carbon losses

Any change in the mechanisms that control soil biogeochemical C cycling, either via climate; i.e. temperature, precipitation, water balance, seasonality and CO_2 stimulation of NPP (Freeman et al., 2004b) or land-use will change that aspect of C cycling. It must be remembered that these processes will not act upon soil C losses in isolation but interact closely with each other. The increases in temperature may initially lead to higher microbial decomposition rates, but over a longer period, these C losses maybe counter-balanced by increases in net C inputs to the soil resulting from higher NPP (Smith et al., 2005a) and an acceleration of temperature-dependent soil physico-chemical processes such as adsorption and other C sequestering reactions (Thornley and Cannell, 2001). Moreover, although climate is recognised as a major factor in soil development, it is important to note that changes to terrestrial ecosystems from land management choices may also influence the future climate (Foley et al., 2003).

While both climate and land-use changes maybe key to understanding future terrestrial C pools and fluxes, new technology will also have significant effects on land management practices, e.g. use of machinery, pesticides, herbicides in agriculture in the future. The uncertainty associated with future technological developments remains large (Metting et al., 2001; Smith et al., 2005a). Table 2 gives an indication of the estimated gains/losses of soil C for a range of land-use changes, as detailed extensively by Freibauer et al. (2004) and Soussana et al. (2004). The degree of uncertainty in this data is either due to lack of relevant studies e.g. forest to grassland pasture or variations caused by contrasting management regimes on the same land-use type, particularly arable and grasslands (Soussana et al., 2004).

3.4.1. Agriculture — cropland and grassland

Carbon losses from agricultural soils occur due to increased soil OM decomposition rates upon cultivation and losses of the higher organic C topsoil through erosion (Freibauer et al., 2004; Ewert et al., 2005). Increased agricultural land-use and changes in management practices may be partly responsible for the losses of soil C recorded recently across the UK by Bellamy et al. (2005). From 1978–2003, manure management practices, such as the spreading of animal manure increasing soil organic C, have declined. This is due to a steady decline in livestock numbers over the same period and intensification of their production leading to increased liquid slurry from animal excreta, at the expense of dry-lot farm yard manure (Burton and Turner, 2003). Moreover, there has been more efficient removal of agricultural products from fields. Agricultural crop yields have increased over the period by about 1.45% per year (Ewert et al., 2005), mainly attributable to improvements in the harvest index (the amount of grain produced relative to straw and stubble). On croplands, this means that less residue is left on the field. Further, improvements in farm machinery may also have led to increased residue removal. Changes in practice over the period, such as increasing production of silage in place of hay, which removes more residue and decreases soil C stocks (Poulton, 1996a), may also have led to more crop residue being removed from managed cropland and

grassland. These changes in technology and practice are predicted to have a large impact on soil organic C stocks in the future (Ewert et al., 2005; Smith et al., 2005a). Furthermore, organic C stocks may continue to decline for many years, as the legacy of land-use changes occurring before a survey period has been shown to have significant impacts on C losses at the national (UK) scale (Milne et al., 2001).

High uncertainties in determination of C gains/losses following changes in grassland management still persist due to a multitude of possible scenarios (Soussana et al., 2004). Under a cutting regime, much of the C is removed via harvesting as hay or silage, which tends to be counter-balanced by applied OM in manure/slurry). Although losses of CO_2 can also increase as manure applications have been shown to increase respiration by 2–3 times (Rochette and Gregorich, 1999). The intensity of grazing is a major factor in determining C losses in the vegetation from pastures; for intensive farming this can be *ca*. 60% of NPP and is either returned to the soil as excreta (25–40%) or back to the atmosphere as CO_2 or to a lesser extent gut fermented CH_4, with proportions dependent on the digestibility of crop (Soussana et al., 2004). Clearing and cultivation will also result in large losses of soil organic C (Johnston, 1973; Smith et al., 1996). Other management processes can indirectly affect microbial activity through modification of soil conditions, such as ploughing, fertilization and liming, which have been shown to increase DOC and DIC concentrations in soil solution. The pH and temperature change, vegetation changes and pasture improvement of these processes contribute to more favourable microbial conditions, as organic solubility increases and hence rates of C decomposition (Rangel-Castro et al., 2004; Worrall et al., 2004a). Nutrient additions can be applied directly or may be deposited as diffuse pollution causing a change in plant and animal community composition (Jensen et al., 2003). However, certain agricultural and grassland management practices do have the potential to increase C sequestration in soils and will be discussed later (Section 5.2).

3.4.2. Forestry

The largest single land-use change in upland UK in the last 100 years was the setting up of conifer plantations, at rates of up to 40,000 ha yr^{-1} in the early 1970s (Farmer and Nisbet, 2004; Stott and Mount, 2004). Increased forestry operations led to serious concerns with regard to water quality as drainage of moorlands, forest harvesting and road building caused initial increases in soil C losses by erosion and sedimentation to upland streams, as well as degrading stream habitats (Farmer and Nisbet, 2004). Anthropogenic disturbances, such as deforestation and natural factors (fire and wind damage) initially cause a loss in terrestrial C but consequently stimulate new forest growth. With time there tends to be a decreased nutrient availability and increased stomatal limitation, which causes decreases in annual NPP as stands mature (Cao and Woodward, 1998). However, long-term trends have suggested that the initial increase in CO_2 production due to ploughing and drainage processes reduces, and eventually is balanced out by increased C sequestration following renewed afforestation (Hargreaves et al., 2003).

