

ECOSSE

ESTIMATING CARBON IN
ORGANIC SOILS SEQUESTRATION
AND EMISSION



Llywodraeth Cynulliad Cymru
Welsh Assembly Government



SCOTTISH EXECUTIVE

**Scottish Executive
Environment and Rural Affairs Department**

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**ECOSSE – ESTIMATING CARBON IN
ORGANIC SOILS SEQUESTRATION AND
EMISSIONS**



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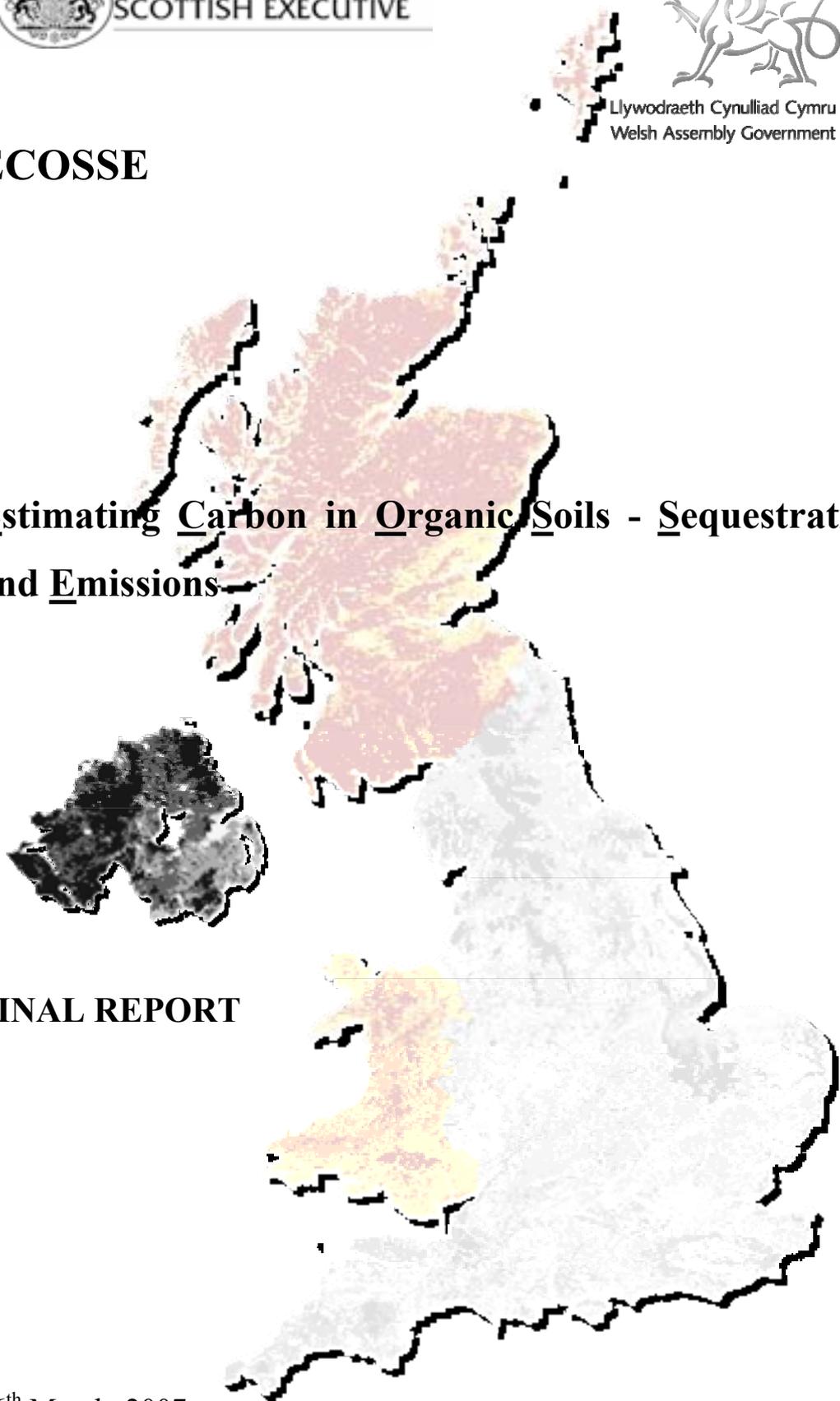
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ECOSSE

**Estimating Carbon in Organic Soils - Sequestration
and Emissions**

FINAL REPORT

16th March, 2007



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EXECUTIVE SUMMARY

Rationale

- Climate change is one of the most serious threats facing our planet, and is of concern at both UK and devolved administration levels. Accurate predictions for the effects of changes in climate and land use on GHG emissions are vital for informing land use policy.
- Land use change and climate change can cause emissions of GHGs; for example, land use change on organic soils is estimated to be responsible for 15% of Scotland's total greenhouse gas emissions.
- Organic soils are abundant in Scotland and Wales. They contain by far the majority of the soil carbon stocks in the UK - combined they contain 3000 Mt carbon. Stock estimates were particularly uncertain below 1 m depth and the project has reduced this uncertainty for the organic soils of Scotland and Wales.
- When carbon is lost from organic soils (as carbon dioxide, the main greenhouse gas responsible for climate change) there are serious consequences for the emissions of greenhouse gases. In addition, methane (an even more potent greenhouse gas), can be emitted from wet organic soils.
- Organic soils behave very differently from mineral soils. This is primarily due to the properties of organic matter and the conditions which favour the accumulation of plant debris. These soils are subject to significant land use changes and the carbon they contain can be affected by these changes.
- Despite the importance of organic soils, prior to this project there were no models able to adequately simulate greenhouse gas emissions from the organic soils of Scotland and Wales.
- In this project, a model - ECOSSE – was developed to predict the impacts of changes in land use and climate change on greenhouse gas emissions from organic soils. ECOSSE stands for Estimating Carbon in Organic Soils - Sequestration and Emissions.

Key Findings

- New estimates have been derived for the amount of carbon stored in organic soils in Scotland and Wales. The data illustrate the huge pool of carbon in the organic soils of Scotland and Wales. Stock estimates have increased by over 30% for Scotland and 20% for Wales with the inclusion of organic material below 1 m depth and the improved estimates of bulk density.
- Some uncertainty remains over soil C stocks and further validation is required to reduce this uncertainty. Remote sensing techniques may potentially be useful to update our knowledge of soil C stocks, particularly in the uplands of Scotland and Wales. It is important to have a reliable estimate for the carbon held in soils in order to be able to monitor and predict the consequences of global change on GHG emissions.
- Measurements of greenhouse gases fluxes from organic soils (carbon dioxide, methane and nitrous oxide) at three sites in Scotland and Wales over the course of the project have provided invaluable data for developing the ECOSSE model, as well as revealing some of the key factors controlling greenhouse gas emissions at each site. New data suggests that emissions from soils will increase when X conditions apply

- Experimental work, albeit on a very limited sample of three sites, has shown for the first time a statistically significant overall loss (13%) in carbon following 25 years of birch growth on heather moorland. Further investigation is merited.
- New land use change matrices for Scotland and Wales for 1950s to 2000s have been derived. This has provided improved land use change information for modelling greenhouse gas (GHG) emissions from land use change.
- A resurvey of the Welsh Bronydd Mawr grassland experiment showed that carbon stocks in the brown earth soil were unaffected by 12 years of pasture de-intensification. The stock of carbon in standing litter and vegetation increased after liming stopped, even under continued grazing.
- New evidence from this project suggests that changes in soil acidity have had a significant influence on the upland carbon cycle. The clearest effect appears to be the doubling of concentrations of dissolved organic carbon in runoff waters since the 1980s. Radiocarbon analysis shows this to be relatively new carbon

Practical Applications

- A review of the effects of forest and woodland planting on the carbon pools and fluxes of upland organo-mineral soils under semi-natural/grazed vegetation has been undertaken. The overall conclusion (as assumed by UK carbon balance models) is that afforestation probably has little net effect on soil organic carbon stores in organo-mineral soils, but this statement is very uncertain. The review findings suggest that these soils are likely to be vulnerable to carbon losses during the tree establishment phase and through erosion losses during ground preparation. A second period of vulnerability may be associated with forest harvesting both through physical disturbance and accelerated leaching losses of dissolved organic carbon. The effect of current forestry best-practice at these vulnerable times has yet to be quantified.
- Suggested guidance for management of new forests and woodlands include minimising ground disturbance, maintaining as much vegetation cover during establishment as possible and encouraging rapid re-vegetation after felling. Evidence from the National Soil Inventory in England and Wales, together with limited evidence from international studies suggests that broadleaved trees maybe more beneficial for soil carbon than conifers. This appears to contradict the results of experimental work undertaken as part of this project. Further research is required before guidance on species choice can be recommended.
- Based on outputs achieved as part of this study an evidence base is provided for the development of guidance on the management of organic soils in Scotland and Wales. Modelled estimates of the consequences of different land management practices will help to choose those management practices that protect organic soils and reduce GHG emissions from organic soils. For example having the ability to predict the effecting of liming and increased stock densities can be useful for decision making

The legacy of ECOSSE

- The funding for this project from the Scottish Executive and the National Assembly for Wales has enabled the development of a model that can simulate and predict GHG emissions from organic soils.
- The project has generated interest from researchers internationally who intend to use the ECOSSE model for application to organic soils outside the UK.

- The UK Meteorological Office (Hadley Centre) and the Natural Environment Research Council (NERC) QUEST programme have recognised the necessity for the ECOSSE model and the importance of GHG emissions from organic soils. Outputs from this project are already being used to understand better the role of organic soils in the global carbon cycle within those ongoing programmes.
- The ECOSSE model will form a central plank in assessing GHG emissions from soils in future UK, Scottish and Welsh inventories. Funding has been secured to develop ECOSSE further for use in future GHG inventories, which are reported to the United Nations under the Framework Convention on Climate Change for the UK.

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Background

Climate change, caused by greenhouse gas (GHG) emissions, is one of the most serious threats facing our planet, and is of concern at both UK and devolved administration levels. Accurate predictions for the effects of changes in climate and land use on GHG emissions are vital for informing land use policy. Models which are currently used to predict differences in soil carbon (C) and nitrogen (N) caused by these changes, have been derived from those based on mineral soils or deep peat. None of these models is entirely satisfactory for describing what happens to organic soils following land-use change. Reports of Scottish GHG emissions have revealed that approximately 15% of Scotland's total emissions come from land use changes on Scotland's high carbon soils; the figure is much lower for Wales. It is therefore important to reduce the major uncertainty in assessing the carbon store and flux from land use change on organic soils, especially those which are too shallow to be deep peats but still contain a large reserve of C.

In order to predict the response of organic soils to external change we need to develop a model that reflects more accurately the conditions of these soils. The development of a model for organic soils will help to provide more accurate values of net change to soil C and N in response to changes in land use and climate and may be used to inform reporting to UK GHG inventories.

Whilst a few models have been developed to describe deep peat formation and turnover, none have so far been developed suitable for examining the impacts of land-use and climate change on the types of organic soils often subject to land-use change in Scotland and Wales. Organic soils subject to land-use change are often (but not exclusively) characterised by a shallower organic horizon than deep peats (e.g. organo-mineral soils such as peaty podzols and peaty gleys). The main aim of the model developed in this project was to simulate the impacts of land-use and climate change in these types of soils. The model is, a) be driven by commonly available meteorological data and soil descriptions, b) able to simulate and predict C and N turnover in organic soils, c) able to predict the impacts of land-use change and climate change on C and N stores in organic soils in Scotland and Wales.

In addition to developing the model, we have undertaken a number of other modelling exercises, literature searches, desk studies, data base exercises, and experimentation to answer a range of other questions associated with the responses of organic soils in Scotland and Wales to climate and land-use change.

Aims of the ECOSSE project

The aims of the study were:

- To develop a new model of C and N dynamics that reflects conditions in organic soils in Scotland and Wales and predicts their likely responses to external factors

- To identify the extent of soils that can be considered organic in Scotland and Wales and provide an estimate of the carbon contained within them
- To predict the contribution of CO₂, nitrous oxide and methane emissions from organic soils in Scotland and Wales, and provide advice on how changes in land use and climate will affect the C and N balance

In order to fulfil these aims, the project was broken down into modules based on these objectives and the report uses that structure. The first aim is covered by module 2, the second aim by module 1, and the third aim by modules 3 to 8. Many of the modules are inter-linked.

Objectives of the ECOSSE project

The main objectives of the project were to:

- Describe the distribution of organic soils in Scotland and Wales and provide an estimate of the C contained in them
- Develop a model to simulate C and N cycling in organic soils and provide predictions as to how they will respond to land-use, management and climate change using elements of existing peat, mineral and forest soil models
- Provide predictive statements on the effects of land-use and climate change on organic soils and the relationships to GHG emissions, including CO₂, nitrous oxide and methane.
- Provide predictions on the effects of land use change and climate change on the release of Dissolved Organic Matter from organic soils
- Provide estimates of C loss from scenarios of accelerated erosion of organic soils
- Suggest best options for mitigating C and N loss from organic soils
- Provide guidelines on the likely effects of changing land-use from grazing or semi-natural vegetation to forestry on C and N in organic soils
- Use the land-use change data derived from the Countryside Surveys of Scotland and Wales to provide predictive estimates for changes to C and N balance in organic soils over time.

Module 1

The Distribution Of Organic Soils In Scotland And Wales And The C Contained In Them

1.1 Types of organic and organo-mineral soils in Scotland and Wales

1.1.1 Scotland

Introduction: In order to allow definitive estimations of the carbon stocks of organic and organo-mineral soils in Scotland to be made, identification of the relevant soils on which to focus attention was required. In addition, modal soil profiles with carbon and bulk density attribute data were needed to estimate carbon stocks. The soil series, i.e. soils with a similar type and arrangement of horizons developed on similar parent material, is the basic soil mapping unit in Scotland and these occur as components of each of the 1:250 000 scale soil map units. Each soil series belongs to a major soil subgroup, which is the reference group adopted in this work.

Methodology: Using expert knowledge and an understanding of the soil classification adopted in Scotland, the major soil subgroups known to occur in Scotland were categorized into:

- 1) Peats
 - 2) Organo-mineral soils
 - 3) Mineral soils
-
- 1) Peats contain more than 60% of organic matter and exceed 50 centimetres in thickness. Peat is an accumulation of partly decomposed plant remains formed under waterlogged conditions where excess moisture has inhibited the normal aerobic process of decomposition. Other conditions conducive to the formation of peat are low average temperature, high acidity and nutrient deficiency, all of which depress microbiological activity. The most important factors that influence peat development are therefore functions of climate, topography and geology. These also control the nature and abundance of vegetation which in turn largely determines the type of peat formed. This grouping includes hill, blanket, and basin peats, all of which are extensive throughout Scotland. In a number of locations, particularly at higher altitudes, the peat is often severely eroded.
 - 2) Organo-mineral soils are classified as soils having organic surface horizons, less than 50 cm thick in Scotland (<40cm in Wales), overlying mineral horizons or rock. This grouping includes humus-iron podzols, peaty podzols, subalpine podzols, alpine podzols, peaty gleys, humic gleys, peaty rankers, podzolic rankers, peaty lithosols and peat alluvium. In most of these soils, the organic surface horizon is a consequence of anaerobic and waterlogged conditions at the surface, due either to slowly permeable subsoil or rock, high ground-water tables in depressions and receiving sites, or to climatic conditions where a combination of low temperatures and high rainfall initiate the formation of the organic surface horizons. The main exceptions to this are humus-iron podzols and subalpine and

alpine podzols where it is considered that the shallow organic surface horizon has developed under aerobic conditions. Many humus-iron podzols are cultivated and no longer have an organic surface horizon, having been mixed and incorporated with the underlying mineral horizons through years of cultivation. 3) Mineral soils have no organic horizons in the soil profile and were excluded from this investigation.

Results: Using this general classification, the relevant major soil subgroups occurring in Scotland have been identified and are shown in Table 1.1.

Table 1.1 Major soil subgroups in Scotland relevant to the study

Classification	Major soil subgroup
Peat (>50cm thick organic surface horizon)	Peat
Organo-mineral (<50cm thick organic surface horizon)	Humus-iron podzol (uncultivated), peaty podzol, subalpine podzol, alpine podzol, peaty gley, humic gley, peaty ranker (including podzolic ranker), peaty lithosol, peaty alluvium

1.1.2 Wales

It is important to define what is meant by peat and organo-mineral soils in the context of this review for Wales. In early soil surveys of Wales, in order to be classified as a peat soil, the layer of accumulation had to contain at least 50% organic matter and extend to a depth of over 12 inches (30 cm) (Roberts 1958) over 15 inches (37.5 cm) (Rudeforth 1970) or 20 inches (50 cm) (Ball 1960). Organic or peat soils, the names appeared interchangeable, were then further classified according to their mode of formation into Basin Peat and Moss Peat.

With the introduction of a unified classification of soils for England and Wales (Avery 1980), horizons described as *peat* contain a minimum of between 20 and 50% organic matter, depending on the clay content. The profile classification has a Major Group of *Peat Soils* in which the thickness of accumulation of organic material exceeds 40 cm depth within the upper 80 cm of the profile. These criteria have been applied in all surveys published since 1978, including the National Soil Map for Wales (Rudeforth *et al.* 1984).

Organo-mineral soils are classified as soils having organic surface horizons less than 40 cm thick (in Wales), overlying mineral horizons or rock. The extent and organic matter content of the organo-mineral soils are well documented in the various soil maps, bulletins and memoirs of the Soil Survey of England and Wales, for example, Rudeforth *et al.* 1984, Rudeforth 1970 and Rudeforth and Thomasson 1970. The desk study and fieldwork in Wales carried out as part of ECOSSE concentrated on the extent and depth of peat soils which were less well known or documented.

1.2 The Spatial Distribution Of Peats And Organo-Mineral Soils In Scotland And Wales

1.2.1 Scotland

Introduction: The 1:250 000 scale soil map of Scotland is the only spatial dataset available in digital format for the whole of Scotland. It uses the mapping concept of landscape units with associated soils. Map units can comprise a range of soil types (major soil subgroups) depending on the complexity of the soil pattern on any particular area of land.

Methodology: Each of the 580 mapping units was assigned to the dominant soil type (major soil subgroup) within it and using GIS, the spatial distribution of the relevant major soil subgroups was plotted and the area of each determined.

Results: The areas of the relevant major soil subgroups are shown in Table 1.2. The distribution of blanket peats and peaty gleys (the most prevalent true peat and organo-mineral soil types) in Scotland are shown in Figure 1.1.

Table 1.2 Area of peats and organo-mineral soils of Scotland

Major Soil Subgroup	Area (km ²)	% of Scotland
Basin peat	837.7	1.1
Blanket peat	7980.0	10.1
Humus-iron podzol	8494.5	10.8
Peaty podzol	12240.4	15.5
Subalpine podzol	3890.6	4.9
Alpine podzol	515.5	0.7
Peaty gley	17156.7	21.8
Humic gley	79.4	0.1
Peaty ranker	696.7	0.9
Lithosol	33.1	0.04
Peat alluvium	No data	No data

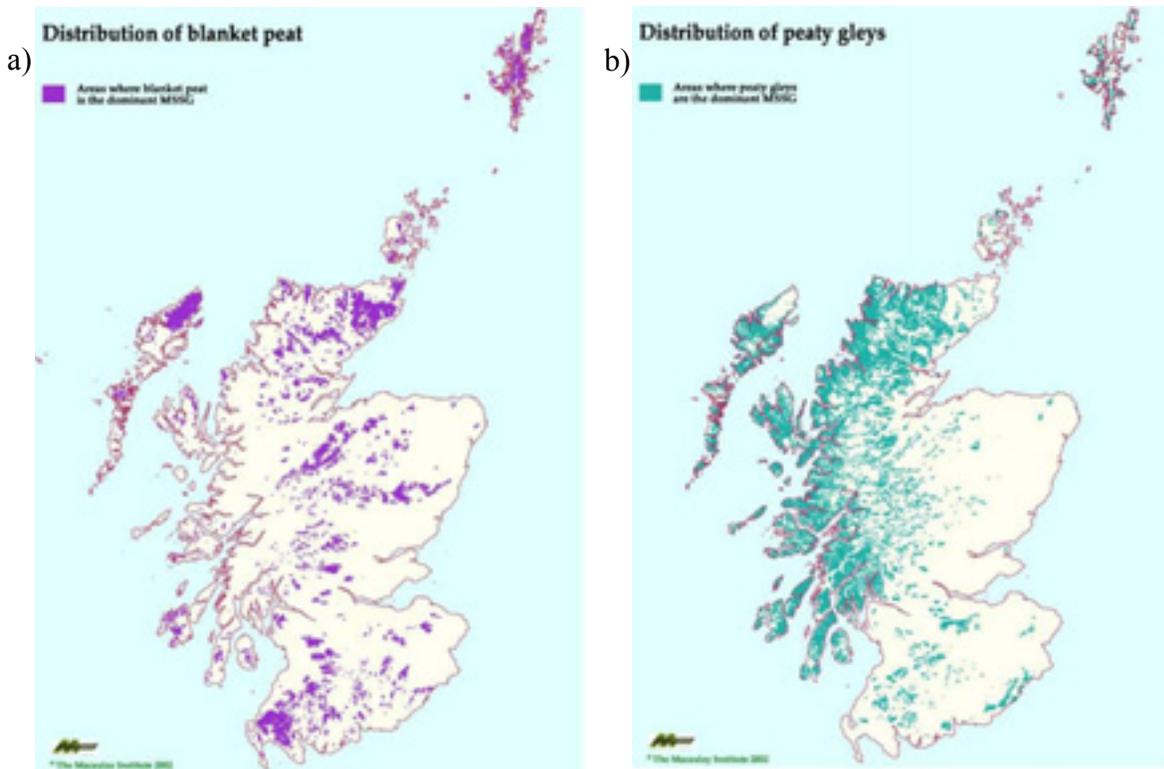


Figure 1.1 Distribution of (a) blanket peats and (b) peaty gleys in Scotland

1.2.2 Wales

Peat soils: Taylor (1975) in the first published figures for Wales estimated that there were 1588 km² of peat soils (9.7% of the Principality based on total area of agricultural land of 16,335 Km²; Welsh Agricultural Statistics, 2005). Comparison with more recent surveys is difficult because no depth or organic matter content were given to distinguish between organic and non-organic soils. Soil associations dominated by peat were mapped over 706 km² (3% of the Principality) during the National Soil Map programme (Rudeforth *et al.*, 1984) (Figure 1.2a). Two soil associations dominate the upland areas, Crowdy 1 (222 km²) and Crowdy 2 (441 km²). Amorphous raw peat soils of the Crowdy series are the most common soil series in both associations. Stagnopodzols of the Hafren series and stagnohumic gley soils of the Wilcocks series are common in the former and less humified peat soils with moss and *Sphagnum* remains (Winter Hill series) are in the latter. Crowdy 1 is mapped discontinuously across the main north to south watershed, generally above 350 m O.D., at the heads of streams and on saddles between the major hills. Crowdy 2 occupies wide upland tracts of blanket bog and scattered peat-filled basins from sea-level at Borth and Tregaron to more than 600 m O.D. in east Wales. In the lowlands some 34 km² of earthy peat soils (Adventurers' and Altcar associations) have been mapped, almost entirely under permanent grassland.

Small areas of less than 1 km² were not identified during the National Soil Map programme, nor were the specific locations of peat soils within the associations. Other associations include small areas of peat soils, for example Wilcocks 2 (721d).

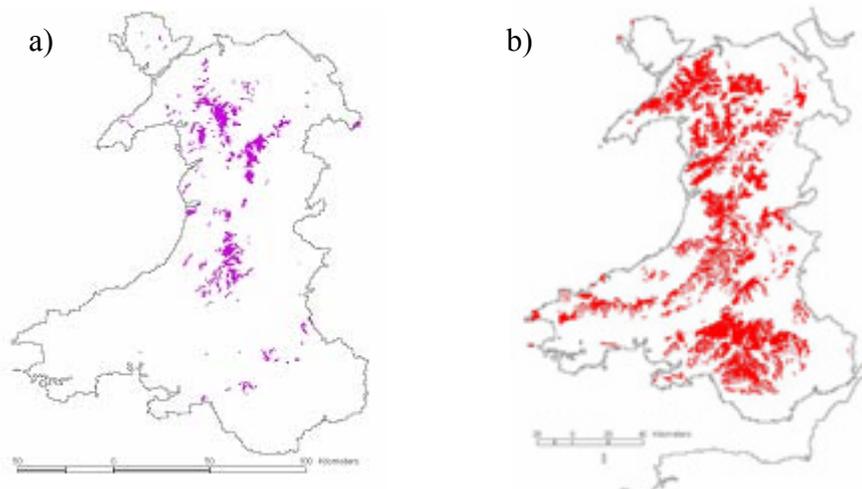
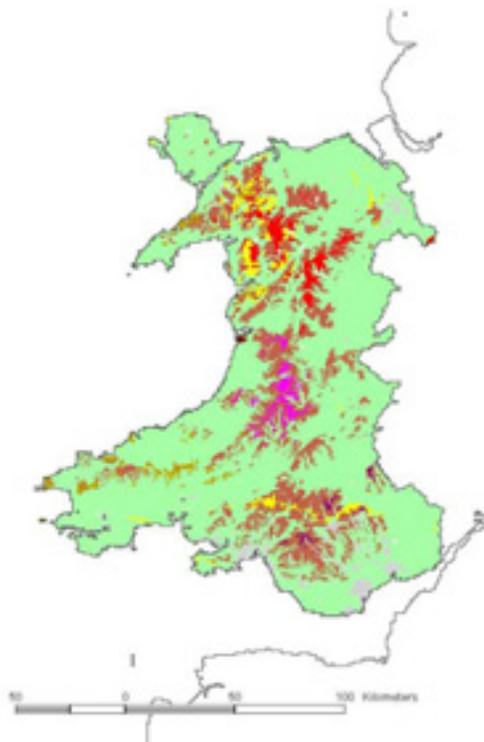


Figure 1.2 Distribution of (a) peat soils, and (b) organo-mineral soils in Wales

Organo-mineral soils in Wales: The extent of organo-mineral soils is given in Table 1.3 and Figure 1.2b. The thickness of their organic horizons is as described in a Defra report (*UK soil database for modelling soil carbon fluxes and land use for the national carbon dioxide inventory – project SP0511*) as summarised in Bradley *et al.* (2005).

Table 1.3 Extent of organo-mineral soils in Wales

	Extent (km²)	% of Wales	Average thickness of organic layer (cm)
Humic gley soils	29	0.1	20
Humic rankers	281	1.3	20
Podzols	166	0.8	15
Stagnohumic gley soils	1552	7.5	25
Stagno-podzols	1564	7.6	20
All organo-mineral soils	3592	17.3	
Peat soils	706	3.0	



	Mineral soils – no organic surface layer
	Organo-mineral soils with less than 15 cm surface organic horizon
	Organo-mineral soils with less than 40 cm surface organic horizon
	Organic soils with between 40 and 100 cm surface organic horizon
	Organic soils with between 100 and 200 cm surface organic horizon
	Organic soils with more than 200 cm surface organic horizon

Figure 1.3 Thickness of organic layers in Welsh soils

1.3 Calculation of organic matter depth for organic soils

1.3.1 Scotland

Introduction: Most recent work assessing the carbon stocks present in Scottish and UK soils (Bradley *et al.* (2005) focused on the top 1 metre of the soil surface, although it was realised that many of the peat deposits were deeper. Assigning depths to the peat polygons occurring on the 1:250 000 scale soil map gives a more realistic estimation of the carbon stocks contained in the peat deposits.

Methodology: The 1:250 000 scale digitized soil map of Scotland formed the framework for the work and the 1328 peat polygons occurring on the map were extracted along with a unique National Grid Reference (NGR), the area of each polygon, ranging in size from 1.25 ha to 53989.3 ha, and the type of peat deposit mapped. The map units and corresponding peat type are shown in Table 1.4.

Table 1.4 Peat map units and type on 1:250 00 scale soil map of Scotland

Unit	Peat Type
3	Basin Peat (>0.5m)
4	Undifferentiated Blanket Peat (>0.5m)
603	Eroded Basin Peat (>0.5m)
604	Deep Blanket Peat (>1m)
605	Eroded Deep Blanket Peat (>1m)
606	Eroded Undifferentiated Blanket Peat (>0.5m)

All sources of peat data available were examined to ascertain peat depths and comprised the following sources:

1) The peat bogs occurring in the peatland database compiled by Sally Ward (circa 1991) which included bog name, eastings and northings of roughly the centre of the bog and mean depth, calculated from the depthing transects across the bog.

2) Depth information from Department of Agriculture and Fisheries for Scotland reports - Scottish Peat (1962), the second report of the Scottish Peat Committee and Scottish Peat Surveys Vols 1-4 (1964-1968).

3) Reports held at The Macaulay Institute by the Scottish Peat Committee (SPC) and International Survey Committee (ISC) Reports of individual bogs were checked and relevant data extracted. Soil Survey of Scotland memoirs were also referred to and data extracted.

4) Peat Survey field notebooks held at the Macaulay Institute were studied and depth information extracted. In most instances this necessitated calculating average depths, sometimes from 300 or more individual recordings.

5) Peat Survey maps were also checked and, although data on most of the bogs had been ascertained from sources mentioned previously, some missing depth data was available, including some additional bogs.

6) More recent reports on survey work, undertaken on the main bogs used for commercial extraction were made available, which allowed more up to date measurements for the estimation of peat depth.

7) Forestry Commission site survey reports.

All bogs with known depth data (278) were entered in a spreadsheet containing bog name, NGR, average depth and source of data. Using GIS, these bogs were matched using the NGR, with peat polygons on the 1:250 000 scale soil map. Remaining polygons without depth data were assigned relevant data using trends from the information gathered above, local knowledge of the landscape and any depth information gathered by the Soil Survey of Scotland during systematic soil mapping.

Results:

1. Spreadsheet of all peat bogs with measured depth data, detailing name, average depth calculated from transect depth data and the source of information.
2. Spreadsheet detailing all peat polygons with NGR, area, estimated depth. The distribution of carbon stocks in soils at depths greater than 1m are show in Figure 1.4.

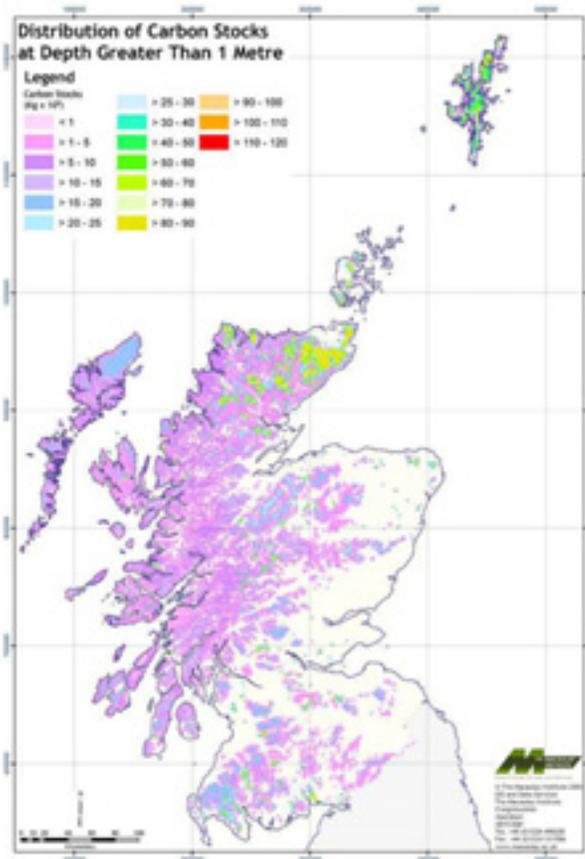


Figure 1.4 Distribution of carbon stocks at depth greater than 1m in Scotland

1.3.2 Wales

Soil patterns of the upper 100 cm of soil material have been mapped at scales from 1:10,000 to 1:250,000. However, there is a considerable gap in the knowledge and understanding of the thickness of peat deposits across Wales (and England) below 100 cm depth. The primary reason for this is that traditional methods of field survey use 100 cm long hand augers. A secondary reason is that considerable additional effort, time and equipment is needed to investigate below 100 cm depth.