3.4.3. Peatlands/wetlands

Many of the alternative explanations of changes in soil C on managed crop and grasslands outlined above do not apply to organic soils, such as those found in upland moorland soils or wetlands. It is possible, however, that changes in management of these land areas could also have influenced the recorded soil C losses by Bellamy et al. (2005). Land-use in UK uplands is mainly restricted to rough and improved pasture for sheep, plantation forestry and moorland for game pursuits as well as providing a major water source. Possible mechanisms for increased soil C losses are instigated by land management practices such as drainage for forestry, higher abstraction, vegetation removal (burning), harvesting of peat as well as an increased amenity value. Practices designed to improve agricultural potential of peatlands can lead to a shift in vegetation (Worrall et al., 2004a). Such measures change the physical, biological and chemical properties of peatlands, leading to major C cycling perturbations (Warburton, 2003; Glatzel et al., 2004).

Localised management practices associated with peatlands show that exports of DOC, CO_2 and CH_4 as well as other nutrients are usually higher from both peats and associated surface waters after disturbance from peat extraction or moorland burning (Hope et al., 1994; Aitkenhead et al., 1999; Billett et al., 2004). Different types of burning can release varying amounts of C; peat (belowground) burning, compared to surface fires where the vegetation alone is burnt, causes higher losses of C, N and organic P, mainly due to volatilisation (Smith et al., 2001b). The drainage and disturbance of peatland/wetland environments has contributed to soil C losses as water table changes cause shifts in C processes and fluxes (Moore et al., 1998; Billett et al., 2004; Cui et al., 2005). Subsurface soil pipe conduits, with their concomitant losses of POC, have also been shown to increase in number following drainage of peatlands (Holden, 2006). The physical disruption caused by drainage also has implications for nutrient release as bare peat undergoes

vertical cracking increasing the total drying area and an absence of roots prevents transport of water to the surface. It also has a lower albedo than moorland vegetation absorbing more solar radiation, resulting in increased surface temperatures (Mitchell and McDonald, 1992). Although peatland restoration can increase C sequestration, it does not appear to compensate for the net C accumulation of the original systems (Waddington and Price, 2000). Most land-use changes to peatlands potentially cause an increase in C loss.

4. Fate of carbon

Addressing the fate of C once it has been lost from soil is a wide subject area. However, this section focuses on the immediate sink of C once lost from the soil and not C processes that occur within subsequent C pools. This initial fate of C is detailed in order to understand the significance land-use management change has on the wider environment. The 3 sinks that C enters, once lost from the soil, are the atmosphere, groundwater and rivers. In terms of the C transfer load, C losses (as CO_2 and CH_4) from the soil to the atmosphere are the most important. However, unlike uni-directional losses from soil to ground and surface waters (atmospheric exchange in CO_2 non-saturated waters and floodplain deposition aside), the losses of gaseous C are part of the GPP equilibrium discussed in Section 3. In addition, the C that has been lost from soil by erosional processes but eventually relocated elsewhere in the terrestrial environment is also discussed.

Recently, to balance C budgets, it was proposed that the losses of C via surface waters equated to the amount of missing C adsorbed from the atmosphere but had not been sequestered in the terrestrial biosphere (Siemens, 2003). Thus, understanding the fate of C is an important component in determing soil C losses, e.g. a total C balance (precipitation, NEE, surface water C losses and degassing of CO_2/CH_4) of a lowland peaty catchment in Central Scotland suggested that the soil may be a net source of C (Billett et al., 2004).

4.1. Soils to groundwater

Decomposition, mineralisation, adsorption and precipitation reactions determine the concentrations of DOC, DIC and gaseous C in pore waters that percolate through soil profiles, some of which will enter the groundwater, depending on geological and hydrological parameters. The main source of DOC in groundwater originates from relatively recent organic C in the soil horizons and tends to consist of labile low molecular weight compounds and humic acids but is chemically dependent on the origin of the terrestrial C (Aravena et al., 2004). Once in the groundwater, microbial activity and the redox potential have a significant role in determining C concentrations and speciation, which can result in further decomposition of DOC to CO_2 and generation of CH_4 (Aravena et al., 2004). Concentrations of DOC can vary spatially and temporally in groundwater but tend to decrease with depth by desorption reactions and heterotrophic respiration along its flow paths (Goñi and Gardner, 2003; Aravena et al., 2004). Supersaturated CO_2 in the groundwater has been shown to occur in the Thames catchment, which contains a broad range of carbonate aquifers common to the UK (Worrall and Lancaster, 2005). Moreover weathering of carbonate-rich parent materials, will also add to the C load (DIC as HCO_3^- and CO_3^{2-}) within groundwater relative to geological characteristics of the aquifer.

4.2. Soils–river–ocean pathway

Rivers act as conduits for the transport of soil C to the oceans and it has been suggested that riverine losses of organic C may regulate future changes in soil C storage (Hope et al., 1997a; Cole and Caraco, 2001; Billett et al., 2006). Moreover, changes in soil C losses to rivers can impact on water quality and hence ecology and biogeochemistry of the freshwater environment (Likens et al., 1977; Romaní et al., 2004) as well as consequences for potable waters (Worrall et al., 2004b).

After C enters headwater streams, in-stream processes continually cycle and recycle C as well as acquiring additional C inputs from bank seepage and streambeds as streams/rivers progress to estuaries, coastal environments and the deep ocean. As stream order increases, terrestrial OM inputs become less significant as upstream inputs become more important (Vannote et al., 1980; Minshall et al., 1985; Dawson et al., 2001). Anthropogenic sources are also important in areas with higher population densities and pollution as well as where land disturbances have occurred (Meyer and Tate, 1983; Armentanao and Menges, 1986; Waddington and McNeill, 2002).

Soils and their associated land-use are the primary influence on the spatial variation in C inputs and other elemental chemistry to surface waters. The main sources of organic C are from vegetation (leaching, rhizodeposition, fragmented plant material and directly from litterfall) and soil (leaching, OM decomposition and erosion) (Thurman, 1985; Meyer et al., 1998; Hope et al., 1994). In headwaters, DOC is assumed to be of mainly allochthonous origin i.e. derived externally from terrestrial organic matter (Vannote et al., 1980; Fiebig et al., 1990; Brooks et al., 1999). This has been supported by work showing a strong

Table 3
Combined carbon flux studies in UK streams and rivers

Details of study catchment (size — km^2)	C-flux (kg C ha^{-1} yr^{-1})				*Reference
	DOC	POC	CO_2–C	DIC	
Scotland					
Brocky Burn — upper (0.68)	174–214	9.0–21	9.8–11	0.6–1.1	Dawson et al. (2004)
Brocky Burn — middle (0.83)	162–206	8.2–19	6.7–8.6	1.0–17	Dawson et al. (2004)
Brocky Burn — lower (1.3)	142–193	5.9–28	2.0–3.2	0.8–1.4	Dawson et al. (2004)
Water of Charr (14.2)	169–262	20–175	2.6–4.3	5.7–8.1	Dawson et al. (2004)
Small Burn (0.41)	95–150	4.6–12	2.5–3.4	0.3–0.4	Dawson et al. (2004)
Burn of Waterhead (3.4)	83–166	4.4–16	2.4–3.5	1.6–1.8	Dawson et al. (2004)
Water of Dye — upper (24.6)	85–145	5.4–19	2.8–4.2	5.2–7.6	Dawson et al. (2004)
Water of Dye — lower (46.3)	103–178	5.9–26	2.7–4.1	5.3–7.0	Dawson et al. (2004)
Auchencorth (3.35)	257–309		9.0–10	10–14	Billett et al. (2004)
River Don, Parkhill (1273)	18±10	5.3±4.0			Hope et al. (1997a)
River Dee, Park Bridge (1844)	22±12	1.9±1.2			Hope et al. (1997a)
Stag Burn (2.4)	34.4		4.4		Dawson et al. (1995)
Glenbuchat (2.0)	24.9			118	Edwards et al. (1985)
Northern England					
N.Pennine, R.Coquet (589)			1.0–9.3		Worrall and Burt (2005)
NE England, R. Tees (818)			1.2–6.7		Worrall and Burt (2005)
N. Pennine, Moor House (11.4)	94–150	27–317	20–38	41–59	Worrall et al. (2003a)
N. Yorkshire Moors (9.0–300)	7.6–24	2.3–5.0			Arnett (1978)
Mid-Wales					
Upper Hafren (0.93)±95% C.I.	84±38	27±19	8.8±3.8	1.3±1.2	Dawson et al. (2002)
Afon Hafren (3.35)	35.7		13.2	0.5	Neal and Hill (1994)
Afon Hore (3.35)	25.8		9.9	2.8	Neal and Hill (1994)
South2-Hore (0.14)	21.2		10.4	0.2	Neal and Hill (1994)
Plynlimon, grassland				8.8–19	Reynolds et al. (1987)
Plynlimon, forest				5.0–16	Reynolds et al. (1987)
Afon Cyff (0.04)	54	8.8			Reynolds (1986)

*Pre-1994 data from Hope et al. (1994).

correlation between soil C pools and streamwater DOC concentrations, particularly in small-scale (<5 km^2) catchments (Aitkenhead et al., 1999; Billett et al., 2006). Isotopic studies have suggested that the majority of headwater terrestrial-derived DOC is from relatively young (decadal — ^{14}C post-nuclear testing) sources (Schiff et al., 1997; Palmer et al., 2001). However, in larger rivers, a combination of young and old DOC and predominantly old POC (>1000 yr) from terrestrial sources is observed. The younger, more labile DOC is selectively respired leaving an older OM component to enter estuarine and oceanic C pools (Raymond and Bauer, 2001). Concentrations and fluxes of DOC and POC are also influenced by other variables such as discharge, precipitation, seasonality, slope and catchment size (Eckhardt and Moore, 1990; Grieve, 1994; Dawson et al., 2002) and vary within and between different streams (Dawson et al., 2002, 2004).

Particulate inorganic carbon (PIC) is mainly derived from weathering and dissolution of carbonate containing minerals (Stumm and Morgan, 1981; Meybeck, 1982). However, DIC in streams originates as three main forms (free CO_2 and dissolved ionic species HCO_3^- and CO_3^{2-}); the proportion being dependent on the carbonate equilibria (Stumm and Morgan, 1981; Dawson et al., 1995). In many UK upland areas, it is insignificant because of the carbonate-poor nature of the underlying parent material. Isotopic dating of DIC and free CO_2 has proved more difficult to assess. Convulated signals in rivers suggest a variety of sources; carbonate weathering in groundwaters, mixing with the atmosphere and decomposition of terrestrially derived DOC (Brunet et al., 2005). Inputs in inorganic C concentrations are dependent on a number of factors such as parent material, soils, topography and hydrology (Jones and Mulholland, 1998b; Hope et al., 2004).

Free CO_2 in rivers is mainly derived from plant root respiration and soil OM decomposition (Edwards and Harris, 1977) but also in-river respiration by aquatic plants or minerilisation of DOC and POC and by

Table 4
Range of CO_2 and CH_4 degassing rates from streams, rivers, lakes and wetland ponds