Given the costs in terms of the time and labour required using an auger to measure the peat below 100cm depth, the approach adopted was to carry out a desk study of existing sources of data from Wales. Although some attempts have been made at shallower depths with radar the technique is not suitable for peats as deep as found here. In the desk study,

many hundreds of paper-based observations were reviewed to derive the tables for Pwllpeiran, the Dee catchment, National Soil Inventory and Forestry Commission sites. The outcome of these investigations gives the best possible estimates available for the amount of organic material below 100 cm that is now available.

For consistency, the same methodology for material below 1 m depth was used as for material above 1 m in the estimate of soil carbon in the UK (Bradley *et al.*, 2005). The one exception was to reduce the Bulk Densities for some of organic layers in the light of work both within the Project, and for from other studies. Within Wales three main blocks of land were identified where there was evidence suggesting that the thickness of additional organic material below 1 m depth would be similar (Table 1.15). CEH then added the stocks from these deep organic layers to the totals derived in the UK study (Bradley *et al.*, 2005) using the same methodology and spatial database to compile the data presented in Figure 1.16.

The main assumption was that the stock of organic carbon, bulk density and layer thicknesses are the same for a soil series wherever it occurs in Wales. This is consistent with the usual assumptions for estimates of soil properties, used across the soils community in similar exercises. Four field-based studies with measured data on the thickness of peat soils from various parts of Wales have been re-assessed. In addition, as part of the ECOSSE project, transects were made across three areas with landscapes typical of south mid and north Wales.

Taylor (1974) presented some estimates of peat depths (Table 1.5) and suggests that the bulk of peat above 300 m OD is characteristically 1.83 ± 0.61 m in depth but up to 3.05 m in places. In the lowland peat areas, he estimated that Cors Fochno (Borth Bog) attained 7.62 m and Cors Caron (Tregaron Bog) 9.15 m depth.

Table 1.5 Estimated thicknesses of peat deposits in Wales (after Taylor, 1974)

Depth range	Assumed average depth	Area (km ²)	Percent of Wales
More than 90 cm	183 cm	842	4.0
Less than 90 cm	46 cm	746	3.5
Total		1588	7.5

Lowland Peat Inventory: Thirty seven lowland peat areas totalling 50.9 km² were identified by Burton and Hodgson (1987) in the Lowland Peat Inventory. The thickness of the peat was estimated, with some measurements in the larger areas.

National Soil Inventory: Examination of the National Soil Inventory data for Wales shows that there are 29 sites described as peat soils with less than 90 cm peat (median thickness 80 cm) and 7 sites with more than more than 90 cm (mainly described as “90+”, with one observed depth of 300 cm) (Table 1.6).

Table 1.6 Summary of results from surveys of thicknesses of peat soils across Wales

	Median thickness of organic soils (soils with 40 to 90 cm peat)	Median thickness of organic soils (soils with more than 90 cm peat)	% of organic soils in study area with more than 90 cm peat
National Soil Inventory (sampled c. 1980)	80	No data	24
Dee catchment (Rudeforth & Thomasson 1970)	51	No data	43
Forestry Commission survey (20 sites in mid Wales sampled c. 1980)	85	187	90
Waun Figlen Felen (sampled 2005)	68	190	78
Pwllpeiran (sampled 2005)	66	99	50
Meigneint (sampled 2005)	63	126	37

Dee catchment study: Rudeforth and Thomasson (1970) reported on survey work across the upper Dee catchment. Their original survey records have been re-examined and the medians for peat thicknesses calculated (Table 1.6 and Figure 1.4). These were based on 65 sites with peat 40 cm or more thick. Sites with more than 99 cm peat were recorded in the field as “90+ cm”.

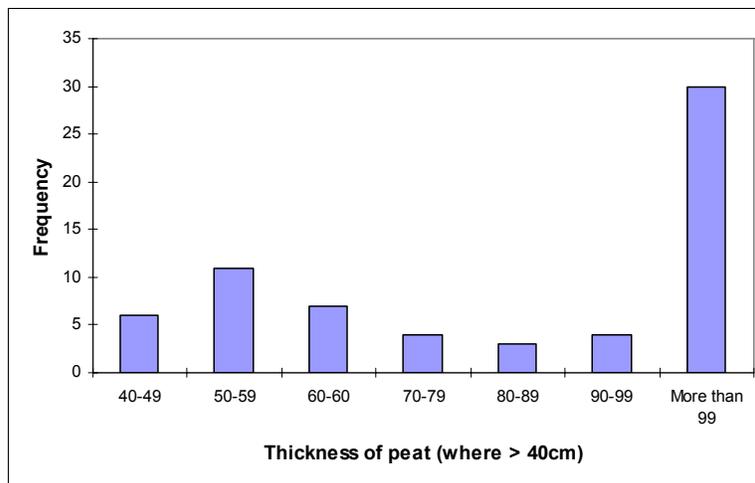


Figure 1.4 Thickness of peat soils in the Dee catchment

Waun Figlen Felen, Powys: Waun Figlen Felen is a small basin bog at 450 m OD in the Brecon Beacons above the Dan yr Ogof cave system (SN825180). Parts of the Waun are suffering from severe erosion where the surface vegetation has disappeared following uncontrolled burning. Three transects were made from south east to north west across the area, the thickness of peat measured with an extendable gouge auger (Figure 1.5) and the median peat thickness calculated (Table 1.6).

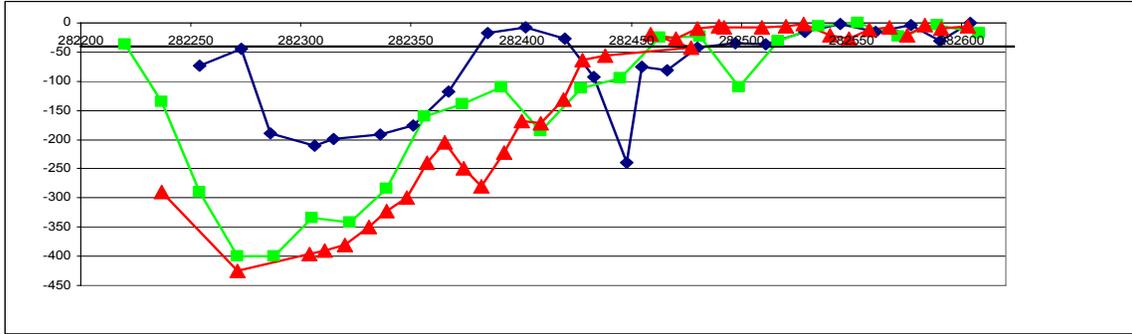


Figure 1.5 Distribution of peat thickness along transects on Waun Figlen Felen

Pwllpeiran, Ceredigion: Pwllpeiran farm extends over 1130 ha south of Plynlimon in the upper Ystwyth valley (SN820780). The four transects were located at 480 to 550 m OD based across areas dominated by peat soils in the grid survey of 1985. The thickness of peat was measured with an extendable gouge auger (Figure 1.6) and the median peat thickness calculated (Table 1.6).

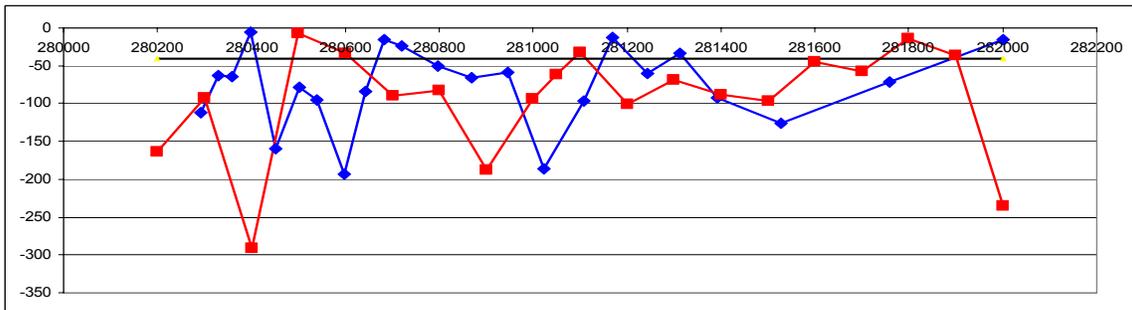
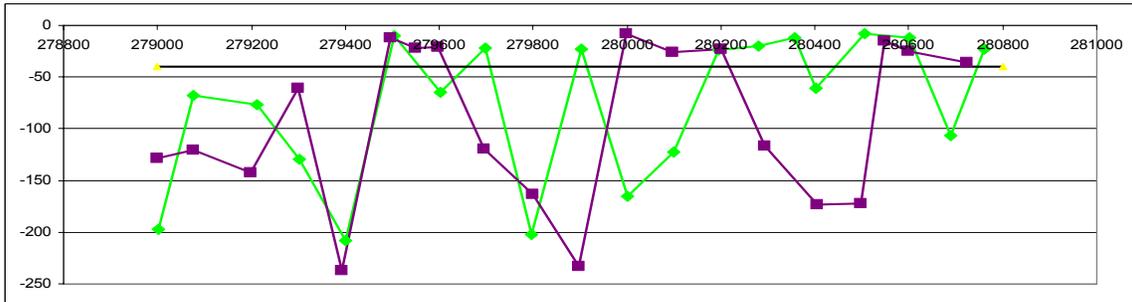


Figure 1.6 Distribution of peat thickness along transects at Pwllpeiran

Meigneint, Ffestiniog, Gwynedd: The moorland below Llyn y Dywarchen west of Mynydd Meigneint (SH760420) rises from 420 to 500 m OD. The thickness of peat was measured with an extendable gouge auger (Figure 1.7) and the median peat thickness calculated (Table 1.6).

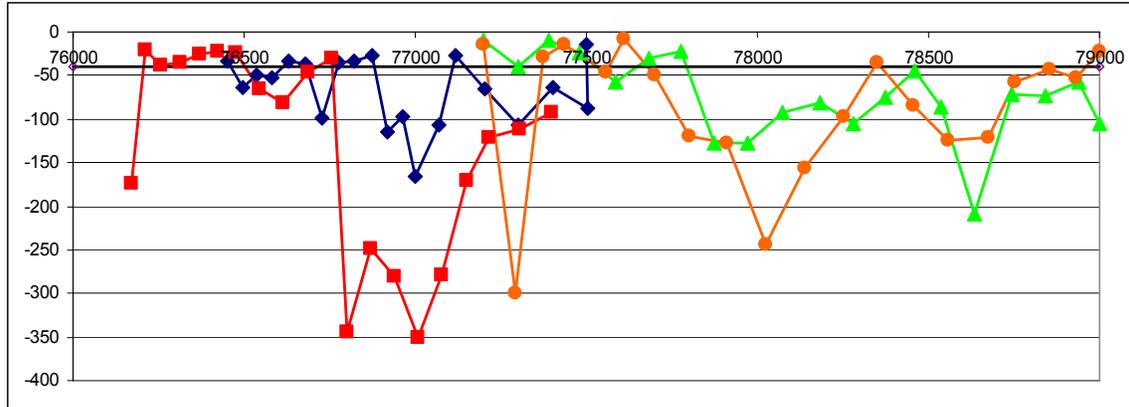


Figure 1.7 Distribution of peat thickness along transects at Meigneint, Ffestiniog

1.4 Validation of mapped estimates of carbon stocks in organic soils in two upland catchments

1.4.1 Introduction

This work aims to assess the accuracy of mapped estimates of soil C stocks in two upland catchments. The first site was located in Plynlimon, mid Wales (SN800860) and the second in Glensaugh near Aberdeen, Scotland (NO650800). In both areas three 1km grid squares were sampled intensively (200m grid), with samples taken from the organic and mineral horizons (where present). Measurements included depth of organic horizon, depth of mineral horizons, bulk density (organic horizon only) and % C (Table 1.7). The sampling design enabled an estimate of the carbon stock to be made, which could then be compared to values contained in the National Soils Map (Natmap).

Table 1.7 Soil properties measured at Plynlimon and Glensaugh

	Variable	Depth/horizon
Organic horizon	Depth of horizon	
	Bulk density	0 – 15 cm
	Bulk density	15 – 30 cm
	Bulk density	50 – 65 cm
	% C	0 – 15 cm
	% C	15 – 30 cm
	% C	50 – 65 cm
Mineral	Depth of horizons	
	% C	A horizon
	% C	B horizon

1.4.2 Calculating carbon stocks

The stock of carbon, in $t\ C\ ha^{-1}$, was calculated for the organic and mineral horizons separately and then added together (Figure 1.8). For the organic horizon, the depth, bulk density for 0-15cm and % carbon for 0-15cm was used in the calculation. The errors associated with using these depths are discussed below (Section 1.4.4). For the mineral soil, the depth of each horizon, % carbon and estimated bulk density was used. Since bulk

density was not measured for the mineral soil, representative figures were used for each horizon based on soil type and land use (Source: NSRI and MLURI).

1.4.3 Differences between measured and modelled carbon

The carbon stock calculated for each area was compared to the carbon stock predicted from the Defra funded GHG inventory project (Defra project SP0511). The results for each 1km square and for the total area are given in Table 1.8. Both Plynlimon and Glensaugh show large differences between the measured and modelled carbon for each 1km square. At Plynlimon, some squares are overestimated and others underestimated. For example, in Square 1 at Plynlimon the amount of carbon determined from the measured data is 23 Kt C, compared to 68 Kt C from the model, a difference of 193%. At Glensaugh, the model underestimates the amount of carbon in each square. The main reason for these differences relates to the proportion of organic soil. At Plynlimon, the model data shows considerably more peat for Square 1 then was observed. Whereas at Glensaugh, more peat was found in each square then predicted. This is discussed in the following section.

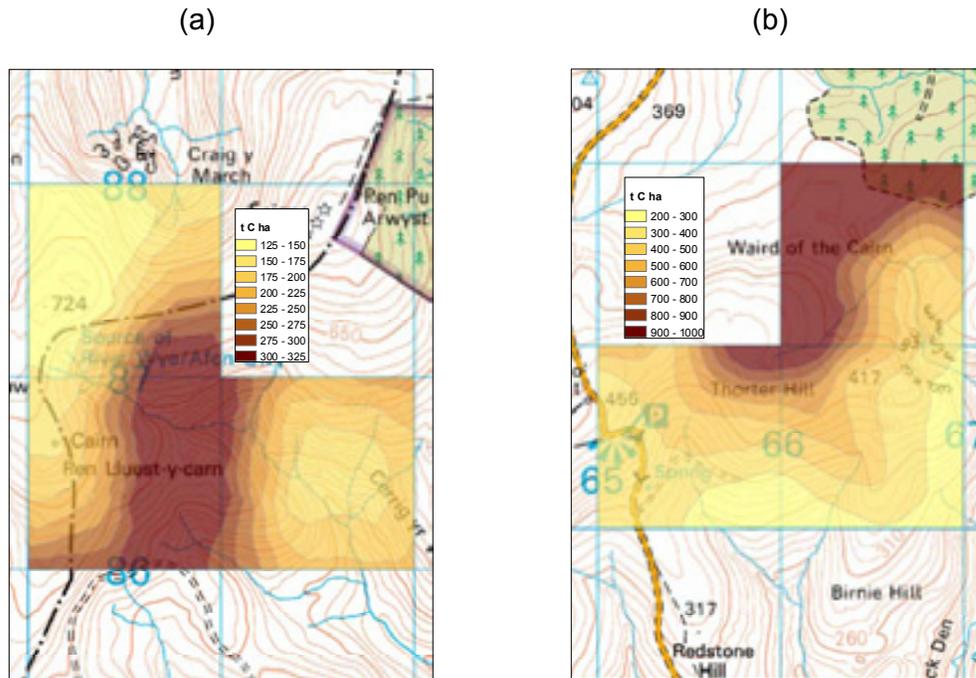


Figure 1.8 Total carbon ($t C ha^{-1}$), including organic and mineral layers for (a) Plynlimon and (b) Glensaugh.

When the differences from each square are summarised as a total, however, they appear to be less, with an error of 28% for Plynlimon and 24% for Glensaugh. Therefore, although there are large differences within individual squares, overall the differences appear smaller. This is, however, only for six 1 km squares and the same might not be true elsewhere.

Table 1.8 Measured and modelled values of Kt C for Plynlimon and Glensaugh.

	Kt C			
	Square 1	Square 2	Square 3	Total
Plynlimon				
Measured	23.27	34.45	23.58	81.30
Modelled	68.10	16.61	19.09	103.80
Difference	44.83	-17.84	-4.49	22.50
% Difference	192.68	-51.79	-19.04	27.68
Glensaugh				
Measured	50.06	99.41	48.52	197.99
Modelled	42.32	70.28	38.11	150.71
Difference	-7.74	-29.12	-10.41	-47.27
% Difference	-15.47	-29.30	-21.45	-23.88

1.4.4 Sources of uncertainty when estimating carbon stocks

Two main sources of uncertainty were identified when determining carbon stock, the value used for bulk density and the accuracy of defining the coverage of different soil types due to differences in the sampling density required for different mapping scales (1:250,00 for Wales as a consistently available scale for the whole Principality and 1:25,000 for the squares mapped in this Module). A further factor could be variations in % C, however, since this changed only slightly down the profile it was not considered here.

Bulk density: At both sites, the bulk density of the organic horizon decreased with depth, and varied depending whether it was a deep or shallow peat. For example, at Plynlimon the bulk density for deep peats at 50-65cm (0.12 g cm^{-3}) was significantly less than 0-15cm (0.20 g cm^{-3}) and less than half that for 0-15cm in shallow peats (0.31 g cm^{-3}). To examine the effect that bulk density has on the carbon stock, different bulk density values were used in the calculation: 1. measured bulk density for 0-15cm, 2. average bulk density for 0-15cm, 3. average bulk density for 15-30cm, 4. average bulk density for 50-65cm. This analysis focuses on the organic horizon only, the mineral horizons were excluded for this purpose. Average bulk density values had to be used for 15-30 cm and 50-65 cm depths due to the fewer sample points. The maps for Plynlimon and Glensaugh are given in Figure 1.9 and Figure 1.10. For Plynlimon, using the measured or average bulk density for 0-15cm produces very similar results. There are large differences, however, when using the average value for 15-30cm and 50-65cm. The maps for Glensaugh do not suggest that these changes are as extreme.

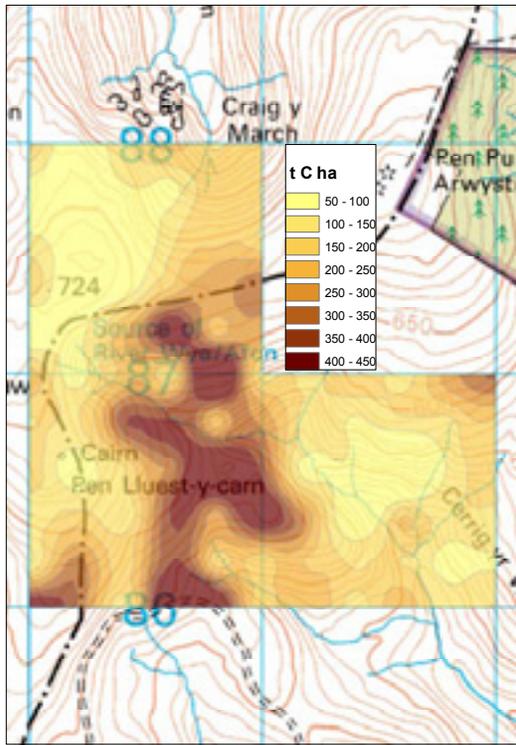
To examine this further, the total carbon for each area, and within a 1km square, was calculated using the 4 different approaches. For Plynlimon, using an average bulk density for 0-15cm and 50-65cm produces differences of around 40% compared to values determined using the measured bulk density at 0-15cm. These depths also have the greatest errors at Glensaugh, but here the errors are around 25% (Table 1.9). This analysis suggests that the bulk density figure should be representative, particularly for organic soils, otherwise it could produce misleading results.

Table 1.9 Comparison of Kt C ha⁻¹ calculated using four different methods (organic horizon only).

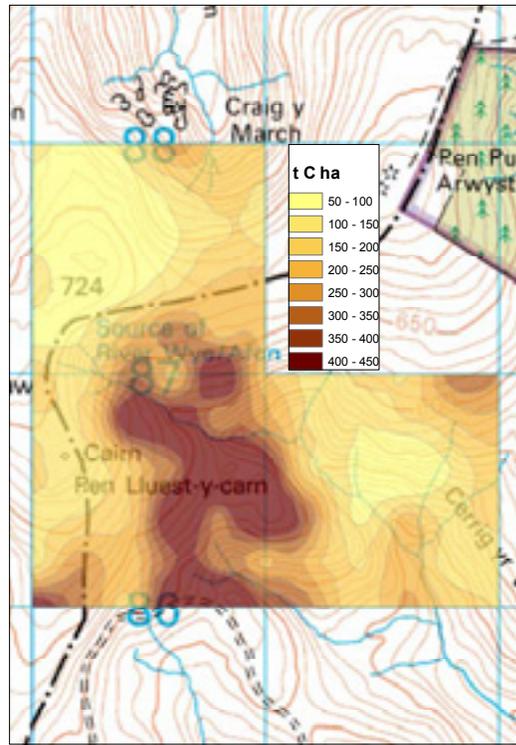
	Square 1	Square 2	Square 3	Total	% difference from method 1
Plynlimon					
1. Measured 0-15cm	18	30	16	65	
2. Average 0-15cm	19	47	25	90	39.57
3. Average 15-30cm	14	35	19	68	4.71
4. Average 50-65cm	8	20	11	39	-39.27
Glensaugh					
1. Measured 0-15cm	35	97	36	168	
2. Average 0-15cm	40	137	38	215	27.69
3. Average 15-30cm	28	97	27	153	-9.20
4. Average 50-65cm	23	81	23	127	-24.81

Soil type: A second source of error is the proportion of each soil series within a square. Many models use this to determine the carbon stock, as each soil type has associated with it a typical bulk density and percentage carbon.

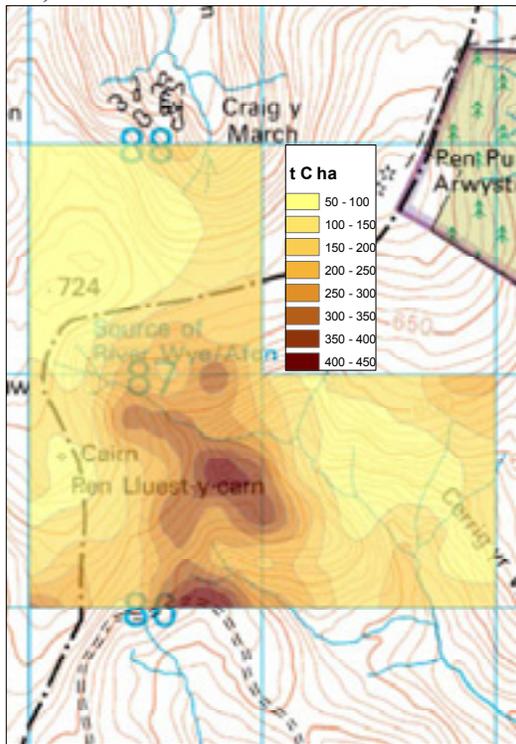
Figure 1.11 shows the differences in coverage between the observed soil and NSRI mapped soils for Plynlimon. There were three main soil types observed at this site (a raw oligo-amorphous peat soil, a ferric stagnopodzol, and a cambic stagnohumic gley), with an intergrade between the two mineral soils also present. The classified soil for Plynlimon, however, shows that in some squares the amount of deep peat (raw peat) has been over estimated and in others underestimated. This has a significant effect on the carbon stocks, as discussed earlier. At Glensaugh, nine soil types were observed, ranging from deep peat to surface ground water gleys (Figure 1.12). This complexity is not reflected in the classified soil. It should be noted that the 1:250,000 map was not designed to be used at this scale, nor does it give the proportions of soils within map units. These were made available by NSRI from un-published data. However, it highlights the importance of using reliable data which is suitable for the purpose intended, particularly where organic soils are concerned.



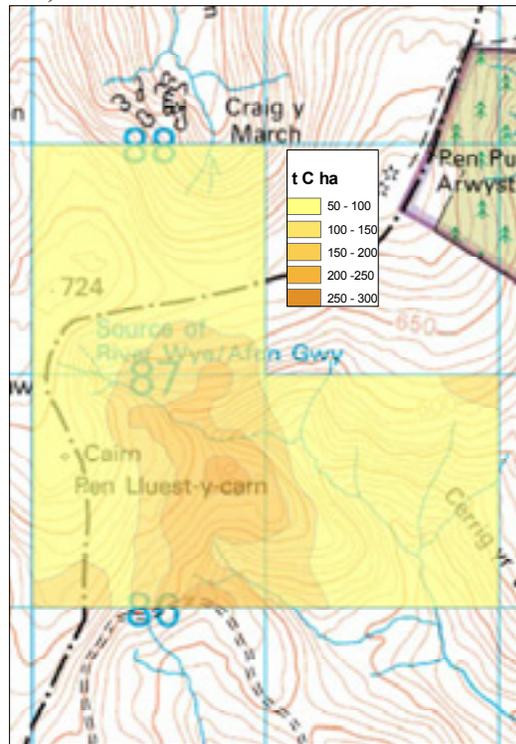
a)



b)

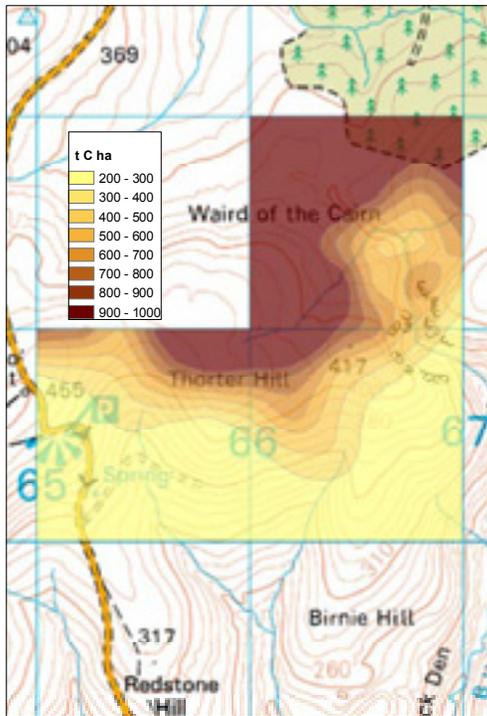


c)

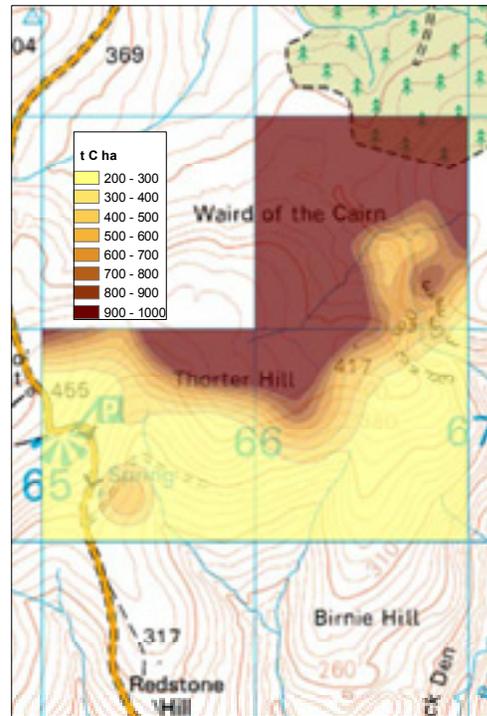


d)

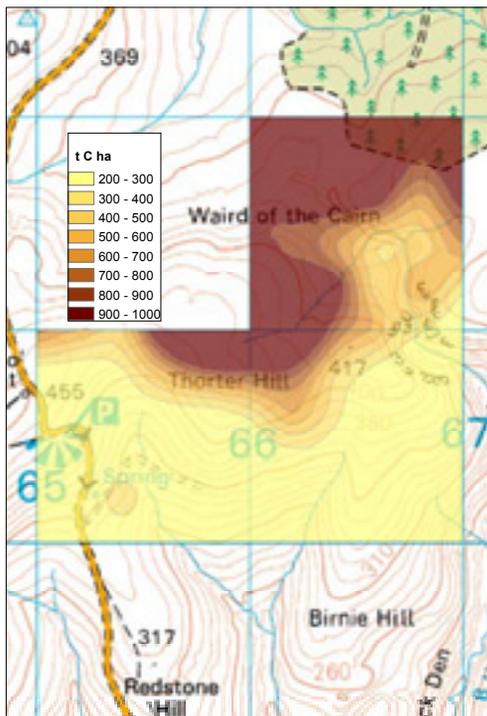
Figure 1.9 Plynlimon $t C ha^{-1}$ for the organic horizon using (a) measured bulk density for 0-15cm, (b) average bulk density for 0-15cm (c) average bulk density for 15-30cm and (d) average bulk density for 50-65cm.



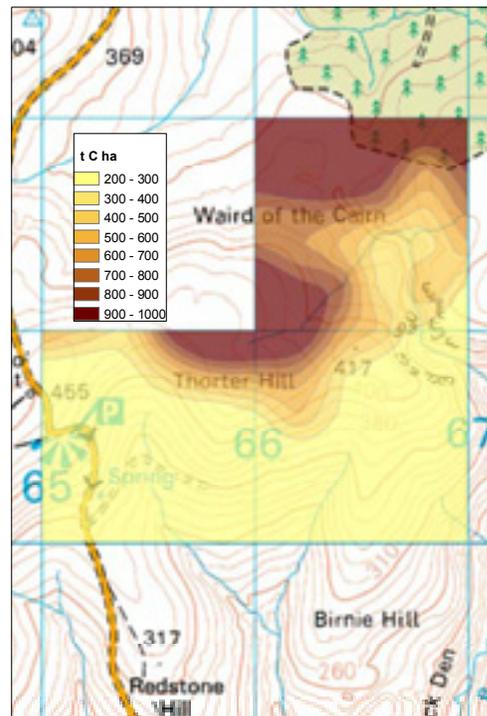
a)



b)



c)



d)

Figure 1.10 Glensaugh $t C ha^{-1}$ for the organic horizon using (a) measured bulk density for 0-15cm, (b) average bulk density for 0-15cm (c) average bulk density for 15-30cm and (d) average bulk density for 50-65cm.

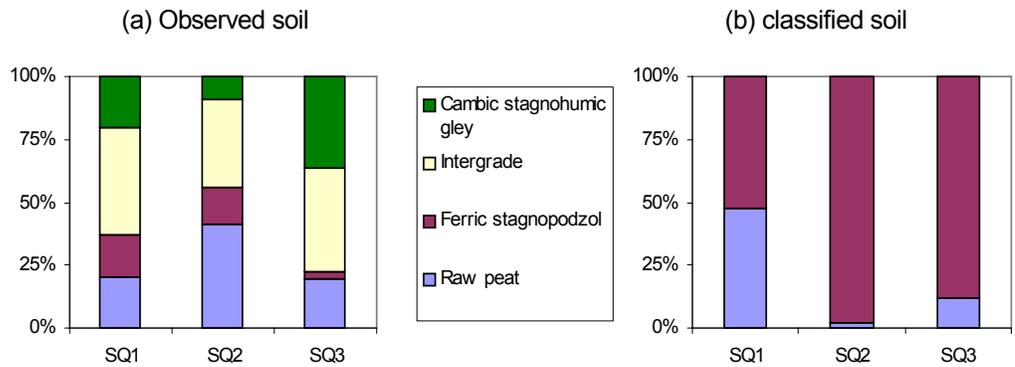


Figure 1.11 Proportions of soil series within each grid square at Plynlimon (a) observed soils (b) Classified soil (Source: NSRI - Natmap data 1:250,000 map).