Details of study	†Evasion fluxes of CO_2 (mg C m^{-2} d^{-1})	*Reference
Streams/rivers		
NE England — R. Coquet	0.0–1973	Worrall and Burt (2005)
NE England — R. Tees	0.0–4082	Worrall and Burt (2005)
NE Scotland — headwater peatland stream	259–45878	Hope et al. (2001)
Alaska, USA — 1st to 6th order arctic rivers	137–1370	Hope et al. (2001)
Tennessee, USA — headwater forest stream	1884–4476	Jones and Mulholland (1998c)
Hudson Bay, Canada — river to estuary	192–444	Raymond et al. (1997)
Alaska, USA — 4 arctic rivers	17–154	Kling et al. (1992)
Lakes/wetland ponds		
Northern Scotland — Loch Ness	148	Jones et al. (2001)
New Hampshire, USA — lake	71–137	Cole and Caraco (1998)
NW Ontario, Canada — beaver pond	1525	Roulet et al. (1997)
Lake District, UK — Esthwaite Water	9.0–92	Maberly (1996)
Quebec, Canada — lake reservoir	500–1100	Duchemin et al. (1995)
Hudson Bay, Canada — lowland wetland ponds	1009–3000	Hamilton et al. (1994)
NW Ontario, Canada — oligotrophic lakes	−51–102	Rudd et al. (1993)
NW Ontario, Canada — lake reservoir	391–411	Rudd et al. (1993)
Alaska, USA — 25 arctic lakes	−66–718	Kling et al. (1992)
Alaska, USA — arctic lake, annual variation	127±7.2–800±240	Kling et al. (1992)
Details of study	Evasion fluxes of CH_4 (mg C m^{-2} d^{-1})	Reference
Streams/Rivers		
NE Scotland – headwater peatland – stream	0.0–345	Hope et al. (2001)
Tennessee, USA — headwater forest stream	0.3–10	Jones and Mulholland (1998b)
Pacific NW — rivers	0.0–238	Lilley et al. (1996)
Hudson Bay, Canada — river to estuary	6.1	De Angelis and Scranton (1993)
Alaska, USA — 1 arctic river	4.3	Kling et al. (1992)
Alaska, USA — surface waters	3.5–98	Whalen and Reeburgh (1990)
Lakes/wetland ponds		
Caithness, N Scotland — wetland pond	4.2	MacDonald et al. (1998)
Fort Simpson, NWT, Canada —wetland pond	188	Moore et al. (1998)
NW Ontario, Canada — beaver pond	70	Roulet et al. (1997)
Quebec, Canada — lake reservoir	5.0–10	Duchemin et al. (1995)
Hudson Bay, Canada — lowland wetland ponds	83–135	Hamilton et al. (1994)
NW Ontario, Canada — oligotrophic lakes	0.3–2.5	Rudd et al. (1993)
NW Ontario, Canada — lake reservoirs	3.0–5.7	Rudd et al. (1993)
Alaska, USA — 10 arctic lakes	1.0–12	Kling et al. (1992)

†Positive values indicate CO_2/CH_4 evasion, *Pre-2001 data from Hope et al. (2001).

exchange at the atmosphere/water interface (Wanninkof et al., 1990; Hope et al., 2004). Streamwater CO_2 concentrations tend to be higher in peatlands (Dawson et al., 2001, 2004), where concentrations of CO_2 in the soil atmosphere are between 10–100 times higher than that of the above ground atmosphere and release of soil-derived CO_2 to the atmosphere is restricted (Skiba and Cresser, 1991; Hope et al., 2004). Following conversion of OM in anaerobic soils, a proportion of the CH_4 produced enters streams. At the soil-stream interface, methanogenic bacteria are only found in areas of anoxia, such as riparian zones and stream sediments (Dahm et al., 1991; Jones and Mulholland, 1998c; Hope et al., 2004).

Extensive measurements of soil organic C losses from a range of terrestrial systems in the UK have been detailed (Hope et al., 1994, 1997b; Worrall et al., 2004a; Billett et al., 2004). Dissolved inorganic C fluxes have also been studied (Neal, 1988; Neal et al., 1997a,b, 2000; Neal and Davies, 2003) and can constitute an important component of C losses to riverine systems (Hope et al., 1994; Neal and Davies, 2003). Studies in the UK that determined the combinations of the components of the total C flux (Table 3) are much rarer and were mainly determined from

upland catchments (Neal and Hill, 1994; Worrall et al., 2003b; Dawson et al., 2004). Organic C fluxes tend to range between 10 and 100 kg ha^{-1} yr^{-1}, although this increases substantially in headwaters of highly organic catchments. However, in more alkali lowland rivers, contributions from mineral soils and groundwater increase DIC fluxes in the form of HCO_3^- and CO_3^{2-}.

Regional losses of organic carbon in British rivers have been reported by Hope et al. (1997b) and total UK C losses were $0.69 \pm 0.07 \times 10^{12}$ g yr^{-1} for DOC and 0.20×10^{12} g yr^{-1} for POC although these are certainly underestimates as they were based on data precluding storm events. Taking account of this discrepancy, Cannell et al. (1999), estimated total organic C flux in UK Rivers as 1.4×10^{12} g of C $yr^{-1} \pm 30\%$.

4.3. Soils–river–atmosphere pathway

Losses of CO_2 (and CH_4) by degassing in headwaters, rivers and wetlands (ponds and lakes) tends to be unaccounted for in most riverine flux measurements and ecosystem C budgets but are potentially important conduits of C transfer back to the atmosphere from the soil C pool (Kling et al., 1991; Cole et al., 1994; Cole and Caraco, 2001; Richey et al., 2002). Measurements in upland Scottish streams have shown that combined gaseous C evasion losses from the stream surface to the atmosphere can be of the same order of magnitude as the total amount of C transported downstream within the water column. However, these fluxes have only recently been determined in the UK in the last 10 years at few sites, mainly in peatland systems (Hope et al., 2001; Billett et al., 2004). Isotopic studies of degassing CO_2 in carbonate free waters suggest that a "small, rapidly cycling pool (2–5 yr) of organic C is responsible for the large C fluxes from land to water to atmosphere" although, where groundwater influences were higher, aged soil CO_2 influxes increased the mean degassed CO_2 age to several decades (Mayorga et al., 2005).

Considerable variations exist in gaseous C efflux data from rivers and lakes in studies from the UK and N. America (Table 4), possibly due to variations in mineralisation rates and the saturation levels of the soils. In UK peatlands, the behaviour of streamwater CO_2 and CH_4 is dominated by areas with significant degassing 'hotspots' in the regions containing C rich soil reservoirs, soil C losses to the atmosphere are reduced and the soil–river linkage is strong. Thus, gaseous C losses from surface waters can vary within the same network where catchment characteristics such as soil type, topography and hydrological flow paths spatially change (Dawson et al., 2001; Hope et al., 2001).

4.4. Movement of soil C within the landscape

Losses of soil OM by erosional processes have been discussed earlier (Section 3.3.2). However, as erosion is spatially dynamic, soils also receive OM and nutrients from upslope (by water erosion or disturbance, e.g. tillage) and aeolian or river flood deposition (Quinton et al., 2006). A combination of erosion and deposition processes suggests that there is an overall net gain of C to some soils (Van Oost et al., 2005). Exposure of less organic soils in eroded areas can be restocked with C as NPP and formation of new soil organic C, while lower down hill slopes, soil organic C increases through deposition of the previously eroded soils. At the catchment scale, it has been estimated that between 30–100 kg C ha^{-1} yr^{-1} could be sequestrated by erosion, translocation and burial processes on sloping arable land (Van Oost et al., 2005). It has been suggested that at these sites, soil aggregation increases as eroded soils containing clay minerals, OM and nutrients are deposited, protecting the OM from decomposition. The net gain of C will mainly be determined by the fate of relocated soil (Stallard, 1998; Liu et al., 2003; Quinton et al., 2006). The subsequent fate of re-deposited soil C could be decomposition or further remobilisation at the next erosion episode.

5. Management of soil carbon loss

Land-use and its management can be used to influence the above fundamental terrestrial processes to increase C stocks (Lal, 2004b). To reverse and limit further C cycle perturbations that have occurred in the past, there is a need that future land-use policies implement management practices that can increase NPP and C sequestration in the terrestrial environment (in soils and vegetation); reduce the rate of OM decomposition and reduce losses of C by soil erosion (Janzen, 2004; Smith, 2004a,b; Smith et al., in press).

5.1. Land-use and C sequestration

Increased C sequestration in soils, as a way to reduce atmospheric CO_2 concentrations, was first proposed in 1977 (Dixon et al., 1994). One appropriate option is to restore a proportion of the historically lost C from soils that have previously been depleted in C, such as agricultural and degraded soils (Smith, 2004a), e.g. revegetation of abandoned arable land may increase soil C by 0.3–0.6×10^3 kg C ha^{-1} yr^{-1} (Freibauer et al., 2004). However, land-use and management of croplands and pastures varies substantially and hence the potential estimates for practices that can increase C sequestration in soils show the greatest

Table 5
Carbon sequestration in soils – measured (M) and potential (P) – under different land-uses in temperate and boreal regions

Details of study	C-sequestered[a] ($\times 10^3$ kg C ha^{-1} yr^{-1})	Reference
Croplands		
Arable management practices (P)		(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Organic farming	>0–0.5	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Organic inputs (crop residues, cover crops, compost, sludge)	0.3–0.8	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Cereal straw and crop residue incorporation	0.69 (0.1–0.7)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Compost	0.4 (0.2–1.5)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Solid animal manure incorporation	0.37 (0.2–1.5)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Slurry application	0.4 (0.2–1.5)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Sewage sludge application	0.26 (0.1–0.3)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
N-fertilization	0.2 (0.1–0.3)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Permanent revegetation of arable set-aside	0.5–1.9	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Permanent shallow water table in farmed peatland	1.4–4.1	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Permanent crops and perennial grasses	0.62 (0–0.62)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Ley-arable rotation (extensification)	0.54 (0–0.54)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Deep rooting crops	0.62 (0–0.62)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Bioenergy crop production	0.62 (0–0.62)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Conservation tillage	>0–0.8	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Reduced tillage	0.2 (0–0.2)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
No till farming	0.39 (0–0.4)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Set aside	0.2 (0–0.2)	(Smith et al., 2000, 2005b, in press; Falloon et al., 2004; Freibauer et al., 2004)
Cropland, temperate climate range, no tillage (P)		Jain et al. (2005)
With changes in climate and CO_2	0.45–0.69	Jain et al. (2005)
Without changes in climate and CO_2	0.40–0.68	Jain et al. (2005)
Cropland, corn and winter rye (M)		West and Marland (2003)
Till and no N-fertilizer	−0.30	West and Marland (2003)
Till and N-fertilizer (low, medium, high)[b]	0.07, 0.08, 0.30	West and Marland (2003)
No till and no N-fertilizer	0.03	West and Marland (2003)
No till and N-fertilizer(low, medium, high)[b]	0.17, 0.20, 0.57	West and Marland (2003)
Grasslands		
Grassland management practices (P)		(Conant et al., 2001; Soussana et al., 2004)
Fertilizer input	0.3	(Conant et al., 2001; Soussana et al., 2004)
Conversion to grass-legume mixtures	0.3–0.75	(Conant et al., 2001; Soussana et al., 2004)
Intensification of permanent grassland	0.2	(Conant et al., 2001; Soussana et al., 2004)
Intensification of nutrient-poor grassland	−0.9 to 1.1	(Conant et al., 2001; Soussana et al., 2004)
Permanent grassland to medium duration leys	−0.2	(Conant et al., 2001; Soussana et al., 2004)
Short duration leys to grassland	0.3–0.4	(Conant et al., 2001; Soussana et al., 2004)
Increased duration of leys	0.2–0.5	(Conant et al., 2001; Soussana et al., 2004)
Improved grazing management	0.35	(Conant et al., 2001; Soussana et al., 2004)
Improved grass species	3.04	(Conant et al., 2001; Soussana et al., 2004)
Introduction of earthworms	2.35	(Conant et al., 2001; Soussana et al., 2004)
Perennial — 60 yr (M)	0.45	Potter et al. (1999)
Species rich — limestone, cutting (M)	1.2±0.5	Fitter et al. (1997)
Species rich — peaty gley, no cutting (M)	6.4±0.6	Fitter et al. (1997)
Temperate upland — lowland grassland (M)		Thornley and Cannell, 1997
-N-fertilization (20–50 kg N ha^{-1} yr^{-1})	0.08	Thornley and Cannell, 1997
-Grazing (5–10 sheep)	0.05	Thornley and Cannell, 1997
Forestry		
88% Mixed Oak, RI, USA (115 yr) (M)		Hooker and Compton (2003)
Woody debris	0.05±0.02	Hooker and Compton (2003)
Forest floor	0.37±0.03	Hooker and Compton (2003)
Mineral soil	0.15±0.06	Hooker and Compton (2003)