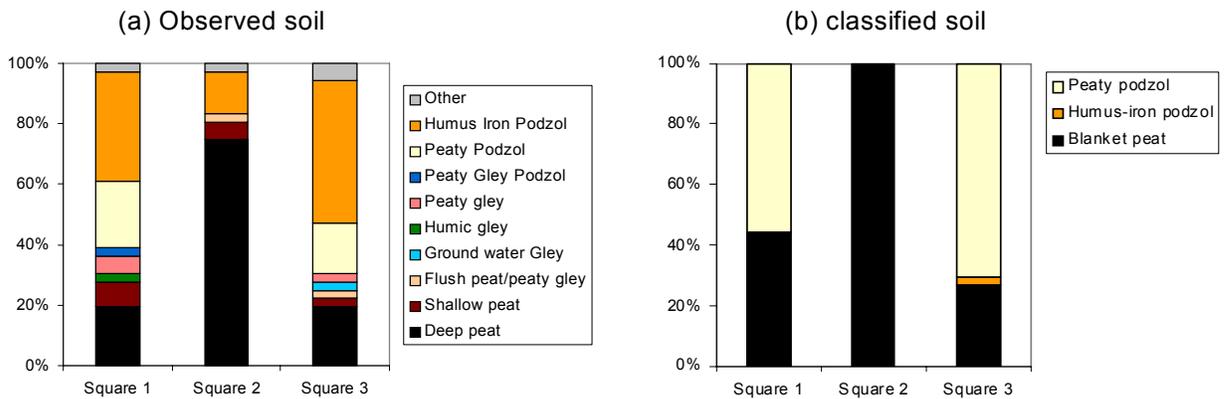


Figure 1.12 Proportions of soil series within each grid square at Glensaugh (a) observed soils (b) classified soil types (Source: Ronnie Milne).

1.4.5 Summary

This work has produced high-resolution maps of measured soil carbon stocks for two representative upland sites: Plynlimon, Wales and Glensaugh, Scotland. Estimates of soil carbon stocks were compared to model predictions, which gave overall errors of 24-28%. Accurate estimates of soil carbon stocks were found to depend on:

- *Bulk density.* The sample data showed that bulk density varied for both deep and shallow organic horizons, and also decreased with depth. Selecting an unrepresentative value gave large errors.
- *Appropriate soil mapping scale.* There were large discrepancies between measured and mapped series. The resultant differences in carbon stock estimation were large at the local scale but somewhat reduced at larger scales. The largest errors occurred where peat soils (relating to the depth of the organic horizon) were misrepresented. It should be recognised that most national soil maps were not designed to be used at such fine scales. Soil maps at more appropriate scales are available for more local areas in Wales, for example Rudeforth (1970).

1.5 Estimation Of The Carbon Within The Peats And Organo-Mineral Soils

1.5.1 Scotland

Introduction: Calculation of carbon stocks using the peat depths determined in Section 1.4, in combination with best estimates of bulk density, enhances the data from Defra work (Defra SP0511 project), giving a better estimate of carbon stocks within the peat lands of Scotland.

Methodology: A spreadsheet was constructed of all peat map units with grid references, peat depth, estimated profile horizons, % carbon of each horizon (from analytical data in the National Soils Database), estimated bulk density of each horizon and estimated carbon mass to profile depth. Bulk density values were predicted using regression equations which are described in Appendix 1. After additional sampling, modifications to the bulk density values used in the Defra project, were incorporated and used in estimations of carbon stocks. As this impacted on the original Defra database, the carbon stocks were re-calculated using the new bulk density values and the revised figures sent to CEH (Bush) for appropriate amendments to the database and national totals.

Results: To allow for consistency with the Defra database and spatial calculations, which assumed peat depths of 1m, weighted average depth values for the 6 types of peat (deep blanket, eroded deep blanket, undifferentiated blanket, eroded undifferentiated blanket, basin and eroded basin) occurring on the 1:250 000 scale soil map of Scotland were used. These were weighted averages taking into account the areas of each polygon and the type of peat (Table 1.10).

Table 1.10 Weighted average peat depth in peat map units

Peat Type	Weighted Average Depth (m)
Undifferentiated Blanket Peat	1.3
Undifferentiated Eroded Blanket Peat	1.3
Deep Blanket Peat	2.3
Eroded Deep Blanket Peat	1.7
Basin Peat	2.9
Eroded Basin Peat	2.7

Using these depths, the carbon stocks were calculated for the peat polygons and using GIS, the carbon stocks present at depths greater than 1 metre were determined and plotted. In addition to the peat polygons, peat also occurs as a component in other mapping units and these were also included. A similar procedure was adopted and peat depth values were assigned to mapping units where peat (3 types) was a component.

Table 1.11 Weighted average peat depth in other map units

Peat Type	Weighted Average Depth (m)
Blanket peat	1.1
Basin peat	2.9
Semi-confined peat	1.3

Weighted average mean depths were determined (Table 1.11) and used to calculate carbon stocks for peat present at depths greater than 1 metre where it occurred as a map unit component. Combining the two datasets the carbon stocks at depths greater than 1 metre were calculated (Table 1.12). The carbon stocks values (>1 metre depth) for the peat soils were calculated for each kilometre square. This data was supplied to CEH (Bush) to amend the total national values as calculated previously and a map showing the carbon stocks >1 metre for Scotland was produced.

Organo-mineral Soils: Using soil profile morphological data in the National Soil Database, which is held at the Macaulay Institute, modal profiles comprising soil horizon type, sequence and thickness for each of the component soil series of the 1:250 000 soil map units had previously been assembled and used in work conducted under Defra project SP 0511. Using analytical data, % carbon values for each soil horizon were assigned and in combination with bulk density values, predicted by regression equations (Appendix 1), carbon stocks were calculated for the depth ranges 0-30 cm, 30-100 cm (Defra SP 0511). With area figures determined by GIS for each soil series, and the profile carbon stock totals, the carbon stocks for each series with an organic surface horizon was calculated. The soil series were then grouped according to major soil subgroups and the carbon stock totals for these major soil subgroups and the organic soils estimated (Table 1.12).

Table 1.12 Estimated carbon stocks (MtC) for peat and organo-mineral soils of Scotland

Major Soil Subgroup*	Stock (>1 metre depth)	Stock (<1 metre depth)	Total Stock
Blanket peat	285.7	799.0	1084.8
Basin peat	65.5	58.2	123.6
Semi-confined peat	134.2	435.0	569.2
Deep peat total	485	1292	1778
Humus-iron podzol (uncultivated)	0	77.7	77.7
Peaty podzol	0	324.6	324.6
Subalpine podzol	0	78.8	78.8
Alpine podzol	0	16.5	16.5
Peaty gley	0	384.6	384.6
Humic gley	0	15.6	15.6
Peaty ranker	0	34.7	34.7
Peaty lithosol	0	3.4	3.4
Peat alluvium	0	21.1	21.1
Organo-mineral total	0	957	957
Total** (peat and organo-mineral soils)	485	2249	2735

* Only those major soil subgroups with an organic surface horizon.

** Excludes mineral soils.

1.5.2 Wales

Bulk density of peat soils: Vernik (2005) visited a number of National Soil Inventory sites across upland Wales previously described as having > 40 cm of peat. Samples of known volume were taken at 10, 30 and 50 cm depth and the bulk densities measured (Figure 1.13). The bulk densities of the samples at 10 cm depth are related to land use in Figure 1.14. This new information on bulk density of peat soils have been used to inform the conversion of weight to volume calculations of carbon stocks (Bradley *et al.*, 2005).

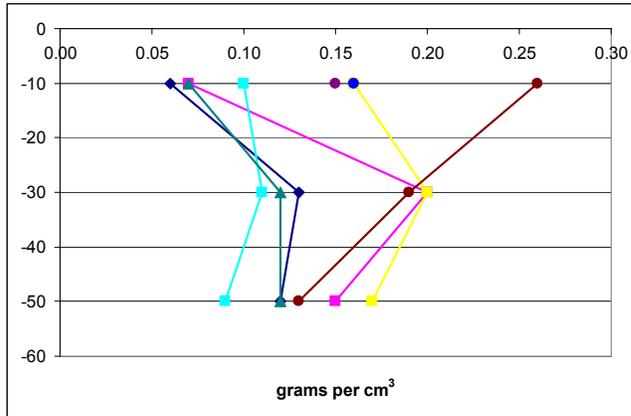


Figure 1.13 Variation in bulk density with depth in six Welsh peat soils

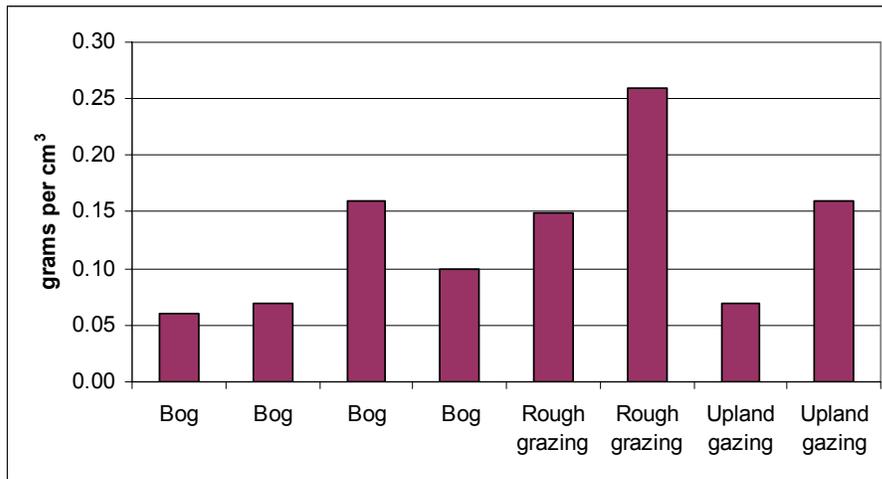


Figure 1.14 Bulk density at 10 cm depth by land use in eight Welsh peat soils

Revised inventory of soil organic carbon: The inventory of soil organic carbon (SOC) for the UK (Milne *et al.*, 2004 and Bradley *et al.*, 2005) quantified the amount of SOC in the profile to a maximum of 100 cm depth. One of the aims of the ECOSSE project was to quantify the amount of SOC below 100 cm depth to complete the picture of soil carbon stocks in Wales and Scotland. The approach adopted for Welsh soils, given that there is no systematic depth data, was as follows:

1. A “bog-wide” depth was allocated to each lowland area dominated by peat soils. Each is associated with Adventurers or Altcar soils and the results summarised in Table 1.10,
2. A “region-wide” depth was allocated to each association mapped in upland Wales,

3. This added a fixed amount of SOC (as kg x 10⁶ per km²) to each series, regionalised as much as is feasible given the nature of the data,
4. Each series has a known distribution from the spatial database compiled by Bradley *et al.* (2005) and therefore spatial distribution of the addition SOC can be mapped (Figure 1.16).

Table 1.15 Additional soil organic carbon below 100 cm (kg x 10⁶ per km²)

Series	South of grid line 240	North of grid line 300	Square SN66	Square SN69	Squares SS69 and SS79	Remainder
Longmoss	NA	NA	NA	NA	NA	NA
Winter Hill	26	90	600	650	550	0
Crowdy	26	90	600	650	550	0
Turbary Moor	NA	NA	NA	NA	NA	NA
Altcar	0	425	0	100	100	425
Adventurers'	0	500	0	100	100	500
Mendham	NA	NA	NA	NA	NA	NA

NA – no peat below 100 cm depth.

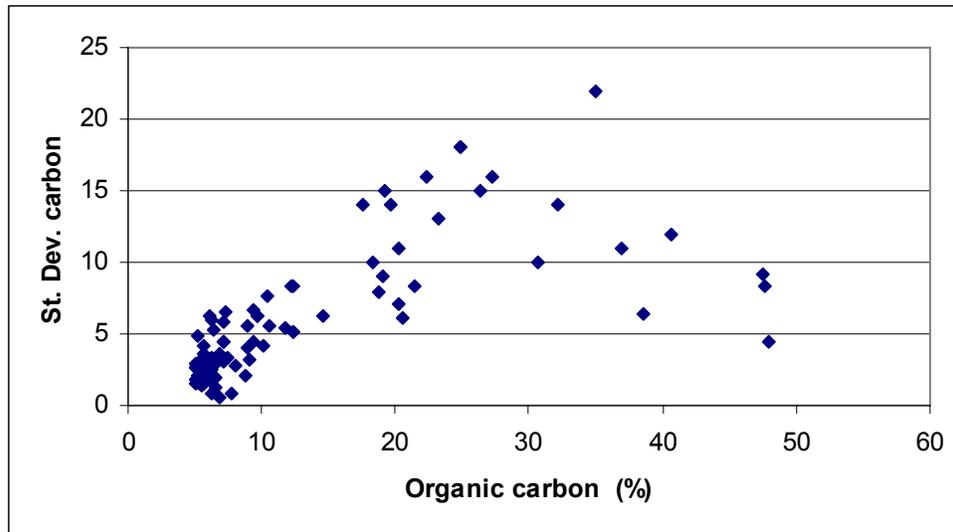
Organic soils (peats and organo-mineral soils) cover about 20% of the area of Wales, yet contain 50% of the carbon because peats are carbon rich all the way down the profile and organo-mineral soils have organic rich top soils. In mineral soils, which cover the remaining 80% of Wales, the maximum soil organic carbon content in the top soil is about 4%, over a subsoil with 1 to 2% organic carbon, whereas the organo-mineral top soils (and peats throughout the whole profile) can be up to 50% organic carbon. Total stocks of carbon in peats and organo-mineral soils in Wales are shown in Table 1.16.

Table 1.16 Estimated carbon stocks (MtC) for peats and organo-mineral soils in Wales

Major subgroup		Stock (>1 depth)	Stock (<1 depth)	Total stock
Raw peat	101n	47.9	67.6	115.5
Earthy peat	102n	3.7	2.1	5.8
Peat total		51.6	69.6	121.3
Humic rankers	311	0	3.8	3.8
Podzols	63n	0	5.3	5.3
Stagnopodzols	65n	0	25.3	25.3
Stagnohumic gleys	72n	0	39.7	39.7
Humic gleys	87n	0	0.3	0.3
Organo-mineral total		0	74.5	74.5
Total		51.6	144.1	195.8

Errors of quantifying organic carbon up to 1 metre depth have been assessed previously. These data formed the basis of the UK study (Bradley *et al.*, 2005). Figure 1.15 shows the organic carbon of topsoils of individual soil series where organic carbon is greater than 5% plotted against the Standard Deviations, showing that the variance becomes greater as the carbon content increases.

Figure 1.15 Organic carbon of topsoils of individual soil series where organic carbon is greater than 5% plotted against the Standard Deviations



1.6 Conclusions

The work in Module 1 provides estimates of the carbon stocks for the peat and organo-mineral soils of Scotland and Wales (organic soils collectively), taking cognizance of previously unused peat depth data and a re-classification of the soils to allow estimates at major soil subgroup level. Estimation of carbon stocks for depths > 1 metre is now possible and in conjunction with previous work (Defra project SP0511), will enable more precise national estimates to be made. Estimates for peats and organo-mineral soils are higher than those quoted in previous work (Bradley *et al.*, 2005), which is due to a combination of factors such as the inclusion of peat greater than 1 metre, modifications to the bulk density and the methods adopted in calculating the areas. The combined map of soil organic carbon stocks in Scotland and Wales are shown in Figure 1.16.

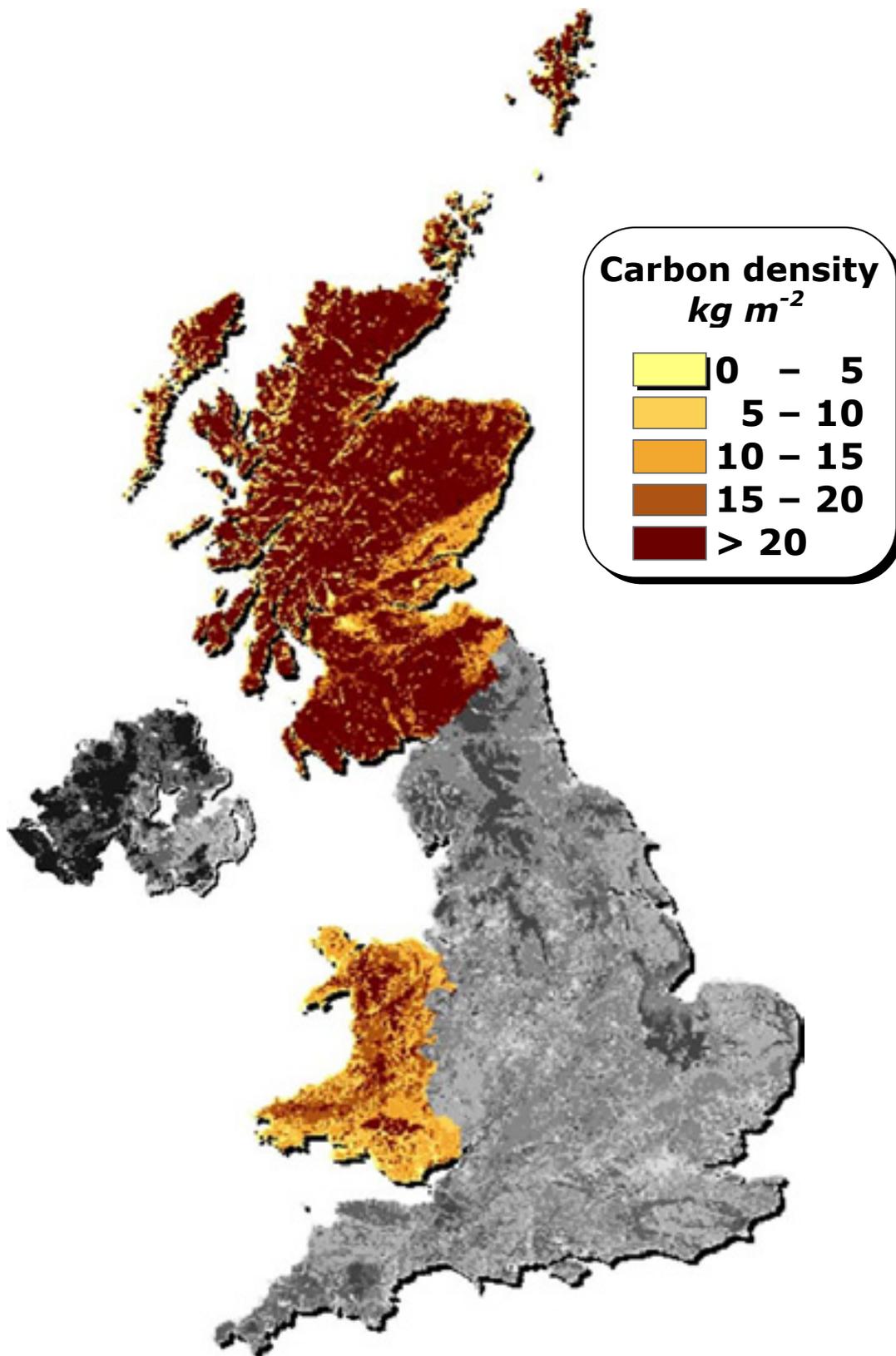


Figure 1.16 Distribution of organic carbon in soils (developed from Milne *et al.*, 2004 and Bradley *et al.*, 2005).

Module 2

Development of a model to simulate C and N cycling in organic soils to provide predictions of their response to land-use, management and climate change

2.1 Modelling soil C and N cycling in organic soils

2.1.1 Where current soil carbon models do not work

The model proposed for possible use in the UK national greenhouse gas inventory (RothC-UK) was run to show soils for which it is unsuitable. These are the areas in which the new organic soil model is needed. Figure 2.1 shows the soils at 0-30cm depth (in red) for which RothC does not provide credible SOC values. Failure is defined by unrealistic high C inputs being required (i.e. $>5 \text{ t C ha}^{-1} \text{ y}^{-1}$) to maintain the measured SOC level. Such high inputs are unfeasible given that net primary production in these areas could not support this input. The large majority of the soils in Scotland and Wales fall into this category.

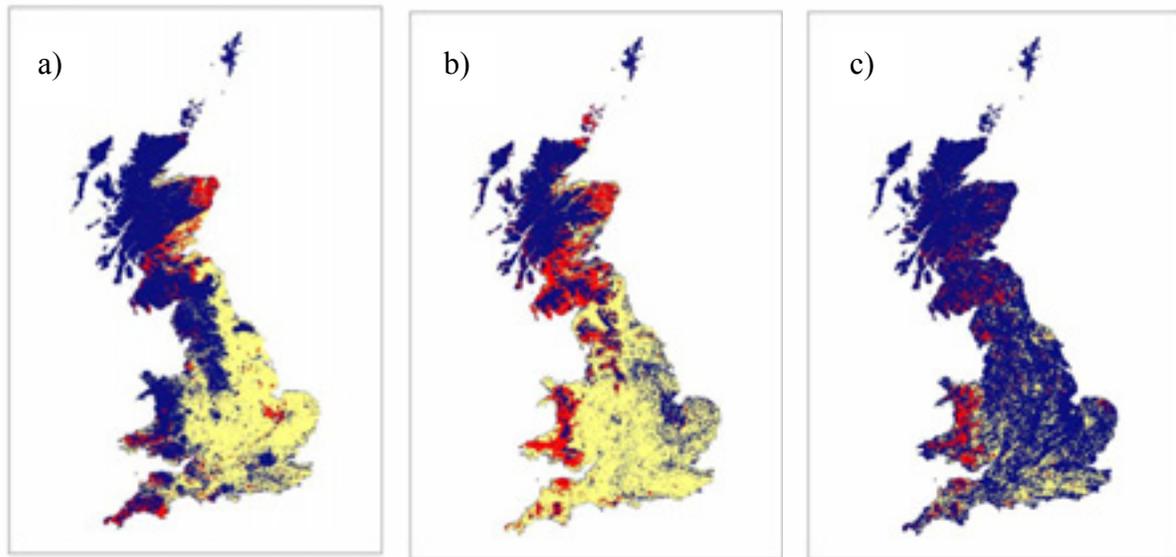


Figure 2.1 Areas (in red) where RothC-UK does not simulate SOC satisfactorily. Areas in blue were not run (peats and other highly organic soils where RothC-UK is known to be unsatisfactory). For soils under arable (a), grassland (b), and semi-natural (c) land use. Woodland is similar to semi-natural (not shown).

2.1.2 Development of the ECOSSE model - summary

Data for preliminary evaluation was collated and used to evaluate the existing models, RothC, SUNDIAL, CENTURY and DNDC. A new model (ECOSSE) was formulated using components of the single point models RothC, SUNDIAL, CENTURY and DNDC, and constructed into a framework that allows spatial datasets to be simulated. Data were collated on greenhouse gas emissions, leaching losses and nitrogen and carbon stocks in the soil to be used for evaluation of the new model. This data has been used to evaluate

the performance of ECOSSE. Simulations are generally within experimental error for both mineral and organic soils, so the model can be used to provide estimates of greenhouse gas emissions from Scottish and Welsh soils.

A database of existing experimental data was constructed, recording all information needed to run the models. These include location, current and previous land use, vegetation type, weather data, soil carbon content, soil pH, lowest summer water table, available soil water, field capacity, bulk density, amount of litter, depth of anaerobic zone, changes in soil C and N content, gaseous C and N fluxes and measurements of labelled C and N. Data on methane, carbon dioxide, nitrous oxide emissions and hydrology are itemised in a database, specifying the type of data, and the time and spatial scales of the data.

Clicking with the mouse on the site code in the metadata sheet locates the actual data in the database. A range of macros has been constructed to download the data from the database in the format required by the model to run. This facility will aid repeated evaluation of the model.

The model comparison has highlighted the most accurate and feasible approaches for simulating carbon and nitrogen turnover in highly organic soils. These are being incorporated into the model development, and are discussed further in the next section.

The new model was constructed from the SUNDIAL-MAGEC model of C and N turnover in soil/plant system (Smith *et al.*, 2000). Characteristics of the ROTH-C, CENTURY and DNDC models have been incorporated into this basic framework, as well as relationships derived directly from experimental data. The new model allows simulation of emissions of all major greenhouse gases from the soil/plant system as well as simulating C and N losses through water movement. The SUNDIAL-MAGEC code was first restructured into a modular format that allows components of the model to be individually evaluated, reformulated and replaced. This new version of the model will be referred to from here on as ECOSSE, Estimator of Carbon in Organic Soils – Sequestration and Emissions. The completed developments of ECOSSE are listed below:

- 1) Modularisation of SUNDIAL/MAGEC;
- 2) Improved description of N₂O production;
- 3) Description of methane production and oxidation;
- 4) Description of dissolved organic matter (carbon and nitrogen): turnover and losses;
- 5) Improved layer structure in the soil profile;
- 6) Initialisation of the size and characteristics of the soil organic matter pools;
- 7) Incorporation of the effect of pH on soil processes;
- 8) Incorporation of the effect of saturated conditions on soil processes;

The detailed description of the model, including the form of equations used to describe the processes included, is given in Annex 1.

2.2 Collection of new data

2.2.1 John Miles Birch Plots

Introduction: The most significant land use change ongoing in Scotland is afforestation (over the last five years, afforestation has averaged 6000ha y^{-1}) and the majority of new planting and natural regeneration (ca. two-thirds) during the 1990s is on organic soils, with most of this on organo-mineral soils (Chapman *et al.*, 2001). New planting is now at a 35-year low of just 8000 ha for Britain including 4000 ha in Scotland (2005/2006) and afforestation in Scotland is now focussed on soil of low organic matter. While the above-ground biomass of woodland may be expected to eventually exceed that of the prior vegetation, often heather-dominated moorland, the effects of afforestation on the below-ground carbon stock are uncertain. Across Scotland two native species dominate natural regeneration: Scots Pine and Birch. These two species are considered to have contrasting effects on soil chemistry. For example, while Scots Pine tends to generate a more acidic environment than even heather moorland (Chapman *et al.*, 2003), birch has been reported to raise the soil pH and increase nutrient cycling (Miles & Young, 1980). SNH have conducted a project to stimulate natural regeneration of native birch in North East highlands by reducing/excluding deer grazing. To further investigate the effects of birch on heather moorland, John Miles (formerly of CEH) set up a fully replicated experiment of birch-heather plots at three upland sites in Scotland. An examination of the chemistry of the organic layer (O horizon) revealed a significant decrease in both mass and % carbon of the O horizon in the birch when compared with the heather (Mitchell *et al.*, 2006). The objective here was to repeat this sampling but to perform a complete carbon (C) and nitrogen (N) budget of the two treatments including C and N in the vegetation and underlying mineral horizon as well as in the litter and organic layers.

Methods: Twelve plots had been established by John Miles at each of two sites (Delnalyne and Craggan) in a paired plot design. In each pair, one plot was left as a heather control plot and the other plot planted with *Betula pubescens*. At the third site (Kerrow) there were 18 plots established in groups of 3, one heather control, one planted with *Betula pendula* and one planted with *Betula pubescens*. The plots were 16×16 m except at Delnalyne where they were 12×12 m. The birch was planted in 1980 at 0.5 m spacing and the experimental area at each site was fenced to exclude grazing by large herbivores. It is important to recognise that these plots have 20- 40 times the usual number of planted seedlings per unit area. Natural regeneration could attain this stocking density but would have a much smaller effect on soil disturbance than planting. Full site descriptions can be found elsewhere (Mitchell *et al.*, 2006).

The sites were sampled in July/August 2005. Two random quadrats, $50 \text{ cm} \times 50 \text{ cm}$, were selected within each plot. For the control heather plots, the total above-ground vegetation was removed by clipping; this was facilitated by installing four vertical bamboo sticks at the corners of the quadrat. For the birch plots, the ground vegetation under the trees was either sparse or absent and was not included; the birch biomass was estimated separately (see below). At all sites, for some of the birch plots, birch growth

was either very patchy or very poor. In the former case, complete random sampling was not possible but restricted to those areas of reasonable establishment (i.e. heather had been ousted). In the latter case (two plots at Kerrow and one plot at Craggan), no samples were taken and the plot was treated as a missing replicate. The underlying reason for this variable birch growth has not been established (Mitchell *et al.*, 2006), but could be related to the mycorrhizal associations. The clumpy distribution suggests that it is associated with isolates introduced with the planting stock rather than the site's own inoculum either at the time of planting or blown in subsequently. This could be an explanation for the site differences described below, and emphasises the need for caution in extrapolating from the results. Three soil cores, in a triangle 15–20 cm apart, were taken from the centre of each quadrat using a 30 cm long, 5 cm × 5 cm square box corer. This sampled the litter (LF) and organic (H or O) horizons and often into the upper mineral horizon (by a few cm). A further mineral horizon sample was taken, if possible, from below the box corer sample using a metal cylinder corer (6 cm diameter × 6 cm depth). Often this was not feasible where large stones or rock were encountered. In a few cases (at Craggan), the litter and organic layer extended below 30 cm; in this case the depth to the underlying mineral horizon or rock was determined with a depthing rod. Care was taken during coring both organic and mineral horizons to facilitate intact cores that could be used for bulk density estimations.

Since the birch was part of an ongoing experiment, it was not possible to harvest trees from the plots in order to measure the tree biomass. However, tree density, tree basal diameter and tree height data were available (R. J. Mitchell, pers. comm.; Mitchell *et al.*, 2006). These had been assessed at the quarter plot level (i.e. the plot divided into four) in order to give some indication of the within plot variability. The sampling quadrats generally fell into different quarter plots but in a few cases were within the same quarter plot. Nine trees from each of the Delnalyne and Craggan sites were harvested at ground level from outside the experimental plots as being representative of the size and shape of those within the plots. Stem and main branches were analysed separately from twigs and leaves for total dry weight and C and N content. Regression equations were developed to enable estimation of the total tree C and N on the plots:

$$\text{Total C (g)} = 5230 - 5212 * (0.9999769^{(D^2*H)}) \quad (r^2 = 0.812)$$

$$\text{Total N (g)} = 71.3 - 70.6 * (0.9999764^{(D^2*H)}) \quad (r^2 = 0.879)$$

Where D is the basal diameter (mm) and H is the height (m).

DBH (diameter at breast height) was not appropriate as many trees were less than this in total. The best fitting relationship was curvi-linear; this is interpreted as being due to the smaller trees having a slightly different form from the larger ones.

In the laboratory, soil cores were divided into LF (litter), organic and mineral horizons. The depth of each horizon was noted and cores weighed for bulk density determinations. These were then air dried for one week and then sub-samples dried overnight at 105°C, ball-milled and analysed for total C and N using a Carlo Erba NA1500 elemental

analyser. Roots encountered in the soil cores were included in the sample for analysis. Mineral samples were first passed through a 2 mm sieve to remove stones. Vegetation samples (heather and birch) were analysed similarly except that they were dried for 24h at 80°C and hammer-milled before being ball-milled. Care was taken to ensure that sub-samples were representative of the whole plant, particularly for the birch where stem/branches and twigs/leaves were analysed separately.

Carbon and nitrogen stocks were calculated for the vegetation, LF (litter), organic horizon and mineral horizon (this being the sum of any mineral horizon sampled with the box corer and with that in the cylinder core), as well as the total. This was done at the core level for the soil horizons but vegetation samples were only available at the quadrat level for the heather and the quarter plot level for the birch; the same mean value was used for each of the cores within the same quadrat. Data was analysed by ANOVA using Genstat 8 (VSN International Ltd, Oxford) with significance being assessed by Fisher's protected least significant difference.

Results: At Kerrow, where two species of birch had been grown, there was no significant difference in total C and N between the *B. pubescens* and the *B. pendula* treatments. Hence, in order to balance the ANOVA for all three sites, only the data for *B. pubescens* was considered further.

All three sites exhibited considerable heterogeneity in soil depth (either depth to underlying stones/rock or depth of organic horizon) and this is reflected in the coefficient of variation for total C (Table 2.1). The site at Craggan was the most variable.

Table 2.1 Coefficient of variation (%) for total C

	Birch	Heather
Delnalyne	27	27
Craggan	46	40
Kerrow	31	26

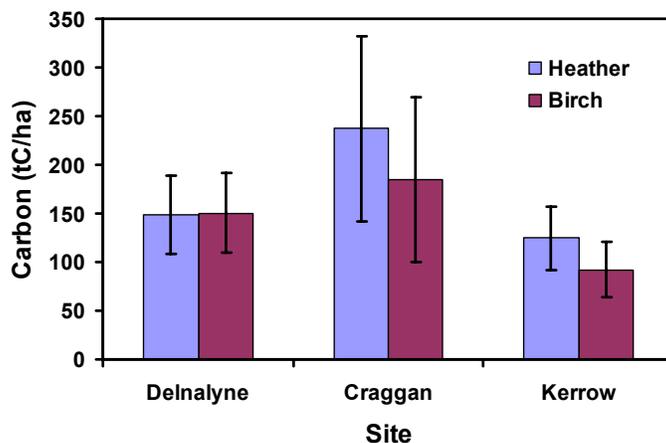


Figure 2.2 Change in total carbon between heather and birch at three sites. Bars show the standard deviation.