(continued on next page)

Table 5 (*continued*)

Details of study	C-sequestered[a] ($\times 10^3$ kg C ha^{-1} yr^{-1})	Reference
Forestry		
Plant biomass	1.53±0.16	Hooker and Compton (2003)
N/NW Europe — forest soil (C in vegetation) (M)		Liski et al. (2002)
Denmark	0.34 (1.08)	Liski et al. (2002)
Scandinavia	0.08–0.09 (0.22–0.23)	Liski et al. (2002)
Belgium	0.24 (0.74)	Liski et al. (2002)
Ireland	0.30 (1.13)	Liski et al. (2002)
Netherlands	1.07 (1.52)	Liski et al. (2002)
UK	0.32 (0.71)	Liski et al. (2002)
Central Europe — forest soil (C in vegetation) (M)		
Austria	0.61 (0.61)	Liski et al. (2002)
France	0.33 (0.55)	Liski et al. (2002)
Germany	0.84 (0.88)	Liski et al. (2002)
Switzerland	0.43 (0.27)	Liski et al. (2002)
Forestry management practices in USA (P)	(units=10^{15} g C yr^{-1})	Metting et al. (2001)
Converting marginal crop/pasture to forest	0.033–0.119	Metting et al. (2001)
Increasing timber growth on timber land	0.138–0.190	Metting et al. (2001)
Growing short-rotation woody crops for energy	0.091–0.180	Metting et al. (2001)
Increasing tree no./canopy cover in urban areas	0.011–0.034	Metting et al. (2001)
UK peatlands — natural accumulation	0.2–0.5	(Clymo et al., 1998; Cannell, 1999)
Eastern Canada — 23 ombrotrophic peatlands	0.73±0.17	Turunen et al. (2004)

[a]–ve value indicates C-emission to atmosphere; grassland data mainly from Jones and Donnelly (2004).
[b]N-fertilizer application=low (84 kg N ha^{-1} yr^{-1}), medium (168 kg N ha^{-1} yr^{-1}), high (336 kg N ha^{-1} yr^{-1}).

uncertainty (usually >50% for most estimates, Freibauer et al., 2004). In order to maximize C sequestration, knowledge of factors such as erosion and the translocation of soil across the landscape must also be considered (Van Oost et al., 2005) particularly on agricultural land where tillage and erosion are strongly related. At present, most temperate grasslands are believed to be C sinks with estimates between 0.03–1.1×10^3 kg C ha^{-1} yr^{-1} as C sequestration is strongly influenced by the productivity of the ecosystems and its management (Soussana et al., 2004) although grassland derived soils do tend to have higher base saturation, enhancing aggregation and increased capability to sequester C (Collins et al., 2000). Moreover, these increases maybe limited over time as a maximum C storage is attained. The uncertainties are large and this may occur as quickly as 10 or up to 100 years. The potentially available terrestrial C sink for forests has been estimated to last for a further 100 years (Dixon et al., 1994; Lal, 2004a; Smith, 2004b). Accumulation of terrestrial C in mature forests slows over decades to centuries but C sequestration on relatively young, disturbed or degraded soils tends to be higher (Dixon et al., 1994). In the UK, forests in NI are sequestering more C ha^{-1} than in GB, because the average forest age is younger (Cannell et al., 1996). While some management practices will produce increases in soil C, this may be counter-balanced by climate changes, although this potential feedback is still unresolved (Jones and Donnelly, 2004).

5.2. Management practices and C sequestration

Table 5 details estimates of the ability of soils to sequester C under different land-use and management regimes. There are 2 main ways that the terrestrial ecosystem can be manipulated to increase its C pool. Land-use change e.g. returning marginal agricultural land back to forest or restoring grasslands on degraded areas will cause a step change in C storage for a given land area whereas 'best practice' land management proposals e.g. sustainable forestry, reduction of intensity of tillage on croplands, prevention of drainage on peatlands optimises the land-use and C storage potential already present (Dixon et al., 1994; Fortin et al., 1996; Janzen, 2004). Although land management to increase C sequestration may only be a limited, relatively short-term response to the reduction of CO_2 in the atmosphere (Cannell, 1999; Scholes and Noble, 2001), a well managed, conserved soil, irrespective of land-use has

other benefits, such as increased productivity, resilience from erosion and biodiversity. Thus the relative ability of the soil to accumulate C can be used as an indicator of ecological and ecosystem performance (Six et al., 2002; Janzen, 2004; Smith et al., 2005b,c).

Additional N inputs may increase the amount of C sequestration, particularly in forests; the applied N being stored in high C:N ratio wood products and soil OM (Schlesinger and Andrews, 2000). Increased N or CO_2 fertilisation have been shown to increase the ability of ecosystems to sequester C in some European forests and forested wetlands (Cui et al., 2005; Vetter et al., 2005) and Canadian wetlands (Turunen et al., 2004). However, the evidence for this is still ambiguous; the definitive increased NPP response may be offset by higher autotrophic and soil respiration removing C from the terrestrial system and additional N can lower the C:N ratio as no extra C is actually sequestered (Schlesinger and Andrews, 2000). Moreover, the CO_2 produced during production, transport and application of fertilizers counteracts much of the potential extra C sequestration that may occur (Schlesinger, 1999; King et al., 2004). One possible area where additional N might be a major benefit is on poorly managed or degraded sites, such as unimproved grasslands (Jones and Donnelly, 2004). However, other potential environmental impacts on increasing the rate of C sequestration may occur, e.g. increased soil N from fertilization in the sequestered OM could lead to increased nitrate leaching or N_2O emissions from soils (Smith et al., 2001a).

Improved agricultural practices that increase C sequestration and reduce soil erosion include minimal tillage across the slope rather than traditional tillage down slope (Quinton et al., 2006). Set aside and field margins management (grass margins, hedgerows, tree strips) can increase soil C stocks as well as the above ground biomass e.g. arable vegetation to set-aside increases vegetative biomass by 2.8×10^3 kg C ha^{-1} (Smith et al., 2000; Falloon et al., 2004). Moreover, agroforestry systems increase above and belowground C (and N) as well as reducing soil degradation (Mutuo et al., 2005) because they can produce more biomass yr^{-1} and have "both forest and grassland nutrient cycling patterns active" compared to its singular land-use (Sharrow and Ismail, 2004). In an 11 yr study from Oregon, USA, an agroforestry system from initial planting, sequestered *ca*. 0.74×10^3 kg ha^{-1} yr^{-1} more C than control forests and 0.52×10^3 kg ha^{-1} yr^{-1} more C than control pastures (Sharrow and Ismail, 2004).

A review of improved grassland management practices and conversions into grasslands showed that the potential for C sequestration varied substantially by grassland type and climate. Soil C stores increased with improved management in 74% of the studies and C sequestration rates ($0.11–3.04\times10^3$ kg C ha^{-1} yr^{-1}, mean $=0.54\times10^{3-}$ kg C ha^{-1} yr^{-1}) were highest during the first 40 yr after treatments began and tended to be greatest in the top 10 cm of soil (Conant et al., 2001). In general, degraded grasslands have the most potential for soil C sequestration as they have lost soil C through previously poor management practices. On intensively managed grasslands, rotational grazing and fertilization sequester the most C; on extensively managed grasslands, grazing intensity and frequency are the optimum management practices. Grazing can increase annual NPP, which increases soil C sequestration as it is redistributed within the terrestrial ecosystem when compared with cutting (Post and Kwon, 2000; Jones and Donnelly, 2004).

Sequestering C in forests is a cost-effective measure to slow increases in atmospheric CO_2 but which could be released back to the atmosphere in future (Cannell, 1999). Thus, good forest management practices include maintaining existing C pools; sustainability can be enhanced by preventing or slowing deforestation and limiting forest degradation, which can be assisted by forest conservation schemes such as the establishment of reserves, national parks and fire protection (Malhi et al., 2002). This will also help the ecological protection of water and soil resources in forests as well as minimising nitrate leaching and enhancing base cation retention by encouraging early revegetation of sites (Reynolds, 2004). To augment existing forest C pools requires expanding afforestation and regrowth (Dixon et al., 1994; Malhi et al., 2002). Not all forests are at a maximum C storage potential and afforestation or revegetation of native species can reduce soil erosion. Upgrading marginal agriculture lands to forestry increases the potential of the land area to sequester more C, both in vegetation as biomass and soils (Dixon et al., 1994; Schlesinger, 1999). Continuous Cover Forestry policies aid C sequestration by reducing the extent of clearfelling and its concomitant erosive potential, diversifying species, stand age and structure in plantation forests and "maintaining uninterrupted woodland coverage" (Reynolds, 2004). Moreover, the increase in planting of broadleaf species, some on brownfield sites, increases the C sequestration potential of 'new' forestry areas. It has been estimated that many marginal, degraded or abandoned areas exist in the UK that could be forested (Farmer and Nisbet, 2004).

Since drainage of peatlands, often necessary for preparing upland sites for forest plantation, may result in large C losses from soils, forests should be preferentially established on shallow organic or mineral soils, resulting in a higher net C sequestration (Cannell et al., 1993; Guo and

Table 6
Land management options that could increase soil C pools

Croplands	Convert marginal cropland to native vegetation, grasslands or forestry; improve crop production and erosion control; improve management of set-aside and field margins; improve farming on eroded soils, erosion control buffer strips, riparian filters; reduced or no tillage; improved residue management; eliminate bare fallow; organic amendments, increased efficiency of animal manure, sewage sludge and composting; inter-sowing and increased duration of grass-leys; improved crop rotations; use perennial crops; use deeper rooting crops; use bioenergy crops; improve water and nutrient (fertilizer) management; increase number of agroforestry systems; do not use highly organic soils for cropping.
Grasslands	Convert cultivated lands to well managed permanent grasslands, species selection; decrease erosion and degradation; eliminate disturbance e.g. fire protection in established pastures; increase forage production by improved fertilization, irrigation, inter-sowing of grasses and legumes; improve grazing and livestock management with controlled light-to-moderate stocking density; moderately intensify nutrient-poor permanent grasslands; introduce earthworms, improve soil structure; maintain a diverse plant community with a dense rooting system.
Forestry	Forest and Water Guidelines by the Forestry Commission, 'best practice' guidelines; increase forest stock; continuous cover forestry to encourage natural regeneration; conserve soil and water resources; improve site preparation and planting techniques to decrease erosion; streamside management with uncultivated buffer zones to stabilize soil and reduce acidification; design of forest roads and network of drains, culverts and sediment catch pits; reduce disturbances from wind and fire; minimise soil and water impacts and reduce clear felling operations to phased felling techniques; minimise nitrate leaching, enhance base cation retention by early revegetation; use species with high NPP or increase number of actively sequestering younger forests; application of nutrients and micronutrients as fertilizers or biosolids; aesthetic planting of previously native trees and shrubs, enhance biodiversity; maintenance of open bog and moorland habitats; extension of guidelines to include conservation, landscape and recreation; plant trees on mineral soils in preference to highly organic soils.
Peatlands and wetlands	Wetland protection, restoration and revegetation on bare peats; prevention of wind and water erosion; reduce peat extraction and disturbance; preserve biodiversity; rehabilitate acidified surface waters; afforestation only in appropriate areas; controlled burning; aesthetic planting of previously native trees and shrubs; where possible block drains and restore water table.