The total carbon varied between the three sites in the order Craggan>Delnalyne>Kerrow (Figure 2.2). At Delnalyne there was no significant difference in total C between the heather and birch, while at Craggan there was a 17.7% loss (P=0.003) and at Kerrow a 22.9% loss (P<0.001). Over all three sites there was a 12.3% loss of carbon (P<0.001).

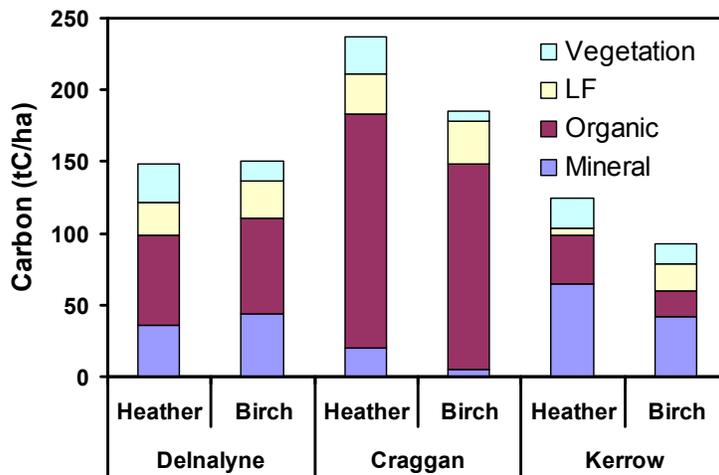


Figure 2.3 Change in total carbon between heather and birch at three sites partitioned between the vegetation and soil horizons.

Looking at the partitioning of the total carbon between the above-ground vegetation and the various soil horizons (Figure 2.3), it was again apparent that the three sites varied. Craggan was dominated by the carbon in the organic (O) horizon, Kerrow was dominated by C in the mineral horizon while Delnalyne was intermediate. It should be noted that the distinction between vegetation and litter for the heather plots was quite difficult to make as the heather was very rank having been ungrazed for 25 years. This caused the stems to fall over and become buried in litter material. Hence the apparently small litter fraction at Kerrow was probably a consequence of greater clipping away of the vegetation-litter complex. No such complication affected the birch plots but for the purpose of statistical analysis the vegetation and litter were combined. Over all sites there was significantly less C in the vegetation plus litter in the birch compared with the heather (16%, P<0.001). This was also true individually for Delnalyne (P=0.012) and Craggan (P<0.001) but Kerrow showed greater in the birch than in the heather (P=0.037).

For the organic horizon, there was an overall loss in the birch (10%) but this was not significant (P=0.088). The greatest loss through birch growth was at Kerrow (46%, P=0.012) but the changes at the other two sites were not significant.

For the mineral horizon, there was again an overall loss in the birch (14%) but this was not significant ($P=0.087$). There was a 47% loss at Craggan ($P=0.025$) and a 28% loss at Kerrow ($P=0.005$). Delnalyne showed an increase but this was not significant.

Table 2.2. Summary of significant changes in moving from heather to birch at the three sites.

	Vegetation + litter	Organic	Mineral
Delnalyne	Loss	No change	No change
Craggan	Loss	No change	Loss
Kerrow	Gain	Loss	Loss

In summary, taking all the sites together, there were C losses from each vegetation and soil component, with about 40% of the loss coming from the organic horizon, and 30% from each of the mineral horizon and litter plus vegetation. However, this varied considerably between the three sites (Table 2.2).

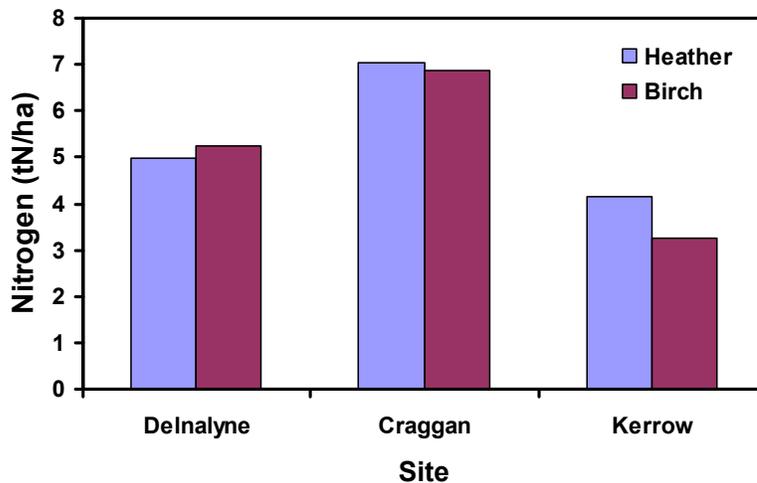


Figure 2.4 Change in total nitrogen between heather and birch at three sites.

When the total nitrogen at the three sites was considered there was no significant differences between the heather and birch at Delnalyne and Craggan but a 22% loss in N ($P<0.001$) at Kerrow. Taking all sites together, there was no significant loss of N (Figure 2.4).

Discussion: The observation of a significant loss in soil carbon following the afforestation of heather moorland is the first of its kind as far as we are aware, apart from the preliminary observation of Mitchell *et al.* (2006). However, such a loss was expected from the results of Mitchell *et al.* (2006) who also recorded significant increases in Truog's phosphorus, mineralizable nitrogen and the decomposition rate of both wooden birch sticks and filter papers, placed *in situ* at the site, in the birch plots as compared with

the heather plots. They also noted a decrease in soil moisture under the birch though this was only observed at the time of sampling.

That part of the total C loss was due to a loss in the vegetation plus litter component was unexpected. However, it should be noted that the birch growth even after 25 years was very variable across the plots and very poor in some areas. It would be expected that the birch biomass would continue to increase with time beyond the 7–14 tC/ha currently seen. Also the heather biomass (ca. 21–27 tC/ha but possibly including some litter) was considerable. However, the heather had been un-grazed for the 25 years of the experiment and had been allowed to become very rank. It is difficult to speculate what would have been the differences had grazing been applied though they might be expected to be less. Equally, it is unknown if the absence of grazing pressure has led to any effects on the below-ground C stocks under either the heather or the birch.

The loss of ca. 20% of the total system carbon at two sites over a 25-year period represents a significant rate of loss, 0.8% per annum, of a similar magnitude to that recorded by Bellamy *et al.* (2005) for all soils in England and Wales (0.6% per annum) though they suggest it may be even greater (2% per annum) in highly organic soils. We have no explanation as to why there was no significant loss at the third site. In most measured properties Delnalyne was intermediate between the other two so we could not point to any particular explanatory factor. In contrast to carbon, nitrogen was generally conserved though one site showed some loss. Since the stocking rates are much higher than those commonly used, the calculated loss rates may not be representative of other sites.

2.2.2 Pinewood site

A secondary objective of the data collection sub-module was to extend the birch findings to a Scots pine system. Unfortunately no parallel experiment on Scots pine with comparative heather moorland plots was set up and none exists elsewhere as far as we are aware. However, field sampling has been completed at Invermoriston, a Forestry Commission experimental site where both birch and Scots pine were planted on prior moorland in 1962. Though no control moorland plots are available, we are fortunate in that the MI has the original soils' descriptions as well as archived samples dating back to the time of planting so that some comparison of changes can potentially be made. Samples have been processed and analysed. Data analysis is in progress.

2.3 Evaluation of model performance

A number of the datasets collected during the project were used for model development and calibration. Testing the model against these datasets does not provide an independent evaluation of the model; for an independent evaluation of the model, external data sets, not used for calibration or development are required. A number of external datasets were used for independent evaluation of the model. The evaluation will be reported in full in forthcoming scientific papers. In this section, we give a brief selection of aspects of the independent model evaluation.

The model has been used to simulate CO₂ emissions from soils treated with litter, nitrogen, litter plus nitrogen and soils alone. Data from laboratory incubations (Foereid *et al.*, 2004) were used and the model tested against these data. Figure 2.5 shows the performance of the model for each treatment.

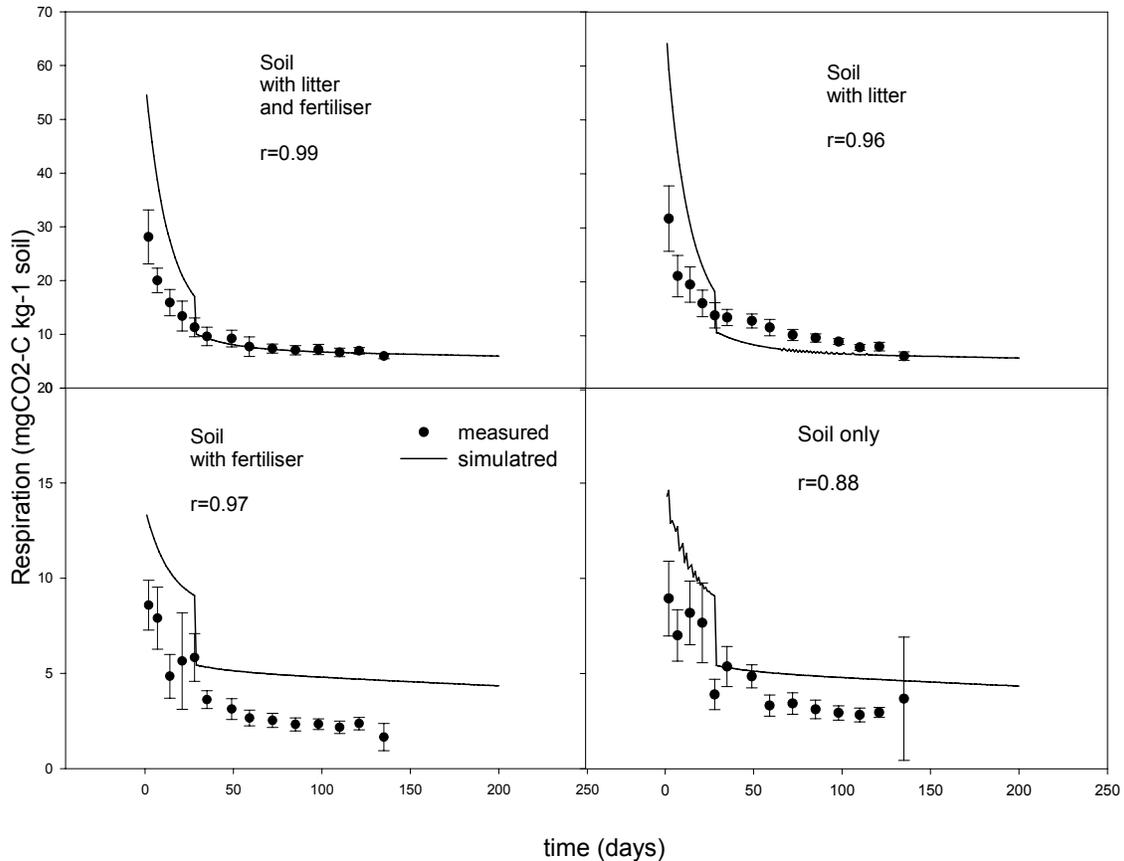


Figure 2.5 Measured (circles with error bars) and simulated CO₂ respiration for laboratory incubated soils either alone, or with additions of litter, nitrogen or litter plus nitrogen (see Foereid *et al.*, 2004 for further details).

The model slightly over-predicts respiration rate, especially early in the incubation, but the rates are comparable with those measured (high correlation, $r > 0.88$ in all cases), especially where litter has been added. The model has also been tested against data from field experiments on organo-mineral soils. Regina *et al.* (2004) published data on N₂O fluxes and soil mineral nitrogen dynamics on a cultivated peat in Southern Finland (60°49'N, 23°30'E) used to grow spring barley. The simulated and measured values of soil ammonium and N₂O emissions are shown in Figure 2.6, a and b, respectively.

The model simulates soil ammonium very well and correctly captures the temporal occurrence of peaks and dips. The model correctly predicts the timing of N₂O peaks, but at this site overestimates the size and duration of these peaks. Reasons for this are under further investigation.

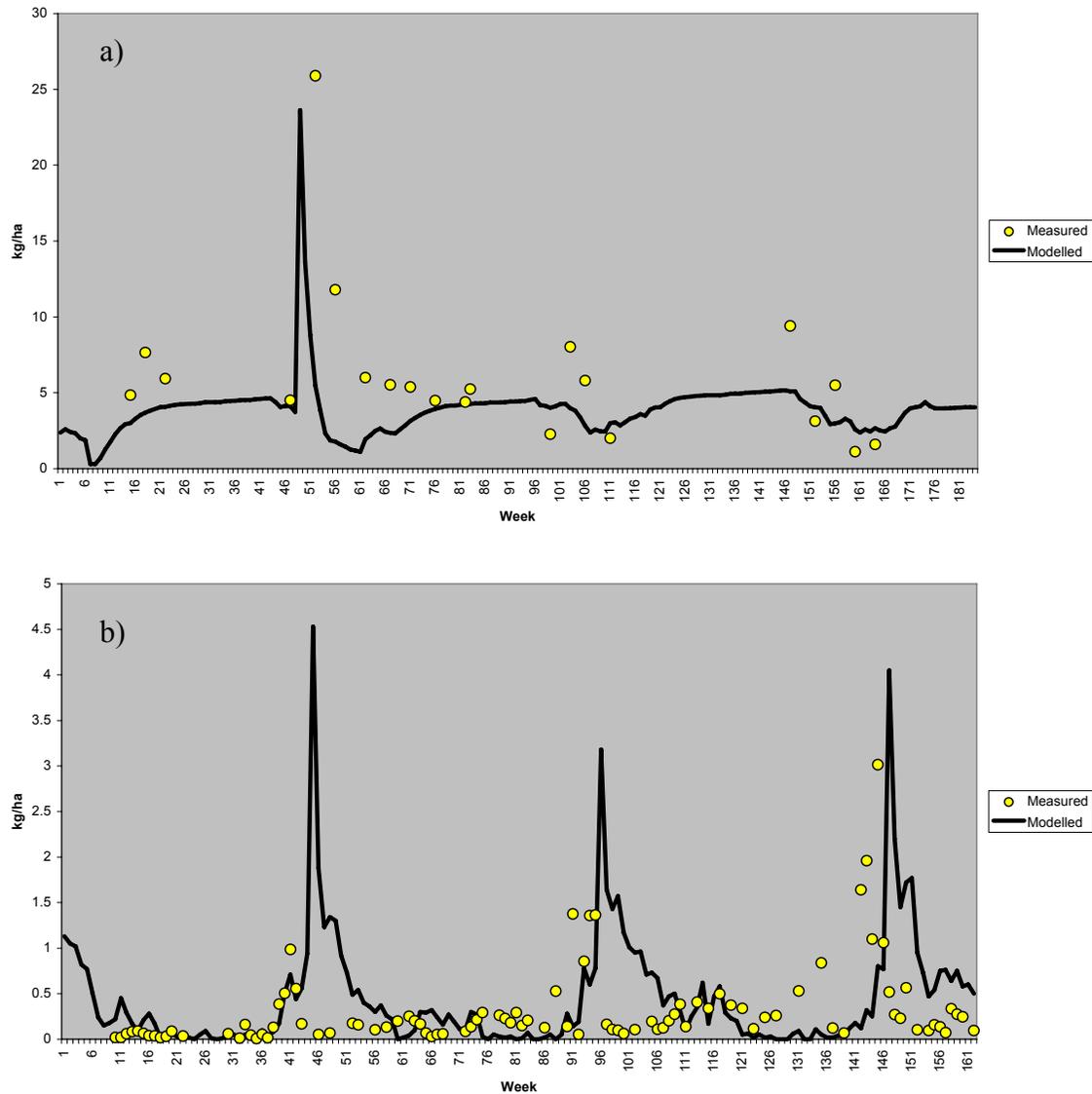


Figure 2.6 Simulated and measured a) soil ammonium and b) N₂O emissions at a cultivated peat in Southern Finland (60°49'N, 23°30'E) used to grow spring barley (see Regina *et al.*, 2004 for further details).

The model has also been tested to evaluate its performance in simulating decomposition and nitrogen dynamics in forest soils, to examine its suitability for simulating land use change following afforestation. Data on mass loss from litter bags (red pine and red maple) and on the nitrogen content of remaining material in the same experiment were simulated with the ECOSSE model. The results are shown in Figure 2.7 a and b.

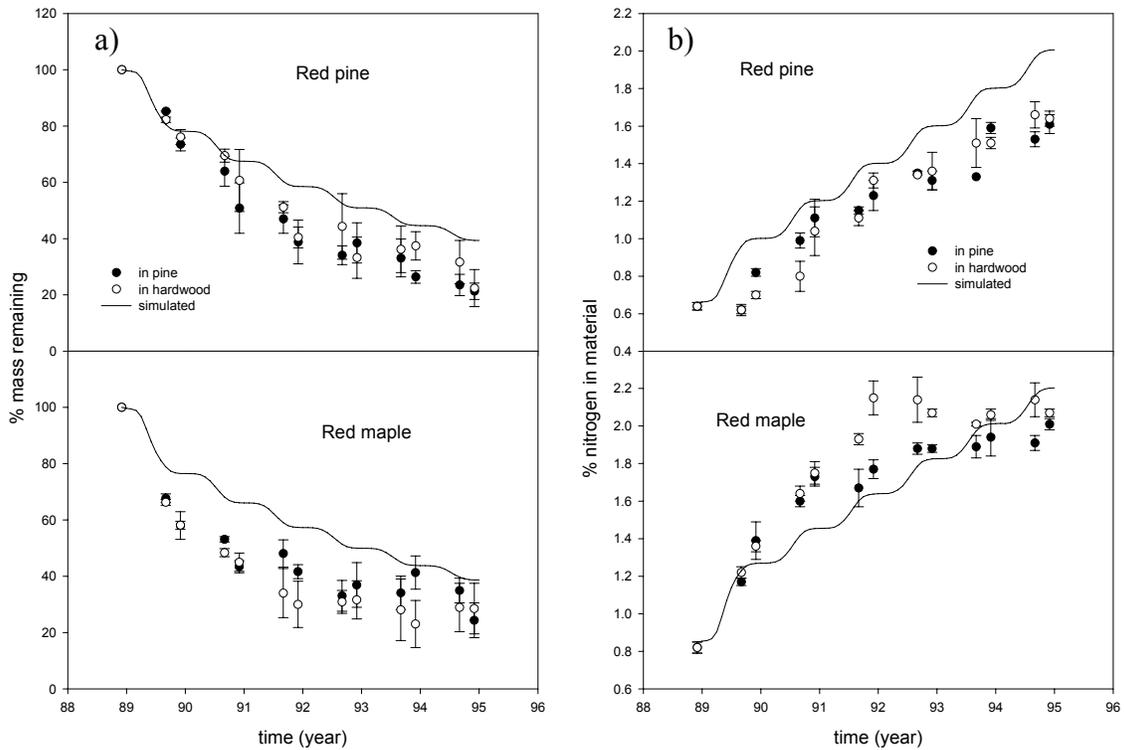


Figure 2.7 Simulated and measured mass loss (a) and N content of the remaining material (b), from litterbag experiments at the Harvard Forest, US. Measured data from Magill and Aber, 1998).

The model predicts trends well but slightly over-predicts remaining mass over time, and slightly over-predicts N in the remaining material for Red Pine.

Other independent evaluations are being analysed whilst others are continuing. These are being prepared for publication and will be submitted for publication during 2007.

Module 3

Land-use and climate change impacts on organic soils and relationships to GHG emissions: new measurements

3.1 Introduction

Study sites: The three sites were chosen to provide typical upland organic soil and land-use conditions but contrasting climates and N deposition rates (Table 3.1). The sites cover the proposed situations of warm & wet with two rates of atmospheric N deposition (Plynlimon - Hafren, Wales), cold and wet, with small rates of N deposition (Ullapool) and cold and dry climate with small N deposition rates (Glensaugh). At Glensaugh and Hafren the sites were the same as chosen for the cores taken for the ¹³C pulse labelling experiment, and first flux measurements were made within one month of the pulse labelling experiment. In addition at the Welsh site we intended to study the influence of N deposition and land use. For this purpose an unimproved land use on a wet peaty gley was chosen at altitude of 380 m asl to be compared with the similar site. Unimproved grassland, at 519 m. These sites should receive different rates of N deposition, due to the more frequent mist development at higher altitude. In addition a small forest plantation in the same valley as the improved and unimproved grassland areas and less than 100 m below the improved grassland site was selected. Deposition rates to a forest can be 2 to 3 fold larger than to shorter vegetation (Fowler *et al.*, 2004).

Table 3.1 The study sites

Sites	Altitude (m asl)	Grid ref (aprox)	N deposition (Kg N/ha/y) ¹	Landuse/ dominant vegetation
<u>Hafren, Plynlimon, N Wales</u>				
Forest	411	SN 824 863	28	SMALL WOODLAND Norway spruce (diameter 14-35 cm). litter layer: 0.5 cm no undergrowth, sheep use this wood as shelter
Improved peaty gley	457	SN819 857	15	SHEEP GRAZED GRASSLAND
Unimproved peaty gley	381	SN 841 882	15	WET, ACID GRASSLAND
Unimproved peaty gley	518	SN 819 863	33	UPLAND GRASSLAND on a very steep slope
Unimproved peat	570	SN 825 874	33	WET BLANKET PEAT Deergrass, calluna, some pockets of sphagnum and Erica
<u>Glensaugh, NE Scotland</u>				
Deep peat	420	NO 661 812	16.5	BLANKET PEAT on Thorter Hill, Eriopherum (dominant), sphagnum, calluna, etc.
Unimproved peaty podzol	360	NO 658 807	16.5	ACID GRASSLAND grazed by cattle & sheep
Improved peaty podzol	330	NO 658 807	16.5	FERTILISED GRASSLAND grazed by cattle & sheep
<u>Ullapool, NW Scotland</u>				
Deep peat	100 ~	NC 073 121	4.2	DEEP BLANKET PEAT Calluna, bog aspedel, spagnum, Eriopherum vaginatum etc
Unimproved peaty gley	100 ~	NC 073 119	4.2	ACID GRASSLAND grazed by cattle & sheep
Improved peaty gley	100 ~	NC 072 119	4.2	FERTILISED GRASSLAND grazed by cattle & sheep

¹ N deposition was estimated from the CEH N deposition map and the presence of trees (Fowler *et al.*, 2004)

3.2 Measurements

Flux chambers were installed in summer/autumn 2004 and remained in position until autumn 2005. Fluxes of N₂O, CH₄ and soil respiration were measured on 7 occasions (6 at one site: Jun/Jul 04 (not Ullapool), Oct/Nov 04, Jan 05, Apr 05, Jun/Jul 05, Aug/Sep 05, Nov 05). Each time air samples were taken from the static chambers, which involved closing the chambers with a plastic dome shaped lid and taking samples after one-hour incubation. Samples were collected by syringe and transferred to Tedlar bags. Samples were analysed at CEH Edinburgh for CH₄, and N₂O by gas chromatography using a flame ionisation detector for the analysis of CH₄ and an electron capture detector for N₂O (Skiba *et al.*, 2006). Soil respiration was measured in situ using a PP Systems EGM soil respirometer, which was placed on plots where the vegetation was cut back and on soil devoid of vegetation. These latter plots were deliberately kept bare by weeding after each measurement as preparation for the subsequent measurements a few months later. Regrowth during this period was minimal. In addition respiration from vegetated plots were measured, but not included in the data analysis. At each visit the soil temperature (at 10 cm depth) and soil surface temperature is measured. Soil cores (0-20 cm depth) were taken for subsequent analysis of soil moisture content and KCl extractable NH₄⁺ and NO₃⁻ concentrations (n=3). On one occasion soil samples were taken at all sites for measurements of bulk density (n = 5), soil pH and total carbon and total nitrogen content (three sub samples from 5 soil cores 0-20 cm mixed well). Species composition in each chamber was recorded at every site visit.

3.3 Results

3.3.1 Soil properties

Soil physical and chemical properties are summarised in Table 3.2 & 3.3. All parameters were in the expected range reported elsewhere for organic soils. The bulk density of the peat ranged from 0.08 to 0.14 g cm⁻³ and did not vary significantly between geographical location, and at Ullapool also not between sites. At Glensaugh and Hafren the peat bulk density was significantly smaller than for the peaty gleys and podzols. Largest bulk densities were measured for the Hafren peaty gleys (> 0.4 g cm⁻³). Total N, pH and bulk density increased over the gradient peat to unimproved peaty gley/podzol to improved peaty gley/ podzol. Soil C content however, decreased along this gradient at Hafren and Glensaugh, and did not change at Ullapool. Soil moisture contents were high throughout the study period, at Glensaugh and Ullapool the peat was always wetter than the peaty soils, however such difference was not observed at Hafren. Soil temperatures ranged from 3 – 15, 1 – 13 and 4 – 14 °C at Ullapool, Glensaugh and Hafren, respectively, air temperatures are shown in Table 3.2.

The KCl extractable NH₄⁺ concentrations ranged from 23 to 37 µg N g⁻¹soil in Scotland, but were smaller in Wales (11 – 26 µg N g⁻¹) and correlated with the soil extractable NH₄⁺ fraction (r²=0.75) (Table 3.3 and Figure 3.1). Nitrate concentrations were below the detectable range of the analyser for all sites, suggesting very small, if any, nitrification rates.

Table 3.2 Soil properties and air temperature at the study sites, n = 3 for total C &N and pH, n=5 for BD, n=3 replica on 7 dates for temperature and moisture

	Total nitrogen		Total carbon		pH		Bulk density		Air temperature		Soil moisture	
	%		%		in 0.01 M CaCl ₂		g/cm ³		°C		g/cm ³	
	average	stdev	average	stdev	average	stdev	average	stdev	average	stdev	average	stdev
Ullapool												
peat	1.31	0.03	50.77	0.12	3.30	0.08	0.12	0.03	7.95	3.92	0.98	0.25
unimproved peaty gley	1.55	0.04	51.20	0.10	3.29	0.10	0.15	0.01	9.18	4.77	0.77	0.10
improved peaty gley	2.21	0.02	51.27	0.15	3.80	0.08	0.14	0.01	8.34	4.99	0.66	0.09
Glensaugh												
peat	1.76	0.02	53.20	0.10	2.80	0.02	0.14	0.02	10.10	8.59	0.88	0.06
unimproved peaty podzol	1.98	0.01	45.87	0.06	3.21	0.14	0.26	0.06	8.94	3.76	0.43	0.11
improved peaty podzol	2.35	0.01	44.40	0.10	4.43	0.54	0.26	0.03	8.90	3.91	0.39	0.06
Hafren												
peat	1.64	0.05	50.70	0.10	3.01	0.06	0.08	0.01	9.50	3.50	0.68	0.04
unimproved peaty gley	1.69	0.03	23.27	0.06	3.50	0.02	0.42	0.03	10.50	3.73	0.57	0.14
improved peaty gley	1.15	0.02	17.20	0.10	4.26	0.05	0.45	0.11	9.60	3.33	0.66	0.20
low peaty gley	2.58	0.01	36.23	0.15	3.56	0.36	0.20	0.04	12.18	5.52	0.59	0.13
wood	2.45	0.03	44.23	0.12	3.05	0.11	0.22	0.04	8.53	3.90	0.86	0.07

Table 3.3 Available soil NH₄⁺ and greenhouse gas fluxes. Average of 3 replica measurements on 7 dates

	KCl extractable NH ₄		Soil respiration Bare soil		Soil & plant respiration Vegetated soil		N ₂ O flux			CH ₄ flux		
	µgN/gsoil		µmolesCO ₂ /m ² /s		µmolesCO ₂ /m ² /s		µgN ₂ O-N/m ² /h			µgCH ₄ /m ² /h		
	average	stdev	average	stdev	average	stdev	average	stdev	median ¹	average	stdev	median ¹
Ullapool												
peat	30.06	8.14	0.90	0.80	2.42	1.63	0.31	6.54	1.25	1053.72	1511.43	467.06
unimproved peaty gley	24.56	9.69	0.37	0.38	1.81	1.23	-0.45	1.65	-0.48	75.01	166.12	0.46
improved peaty gley	37.39	15.65	0.47	0.19	2.02	1.02	-5.50	9.63	-0.05	43.87	53.12	17.29
Glensaugh												
peat	33.73	9.30	0.93	0.70	1.82	1.44	-1.88	9.89	0.78	448.35	474.59	378.45
unimproved peaty podzol	23.08	13.66	0.96	0.37	2.27	1.11	-7.46	33.62	3.61	-23.60	29.67	-14.77
improved peaty podzol	32.37	12.11	2.17	1.61	4.53	2.48	5.83	25.65	2.60	-35.48	87.96	-12.44
Hafren												
peat	23.90	8.47	0.74	0.38	1.83	1.10	-19.31	47.49	-1.36	-8.22	261.84	14.60
unimproved peaty gley	14.63	2.43	0.36	0.12	2.23	0.42	-7.56	21.37	-0.01	-113.76	229.67	-4.10
improved peaty gley	11.44	3.12	0.57	0.71	3.32	2.54	-7.68	20.37	-0.39	-63.05	261.62	0.10
low peat	25.75	5.57	1.78	1.10	4.76	1.68	6.78	19.60	3.51	-87.30	154.87	1.91
wood	21.11	3.33	1.98	1.18	1.98	1.18	32.00	48.55	17.90	-70.82	280.05	3.98

¹ Median values are reported for N₂O and CH₄ fluxes, as these are very different from the average values

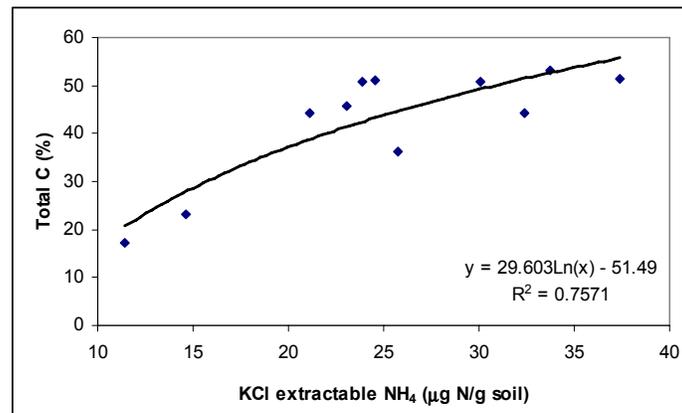


Figure 3.1 Relationship between soil available NH₄ and total C content in organic soils

3.3.2 Soil respiration, CH₄ and N₂O fluxes

Soil CO₂ (respiration), methane and nitrous oxide fluxes are shown for each site in Table 3.3. Temporal and spatial variations were typically large for all gases measured, as shown here for the CH₄ fluxes from peat at the three study sites and the N₂O fluxes from the fertilised improved grassland soils (Figure 3.2).

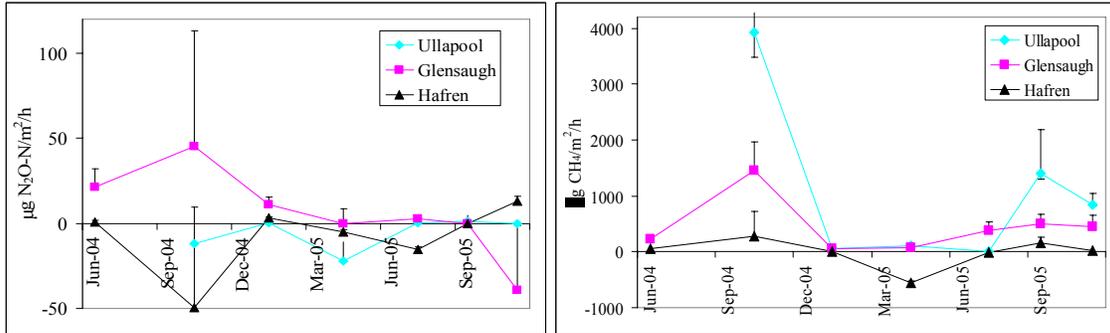


Figure 3.2 Temporal variations of N₂O fluxes from the improved peaty podzol/gleys (left graph) and of CH₄ from unimproved peat (right graph). Data are averages from 4 chambers and standard error of the mean. The SE is occasionally too small to be visible.

Nitrous oxide fluxes: All sites showed periods of N₂O emission and uptake (Table 3.3, Figure 3.2). Largest and most frequent N₂O emissions were measured from the improved peaty podzol at Glensaugh, the wood in Hafren and the unimproved peaty gley at the foot of the hills at Hafren. Uptake of N₂O occurred most often at Hafren on the peat, unimproved peaty gley, improved peaty gley. Contrary the peat at Ullapool emitted N₂O when the peaty gleys here absorbed N₂O. It was not possible to explain the switch from N₂O emission to N₂O uptake on hand of the variables measured here.

The large spatial and temporal variations in N₂O flux made it very difficult to provide correlations that would be useful to the ECOSSE model for either, the entire data set, or subsets of measurements at the same site, or for the same soil type. Average N₂O fluxes and log transformations of the fluxes only correlated well with total N and soil respiration (Figure 3.3).

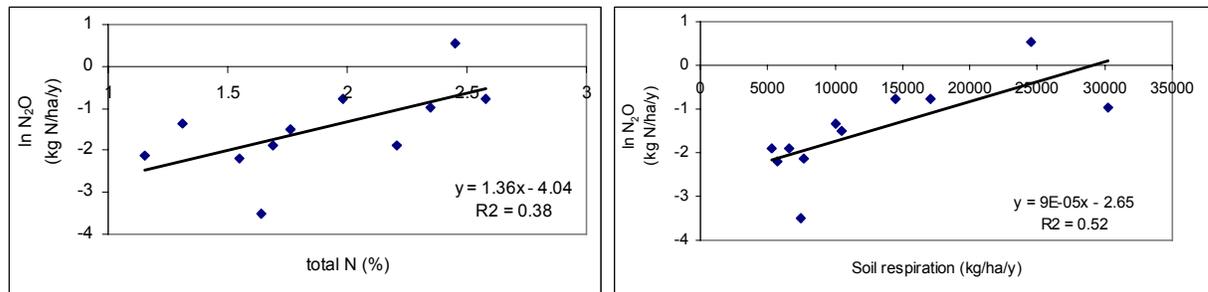


Figure 3.3 Relationship of N₂O flux with total soil N (left graph) and soil respiration (right graph). Data are averages of 4 measurements on 7 occasions. In order to transform the negative N₂O fluxes, all data were increased to positive values by adding 0.15.

The range of fluxes measured in this study is similar to those measured from organic soils and similar landuse elsewhere in Britain by CEH. The 62 measurements (average values various time periods) to date (including this study) suggest an average N₂O emission of 0.25 kg N₂O-N/ha/y, ranging from -0.99 to 3.7 kg N₂O-N/ha/y (n= 62).

N deposition and nitrous oxide fluxes: The two ‘extra’ sites at Hafren (unimproved peaty gley at the bottom of the hill and the forest) were chosen to establish a relationship between atmospheric N deposition and N₂O emission. The forest site is adjacent to the improved grassland, both are on a slope but differ in N deposition rate, due to the scavenging effect of the forest (Table 3.1). Forest total C&N, soil available NH₄, N₂O and soil respiration rates are all larger compared to the adjacent improved grassland. Due to large spatial and temporal variability these differences are not significant. It is not possible to blame the enhanced deposition rate to the forest entirely for these increases in N & C accumulation and fluxes. The wood is the only shelter for sheep on this exposed hill site, and evidence of livestock excreta as a source of N input is plentiful.

Comparison of the unimproved peaty gleys in a hollow at the bottom of the hills, surrounded by forests, with the unimproved peaty gley on a steep slope at higher altitude, where N deposition rates are larger due to enhanced occult deposition, provided not the expected results. Total N, available N, soil respiration and N₂O fluxes were larger at the low N deposition site. The explanation for this is possible runoff of N from the above forests.

Soil respiration: Soil respiration rates were roughly of the same order at all three locations, however 4 times smaller than average annual soil respiration rates measured from improved, but unfertilised grassland on mineral soil in SE Scotland (Jones *et al.*, 2006). There were not always four replicates per site, due to problems with the analyser and availability of bare, vegetation free patches of peat needed to make the measurements. Soil respiration rates correlated significantly with the N₂O flux (Figure 3.3) using average data for each site and also with total C& N content and the bulk density of the soil (Figure 3.4).

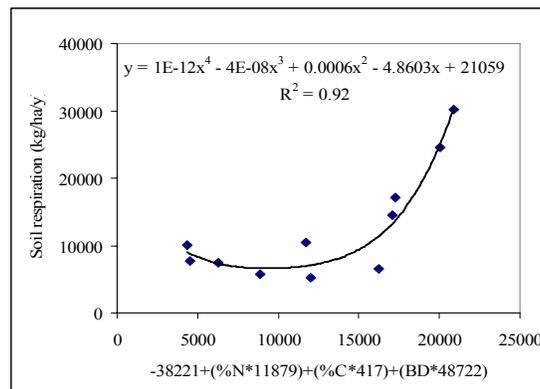


Figure 3.4 The dependence of soil respiration on soil N,C and bulk density, calculated by multivariate regression analysis. The data were fitted by polynomial equation shown above the graph. Data points are averages of 4 measurements .

At individual sites, good correlations between soil and plant respiration and soil temperature and soil moisture was observed at Glensough ($\ln \text{CO}_2 \text{ (plant\&soil)} = -0.61 + (\text{soil Temp} * 0.061) - (\ln \text{soil moisture} * 0.8)$, $r^2 = 0.58$). At Hafren a weak relationship between (the natural log of) soil respiration and soil available NH_4^+ was found. For the entire study soil respiration rates were 14.89 t/ha/y (range 0 to 64.4) for bare soil and 37.18 t/ha/y (range 7 to 24.9). These respiration rates were smaller by a factor of 2 compared to those measured from bare patches of mineral soil on unfertilised and fertilised grassland (Jones *et al.*, 2006).

Methane: Average CH_4 fluxes for each site showed that Ullapool was a net source for CH_4 , but Hafren was a net sink and at Glensough the peat was a source and the peaty podzols a sink (Table 3.3). Fluxes were very variable and frequently changed between uptake and emission. For the Hafren soils this variability meant that the above general conclusion very much depends on calculating median or averages of the 7 measurement dates (Table 3.3, Figure 3.5). The 7 measurements are averages of flux measurements from 4 chambers, which fortunately were similar to median values.

The woodland at Hafren was a sink for CH_4 only in June 2005 ($-624 \mu\text{g CH}_4 \text{ m}^{-2} \text{ h}^{-1}$), a small source ($+2.7$ to $10.8 \mu\text{g CH}_4 \text{ m}^{-2} \text{ h}^{-1}$) in January, March, Sept and Nov 2005, and a substantial source in July and Sept 2004. The September 2004 data are a bit dubious, the 4 chamber fluxes are all positive, but some are unusually large (22060, 9820, 2591 and $63 \mu\text{g CH}_4 \text{ m}^{-2} \text{ h}^{-1}$), and therefore are removed from the general interpretation of the summary data (Figure 3.5). However these data show that under certain conditions forests can be a source rather than a sink of CH_4 .

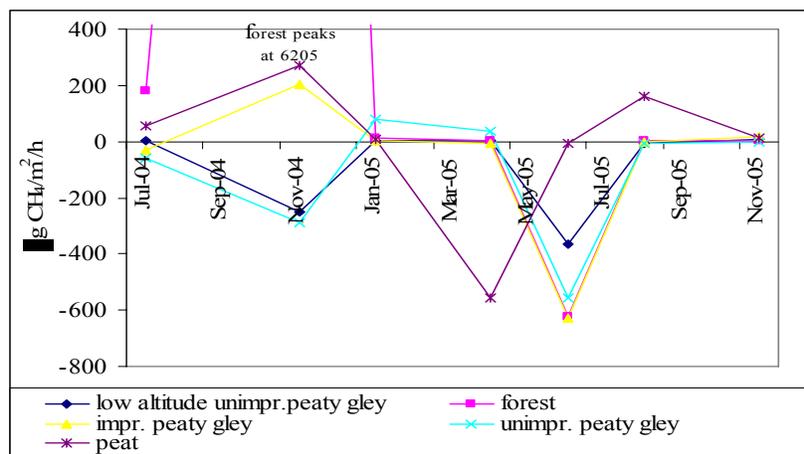


Figure 3.5 Temporal variability in CH_4 flux at the Hafren sites

Methane fluxes depend on soil wetness. Soil wetness was expressed here as water filled pores space, which was calculated from the volumetric moisture content (the product of gravimetric moisture content, measured every time fluxes were measured, and the bulk density, measured once at each site) and the porosity. Water filled pore space (WFPS) was calculated by dividing the volumetric water content by total soil porosity. Total soil

porosity was calculated according to the relationship: soil porosity = $(1 - \text{soil bulk density}/1.4)$, assuming a particle density for peat of 1.4 g cm^{-3} (Rowell, 1994). It is interesting that at a WFPS < 50% CH₄ flux was very small and that at WFPS > 50% both CH₄ uptake and emission occurred (Figure 3.6). Maximum uptake occurred at the WFPS range of 70-90 %, and maximum emission at the range 90-100%. These differences can partly be explained by differences in soil available NH₄ concentration. Inclusion of these in a multi-regression equation provided a good fit of the CH₄ data (Figure 3.7).

This is in agreement with the relationship of soil available NH₄ and total C content (Figure 3.1). The larger the C content, the wetter the soils, the smaller the chance for nitrification, but the larger the opportunity for CH₄ oxidation to occur.

The magnitude of the CH₄ fluxes reported here reflects the range of those measured by CEH in previous studies. Combined average CH₄ fluxes from organic soils are: 3.9 kg CH₄/ha/y (range -3.95 to 62.4, n = 44).

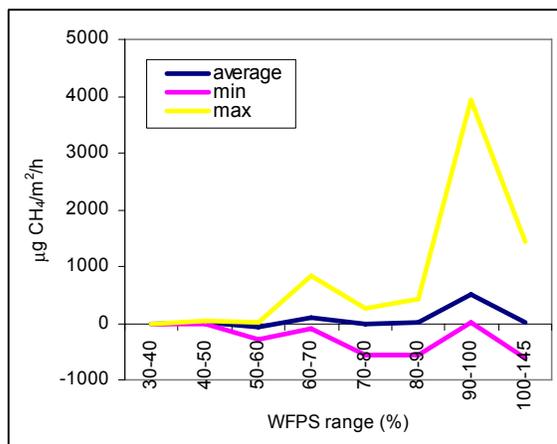


Figure 3.6 Average, minimum and maximum CH₄ concentrations at various ranges of WFPS. The data are calculated from average fluxes from 4 chambers measured on 6/7 dates at all sites.

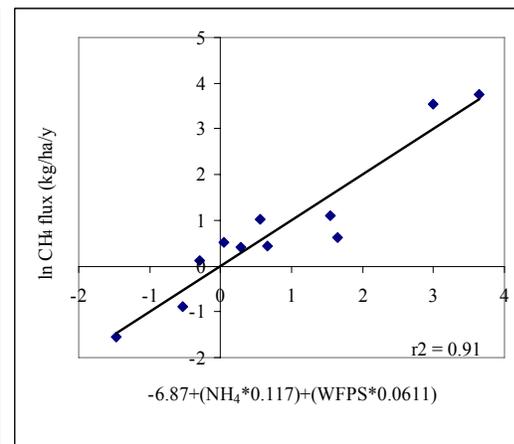


Figure 3.7 The dependence of the (the natural log of) CH₄ flux on soil available NH₄ concentration and water filled pore space of the soil. Data are averages of 4 measurements. In order to transform the negative CH₄ fluxes, all data were increased to positive values by adding 1.5.

3.4 Summary

- 1) Typical large spatial and temporal variability was observed for all fluxes measured.
- 2) Methane and N₂O fluxes were bi-directional, fluxes were of the same order of magnitude as observed in previous studies.
- 3) Soil respiration rates were smaller than those measured from bare patches on mineral unfertilised and fertilised grasslands soils. Respiration rates measured from vegetated plots were larger than from bare plots, it is therefore important to exactly report the measurement surface to modellers so that this can be accounted for when using the results for calibration or model testing.

- 4) Relationships of fluxes with measured parameters were better for the averages of the entire dataset than for individual sites or dates.
- 5) The influence of N deposition on N₂O fluxes was not clear, due to possible overriding influence of other variables (livestock and runoff).
- 6) N₂O fluxes from organic soils correlated with total N content in soil and soil respiration rates.
- 7) Soil respiration could be calculated from the total soil C and N content and the bulk density of the soil.
- 8) Methane fluxes were depended on the WFPS and could be calculated from the soil available NH₄ concentration and the WFPS.

3.5 ¹³CO₂ pulse labelling

Methods: ¹³CO₂ pulse labelling of cores was carried out at CEH Lancaster to ensure sufficient enrichment of photoassimilate (carbon fixed from the atmosphere by the plant by photosynthesis) to produce a strong ¹³C rhizosphere C flow pulse into the soil and enable chasing of the pulse to characterise turnover. After pulse labelling, sampling of the cores was carried out at 0, 3, 16h, 6d, 1 month and 3 months to determine ¹³C fate in vegetation, soil pools and microbial biomass All ¹³C analyses were carried out using an Infra-Red Mass Spectrometer (IRMS).

Results: Plynlimon data only are presented here as similar patterns were found at both sites. When vegetation on 3 soil types was pulse labelled with ¹³CO₂, a weak ¹³C signal could be detected in the soil within a few hours of the pulse, and this was detectable up to three months (the last sampling point). This agrees with earlier data from pulse labelling at Sourhope on the NERC Soil Biodiversity programme (Rangel-Castro *et al.*, 2004; Rangel-Castro *et al.*, 2005a,b). Only the improved podzol, with its high above ground biomass and its associated high photosynthetic activity, showed a marked peak in ¹³C in the soil (Figure 3.8).

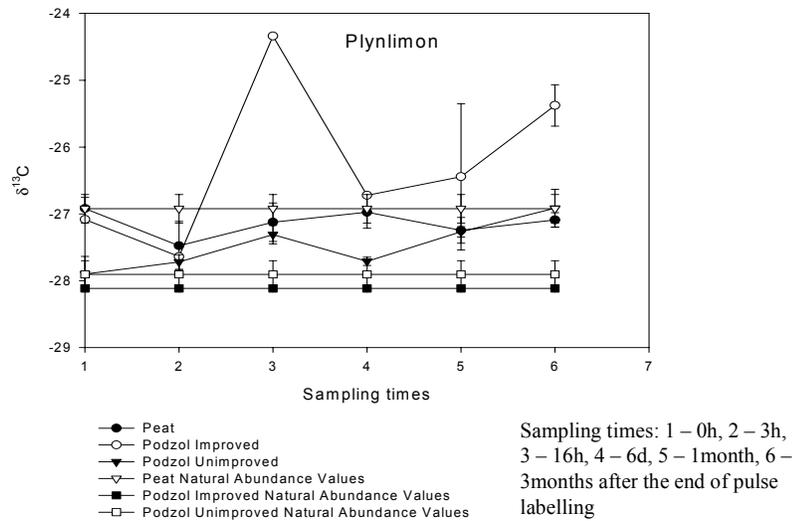


Figure 3.8 Soil ¹³C levels for Plynlimon

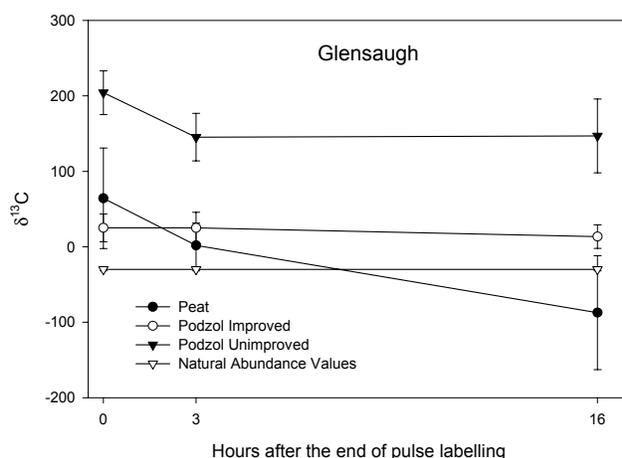


Figure 3.9 ^{13}C soil microbial biomass levels for Glensaugh

The greatest enrichment was seen for ^{13}C soil microbial biomass at Glensaugh for the podzols with unimproved treatments (Figure 3.9). This may be due to the slower turnover of ^{13}C in the microbial biomass of the unimproved versus the improved podzol, rather than greater assimilation of ^{13}C . The peat microbial ^{13}C showed the low levels typical of such low productivity systems.

It is interesting to note that the microbial biomass demonstrated very rapid assimilation of ^{13}C . This is a strong aspect of this study as it has often been missed by other studies because the sampling has been too slow to catch the very rapid pulse of ^{13}C from the plant roots, primarily through exudation. Stable isotope probing of the nucleic acid pools of the Glensaugh microbial biomass is currently underway and will be published during 2007.

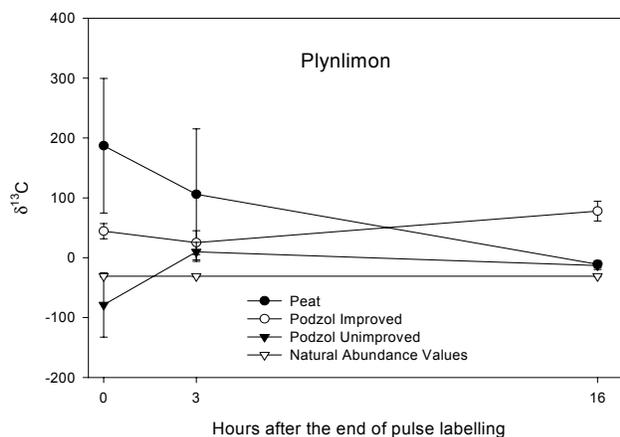


Figure 3.10 ^{13}C soil microbial biomass levels for Plynlimon

In contrast to Glensaugh, higher enrichments of ^{13}C in the soil microbial biomass were associated with the peat at Plynlimon (Figure 3.10). This is a surprising result as one would expect greater supply of ^{13}C from the vegetation associated with the podzols. It

could be, however, that the relatively high substrate assimilation efficiency of an almost entirely fungal population (often around 40-50% compared to around 20% for many bacteria) explains the greater enrichment in the peat. Further microbial investigation, based upon selective probes of DGGE (Denaturant gradient gel electrophoresis) also revealed a high actinomycete population in the peat and this could also be a factor in the high ^{13}C enrichment compared to the podzols at Plynlimon.

Mean residence time calculations for shoots (Table 3.4) revealed a broadly similar turnover across the different soil/vegetation systems, with a mean residence time of between one and two months.

Table 3.4 Mean residence times – shoots

Comparative turnover (Tmrt) of plant fixed ^{13}C at Glensaugh vs Plynlimon		
Site	Soil	Mean residence time (y)
Glensaugh	Peat	0.14
	Podzol Improved	0.16
	Podzol Unimproved	0.16
Plynlimon	Peat	0.13
	Podzol Improved	0.15
	Podzol Unimproved	0.11

Table 3.5 Mean residence times - soils

Comparative turnover (Tmrt) of plant fixed ^{13}C at Glensaugh vs Plynlimon		
Site	Soil	Mean residence time (y)
Glensaugh	Peat	0.80
	Podzol Improved	0.21
Plynlimon	Peat	2.23

* ^{13}C turnover could not be detected for other soil types at Plynlimon.

** Data are preliminary due to level of replication for natural abundance determination.

Mean residence times for the plant fixed ^{13}C pulse in soils highlighted the very slow turnover in peat compared to the improved podzol (Table 3.5). The data in this matrix are not yet complete due to problems in obtaining reliable natural abundance data for some systems.

The final strand of the research in this work package involved using a combined stable isotope and molecular approach (stable isotope probing of the nucleic acid and fatty acid pools of the soil microbial communities of the study sites) to identify the key players in terms of carbon processing in the soil. In the two figures (stable isotope probing of denaturation gradient gel electrophoresis) below, the incorporation of ^{13}C into soil actinomycete RNA from pulse labelling of the vegetation sward on the improved and unimproved podzols at Glensaugh is illustrated as an example. The figures firstly demonstrate the complexity of the active microbial community (or, in this case, a sub-set of it) in terms of carbon processing. They also illustrate, because of the changing band profile at time 0 and time 3h, that there is a broader community response to the low

molecular weight carbon compounds (dominated by glucose and simple organic acids and amino acids) first ^{13}C labelled than there is with time when there is activation of a more restricted, specialised microbial community degrading the higher molecular weight, labelled carbon. In terms of a comparison between unimproved and improved podzols, the figures show that the active soil community is less complex in the unimproved compared to the improved podzol. This agrees with the findings of Rangel-Castro *et al.* (2005a, b) for soil bacterial and fungal communities at a similar site in the Scottish borders. Furthermore, the active microbial community at time 0 is simpler in the unimproved compared to the improved podzol.

These findings of the nature of the active microbial communities processing carbon in upland, organic soils using state of the art ^{13}C stable isotope probing are exciting and new and are currently being more fully evaluated through confirmatory application of ^{13}C phospholipid fatty acid probing.

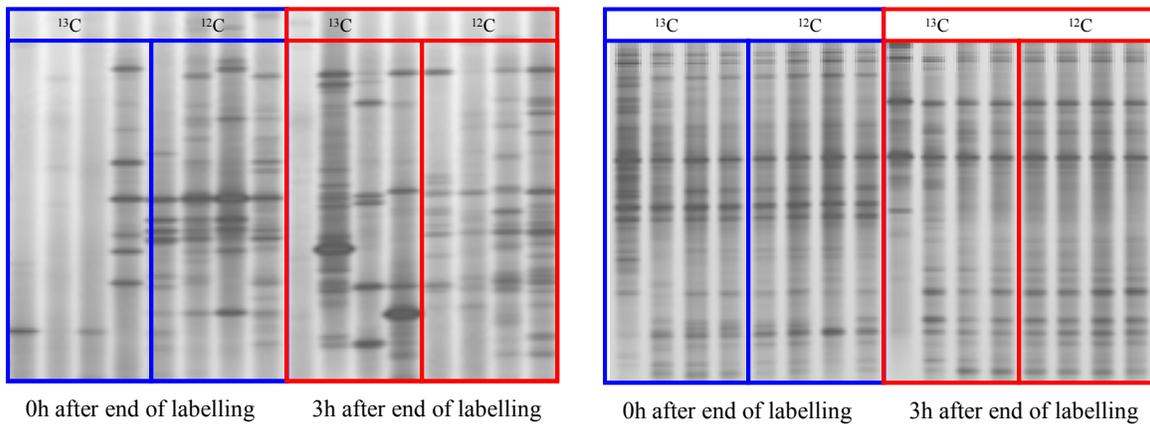


Figure 3.11 SIP-DGGE of actinomycete community analysis – podzols at Glensauagh, unimproved (left) and improved (right)

Conclusions: The ^{13}C data presented in this report are particularly exciting and reveal a great deal about C cycling and turnover in upland acid systems for the first time. They highlight the rapid capture of the plant derived ^{13}C in the soil microbial biomass and the slow turnover of this material in the acid soils of this study, particularly in the blanket peats. The ^{13}C SIP data reveal the active components of the soil microbial community in terms of carbon processing, and highlight that a greater proportion of the community is involved in processing of simpler C compared to more complex C and how the complexity of the active community changes under soil improvement. This shows that carbon is captured quickly in upland systems, but that when in the soil, carbon can take a long time to turn over. These data will be used to calibrate the processes of C capture and soil C turnover in versions of the model that have an explicit representation of plant growth, which are scheduled for development in future projects.

Module 4

The effects of land-use change and climate change on the release of Dissolved Organic Matter from organic soils

4.1 Introduction

4.1.1. Background

Dissolved organic carbon (DOC) is the carbon contained within organic matter in solution, and represents a potentially important pathway for carbon loss from organic soils. Previous studies have shown that riverine DOC fluxes are closely related to the extent of peat within the catchment (e.g. Hope *et al.*, 1997). There is growing evidence that DOC concentrations in UK rivers have increased in recent decades (Freeman *et al.*, 2001a; Worrall *et al.*, 2004; Evans *et al.*, 2005), and that rates of increase are greatest in the most organic-rich catchments (Freeman *et al.*, 2001a). The recent paper by Bellamy *et al.* (2005) suggests that these same organic soils have experienced a loss of carbon over the same period. This desk study provides a review of existing literature, supported by some unpublished CEH data, to assess 1) the significance and sources of the DOC flux term within the overall C budget of organic soils; 2) the external environmental drivers potentially responsible for observed increases in DOC; and 3) the extent to which land management may accelerate or ameliorate rates of DOC loss.

4.1.2. The importance of DOC fluxes in the C budget

Many studies of the carbon balance of organic soils focus on the exchange of C between the atmosphere and the land surface. Net ecosystem CO₂ exchange represents the balance between two large gross fluxes (primary productivity and ecosystem respiration), and may be positive, negative or zero. Riverine DOC fluxes are typically one to two orders of magnitude smaller than gross land-atmosphere C fluxes (Hope *et al.*, 1997), and have therefore been neglected in some studies of ecosystem carbon balance. However, DOC fluxes are unidirectional (i.e. always represent a C loss), and comparisons between riverine DOC export and net ecosystem CO₂ exchange, both globally (Hope *et al.*, 1994; Schlesinger, 1997) and in the UK (Cannell *et al.*, 1999) suggest that they are actually of similar magnitude. A peatland catchment carbon flux study by Worrall *et al.* (2003a) recorded a DOC flux at Moor House, Northern England, of 94 g C/ha/yr, compared to a (literature-based) estimated net CO₂ exchange of 400-700 g C/m²/yr. For Auchencorth Moss in Scotland, Billett *et al.* (2004) measured a DOC flux of 283 g C/m²/yr, and a net CO₂ exchange of 278 g C/m²/yr. At Auchencorth Moss, the inclusion of the DOC (and other riverine) loss terms was sufficient to turn the catchment from an apparent C sink to a substantial (83 g C/m²/yr) C source. In general, it is clear that the omission of DOC from studies of carbon in organic soil ecosystems will lead, at best, to substantial errors in the estimate, and at worst to the misattribution of systems that are either stable, or losing carbon, as carbon sinks.

4.1.3. Sources of DOC

DOC may derive from plant litter, root exudates, microbial biomass or soil organic matter. Identifying the origin of (rising) DOC losses from organic soils is important in

determining whether they are indicative of the destabilisation of ‘old’ soil carbon (as suggested, for example, by the results of Bellamy *et al.*, 2005), or of an accelerated throughput of ‘new’ carbon which has been assimilated relatively recently from the atmosphere. Measurements of radiocarbon, ^{14}C , provide a means to estimate DOC age. Previous studies show mixed results; some suggest a predominantly recent origin for DOC (Palmer *et al.*, 2001; Benner *et al.*, 2004), others a substantial older component (Raymond and Bauer, 2001), or a shift from older to newer material with increasing flow (Schiff *et al.*, 1997). In a study linked to the current project, DOC^{14}C was measured in streams and soil solutions draining a range of soil and land-use types in North Wales. Results suggest that most of the DOC exported from organic soils is ‘new’ (post-1950s), especially at high flow, when the majority of DOC flux export occurs (Figure 4.1). These data are difficult to reconcile with possible large-scale peatland destabilisation (e.g. Freeman *et al.*, 2001a,b; Bellamy *et al.*, 2005), at least if riverine DOC represents a significant loss pathway for soil C. Instead, they suggest that rising DOC in surface waters must be explained by either 1) rising production of ‘new’ DOC from recent plant material, or 2) decreasing retention of this DOC within the soil profile. It is worth noting that ^{14}C data for mineral (upland and improved grassland) soils do suggest an older soil origin, and may therefore be more consistent with suggested long-term soil organic matter depletion from these systems.

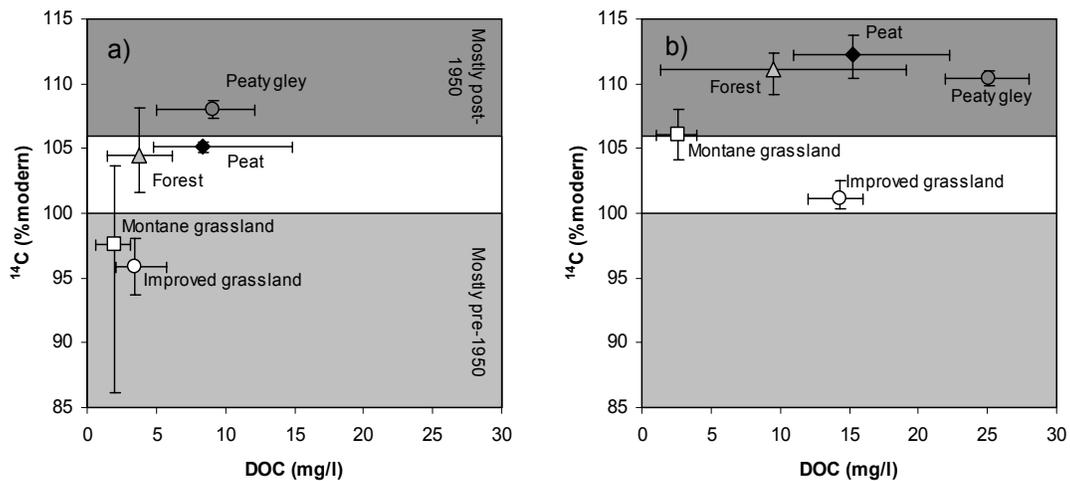


Figure 4.1 ^{14}C Isotopic composition and concentration of stream DOC at a) summer low flow, and b) autumn high flow for five major soil/land-use types in North Wales. Error bars show range of observed values. Samples in upper shaded area have ^{14}C composition indicative of a probable post-1950s source, lower shaded area indicates pre-1950s source. From Evans *et al.* (in review).

4.1.4. Causes of rising DOC concentrations

DOC concentrations have been rising in the 22 lakes and streams of the UK Acid Waters Monitoring Network, including all sites in Scotland and Wales, since 1988 (Freeman *et al.*, 2001; Evans *et al.*, 2005; Figure 4.2). These increases have been replicated across a wider range of UK surface waters (Worrall *et al.*, 2004), in some water colour records going back to the 1970s (Worrall *et al.*, 2003b); and across large areas of Scandinavia and Northeast North America (Skjelkvåle *et al.*, 2005). The geographical extent

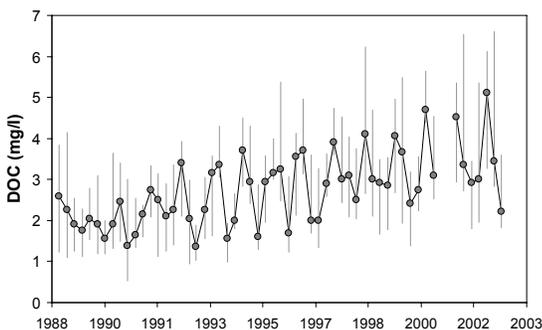


Figure 4.2. DOC concentrations at 10 lakes in the UK Acid Waters Monitoring Network. Circles and bars represent the median and interquartile range of DOC concentrations across all sites at each sampling interval. (From Evans *et al.*, 2006)

of observed trends, from heathland, grassland and plantation forest catchments in the UK, to unmanaged mixed forests in North America, argues against local management factors as the fundamental cause of these increases, and this section therefore focuses on large scale environmental drivers that might account for observed changes. However, there is good evidence that management and land-use can have a significant impact on DOC export, either contributing to observed increase in some areas, or providing the means (through management or land-use change) to ameliorate changes; these issues are discussed in the following section.