Gifford, 2002; Forestry Commission Scotland, 2006). Given the importance of peat-dominated regions to terrestrial C storage in northern latitudes (Gorham, 1991; Cannell et al., 1999) and likely effects of climate change on precipitation, hydrological regimes and OM transport (Freeman et al., 2001b; Bragg, 2002), good management that conserves or restores these systems is important.

5.3. Amelioration of soil C losses

The terrestrial C cycle is closely linked with the atmospheric concentrations of CO_2 and induced climate change. As part of the global climate change C feedback on terrestrial systems (Cox et al., 2000), a reduction in atmospheric CO_2 is required to ameliorate losses from the terrestrial ecosystem. Therefore, a reduction in the consumption of fossil fuels and deforestation are the first steps in reducing CO_2 inputs into the environment. However, increased C sequestration in the terrestrial biosphere is a small but important component of this process (Scholes and Noble, 2001; Smith, 2004b). The replacement of fossil fuels with biofuels is preferable as the C consumption from these fuels is recycled more quickly (10–100 yr) back into the terrestrial store than peat (10,000 yr) and fossil fuels (geological time). Conversion to land-uses that have the capability to sequester more C under current climate change scenarios, i.e. forests and permanent grasslands from croplands (Guo and Gifford, 2002; Janssens et al., 2003; Smith, 2004a), could be a major driver in amelioration of C losses. However, this is not always practicable on a national scale, thus, preventing further degradation of existing landscapes such as erosion and drainage of peatlands and 'best practice' management protocols remain important in reducing C losses from soils. As of 2003, the "conservative achievable" (0.7–1.3%) estimates of increasing C sequestration to offset UK fossil fuel emissions were less than their "realistic potentials" (2.0–3.4%) and *ca*. 10 times less than theoretical potentials (Cannell, 2003). The potential C sequestration in soil could double under improved management practices (Cole et al., 1997; Lal, 2004b; Post et al., 2004; Smith, 2004a). These aims require utilisation of processes that increase the input rates of OM; place OM deeper in the soil; enhance physical protection through either intra-aggregate or organo-mineral complexes; partition C to longer-lived pools; increase longevity of existing C pools; use species more resistant to decomposition as well as revegetation of bare land and awareness of erosion mitigation measures during disturbance processes, such as clearfelling. Table 6 describes a number of small and large-scale measures that may be considered when deriving "best practice guidelines" under varied land-uses prevalent in the UK (Post and Kwon, 2000; Carling et al., 2001; Conant et al., 2001; Farmer and Nisbet, 2004; Forestry Commission Scotland, 2006;

Freibauer et al., 2004; Jones and Donnelly, 2004; Lal, 2004b,c; Smith, 2004a; Soussana et al., 2004; Stott and Mount, 2004).

These guidelines can form a central role in any portfolio of measures to reduce atmospheric CO_2 concentrations over the next 20–30 years, while new energy technologies are developed and implemented. However, irrespective of 'best practice', in order to realise the future potential of C sequestration may require incentives, financial or otherwise, to enable maximisation of UK C stocks under each land-use criteria (Smith et al., 2005b). Policies that protect and enhance terrestrial C stocks at the same time as improving other aspects of the environment show the greatest promise in meeting our ambitions.

One of the main problems in the future will be to verify that small 'measured increases' in terrestrial (both soil and vegetation) C stocks are due to the implementation of the management practices that enhance C sequestration and not just from natural and analytically-derived variation of C pools and fluxes (Scholes and Noble, 2001). This can only be achieved when terrestrial C budgets are accurately measured and errors reduced substantially in order to provide a "baseline to gauge future climate and land-use induced changes" (Garnett et al., 2001). This will include a more consistent methodology for the complete analysis of soil C to depth. Moreover, there are still gaps in our knowledge with regard to the dynamics of C cycling processes within soils and its subsequent fate for the wider environment.

6. Conclusions

Throughout this review, we have noted uncertainty in our figures of C stores, soil processes or C fluxes from soils. Uncertainty in estimates mainly arises from variability in measurements and analyses undertaken, a lack of data in specific areas, or a lack of knowledge of the processes involved in C cycling. Moreover, coupled with future climatic scenarios, the heterogeneous nature of soils, land-uses, and management practice on each particular land-use, inevitably means that assumptions and generalisations are made, leading to further uncertainty in our figures. Differences between observed and predicted (modelled) changes in soil C processes and fluxes highlight "uncertainty in predicting soil C losses and again emphasises the need for a wider range of repeated soil C inventories to constrain modelled soil C losses" (Janssens et al., 2003).

Future research might address the following outstanding questions: What is the long-term potential of C sequestration? How will C sequestration be affected by future environmental change? How much more C can be sequestered in terrestrial ecosystems? At what rates? What are the most cost-effective ways to achieve and promote C gains? How secure is the newly stored C? How are C gains linked to other nutrient and water cycles? (Scholes and Noble, 2001; Janzen, 2004) What are the interactions between human-induced and naturally occurring effects on soil C sinks (Smith, 2005)? Finally, how accurately can we measure small increases in soil C? If this is not possible then how can we show that gains are due to optimal management practices or simply, the natural buffering/variability of the ecosystem?

Answers to these questions will require the collation of data and necessitate new research to better quantify and integrate fluxes within whole-catchment carbon budgets and understand processes in a range of key ecosystems, particularly those holding large amounts of C (peatlands), or those under human management that can be manipulated to store more C (croplands, grasslands and forests).

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