Possible drivers of DOC change operating at a large scale include climate (e.g. temperature, rainfall, drought), changes in atmospheric chemistry (CO_2), and deposition (of sulphur and nitrogen compounds). Each of these mechanisms has been reviewed in Evans *et al.* (2005), a paper supported by the project. The following summarises and updates the published review:

Decreasing acid deposition: Sulphur (S) deposition has more than halved across the UK since the mid-1980s, triggering widespread increases in soil solution and runoff pH, and decreases in ionic strength (IS). Low pH and high IS reduce the solubility of weakly acidic humic substances (Tipping and Hurley, 1988; Kalbitz *et al.*, 2000). Although buffering of mineral acid changes by organic acids in peaty soils was proposed as early as the 1980s (Krug and Frink, 1983), it has only relatively recently been considered as a likely driver of rising DOC (e.g. Stoddard *et al.*, 2003; Evans *et al.*, 2005). Experimental studies appear to confirm that soil solution DOC is highly sensitive to changes in acidity and IS (Mulder *et al.*, 2001; Clark *et al.*, 2005). The DOC model of Lumsdon *et al.* (2005), and the DyDOC model (Michalzik *et al.*, 2003) both incorporate pH effects on DOC solubility. In a second paper supported by the project (Evans *et al.*, 2006), decreasing S deposition was argued to be the dominant cause of rising DOC, in the UK and elsewhere.

Rising temperature: Due to the temperature-dependence of organic matter decomposition rates, when DOC increases were first noted in UK waters it was suggested that this could be a response to long-term warming (Freeman *et al.*, 2001a). A large number of studies have observed an increase in DOC production with rising temperature (see e.g. Kalbitz *et al.*, 2000; Evans *et al.*, 2005, and references therein). In addition, seasonal DOC variations are closely related to temperature (e.g. Scott *et al.*, 1998; Clark *et al.*, 2005; Billett *et al.*, 2006), and the DOC model developed by Lumsdon *et al.* (2005) required a temperature-dependent term describing biological activity, in addition to geochemical controls, to reproduce seasonal variations in soil solution DOC. However, the near-doubling of UK surface water annual average DOC concentrations since the 1980s, for an approximate 0.6 °C average warming over the same period, would require a Q_{10} for DOC production (the amount by which the rate of a process changes over a 10°C change in temperature – a doubling of a rate over 10 °C would be a Q_{10} of 2) well in excess of those recorded in peat warming experiments (1.33 to 2.13 under anaerobic conditions; Freeman *et al.*, 2004; Evans *et al.*, 2006). In addition, both decomposition rates, and the partitioning of decomposition products between CO₂ and DOC, are sensitive to factors other than temperature. In peatlands especially, the maintenance of anaerobic conditions due to waterlogging may restrict decomposition (e.g. Freeman *et al.*, 2001b; Davidson and Janssens, 2006). In aerated peat, decomposition rates are higher and temperature sensitivity increases (Q_{10} for DOC production 3.66, Evans *et al.*, 2006). Therefore a combination of warming and drying may have made a significant (but probably partial) contribution to observed DOC increases (e.g. Worrall *et al.*, 2004; Evans *et al.*, 2006).

Hydrological changes: Several possible mechanisms exist by which hydrological factors may affect DOC export. These include shifts in water flowpath from (low-DOC) mineral horizons into (high-DOC) organic horizons during rain or snowmelt events (e.g. McDowell and Likens, 1988; Hongve *et al.*, 2004), and increased DOC production rates under aerated conditions during droughts. The former is more important in organic-mineral soils, the latter in peats. The effects of drought in peats may be complicated by constraints on enzyme diffusion at low soil moisture levels (see e.g. Davidson and Janssens, 2006); more complete mineralisation of organic matter (i.e. to CO₂ rather than DOC; Scott *et al.*, 1998) and concurrent chemical changes (e.g. oxidation of reduced sulphur compounds) which reduce DOC solubility in soil water (Clark *et al.*, 2005). Although there is no question that hydrological factors have a major influence on DOC export in the short-term, there is less consensus as to whether long-term changes can explain DOC trends. Consistent UK-wide hydrological trends are difficult to identify, and there have certainly not been consistent hydrological changes across the wider European and North American area throughout which rising DOC trends have been

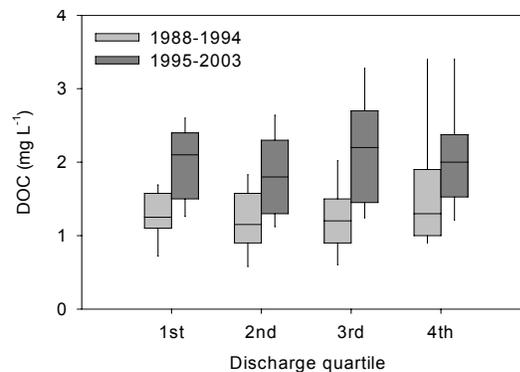


Figure 4.3. Increasing DOC under all flow conditions at Dargall Lane AWMN site (from Evans *et al.*, 2006)

recorded. Analysis of UK monitoring data suggest that DOC increases have occurred under all flow conditions rather than due to any long-term change in the dominance of low- or high-flows (Figure 4.3). Results from an ongoing CEH experiment on an organo-mineral soil in North Wales (Figure 4.4) provide clear evidence of the short-term impact of annual induced droughts, but evidence of a long-term response is equivocal.

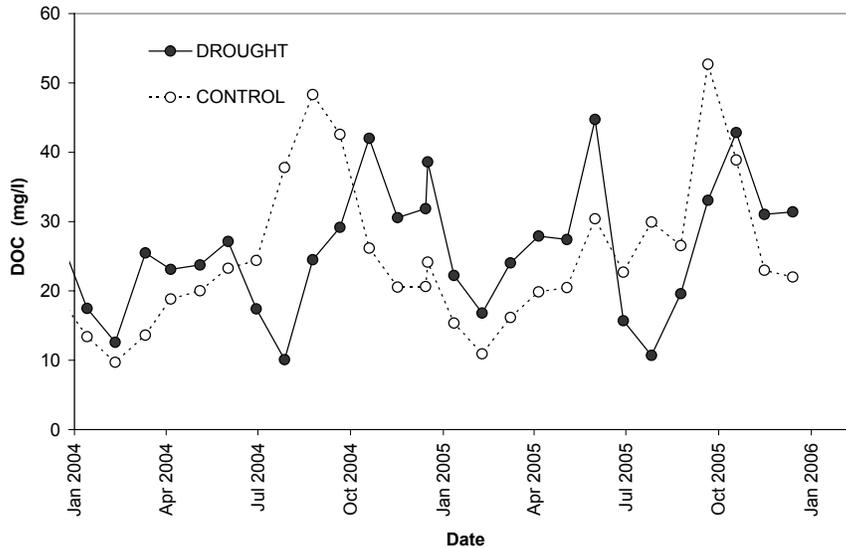


Figure 4.4 The effect of experimental drought on heathland organic horizon DOC concentrations at the Clocaenog site, North Wales. DOC concentrations are reduced during experimental summer drought, and increased during winter (CEH Bangor, unpublished data).

Productivity changes: As noted above, much of the DOC exported from organic soils appears to comprise carbon assimilated from the atmosphere relatively recently, rather than from decomposition of ‘old’ soil organic matter. Changes in ecosystem productivity have the potential to increase DOC export through accelerated production of labile litter or root exudates. Freeman *et al.* (2004) found that vegetated peat cores exposed to elevated atmospheric CO₂ underwent an increase in productivity (in part due to vegetation changes) and that the proportion of recently-assimilated carbon in the DOC pool increased greatly. The resulting increase in the bulk DOC pool, relative to an experimental doubling of atmospheric CO₂, appears insufficient to explain the full magnitude of surface water DOC increases (Evans *et al.*, 2006), but may be a partial contributor to observed trends. In addition, productivity of most organic soil ecosystems is N-limited, and productivity increases due to elevated anthropogenic N deposition have been proposed as a cause of rising DOC (Findlay, 2005). While evidence for recent productivity changes on the required scale in UK upland ecosystems has not been identified, this is considered to be an area requiring further investigation. A review of existing N manipulation experiments was carried out, with inconsistent results: some experiments show a substantial DOC increase (e.g. Pregitzer *et al.*, 2004) but many others show no change. Because experimental N addition can potentially change many

factors other than productivity (most notably soil acidity), considerable caution is required in the interpretation of these results.

4.1.5. Influence of land use and management on DOC loss

At the scale of Northern Europe and Northeast North America across which DOC increases have been recorded, no single explanatory land-use change factor can be identified. Within the UK itself, DOC increases have been observed in waters draining grassland, heathland and plantation forest catchments, subjected to varying types and intensities of management (Evans *et al.*, 2005). However, at a local and regional level, changes in land use and management have the potential to significantly alter carbon cycling in general, and DOC export in particular. This section therefore summarises both the extent to which past changes may have influenced DOC losses from managed upland systems, and the extent to which future management might be adapted in order to minimise DOC losses.

Drainage: Peatland drainage has been widely practiced in the UK since the 1940s, e.g. to improve grazing quality or to permit afforestation. By lowering water tables and providing a more intensive drainage network, this may both increase decomposition rates (due to increased aeration), and reduce the potential for DOC retention within the soil (e.g. Holden *et al.*, 2004; Worrall *et al.*, 2004). However as a mechanism to explain recent DOC increases this is unlikely to be important, because since the 1990s there has been an increasing tendency towards peatland restoration through drain blocking. A recent study by Wallage *et al.* (2006) confirms that DOC losses were higher from drained peats, and suggests that drain blocking will reduce long-term DOC losses to levels below those of undrained sites. Worrall *et al.* (in review) found that drain blocking actually led to a short-term (1 year) DOC flush, but a modelling study (Worrall *et al.*, submitted) does suggest that, in the longer term, DOC losses will be reduced. Drain-blocking may therefore represent an effective amelioration strategy for reducing DOC loss in drained peatlands.

Burning: Heathland burning is practiced in many areas, both for habitat management, and (more intensively) for grouse shooting. For England, Yallop *et al.* (2006) found evidence of burning in around 70% of the heathland area, with a near-doubling in the extent of new burns since the 1970s. In the Pennines, a strong correlation has been shown between surface water DOC concentrations and extent of burning within the catchment (A. Yallop, pers. comm.), suggesting that increased burning could be a significant contributor to increased DOC loss. For a long-term burn manipulation at Moor House, however, Worrall *et al.* (in press) recorded reduced DOC concentrations in soil solution beneath burnt plots. Although the reduction or cessation of burning in managed heathlands might be expected to reduce catchment C losses, then, further work is required to confirm this. A paired-catchment experiment in the Peak District (O'Brien *et al.*, 2005), in which burning, drainage and grazing have all been manipulated, may provide new information, but results are not yet available.

Afforestation: Establishment of plantation forestry on grazed or semi-natural organic soils could alter DOC export in several ways. Drainage will increase soil aeration, as described above, and enhanced transpiration will have the same effect. Changes in carbon cycling, including the development of a forest litter layer, might be expected to increase the production of DOC. However, the evidence that afforestation alters DOC export is surprisingly weak; although Grieve and Marsden (2001) recorded higher DOC in forest versus moorland soil solutions, Hope *et al.* (1994) actually found slightly lower export rates for temperate forests versus moorland and grasslands. Comparing CEH data from moorland and forest catchments on peaty podzols, no effect of land-use on DOC loss can be identified for either streams (Reynolds, in press) or soil solutions (CEH unpublished data - see 2004 report). The strongest forest impact appears to accompany clearcutting, when large DOC pulses may occur (Hughes *et al.*, 1990).

Grazing: Grazing of the uplands, primarily by sheep, can affect vegetation cover, compaction and drainage (Milne, 1996), and grazing intensity has increased widely in recent decades. Available data, however, show little evidence that grazing changes affect DOC export. Worrall *et al.* (in press) found no significant difference in soil solution DOC between grazed and ungrazed heathland plots at Moor House. Data from the CEH Pwllpeiran experiment (B. Emmett, unpublished data) also showed similar soil solution DOC in a podzolic grassland under low and high grazing intensity. However, complete cessation of grazing in grasslands may lead to a relatively rapid transition to heathland (e.g. Hill *et al.*, 1992), with potentially greater consequences for DOC losses, and C cycling in general.

Liming: Liming of acid soils may increase DOC production, by accelerating microbial activity, or reduce DOC retention by raising pH, a process analogous to that described earlier with regard to falling S deposition. Increased DOC loss due to liming has been shown in studies of moorland podzols (Hornung *et al.*, 1986; Reynolds *et al.*, 1994), peats (Soulsby and Reynolds, 1995) and forests (Andersson and Nilsson, 2001). Liming was widely practiced in the UK to improve the productivity of acid grassland, but has become less widespread since the 1980s due to the removal of subsidies. However, recent restoration work to revegetate degraded peats in the Pennines has included application of lime. It is suggested that this will lead to elevated DOC losses, potentially over a sustained period.

In summary, if future climate change results in warming and drying, DOC losses would be expected to increase. Further, climate induced changes in hydrology and plant productivity could increase C losses (at least in the short term) though the longer term impacts are uncertain. A number of land management practices, including drainage, burning, afforestation, grazing and liming, might also influence DOC losses. This is significant also from a water quality perspective as there are significant costs associated with removing colour from drinking water.

Module 5

Estimates Of Carbon Loss From Scenarios Of Accelerated Erosion Of Peats

5.1 Introduction

This module reviews the processes that have been implicated in the erosion of peats and organo-mineral soils and how these may change into the future. Estimates of the extent of erosion are also presented and discussed and areas requiring further research have been identified.

The literature, and indeed evidence, pertaining to the erosion of organic-rich soils focuses heavily on blanket peat or mire soils rather than organo-mineral soils such as peaty podzols, peaty gleys or peaty rankers. All the available evidence suggests that the latter group of soils are much more stable than blanket peats and there are few instances of erosion on them. More details are given in Table 5.1. Against this background, the remainder of this section will focus on peat erosion and it is likely that the processes that contribute to peat erosion are also applicable in some degree to organo-mineral soils.

This report covers a brief review of the processes and causal factors of erosion of peats, assessments of the extent and likelihood of erosion of peats, approaches to define, identify and quantify peat erosion and losses, an appraisal of potential changes to the causal factors and an assessment of likely changes to erosion risk and ends with conclusions and a summary.

5.2 Background

Soil erosion is a natural process which occurs in all soils to a greater or lesser extent. Soil erosion becomes of concern when the rate exceeds “natural” or “background” rates which can be considered as broadly equal to the rate of formation of new soil material by weathering processes. Soil erosion at rates exceeding background values is termed “accelerated erosion” and, in humid temperate climates such as that of Scotland, much of the erosion is usually the result of human activities that lead to removal of the protective vegetation cover. However, mass movements such as landslips also occur in the over-steeped, glaciated hillslopes of Scotland (Ballantyne, 1991). The material eroded from soils is a major contributor to the sediment load carried by streams and rivers and can disrupt transport links, therefore erosion has implications both on- and off-site.

The major processes of soil erosion are:

- i) Water erosion including gullying, rilling and sheet erosion
- ii) Mass movements such as landslides
- iii) Wind erosion.

Erosion of cultivated mineral soils has been the subject of considerable research efforts and the review of evidence within Towers *et al.* (2006) provides an up-to-date summary of our current understanding of the processes involved. These events tend to be localized and, although their short term impact can be dramatic, in most circumstances they are

relatively easy to rectify. In contrast, upland erosion can cover large areas and is more difficult to repair.

Erosion of peat ranges from a few scattered rills, hags or gullies within an otherwise pristine bog, to bogs where there has been complete removal of the surface vegetation with large areas of bare exposed peat. Two examples are given in Figures 1 and 2. In Scotland, the pattern represented by Figure 5.1 is much more common whereas that represented by Figure 5.2 is restricted to specific areas of serious erosion. In some places, the original peat cover has been almost completely removed and only remnant “hags” remain (Figure 5.3)

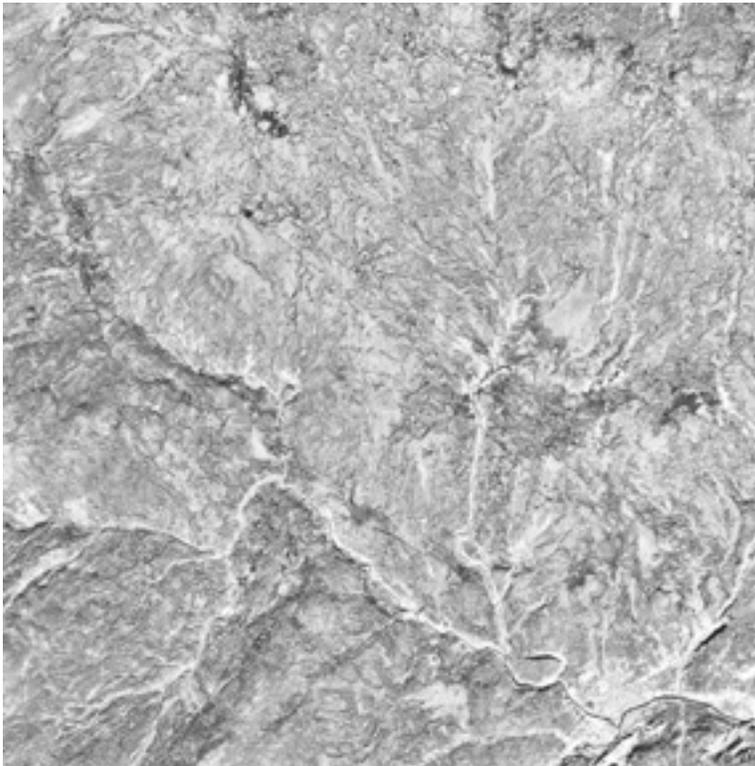


Figure 5.1 Typical pattern of gully erosion in blanket peat



Figure 5.2 Severely eroded peat with extensive bare ground (black areas)



Figure 5.3 Remnant of peat, indicative of a much more extensive cover in the past.

5.3 Causes of peat erosion

5.3.1 Erosion processes

There are various ways to describe different erosion processes, features and patterns in peat. Perhaps the most influential was that of Bower (1960) who identified five main types of erosion system on the basis of morphology and pattern. These were considered to result from the operation of two main processes: water erosion and mass movements. Water erosion produces dissection systems which develop onto and into the peat mass whereas sheet erosion occurs on the peat surface as the vegetation breaks up. On steeper slopes mass movements may result and under very wet conditions bog bursts and peat slides can occur.

The dissection process is much more prevalent and extensive compared to bog bursts and slides, produces a very distinctive landscape, but operates over very long time periods. Any changes to that landscape happen very slowly compared to peat slides and bog bursts and for that reason, the latter are studied much more intensively as the evidence of the erosion event and its associated impact is much more immediate. Brief reviews of both processes are given below.

5.3.2 Dissection processes

Drainage dissection was considered by Bower to be the more important erosive process both spatially and in terms of volume of peat removed (Bower, 1960). Indeed understanding drainage dissection is arguably the key to understanding peat loss in the uplands (Holden *et al.*, 2006). Two types of dissection system associated with water erosion were identified by Bower, which produce different types of gully patterns. The key differences in the pattern of gullies produced by the two dissection processes are a function of peat depth and slope angle, both of which themselves are inter-related. The frequency and complexity of the gully pattern is highest on high level, very gently sloping upland plateau. Bower's classification has been widely employed in subsequent studies on peat degradation albeit with some suggested modifications to it, e.g. Tomlinson (1981).

Although peat erosion produces a quite dramatic landscape of hags, gullies and shallower channels, poor connectivity between them often means that peat that is removed does not always find its way to the drainage system but is redistributed within the peat mass (Evans & Warburton 2005). The connectivity between these different components can change with time and can alter the amount of sediment delivery. Extensive revegetation of the gully floors can have similar effects.

The trigger for erosion of organic and organic-mineral soils has been the subject of much debate and research for a number of decades. Burning, either accidentally or deliberately, grazing pressure by sheep and/or deer, air pollution, peat cutting, drainage and recreational pressure have all been cited in the literature as either causing or exacerbating erosion of upland soils. Exceptional weather events are also contributory factors and there are a number of studies that suggest that climatic perturbations during the previous millennium (Burt *et al.*, 1997. Holden *et al.*, 2006) caused desiccation at the peat surface

and provided the initial trigger for peat erosion. Subsequent anthropogenic factors such as burning, grazing and atmospheric pollution have then impacted on the initial disturbed surface and exacerbated the erosion effect.

A number of factors have been implicated as the causes of peat erosion but most workers recognise the complexity and inter-relationships between them. By far the most work on peat erosion is from the Southern Pennines and there must be care in how the findings from these studies are extrapolated to other areas; the Southern Pennines receives a lower annual rainfall than much of the remainder of blanket mire in the UK so climate influences differ but, equally important, they are in close proximity to a population of more than 5 million people and the recreational and air pollution pressures that that entails (Tallis, 1997).

5.3.3 Climate

There is still considerable debate about the balance between whether peat erosion is a man-induced process or whether it is also partly a natural process representing the end point of a cycle of accumulation and build up of organic matter. There is some evidence from the Southern Pennines suggesting that climatic perturbations at key points of the last millennium provided the initial ‘trigger’ for peat erosion and that the much of the present landform stems from that period. Between AD 1150 to AD1300 (the early medieval warm period), there was a drier than average period causing drying out of pools, shrinkage and cracking of the peat and crusting of bare peat surfaces. From AD1300 to AD 1740 during which time the Little Ice Age occurred, there were generally colder winters and wetter summers with resultant volumes of run-off increasing erosion risk (Burt *et al.*, 1997). Wishart & Warburton (2001) also suggest that some of the gully systems on the Cheviot Hills may be at least 500 years old.

Presently much of the eroded peat resource is at high altitude and in general terms, the higher the peat is above sea level the greater is the likelihood of it being eroded and the proportion of it that is eroded (Grieve *et al.*, 1994). This suggests that the current climate is also a factor in peat erosion. The positive correlation between altitude and severity of gully erosion, suggesting that climatic factors, including the incidence of frost, high winds and more frequent and heavier rainfall events are important, was recognised as long ago as 1949 by Osvald and also by Bibby (1982). Although some areas of eroded peat are found as low as sea level, it is usually in very exposed situations such as Lewis, Shetland and the western seaboard of Scotland. Analysis of a number of sites from throughout the UK and Ireland suggest that climatic influences may be the most important influence in causing erosion (Rhodes & Stevenson, 1997).

Summer drought has the potential to bring about large and possibly irreversible changes to peats through desiccation of the soil surface, leading to cracking and increased vulnerability to erosion (various studies cited in Holden *et al.*, 2006). Such drying also makes the peat more vulnerable to wildfires.

5.3.4 Burning

Burning is a commonly used management tool to control semi-natural vegetation on heathlands and to a lesser extent on blanket mires. If these fires take place at the appropriate time (October to April), are well managed and not allowed to become too hot and destroy the root mat, they should not expose the peat to erosive forces (Rhodes and Stevenson 1997). However in some areas, damage is caused by fires as a result of heavy recreational use, often in summer and these fires are unmanaged (Anderson 1997). Rhodes and Stevenson (1997) found that vegetation burning was implicated on only one site out of seven sites studied although the authors do speculate that the burning may have helped perpetuate and enhance erosion. Imeson (1971) and Yallop (2006) contend that heather burning has been a causal factor in soil erosion, and Yallop (pers. comm.) suggests that there is a lot of additional evidence, admittedly most of which is apocryphal, to support this view. The balance of opinion is that fire in general is not a key trigger in initiating peat erosion although accidental (or deliberate) sporting- and recreation-related fires do have the potential to cause serious damage in single events. More work is required to determine the role that fire plays in soil erosion, but if practiced properly.

5.3.5 Grazing

Grazing by domestic (largely sheep) and wild animals (largely red deer) can alter the ground vegetation on peat (which in itself is viewed as degradation of the resource) and when carried to its ultimate extreme, create bare patches of peat which are exposed to climatic and other influences to promote further erosion. This process is very similar to the poaching of wet soils by grazing animals within fields but a key difference is that the latter can be rectified reasonably well and quickly whereas restoration on peat is much more difficult, expensive and takes much longer.

Heavy grazing by sheep causes a decline in the cover of heather and other ericoids and replacement by tussock forming graminoid species such as *Eriophorum Angustifolium*, *Molinia Caerulea* and *Nardus Stricta*. *Nardus* in particular has short rhizomes and hence has poor soil binding qualities. Thus peat or other soils with high organic rich surface horizons supporting a *Nardus*-dominated vegetation become more prone to erosion (McKee & Skeffington, 1997). Birnie & Hulme (1990) noted that much of the peatland vegetation in Shetland showed evidence of grazing modification and peat erosional features are widespread. Biologically unsustainable grazing levels have been identified as the cause of the continuing degradation of both the vegetation and peat resource in Shetland. The balance of evidence is that while heavy grazing may exacerbate and accelerate peat erosion the original trigger for its initiation probably predates the intensive use of the hills for grazing, around two centuries ago (Rhodes & Stevenson 1997).

5.3.6 Atmospheric pollution

Skeffington *et al.* (1997) suggest that acid deposition since the start of the Industrial revolution about two centuries ago may be one of the causal factors for peat erosion in the Southern Pennines. *Sphagnum* has almost completely disappeared from these bogs thereby preventing further accumulation of organic matter at the soil surface; although

the reason for this loss is not well understood, it is thought to be linked to acidification. Another aspect of acid deposition on blanket bogs and whether it has caused a reduction of fertility of these systems is the leaching out of base cations. In a comparison of 8 bogs across the UK from NW Scotland to SW England, there is some evidence, but no more than a trend, that the two Southern Pennine peats have lower exchangeable Mg and K than the others. Exchangeable Ca is lower at one of the sites. However, none of the levels of base cations correlated significantly with sulphur deposition. It may be that atmospheric deposition does not cause erosion *per se*, but the huge loss of *Sphagnum* spp. due to acid deposition, provided the impetus for other factors to take effect. Cresser *et al.* (1997) suggests that further work is required on the possible effects that acid deposition has had on the physical properties of peat.

Nitrogen deposition onto organic soils is harmful to them in terms of altering species composition and hence the habitat value of the resource but is unlikely to increase the risk of physical erosion. Indeed the decline in sulphur deposition and increase in N deposition has been linked to revegetation of parts of the North Pennines and hence providing a stabilizing influence (Evans and Warburton 2005) and may be responsible for increased DOC loads in streams (see Module 4).

5.3.7 Drainage

During the 1950s and 1960s, there were extensive efforts to drain blanket peat for agricultural purposes, although their effectiveness was never properly assessed (Holden *et al.*, 2006). Artificial drainage may have altered the natural hydrology of peatland systems and thereby potentially making them more susceptible to erosion forces, but there is no clear evidence that this is the case. Certainly in Scotland, areas of eroded peat have not been subjected to artificial drainage and areas that have, based on visual observation in the field and from air photographs, do not show any evidence of erosion.

5.3.8 Other factors

Previous studies (Tallis *et al.*, 1997) have identified factors that cause peat 'degradation' but in a different sense from peat erosion. These include the domestic and commercial extraction of peat as a fuel, the fertilization of peat for agricultural improvement, forestry development, reseeded, and the increase in the use of ATVs (all terrain vehicles) by land managers. Peat extraction and trafficking by vehicles have the potential to increase the amount of bare soil exposed to other agents but there is little hard evidence that this has happened to any great extent. Forestry operations in the past, particularly crop establishment, were very intrusive, but there is no literature to suggest that this caused any increase in the physical movement and erosion of peat. Peat is also very susceptible to damage by trafficking either by walkers or by the use of All Terrain Vehicles (ATVs) and this is a growing threat to the peat resource, albeit in specific local contexts.

5.2.9 Peat slide and bog bursts

Bog bursts and slides are generally relatively small in extent and Warburton *et al.* (2004) has produced a comprehensive review of the processes involved. Sites of peat mass movement share a number of common characteristics which predispose them to failure:

- a peat layer overlying an impervious or very low permeability base
- a convex slope or a slope with a break of slope at its head
- proximity to local drainage either from seepage, groundwater flow, flushes, pipes or streams
- connectivity between surface drainage and the peat/impervious interface

There have been a number of bog bursts and slides reported in Scotland, e.g. Bower (1960) and Acreman (1991), but most have been reported from the Pennines (Nolan and Birnie 2006, Holden *et al.* 2006). An example is illustrated in Figure 5.4. There is evidence of a marked increase in the frequency of landslides recorded in British peatlands and the associated frequency of triggering events such as intense rainstorms. There are obvious implications for this trend continuing should predicted climate change scenarios become reality.



Figure 5.4 A bog slide in Skye, September 2005. The human figure on the left hand side, near the top of the slide, indicates the scale. Photo courtesy of A J Nolan

Nolan & Birnie (2006) in their report on a peat slide on Skye in 2005 discussed whether this type of event was possible to predict. As Warburton *et al.* (2004) concluded that ‘the prospect of predicting the location and timing of such events is still a long way off’, they suggest a more practical approach might be to develop a field assessment method which will provide an indication of the vulnerability of any slope to failure. It is the actual risk of failure – the product of inherent vulnerability and the exposure to trigger events – that makes it very difficult to predict. Vulnerability is a relatively static factor and is therefore easier to assess, exposure is not. They suggest a set of easily observable factors that should be considered to assess vulnerability of any slope to failure. These include:

- the presence and depth of a peat surface horizon
- the steepness and form or slope
- degree of humification of the peat
- general hydrological characteristics of the site
- if the peat does fail, are there natural anchoring points to allow it to stop?

Whilst not suggesting that this is a fully developed rigorous decision support tool, further development and refinement of the underpinning logic provides some promise.

5.4 Extent of peat erosion

5.4.1 Scotland

The extent of peat erosion has been quantified to a much greater extent in Scotland than in Wales, but there are difficulties in defining, identifying and delineating peat erosion consistently. Without a systematic image analysis approach, it is virtually impossible to get a completely consistent definition or delineation of 'eroded peat'. For this reason, there is no defined density of hags or gullies for an area to qualify as eroded peat within either the 1: 250 000 scale soil map of Scotland (MISR 1984) nor the Land Cover of Scotland 1988 (LCS88) (MLURI 1988) datasets. Areas delineated as eroded peat in both these datasets include large areas that are not actively eroding, i.e. the areas that lie between the gullies that in most instances comprise most of the area. In essence what was delineated was a recognisable area of land that exhibited erosion features but with no hard and fast definition.

A number of data sources have been used (MISR 1984, MLURI 19995 and Grieve *et al* 1994, 1995) to assess the extent of peat erosion and have been summarized in Towers *et al* (2006). The LCS88 dataset shows that just less than 6% of Scotland had eroding blanket bog which is approximately 34% of all areas of blanket bog identified. This compares with around 7.5% of Scotland or 31% of all peat categories as calculated from the NSIS. Given the differences in the datasets (point vs polygon), there is good correspondence between these figures. Given that a map unit of eroded blanket bog will have substantial areas of both bare and vegetated (that is, uneroded) peat, in many cases, the vegetated area will dominate, the actual area of eroded and bare peat will be considerably less than 6%. This also holds true for calculations based on the NSIS. The largest areas of eroded peat are along the Caithness/Sutherland border, on Lewis and Shetland and in the Monadhliaths between Loch Ness and Strathspey.

A partial resampling of the NSIS is due to start in 2007, and this may provide an indication of whether there has been any change in the incidence of erosion on different soils since the initial survey. However the intensity of erosion was not recorded and the lack of strict definitions for each erosional feature suggest that any new information may be of limited value.

Grieve *et al.* (1994, 1995) quantified the area of erosion from aerial photographs in a 20% sample of the Scottish uplands. Peat erosion accounted for the greatest extent, 6% of the sample area, and the extent is very similar to that derived from analysis of the LCS88

and NSIS data. It must be emphasized again however that these data do not indicate that 6% of the blanket peat area of Scotland has been lost to erosion, but simply that erosion has affected peat soils in 6% of the area.

5.4.2 Wales

The total area of peat in Wales is between 63 500 (Rudeforth *et al.* 1984) and 78 000 (Taylor and Tucker 1983) hectares. The two surveys use slightly different criteria to define peat but both estimates exclude mineral soils with peaty surface horizons (organo-mineral soils). Both equate approximately to around 3.4% of the total area of the country, the figure given by CEH (2003). The distribution of peat in Wales is shown in Figure 1.2.

There is no indication of the extent of eroded peat within Wales (Rudeforth *et al.* 1984, CEH 2003) although both publications allude to the existence of eroded peatlands. Holden *et al.* (2006) contend ‘that many organic soils in England and Wales are severely degraded and that most organic soils in England and Wales are degraded in some way, even if not severely’. McHugh *et al.* (2000) state that most erosion on upland soils was associated with peat soils, with increasingly lower incidences of erosion on wet organo-mineral and dry mineral soils similar in broad terms to that observed in Scotland (Table 5.1). It is recommended that a systematic assessment be made of the Welsh peat resource in order to get a robust estimate of the area of eroded peat.

The recent DEFRA-commissioned study of the extent of soil erosion in the upland areas of England and Wales (McHugh *et al.*, 2002) was a ground-based survey measuring similar parameters to those measured in the SNH-commissioned air photo study of upland Scotland (Grieve *et al.*, 1994, 1995). McHugh *et al.* (2002) found that the extent of degraded soil represented around 2.5% of the area surveyed, a smaller percentage than that reported by Grieve *et al.* (1995). McHugh *et al.* (2002) quantified only the area of degraded soil, and thus the results are not directly comparable with the area of soil affected by erosion computed in the Scottish study. However, both studies provide clear evidence that erosion is most extensive on organic (peat) soils. Organo-mineral soils contributed about one third of the sites that displayed erosion.

5.5 Approaches to define, identify and quantify peat erosion and losses

5.5.1 Modelling of erosion susceptibility

The risk of soil erosion occurring in Scotland has been modelled following two different procedures.

- Lilly *et al.* (2002) used a rule based approach to identify the inherent geomorphological risk of soil erosion by overland flow
- Anthony *et al.* (2006) adopted a more process based approach integrating water balance models with understanding of soil erosion processes at a field scale and was designed to predict sediment and phosphate movement to waterbodies as part of a diffuse pollution screening tool.

Both approaches relied to varying degrees on national datasets of soil texture, Hydrology of Soil Types (HOST) class and digital elevation models (DEM) and produced output at a

fairly crude scale (1km grid cells). These approaches are described in detail in Towers et al (2006).

Lilly *et al.* (2002) developed an erosion classification system that assumes that all soils are bare and that erosion can be modeled using the inherent characteristics of the soil to absorb water from rainfall or snowmelt. Separate decision rules were developed for mineral and organic soils and applied to each 1km grid cell in the national scale 1:250 000 soil map of Scotland which was overlain with a GIS coverage of slope categories derived from a DEM. A summary of the results are shown in Figure 5.5.

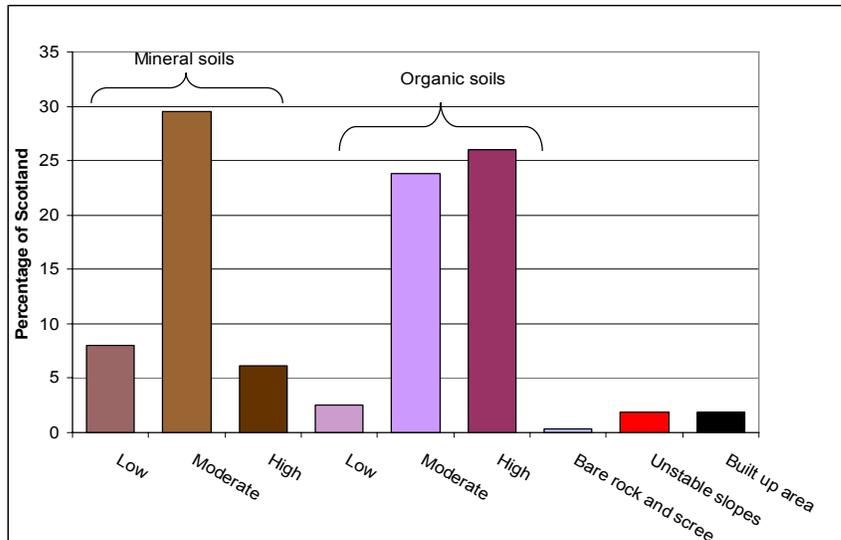


Figure 5.5 Proportion of inherent geomorphological erosion risk.

Figure 5.5 shows that the majority of mineral soils fall into the moderate risk category (just under 30% of Scotland) while organic soils are predominantly in the moderate or high risk categories (≈ 24 and 26% respectively). Much of the mineral soils at moderate risk of erosion by overland flow are to be found in the agricultural lowlands (Figure 5.6). However, it has to be stressed that this classification is the inherent risk and not the actual risk of erosion.

This work has been extended for upland soils where the frequency of disturbance to the vegetation cover was added. Thus soils with a semi-natural vegetation cover that was unlikely to be removed would be deemed to have a low risk of erosion even if the soils were highly susceptible. Land uses such as forestry or those involving muirburn would have a greater likelihood of bare or disturbed soil and be at risk. There is a very marked reduction in the risk of erosion occurring in upland Scotland when vegetation cover is taken into account. Areas with the highest susceptibility to erosion fell from 26 (from Figure 5.5) to 8% while the area with the lowest susceptibility rose from 2.5 (from Figure 5.5) to 42%.

The diffuse pollution screening tool (Anthony *et al* 2006) was developed for SEPA and EHS Northern Ireland and covered a wide range of potential pollutants amongst which

was suspended sediment. This model differed from that of Lilly *et al.* (2002) in that it attempted to predict sediment yields. Most of the area under organic or organic-mineral soils have predicted sediment losses of less than 50 kg/ha/annum and on closer examination of the database, 84% of this area has predicted sediment losses of less than 10 kg/ha/annum and 65% has predicted losses of less than 1 kg/ha/annum. It may be that the particulate organic carbon loss is being underestimated through a lack of understanding of the processes involved compared to the removal and transport of mineral sediment.

Estimates of particulate organic carbon (POC) release from UK uplands vary from 0.1 tonne/sq km/annum in intact Scottish peatlands (Hope *et al.* 1997) to about 100 tonne/sq km/annum in the southern Pennines (Evans *et al.* 2006). POC flux is largely controlled by the geomorphic processes that determine the transport and deposition of eroding organic material. There is some debate about the fate of POC (or eroded carbon). If it is rapidly altered to gaseous and dissolved forms in the fluvial system the carbon may be made available in a climatically active form (Pawson *et al.* 2006). Holden *et al.* (2006) argue that POC loss in severely eroding peatlands is a major, if not the single largest component of the carbon budget. Most sampling programmes have ignored POC removal from organic soils and Holden *et al.* recommend that further research is required on POC fluxes in degraded peatland systems.

Overall, erosion models provide a useful indication of the geographical variation in erosion susceptibility and highlighting areas potentially at risk. They also provide a framework for evaluating the effects of changing rainfall patterns or intensities. However, the predictions from soil erosion models are often insufficiently verified against field data (Brazier, 2004) and model predictions cannot therefore be seen as a substitute for measurements of actual erosion rates or of the occurrence of erosion events.

5.5.2 Surveys vs. sediment budget analysis

There have been a number of surveys, using subtly different methodologies, with the objective of quantifying the extent and severity of peat erosion. They are sampled based and have variously used National Soils Inventory (NSI) sites in England and Wales (McHugh, 2002) and 5 km X 5 km squares in Scotland (Grieve *et al.*, 1995). It should be emphasised that both surveys were seeking to identify and quantify all types of upland erosion including debris flows, landslides and sheet erosion as well as different forms of peat erosion.

There has been some relatively recent debate about the relative usefulness of such surveys specifically by Warburton *et al.* (2003) on the McHugh survey. They argue that the methods used lacked adequate definition, that there was a lack of geomorphological context to the results in that there was a lack of appreciation of the context of the 50m plots within the larger sediment system and no linking of erosion loss to timescale. They argue that an approach based on sediment budget analysis provides reliable estimates of erosion and deposition, an indication of how upland sediment systems operate, and the relative significance of slope and channel processes. They are also critical of the emphasis on the assessment of the amount of bare ground and a lack of awareness of the

significance of stream channels as the main factors controlling the sediment flux from upland areas. They advocate a number of appropriately placed sediment budget assessments in the uplands rather than a rigid countrywide survey.

5.5.3 Use of air photographs/remote sensing

There is a strong temporal dimension to peat erosion processes and there is some strong evidence that the current pattern of gullies and hags may have been initiated several centuries ago. Two studies have used air photographs from two different time periods to compare and contrast and whether any significant changes in erosion extent could be tracked over that time period. McHugh (2000) found that between 1946 and 1989 both peat and mineral erosion appeared stable with approximately 60% of the erosion features in both categories showing no change in erosion. Conversely, between 55 and 60% of erosion caused by landusers and grazing animals deteriorated further in the same period. When subdivided into individual five-year periods, it was clear that observations of recovered erosion consistently equalled or exceeded those of deteriorated erosion. Air photographs also indicated that peat and mineral erosion were not important sources of continued erosion or of newly initiated degradation between 1946 and 1989. However the report acknowledges that one of the main difficulties associated with assessing losses from eroded peat may be the continued removal of material but without an associated increase in the areal extent of the feature.

Wishart & Warburton (2003) compared oblique photographs from the 1920s to those taken at the present day and a 1951 vertical air photograph with one taken in 1983. The former demonstrate that little change had occurred in the form and position of the peat margin during this period with only small areas around the projecting areas of the margin being lost. Comparison of a second sequence of oblique photographs of the same scene with an 80 years gap indicates considerable re-vegetation of previously bare peat with no increased erosion features. The comparison of vertical air photography between 1951 and 1983 showed no major changes in the pattern and dimensions of dissections systems. The only significant change is some limited re-vegetation and vegetation destruction and erosion along public rights of way. Apart from local adjustments, general changes in gully development are not detectable by this method over these timescales and strongly suggest that gullies must be much older and develop either over long time periods or during periods of greater landform change. Field work carried out in 1999 showed that the gully pattern had not changed from 1983, implying little or no change for 48 years. The authors concluded that blanket mire geomorphology in their study area, the Cheviot Hills, had remained largely unchanged for over 100 years with peat loss occurring steadily but at a rate imperceptible in terms of landform change. This concurs strongly with the observation and conclusion made by McHugh: the spatial extent of erosion is not changing, but it is highly likely that erosion has continued where it was originally initiated. Based on comparisons with the Southern Pennines, the authors speculate that the gully systems on the Cheviot may be at least 500 years old.

LIDAR (Light Detection and Ranging) technology is a remote sensing technique that uses laser light in much the same way that sonar uses sound, or radar uses radio waves. The journey time of the laser beam, from leaving the instrument to its return after

reflection, is measured and – knowing the speed of light – a distance can be computed between the aircraft and the ground. The technology has been used to evaluate peat morphology and hydrology on National Trust Properties (Haycock 2003) and it provides a potentially powerful tool for monitoring our peat resource more widely. LIDAR potentially provides a much more sensitive method of measuring and monitoring peat depths within gully systems and hence rates of removal and also subtle changes in gully dimension. Neither of these have proved possible by conventional remote sensing techniques.

5.5.4 Field measurements of rates of peat loss

There have been rather few systematic studies into the causes of blanket mire degradation in Scotland (Coupar *et al.*, 1997) resulting in a serious deficit of data relating to rates of peat erosion. Birnie (1993) provides the only measured loss rates in Scotland and they range from between 1-4 cm/annum. Measurements of erosion rates were made on bare peat surfaces at two hill top sites in Shetland between 1982 and 1987 and represent the worse case scenario for maximum erosion risk in the Shetland context. Hulme & Blyth (1985) suggest that up to 20 mm of peat can be removed during one storm, so the results produced by Birnie concur, in broad terms, with this estimate. It must be emphasised that this rate of peat loss only refers to situations where the surface of the peat is already bare and where there is no surface vegetation to bind the upper surface. In parts of Shetland, bare peat surfaces can extend to 15% of the blanket peat area (Birnie & Hulme, 1990), so the problem can be locally serious. Nevertheless, the vast majority of the peat resource is not bare and is not actively eroding.

In Wales, rates of peat erosion have only been calculated on Pumlumon (Yeo, 1997). Francis (1990) recorded average rates of 16 mm/yr whilst Robinson & Newton (1986) recorded rates of 30 mm/yr. Francis estimated that only between 19 and 44% of the observed surface recession could be attributed to sediment loss, the majority being caused by peat wastage due to oxidation, shrinkage, consolidation and compaction of the peat. Desiccation during summer was identified as a more important erosive agent than frost heave during the winter.

Other studies have taken place in the Pennines and they all show similar scales of loss to the Shetland and Wales studies. Tallis (1998) states that peat can be lost at rates up to 5cm/year either from the whole surface or along lines of preferred water flow. Here again, losses of this scale can occur only once bare peat is exposed so it is important to maintain vegetation cover to prevent new erosion. Anderson *et al.* (1997) have differentiated between rates of surface-lowering (mm/year) on gently sloping peat flats (28.7, range 16.0 -53.1, n=10), gully sides (11.9, range 7.8 - 20.4, n=5) and gully floors (5.5, range 2.5 - 8.1, n= 3). Warburton (2003) studying the combined effect of water and wind ('wind-splash erosion') on a sparsely vegetated peat surface calculated losses of 0.46 and 0.48 tonnes/hectare in two different years, corresponding to an annual average annual surface lowering of approximately 0.5 mm, equivalent to a surface lowering of approximately 3mm in the field when the density of the surface peat is taken into account. These losses are much lower than those recorded in other studies, but most of those took place in gully situations where the erosive energy of flowing water is much

greater. Warburton's study took place on a 3 ha relatively flat, sparsely vegetated peat with relatively few hags.

Holden *et al.* (2006) have tabulated the results from sixteen studies from England and Wales, including those cited above, that have measured peat erosion rates using erosion pins. The majority (9) report rates between 10 and 30 mm/annum and three report rates below 10. The highest rate of 73.8 mm/annum is based on a one year study. Whilst we cannot be definitive about peat losses from bare peat surfaces due to small scale site heterogeneity and flow patterns and local anthropogenic pressures, losses of between 10 and 30 mm/annum is a reasonable figure based on evidence from throughout the UK. Because of the complexity of the gully systems and the fact that some removed peat is simply re-distributed, it is very difficult to estimate carbon losses with any confidence.

There is considerable qualitative evidence indicating that in recent decades the rate of peat loss has declined for many upland UK sites. Evans & Warburton (2005) have reported a three fold reduction in sediment loss compared to 40 years ago in a small catchment in the Pennines. Similar trends were by Wishart & Warburton (2003) and Clement (2005).

5.5.5 Long-term river records

Data from harmonized monitoring provides useful insights from which long-term trends in net soil erosion losses from catchments may be inferred. Mean net erosion rates within river catchments can be determined from sediment loads carried by rivers. The harmonised monitoring data maintained by the Scottish Environmental Protection Agency (SEPA) includes records of river flow and suspended sediment concentration measured at approximately monthly intervals since the early 1970's. These form the basis of an assessment of changes in net erosion rates over the last 3 decades, but do not quantify soil redistribution within catchments by erosion and deposition.

Figure 5.6 below demonstrate the trends of suspended sediment concentrations in a selection of five catchments where a large proportion of the catchment comprises soils with an organic surface horizon. There is no real evidence of a trend of rising sediment loads in any of the catchments. Each catchment has a few abnormally high readings, but there is no real indication that they are increasing in number or intensity. If any trend is apparent it could be interpreted as one of decreasing sediment loads on the river Nairn, but this could be down to more careful management of the predominantly sandy soils on the lower cultivated land rather than any influence that the more organic soils might be exerting.

The River Findhorn is a predominantly upland catchment in eastern Scotland which was the one of the areas identified by Grieve *et al.* (1995) and in the LCS88 dataset with the most significant evidence of severe peat erosion but even here there is no evidence of a rising trend. These data are imperfect, in that they do not capture every extreme event, they do not differentiate between the mineral and organic carbon particulates within suspended sediments and most of the rivers in the Harmonised monitoring data are relatively large (in a Scottish context) and some of the sediment removed near the head

waters is likely to be redeposited before the measuring gauge near the estuary. More gauging stations are required in subcatchments dominated by organic soils including measurements of particulate soil organic carbon.

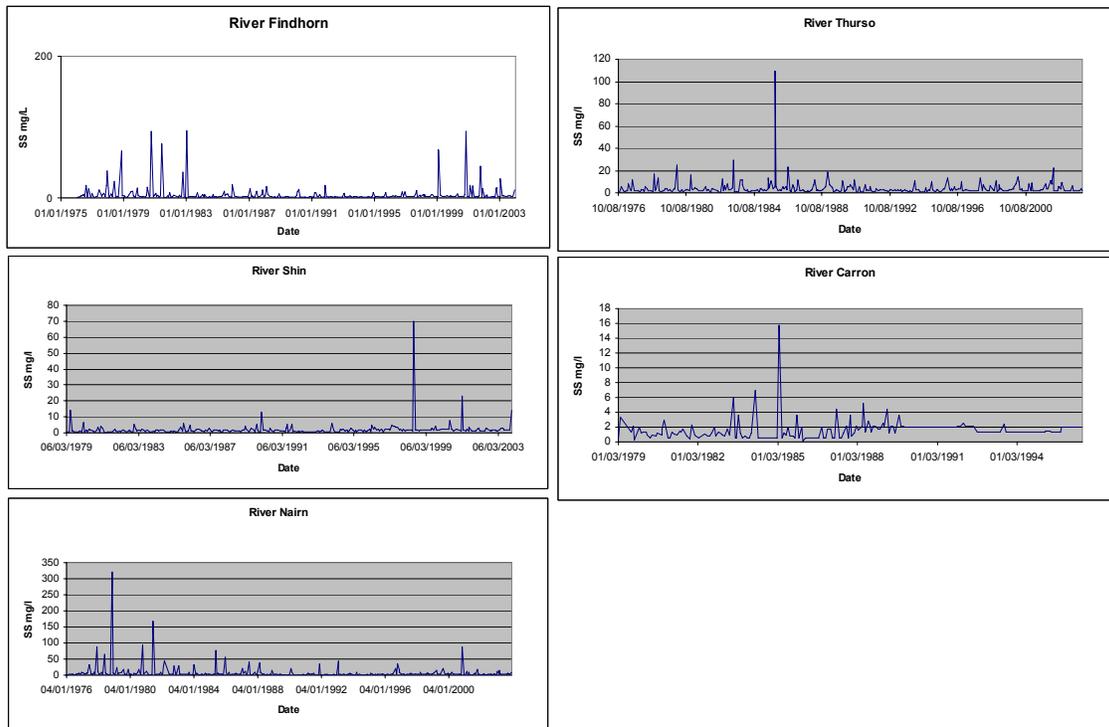


Figure 5.6 Trends in suspended sediment concentrations (SS) from selected catchments.

Other methods include the reconstruction of erosion rates from sediment storage sites such as lakes (Yeloff *et al.*, 2005) or using tracers such as Cs-137 (Walling *et al.*, 2005).

Holden *et al.* (2006) argue that what is missing in all these approaches is an attempt to provide a national picture of peat loss. They recommend a spatial approach to this issue to gain a better understanding of the erosional dynamics of peat systems in different areas each defined by their topography, hydrology and vegetational characteristics.

5.6 Future trends

Given the uncertainty over the initial causes and subsequent factors involved in peat erosion, it is difficult to be too definitive about potential future trends. However, some commentary is offered on the potential future role of the key triggers into the future. There is a much greater environmental awareness of the role of blanket peat in terrestrial carbon budgets and climate change mitigation strategies and there has been significant changes made to policies and to management guidelines on activities that might adversely affect our peatland resource. These are outlined below but there may also be influences that are much more difficult to control.

5.6.1 Burning

If best guidance is adhered to, the Muirburn Code (Scottish Executive 2004), the Heather and Grass Burning Code (MAFF 1994 on behalf of the Welsh Office Agricultural Department), and the 2005 consultation on it, should prevent any future burning of blanket mires or peats. The exception to this is peat where heather constitutes more than 75% of the vegetation cover but these are extremely rare. Heather burning is normally carried out on peaty podzols that support heather-dominant vegetation, and one of the objectives of the guidance is to prevent or minimise damage to the underlying soils

The main future threat to blanket peat from fire are likely to be those related to accidental or deliberate fires started by recreational users. Recreational pressure is growing constantly and although the pressure may be greatest in the Pennines due to the proximity of large urban centres, the value of peatlands for bird watching and a 'wilderness experience' is growing. This may lead to increased visitor numbers even on more remote bogs and an increased risk of wildfire occurrences. Fire risk may also be increased if climate change produces long warm dry spells causing desiccation of the peat surface.

5.6.2 Grazing

There is increasing recognition that many of Britain's uplands have suffered from overstocking by domestic livestock and more recently by wild herbivores such as red deer. In recent years, the number of sheep in Scotland has declined quite dramatically from a peak of almost 10 million in 1991 to 7.6 million in 2006, representing a drop of almost 25%. Similar trends have been found in Wales. In 1993, there were 11, 256,000 sheep, rising to 11, 768, 500 in 1999 but falling sharply to 9,475,000 in 2005, representing a fall of around 20%. Cattle are confined predominantly to enclosed land and mineral soils so equivalent numbers have not been presented. It is very difficult to predict whether these trends will continue but as farming continues to move from an industry focused on production to one that is expected to deliver a range of environmental benefits, certainly any reversal of this trend is highly unlikely.

Estimates of the number of red deer in Scotland are difficult to obtain and often contentious but an estimate of 'wild deer' from the Deer Commission for Scotland's website is 600 000. In their Vision Document (DCS 2000), the estimate of 350 000 red deer has been made. This is lower than a figure of 450 000 made by the RSPB and WWF in a report published in 2003, although it is made clear that this figure is an estimate. The same document indicates that this represents a trebling of red deer numbers over a 40 year time period at the same time as their range has declined in extent due to afforestation. The Deer Commission for Scotland is committed to achieving sustainable levels of deer in the uplands and the best prediction that can be made is that deer numbers will decline, perhaps modestly and slowly, from their current levels. More information can be found at the DCS and WWF websites below.

<http://www.rspb.org.uk/scotland/policy/deer/index.asp> (accessed 8th November 2006)

http://www.dcs.gov.uk/i_faq.htm (accessed 8th Nov 2006)

5.6.3 Acid deposition

Acid deposition has been implicated as one of the causal factors of peatland erosion, in particular in the Southern Pennines and might be one explanation for the severity of the problem in that area. The UK's emissions of SO₂ peaked in 1970 and have declined by almost 70% since 1990 (Fowler 2001). In the context of International transboundary agreements and protocols for acid emissions, this causal factor has decreased considerably and is likely to continue to do so. Although N deposition has yet to show similar reductions, it may as car engine technology improves and the use of N fertilisers decreases (Holden *et al.* 2006).

5.6.4 Climatic factors

Climatic perturbations have been implicated in the past as a causal trigger of peat erosion and a factor in current erosion processes and there appears to be a growing consensus that many of the gully and hag systems seen today are not recent phenomena (say within the last 100 years), but may have been initiated towards the start of the last millennium, long before the bogs would have subjected to the anthropocentric influences since the Industrial Revolution. Prolonged periods (several decades to a century or two) of climatic conditions that differ from the 'norm' over the last millennium, have been identified as key trigger periods for the initiation of peat erosion (Tallis, 1997; Burt *et al.*, 1997). There is also a strong correlation between areas of peat erosion, altitude and current adverse climate (Osvald, 1949; Rhodes & Stevenson, 1997).

Current thinking suggests that we may be entering into a period of unprecedented warming. Current models suggest a possible warming of up to 3.5 °C in summer and 2.5 °C in winter by the 2080s (Hulme *et al.*, 2002 quoted in Scottish Executive, 2006) in a high emissions scenario. In addition, summers are predicted to become drier across Scotland, particularly in the south and east, with more severe extreme rainfall events. Similar scenarios are presented for Wales (Hulme *et al.*, 2002). If this combination of events were to become reality, it is feasible to speculate an increase in the processes observed by Hulme & Blyth (1985) and described by Warburton (2003). Essentially these processes involve the physical removal of exposed, bare dry crusted peat surfaces, often within peat hags but also on more open exposed bare ground by wind and/or water. The combination of a warmer climate causing increased desiccation of the peat surface and an increase in severe rainfall events is likely to be the principal cause in any increase in peat erosion.

This view is supported by Holden *et al.* (2006) who state that the most significant risk to the physical integrity of peatlands in England and Wales is posed by climate change and they also make the same link to previous climatic perturbations. Peat bogs are located where they are largely governed by climatic influences and the major influences that cause physical erosion, i.e. water and wind, are also climatically controlled. Holden *et al.* also suggest that climate change may lead to a decrease in frost frequency and an extended growing season, which should promote revegetation and reduce sediment supply. There are therefore both potentially positive and negative effects of climate change on peat erosion trends and there is huge uncertainty which, if either, might predominate. Bragg & Tallis (2001) also conclude that particular care should be given to

the management of our blanket peat landscapes at the present time of climatic uncertainty.

Intense rainfall events have been cited as the key trigger for bog bursts and slides (Nolan & Birnie, 2006). Most climate change scenarios suggest an increase in the frequency and duration of such events but the actual timing of these is impossible to predict except within short (days and hours) timespans and even then there is no certainty that any burst or slide will actually occur. Based on available evidence, there is potential for an increase in bog bursts and slides in the context of future climate scenarios.

5.6.5 Other factors

Blanket bog is protected under Natura 2000 and many of the major bogs are protected by a number of designations including SSSI, NNR, SAC and SPAs. Such protection should prevent potentially damaging management practices such as overgrazing, injudicious burning, extensive peat extraction (except for domestic use in certain parts of Scotland) and ploughing that may trigger or exacerbate erosion processes. Past practices that may have been making peat more vulnerable to erosion processes such as drainage is no longer being carried out and indeed is being actively reversed in many areas including the extensive peatlands of Caithness and Sutherland. Liming was practiced in the past to increase biomass production for grazing sheep, but is strongly discouraged now for the effect that it has on the blanket bog vegetation. It might have a role in encouraging vegetation growth in peatland restoration projects but as higher soil pH encourages faster organic matter turnover, it should not be used on blanket bogs.

5.7 Conclusions

It is not possible to arrive at a definitive rate of peat erosion due to hydrological and topographical differences between bogs. However, of the studies conducted, losses in the range of between 10 and 30 mm/ha/annum from bare peat surfaces can be taken as a reasonable estimate.

At present most of the factors implicated in peat erosion processes are moving in a direction favourable to reducing the rate of peat erosion. However there is huge uncertainty over the potential impact that a changing climate might have on peat erosion. Current scenarios suggest a warmer climate generally with more extreme conditions; a hot dry spell causing desiccation at the peat surface followed by an intense rainfall event provides one of the key triggers for enhanced peat erosion. Peat erosion has been linked to significant climatic perturbations in the past and there is a consensus that we may be on the cusp of a similar climatic shift at the present time. Past evidence suggests that climate change may have a significant impact on the physical integrity of organic and organo-mineral soils and in particular, on deep blanket peat.

This review has also identified gaps in our understanding and recommends additional research in the following areas:

- An estimate of the extent and location of peat erosion in Wales
- A better understanding of the physical processes that causes peat erosion

- The role of particulate organic carbon in carbon budgets in eroded peatland systems
- A spatial approach to gain a better understanding of the erosional dynamics of peat systems in different areas each defined by their topography, hydrology and vegetational characteristics.
- The potential role of LIDAR technology in monitoring peat erosion rates in a number of different bog systems

5.8 Summary

- Most work has taken place on erosion of blanket mires or peat compared to soils with shallower organic rich surface horizons such as peaty podzols or peaty gleys (organo-mineral soils)
- Peat erosion results from the complex interaction of climatic and anthropogenic influences acting over a long period of time
- Considerable debate surrounds the processes that underpin the development of peat erosion, but there is a growing consensus that it is not a recent phenomenon and that climatic shifts in the last millennium may have initiated the erosion. Other factors now operate to perpetuate the erosion processes
- There have been relatively few studies on the rates of peat erosion from bare surfaces with most having taken place in the Pennines; the few results from Scotland and Wales are reasonably consistent with these. Rates vary according to local circumstances but the majority of studies suggest losses of between 10 and 30 mm/annum. It is not possible from current knowledge or data how this translates into carbon losses due to peat erosion
- There is no hard evidence that the concentration of suspended solids in Scottish rivers have increased between the mid 1970s and the present day.
- Currently most of the factors implicated in peat erosion are such that erosion should halt or decline. The exceptions to this are national and local climatic change that may lead to an increased risk.
- More strategically placed monitoring stations in sub-catchments dominated by organic and organo-mineral soils, would inform us of trends in particulate organic carbon displacement and movement. Separation of the organic and mineral particulate within suspended sediments would indicate whether there was any change in their proportions.
- More studies are required in order to develop an understanding of the mechanics of erosion and the production of a model to predict erosion risk of carbon rich soils.
- Studies on monitoring the extent of peat erosion over time have found the rate of change is too sensitive for the use of aerial photography.
- Lidar technology, coupled with ground truthing surveys, offers potential for future monitoring of peat erosion rates.

Module 6

Options For Mitigating C And N Loss From Organic Soils Used For Agriculture

This module summarises the impacts of various aspects of agricultural management on organic soils, to aid in developing policies to minimise the loss of stored carbon and the emission of greenhouse gases. Where management strategies cause trade-offs between different greenhouse gases, Global Warming Potential (GWP) is used to determine the best practice for minimising climatic effects. This is generally expressed as the amount of carbon as CO₂ which would have the equivalent heating effect over 100 years, and therefore allows comparison of different GHG emissions in terms of their impact on climate change. Practices covered include the management of natural heathland by grazing and burning, and practices associated with using the soils for improved grassland or cropland, such as drainage, fertilisation, and tillage. Good practice guidance is provided for each management practice within the review and then summarised by land use in section 6.2. Conversion to forestry is covered in module 7 of this report.

6.1 Literature review of impacts of agricultural management on organic soils

6.1.1 Introduction

Reports have revealed that around 30 % of Scotland's greenhouse gas (GHG) emissions come from the agriculture, forestry and land use section, of which roughly half originate from land use changes on high carbon soils (Milne *et al.*, 2004). In Wales these figures are lower, with agriculture producing 12.1% of total GHG emissions, and land use change sinks and sources being almost in balance (Baggott *et al.*, 2006). These GHG emissions are now recognised as affecting our climate, posing one of the most serious

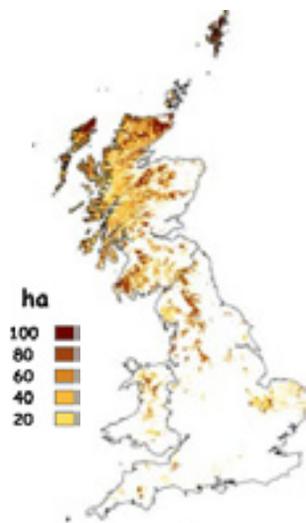


Fig 6.1 Area of organic soils in 1 km grid cells from the DEFRA soil carbon database

environmental threats to the planet. Therefore, issues surrounding the management of these soils is of immediate concern to both UK and devolved policy makers. Organic soils, when underlying natural vegetation such as heathland, tend to be sources of methane (CH₄), sinks for carbon dioxide (CO₂), and neutral as regards nitrous oxide (N₂O). When managed for agricultural purposes however, activities such as drainage and the application of fertiliser, increase the rate of decomposition and produce significant emissions of both CO₂ and N₂O, but also reduce emissions of CH₄, although not sufficiently to counteract the increase in other gases. Assessing the effect of changes in land management on organic soils is a major area of uncertainty in the estimation of GHG emissions.

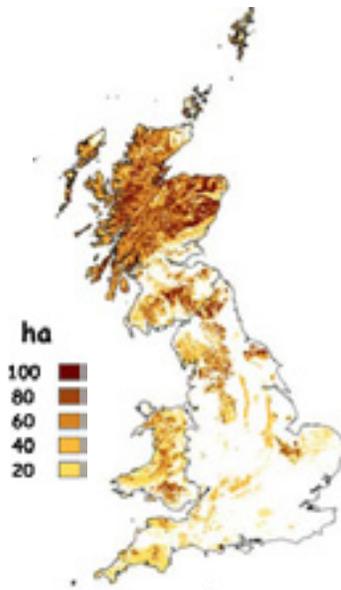


Fig 6.2 Area of organo-mineral soils in 1 km grid-cells from the DEFRA Soil Carbon Database

Organic soils form when decomposition processes are slower than the input of organic matter from plants, allowing a build up of litter and partially decomposed matter. This occurs when conditions are unfavourable for the microbial populations which break down the litter, typically due to a combination of factors such as low pH, and anaerobic conditions due to water logging. These conditions are also unfavourable for agricultural usage, especially the growing of crops, but 'improving' the soil leads to more rapid turnover and tends to result in the loss of stored carbon. The organic soils considered in here include both truly organic soils, such as deep peats, and organo-mineral soils, where a shallower (> 50 cm) but highly organic layer overlays a mineral soil, such as podzols and gleys. While true peats are relatively well understood and conserved rather than converted to forestry or other land uses, organo-mineral soils are widespread and are susceptible to land use change as well as to the impacts of climate change.

In Scotland, organic soils cover 18 000 km² while organo-mineral are more common than mineral soils, covering 32 000 km² (38 % of total land cover) (Bradley *et al.*, 2005). Recent revision of estimates of carbon stocks stored within these soils has indicated that carbon in deep peat has been overestimated in the past, but in the 0-30 cm and 30-100 cm soil layers, carbon content has been underestimated by 320 Tg and 261 Tg respectively, representing 22 % and 20 % of the newly calculated total stocks at these depths (Bradley *et al.*, 2005). In Wales, organic and organo-mineral soils are less extensive, covering 3% and 17.3 % of the land surface, and total revised stocks amount to 396 Tg (Bradley *et al.*, 2005, Module 1). Appropriate management of these soils to retain soil carbon levels is becoming increasingly important, as research into carbon levels in soils in England and Wales shows that soils with carbon content greater than 100 g kg⁻¹ lost carbon at an average rate of more than 2 % yr⁻¹ between 1978 and 2003, and suggests a link to climate change among possible reasons (Bellamy *et al.*, 2005). It is however, important to note that these represent total losses and it is unclear what proportion is lost to water as dissolved or particulate carbon, as opposed to gaseous losses to the atmosphere.

Large areas of organic and organo-mineral soils have been drained for agricultural usage over the years. In the UK, peat drainage was at it's height between the 1940s and 1970s but still continues today on a smaller scale (Holden *et al.*, 2004), although the UK GHG inventory assumes no fenland (lowland) drainage has taken place since 1990 and that emissions due to drainage continue only in England (Milne *et al.*, 2004). Land use change matrixes developed for this project (see Module 8) indicate that land use change has slowed since the 1980s but the most significant change on these soil types has been change to cropland and improved grassland on the East coast of Scotland, change to improved grassland throughout Wales, and change to cropland in South West Wales (see below). Current land use can also be a problem.

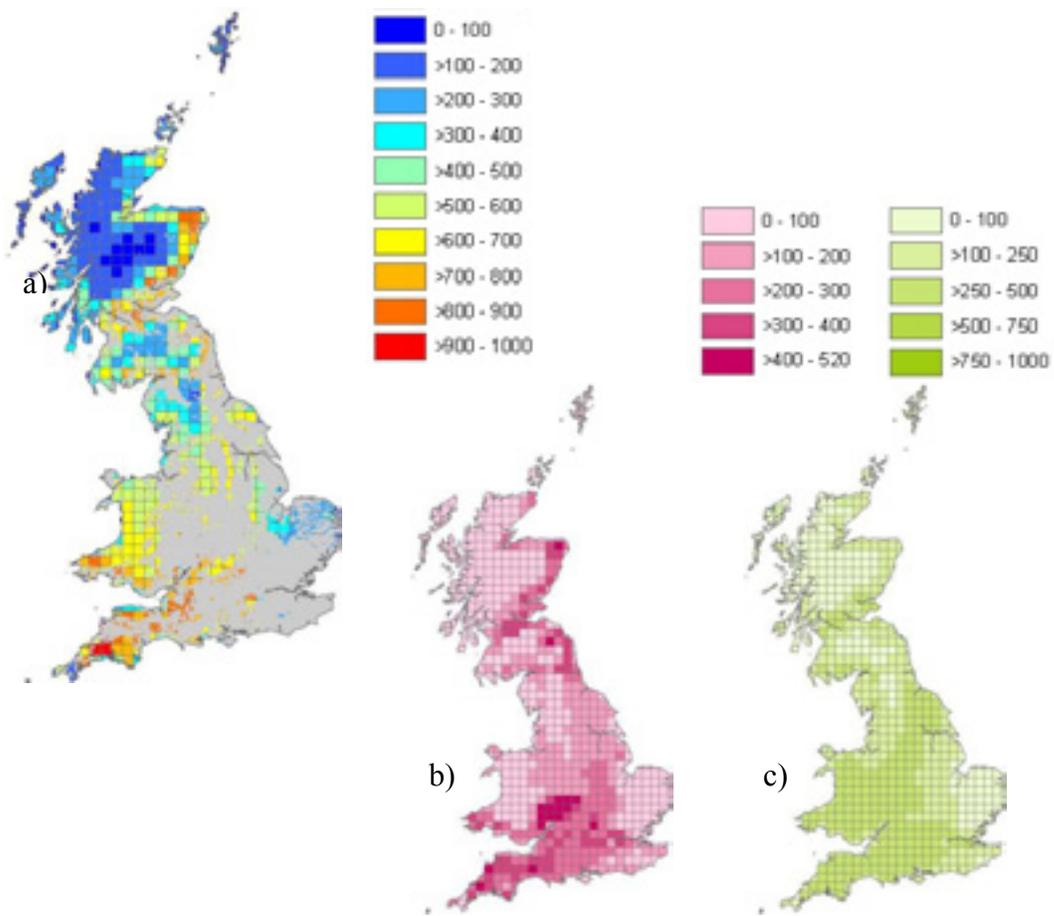
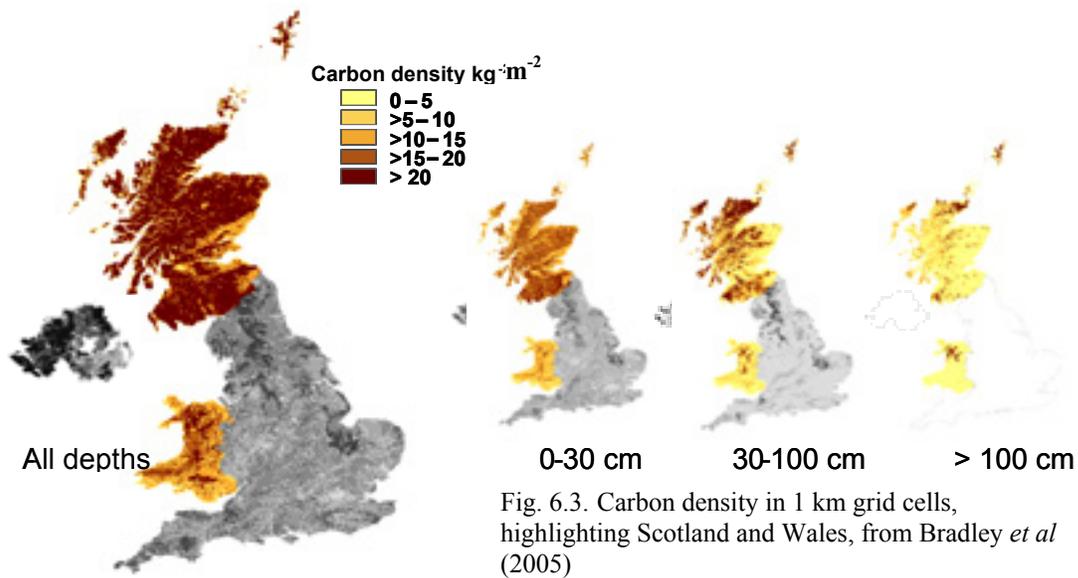


Fig. 6.4 Land use change in the UK a) total LUC 1980s to present on organic and organo-mineral soils (rainbow colours) b) change to cropland 1990s and 2000s (purple) and c) change to pasture (improved grassland) 1990s and 2000s (green). All units ha/decade/20 km square. (see module 8 in this report)

For example, the South West of Scotland has been identified as a potential hotspot of N₂O (GHG) emissions due to high concentrations of cattle grazed, fertilised pastures combined with a mild wet climate (Flynn *et al.*, 2005). The presence of large carbon stocks in the local soils mean any mitigation methods should be tailored to more organic soils. In Wales, concern has been raised about overgrazing degrading montane vegetation and reducing soil carbon stocks (Britton *et al.*, 2005), while in Scotland, burning of heathland may be on the increase (W. Towers, pers. obs.) with potentially damaging consequences.

6.1.2 Agricultural Management Practices

Drainage: The rate of organic matter decomposition is dependent on soil water conditions (Goncalves and Carlyle, 1994; Leiros *et al.*, 1999; Rodrigo *et al.*, 1997) and therefore altering the natural water table position in an organic soil has knock-on effects on carbon and nitrogen cycling and associated emissions of greenhouse gases. While natural wet peatlands tend to be sources of CH₄ emissions, drainage leads to increased loss of carbon and a greater contribution to total GHG emissions and climate change. Decomposition is faster above the water table and lower water tables have been shown to enhance CO₂ production in a number of studies, including laboratory studies (Moore and Dalva, 1993), drained peatlands (Silvola *et al.*, 1996), natural variability in water table position (Alm *et al.*, 1997; Kim and Verma, 1992) and drier than average years (Bellisario *et al.*, 1998; Carroll and Crill, 1997). In Germany, where a high proportion of peatlands have been drained for agricultural usage, peat mineralization has been reported to be highest where the water table is 0.8-1.2 m below the surface during the growing season (Hoper, 2002).

The position of the water table controls how much of the soil profile is anaerobic, as oxygen is quickly depleted in the lower waterlogged portion. Anaerobic conditions are perhaps the most important factor in reducing the rate of decomposition to allow the formation and maintenance of highly organic soil layers. This reduced rate of decomposition is illustrated by measurements of carbon dioxide emissions from peatland soils, which have been reported to be around 2.5 times lower under anaerobic than aerobic conditions in short-term laboratory incubations (Bridgham and Richardson, 1992; Moore and Dalva, 1997), and much lower, an average of 14 times less, in long-term incubations, suggesting anaerobic CO₂ production is inhibited by the accumulation of repressing agents (Magnusson, 1993).

The effect of a drop in water table depth may however, be dependent on the usual conditions of the system. Laiho (2006) notes that the increase in CO₂ emission generally tails off when the water table drops below a certain depth (Chimner and Cooper, 2003; Silvola *et al.*, 1996) and argues that deeper layers may lack easily oxidizable labile C (Chimner and Cooper, 2003; Hogg *et al.*, 1992). Peatlands where the water table is generally 20 cm or more below the surface over the summer will already have been exposed to aerobic decomposition for extended periods, leaving only more resistant organic matter (Bridgham and Richardson, 1992) and therefore further drawdown due to drainage may not lead to increased C turnover. In contrast, peatlands such as wet fens which are generally continuously inundated, may exhibit strong responses to a drop in

water table as there may be a large pool of labile C which has not been decomposed due to the waterlogged, anaerobic condition.

This clearly demonstrates that drainage, particularly of soils which are generally inundated, increases CO₂ emissions and leads to a loss of stored carbon. However, oxidizing decomposition processes are not the only processes involved in nutrient cycling and GHG emission production in organic soils. Methane (CH₄) is produced by anaerobic fermentation processes below the water table and is either emitted from the soil profile, or oxidized into CO₂ by aerobic methanotrophic bacteria above the water table. Therefore, when the water table is below the soil surface, oxidation becomes a major controlling variable for methane efflux (Christensen *et al.*, 2000) and a lower water table decreases methane emission (Blodau *et al.*, 2004). This decrease in emissions with increasing water table depth has been reported using soil cores incubated in the laboratory by a number of workers (Daulat and Clymo, 1998; MacDonald *et al.*, 1998; Moore and Dalva, 1993) but has also been confirmed using field measurements over a peat wetland in Caithness (Hargreaves and Fowler, 1998). Wetlands are not always a significant source of CH₄ however; where the water table is above the surface, oxidation can also reduce methane emissions, particularly if light availability allows benthic photosynthetic activity (Le Mer & Roger, 2001). The highest emissions of CH₄ in temperate systems are associated with wetlands with significant reed or sedge cover as these plants have stems which can transport the gas to the surface and reduce the likelihood of it being oxidised before it reaches the atmosphere. Indeed, sedge cover has been identified as the single most important single variable affecting CH₄ flux in several studies (Bellisario *et al.*, 1999; Nilsson *et al.*, 2001; Schimel, 1995; Tuittila *et al.*, 2000; van den Pol-van Dasselaar *et al.*, 1999).

Although CH₄ is a potent greenhouse gas, it makes up a small proportion of total C cycling. CH₄ flux generally accounts for between 1 and 4.7 % of carbon respired, and <2 to 7 % of net primary production in northern wetlands (Christensen *et al.*, 1996). Other workers have reported that around 2 % of total C cycling in a high arctic fen (Friborg *et al.*, 2000), and 4 % of net annual CO₂ assimilation and 9 % of net carbon fixation from a reed-dominated wetland in Denmark (Brix *et al.*, 2001), is returned to the atmosphere as CH₄. Whiting & Chanton (1997) report that for wetlands in general, the ratio of methane released to annual net carbon fixed varies between 0.05 and 0.13 on a molar basis. While draining peats would reduce CH₄ emissions and may result in producing net sinks for CH₄ (Blodau & Moore, 2003; Huttunen *et al.*, 2003; Maljanen *et al.*, 2002), these are generally very small, particularly for upland soils (Le Mer and Roger, 2001), and it would not reduce total carbon emissions. Measurements from sites across Europe suggest that ombrotrophic bogs drained for forestry or peat cutting have an average Global Warming Potential (GWP) of 1253 CO₂-C equivalents kg ha⁻¹ yr⁻¹ in comparison with an average GWP of 192 CO₂-C equivalents kg ha⁻¹ yr⁻¹ (entirely due to CH₄ emissions) for undrained sites with natural vegetation (Byrne *et al.*, 2004).

Nitrogen dynamics are also affected by soil water conditions. The greenhouse gas nitrous oxide (N₂O) is formed through both nitrification and denitrification, and high N₂O

emissions often only occur when soil water content is high or immediately following rainfall (Aulakh *et al.*, 1984; Skiba *et al.*, 1992; Hansen *et al.*, 1993; Smith *et al.*, 1998; Dobbie *et al.*, 1999; Ruser *et al.*, 2001). These peak emissions are generally associated with denitrification which, as an anaerobic process, becomes dominant at higher water contents, generally reported to be above 60-65 % WFPS (water-filled pore space) (Clayton *et al.*, 1997; Lin and Doran, 1984). In a peat soil, total N₂O emissions have been reported to be nearly constant at 40-80 % WFPS but significantly higher at 100 % (Pihlatie *et al.*, 2004). Peat soils have also been shown to emit more N₂O in drier conditions than sandy loam soils, but less when wet; in dry soils (40% and 60 % WFPS), N₂O production was an average of 2.0 % (peat), 0.3 % (loamy sand) and 0.5 % (clay) of that produced in wet soils (80-100% WFPS), where production was highest in the loamy sand and lowest in the clay (Pihlatie *et al.*, 2004). However, N₂O production is a complex balance of processes with different drivers and when nutrient levels are not limiting, drainage does tend to increase emissions in general, while nutrient-poor organic soils tend to be very small sources of N₂O even when drained (Byrne *et al.*, 2004; Martikainen *et al.*, 1993).

This is a pertinent issue, particularly in the South West of Scotland where managed grassland is a common land use and organic soils are more prevalent than in the arable areas of the East coast. This area has already been identified as a hotspot for N₂O emissions which may increase further as the climate is predicted to become wetter and milder, and drainage has been recommended to mitigate this (Smith *et al.*, 2004). For organic soils, since drainage will reduce C storage and increase CO₂ emissions, N₂O emissions could be reduced by reducing N inputs instead, since production processes rely on available ammonium or nitrate (see fertilisation section below).

As well as affecting soil biochemical processes, drainage may also have a physical impact on C stocks by increasing vulnerability to erosion. Subsidence rates of 2-4 cm per year have been reported for drained fens in England (Bradley, 2000; French and Pryor, 1993; Hutchinson, 1980). Increases in decomposition may also lead to increased losses of dissolved organic matter (DOM) (Tipping *et al.*, 1999) which can lead to problems when it colours drainage water used for water supply. Particulate losses of carbon from the development of drainage pipes is also a significant problem, Holden (2006) calculates that the loss from a peat slope drained for 40 years would be sufficient to approximately halve the size of the net carbon sink of a healthy blanket peat catchment. Modelling studies have suggested that although restoring a high water table increases CH₄ emissions, it may reduce total GHG emissions from peat soils (van Huissteden *et al.*, 2006). This is supported by measurements from European sites suggesting that restored bogs have a GWP of 517 CO₂-C equivalents kg ha⁻¹ yr⁻¹ less than those drained for forestry or cutting, and restored fens have a GWP of 368 CO₂-C equivalents kg ha⁻¹ yr⁻¹ less than those drained for forestry (Byrne *et al.*, 2004). Blocking old drains may also be worthwhile as research suggests that subsurface piping increases over time causing particulate carbon loss from drained peat slopes to increase exponentially (Holden, 2006).

Good Practice Guidance for drainage of organic soils

- Any new drainage of organic soils should be avoided
- Existing drains should be blocked to reduce erosion, especially in catchments with reservoirs where colour in drainage water is a problem. Resources for this should be focused on slopes where the drainage is oldest.
- Maintaining as shallow a water table as possible should be encouraged.
- Where drainage is a necessity, areas where the water table is generally 20 cm or more below the surface in the summer should be drained in preference to constantly waterlogged areas
- Drainage should not be used to mitigate N₂O emissions on non-mineral soils, instead options such as reducing N inputs and grazing intensity should be explored

Grazing: Grazing is probably the most common land use on organic soils in the UK and its effects are dependent on livestock type and stocking density. Direct impacts on soil are caused by a combination of trampling and nutrient addition via deposition of dung and urine. Trampling effects tend to be concentrated around fence lines and feeding stations, are more pronounced with larger, heavier animals, and greater stocking densities, and are also exacerbated by wet conditions (Shaw *et al.*, 1996). Trampling has been shown to stimulate denitrification due to reduced soil aeration and plant N utilisation (Menneer *et al.*, 2005) and may therefore increase N₂O emissions, especially under wet conditions. Under dry conditions, trampling disturbs the soil and may act like tillage to increase aeration and stimulate decomposition and associated CO₂ emissions. Both of these effects may be amplified by the addition of nutrients. It is difficult to quantify soil emissions resulting from grazing however, as field measurements are usually taken using chambers and grazing animals are often excluded from field sites to prevent damage to equipment, and also the effects of nutrient deposition are highly localised. Data collected for this study indicates that N₂O emissions from organo-mineral soils are highest where grazing pressure is highest. Effects on CO₂ emissions are even harder to pin down as net ecosystem exchange (NEE) is usually measured and this includes plant and rhizosphere exchange from respiration and photosynthesis as well as soil respiration from decomposition processes (see for example, Nieveen *et al.*, 2005).

Overgrazing can cause severe degradation, not only damaging vegetation but also reducing soil organic matter and removing upper organic layers. In Wales, there was a 71 % increase in sheep numbers between 1974 and 1998 (Countryside Council for Wales and Forestry Commission, 1999) and an upland area of North Wales has seen an increase in grazing density from around 1.2 sheep ha⁻¹ in the 1950s to an average of around 5-6

sheep ha⁻¹. This has had a detrimental effect on the ranker and peaty podzol soils, with degraded areas containing significantly less carbon and nitrogen, means of 5 % C and 0.4 % N in comparison with 24-27 % C and 1.1-1.4 % N in intact heathland ecosystems at the same site. Acidity has also significantly reduced (Britton *et al.*, 2005). These effects can be considered typical of sustained heavy grazing pressure (Milne *et al.*, 1998; Rudeforth *et al.*, 1984). In recent years, numbers of sheep in both Scotland and Wales have declined. In Scotland however, red deer numbers have increased sharply, although estimates are subject to debate, and it may become increasingly important to control grazing pressure by taking account of these wild grazers as well as sheep numbers.

Properly managed for site conditions however, grazing can be beneficial. There is some evidence that carefully managed grazing can aid carbon storage in organic soils. Studies of the Moorhouse NNR in the Pennines have indicated that grazed plots tend to accumulate carbon slightly faster than ungrazed areas and certainly light grazing did not cause any reduction in C accumulation in comparison with no grazing (Garnett *et al.*, 2000). Grazing effects can also be minimised by not allowing animals access all year round, for example by moving them to pastures on mineral soils in winter when wetter conditions will exacerbate trampling effects. This has been shown to help maintain heathland vegetation (Grant *et al.*, 1982). It should also be recognised that most heathlands can support higher stocking densities during summer than in winter. Finally, different types of grazing animals may have different preferences for different types and ages of vegetation and can therefore be selected on the basis of the type of plant community land managers wish to promote.

Good Practice Guidance for grazing of organic soils

- Overgrazing causes significant damage to soil organic layers, reduces carbon storage and increases greenhouse gas emissions so stocking densities need to be carefully managed to minimise this, while limiting succession to protect vegetation communities.
- Heavier animals such as cattle should only be used in very limited numbers, where their less selective feeding will aid vegetation management, and not on wetter sites where they are more likely to cause significant damage.
- Stocking densities should be reduced in winter or animals removed completely, particularly on wetter sites.
- Stocking densities must take into account grazing pressure from wild animals such as deer and a reduced number of domestic livestock should be used on sites where deer are numerous.
- Deer control measures should be concentrated in areas with organic soils which are at risk of overgrazing.

This is discussed in some detail in a report produced for English Nature, which describes dietary preferences and reported effects of cattle, sheep, ponies, goats, deer and other grazers, such as grouse and mountain hares (Shaw *et al.*, 1996). Cattle may have a role to play as they tend to graze less selectively than other grazers, however, this needs to be balanced against the increased risk of trampling from these heavier animals, and they also deposit more nutrients through dung and urine than smaller ruminants.

Burning: Burning is a traditional management technique for maintaining open shrub (usually heather) or grass dominated vegetation in upland areas and preventing succession to scrub and woodland. It is however, a source of CO₂ emissions to the atmosphere. An investigation into wildfire on peatlands in North Carolina using remote sensing reported that carbon emissions ranged from 0.2-11 kg C m⁻² and total emissions were between 1 and 3.8 Tg (Poulter *et al.*, 2006). Burning also has long term effects on soil properties; investigations into a drained peatland in Turkey have reported significantly reduced soil organic carbon, even in areas which were burnt as long ago as 1965 (Dikici and Yilmaz, 2005). Organic matter accumulation at the Moorhouse NNR in the Pennines has been reported to be lower for plots which have been burnt every 10 years than for those which have not been burnt since 1954 (Adamson, 2003; Garnett *et al.*, 2000), and data from a Finnish mire also suggests that frequent fires reduce carbon accumulation (Pietikäinen *et al.*, 1999).

Organic matter may be physically broken down by burning but evidence also suggests that further mineralization may occur after the fire due to increased microbial activity; microbial respiration has been reported to be three times higher following burning, in response to higher nutrient and substrate levels in remnant soils and enhanced soil temperature (Kim and Tanaka, 2003). There is also some evidence that burning increases the pH of organic soils, which would also favour increased rates of decomposition (Allen, 1964; Stevenson *et al.*, 1996). Measurements taken in the Everglades to compare soil properties pre- and post-peat fire also confirm that total C and N contents are significantly reduced and not just by physical processes (Smith *et al.*, 2001).

As burning removes vegetation, it also increases vulnerability to physical erosion, especially as charred peat can form loose crusts which are easily washed away, and increases losses of DOC (Tucker, 2003). This erosion is arguably the most important impact of burning on peatlands, and is most severe when all the vegetation cover and root matt is removed, leaving bare soil exposed to sun, wind, water, and in winter, more intense freeze-thaw cycles. A study of post-fire erosion of podzols and peaty gleys in the North Yorkshire Moors concluded that vegetation cover was the main determinant of erosion rates and reported relatively slow rates where the burnt remains of heather made up the ground surface. More severe burning which exposed the peaty or mineral subsoils however, caused much more rapid erosion, and in winter (September to April) erosion rates on bare soil were up to 10 times higher (Imeson, 1971). This is particularly significant as current UK law only allows for moorland burning between October and April to reduce the likelihood of dry peat being consumed along with the heather stand. Vegetation recovery limits erosion (Kinako and Gimingham, 1980) and it has been suggested that heather recovers better from autumn rather than spring burns (Miller and

Miles, 1970; Phillips, 1991), although this may not be the case if regeneration is dependent on seedling establishment (Tucker, 2003). Autumn burning is generally considered to be less intense due to greater moisture content in vegetation and litter (Hobbs and Gimingham, 1987) but burning after a wet period, post thawing should limit fire severity and avoid the worst effects of erosion. The best way to avoid detrimental effects from burning however, is not to use it at all, as heather cover in many cool, moist upland European heathlands can probably be maintained without burning (MacDonald *et al.*, 1995) provided grazing is used to prevent succession to woodland (Marrs, 1988).

Good Practice Guidance for burning on organic soils

- Serious consideration should be given to halting the practice of burning on soils with significant stores of carbon, and the use of grazing alone to control community succession and encourage vegetation renewal, although this may require reductions in grazing density.
- Burning should not be allowed in areas with a high risk of erosion such as exposed areas or areas with extensive drainage gullies or ditches.
- Fires should be as small and controlled as possible so that only vegetation is consumed and litter and root mats remain to protect the soil from exposure
- The burning season should be confined to after wet periods in early spring, towards the end of the current legal season, as winter freeze-thaw cycles increase the risk of erosion when soil is no longer protected by vegetation.

Fertilisation: Fertilisation is generally considered to aid carbon storage in mineral soils (Gregorich *et al.*, 2005). In organic soils however, this assumption may not hold true. Tundra organic soils which had been fertilised annually for 8 years showed different effects depending on ecosystem type, with moist acidic soils from between sedge tussocks containing significantly less carbon and nitrogen in comparison with unfertilised plots, soils from beneath the tussocks containing significantly more carbon but no significant effect on nitrogen, and the drier heath soils with a shallower organic layer showing no significant effect of fertiliser on carbon or nitrogen stocks (Shaver *et al.*, 2006). In general however, fertiliser would be applied to drier organic soils rather than waterlogged bogs, and added nutrients combined with aerobic conditions accelerate decomposition and increase CO₂ emissions (Byrne *et al.*, 2004). This effect may be particularly acute when lime is also applied, making conditions more favourable for decomposition as well as supplying extra nutrients.

Emissions of N₂O have a more straight forward relationship with fertiliser inputs. Like mineral soils, farmed organic soils have been shown to emit high levels of N₂O

temporarily after fertilisation (Augustin *et al.*, 1998), especially in wet conditions and early in the growing season when vegetation competes less for the available N (Kettunen *et al.*, 2005; Nykänen *et al.*, 1995). However, emissions cannot be calculated from fertiliser N input in the same way as for mineral soils, as existing stocks of N in the soil mean that emissions are not directly related to input levels (Mosier *et al.*, 1998). Different types of fertiliser may also have different levels of impact; the application of solid manure has been reported to produce substantially lower N₂O emissions than mineral fertilisers or liquid manure (Gregorich *et al.*, 2005), while sewage sludge pellets and poultry manure resulted in emissions 12-26 times larger than mineral NPK when applied to a Scottish grassland (Jones *et al.*, 2005).

Aerobic farmed organic soils have been reported to be negligible sources or sinks of CH₄ with low production and oxidation potentials (Kasimir-Klemedtsson *et al.*, 1997; Kettunen *et al.*, 2005) and are therefore largely neutral in terms of the CH₄ budget. There is some evidence however, that fertilisation may reduce any small oxidation potential they may have, as Crill *et al.* (1994) report that inorganic NH₄⁺ can inhibit CH₄ oxidation in drained peat soil.

Good Practice Guidance for fertiliser usage on organic soils

- Fertiliser applications should be kept to a minimum, especially when the soil is also limed.
- Applications should be carefully timed to avoid wet conditions as far as possible and ensure that nutrients are applied when the vegetation is in a growth phase and can make best use of them.
- The application of solid manure is preferable to either mineral fertilisers or liquid manure/slurry.
- Where applicable, nutrient input from manure and urine deposition by grazing animals should be accounted for when assessing further nutrient requirements.

Liming: Many organic soils are naturally acidic and this is generally considered to limit the activity of decomposers, which favour a neutral environment, aiding the build up of organic matter. A significant effect of soil pH on decomposition has been observed in many studies (eg. Andersson and Nilsson, 2001; Hall *et al.*, 1997; Situala *et al.*, 1995) and there is a positive correlation between soil biological activity levels and soil pH (Motavalli *et al.*, 1995). Acidic irrigation studies have shown that reduced pH leads to reduced CO₂ fluxes; Situala *et al.* (1995) reported that pH 3 produced CO₂ fluxes 20 % lower than those at pH 4 and 5.5, and Persson and Wiren (1989) reported increasing the acidity of forest soil from pH 3.8 to 3.4 reduced CO₂ production by 83 % and from pH 4.8 to 4 by 78 %. This clearly suggests that increasing pH towards neutral by the addition

of lime to a naturally acidic soil will lead to an increase in CO₂ emissions and a reduction in C stocks. This is supported by a study of carbon flow in an upland grassland, which showed that liming caused more rapid C turnover (Rangel-Castro *et al.*, 2004).

Soil pH may also have a varying impact depending on aeration and water logging. Production rates of CO₂ have been shown to increase more with increasing pH under anaerobic conditions; Bergman *et al* (1999) compared CO₂ production rates at pH 4.3 and 6.2, and found that under anaerobic conditions rates were 21- 29 times greater at the more neutral pH (depending on temperature), while under aerobic conditions rates were 3 times greater at 7°C but pH had no significant effect at 17°C. This suggests that liming will have a greater impact on wet organic soils.

Just as neutralising low pH tends to increase CO₂ production, as a general rule it also increases rates of organic N mineralization, ammonification, nitrification and denitrification. Cumulative N₂O emissions have been shown to be positively correlated to pH between 4.14 and 7.53 in arable soils incubated under flooded conditions (Wlodarczyk *et al.*, 2002) and experiments carried out at the Rothamsted Experimental station have indicated that lime treated grassland plots have significantly higher N₂O emissions; Yamulki *et al* (1997) reported that emissions from a pH 3.9 plot were around 36 % of those from a pH 7.6 plot. In the short term, there is evidence that neutralising a naturally acidic soil may temporarily reduce N₂O emissions as the indigenous microbial community is adapted to the acidic conditions. For example, Yamulki *et al* (1997) found that treating a pH 3.9 soil to increase the pH to 7 reduced N₂O emission rates by more than 40 % and Brumme and Beese (1992) found that liming from pH 4.5 to 6.5, reduced N₂O emissions by 74%. However, in the longer term, the microbial population will adapt to the new conditions and organic agricultural soils generally emit more N₂O than mineral agricultural soils (Duxbury *et al.*, 1982; Martikainen *et al.*, 2002).

More neutral conditions may also favour CH₄ production as methanogenic bacteria are generally reported to exhibit maximum activity under neutral or slightly higher pH conditions (Garcia *et al.*, 2000). An investigation of CH₄ production in peat soil samples from temperate and subarctic areas (pH 3.5–6.3) reported an optimum pH of 5.5 to 7.0 (Dunfield *et al.*, 1993), and a positive correlation between CH₄ production activity and soil pH has been reported for peat soil samples from a temperate wetland with a pH range of 5-7 (Inubushi *et al.*, 2005). Experiments which increase the pH of peat samples towards neutral have also shown an increase in CH₄ production;

Williams and Crawford (1984) found that a pH increase from 3.2 to 5.8 increased the methane production of an incubated peat from a Minnesota peatland by 1.5-2.2 fold, depending on depth, and Murakami *et al* (2005) also report that liming acid peats increases CH₄ production potential. However, this increase in CH₄ production activity may not necessarily lead to increased CH₄ emissions, as more neutral conditions may also favour CH₄ oxidising bacteria (methanotrophs), which have a reported optimum pH of 5.0 to 6.5 in temperate and subarctic peats (Dunfield *et al.*, 1993). Hutsch *et al* (1994) reported that, in a non-fertilised permanent grassland at the Rothamsted experimental station, a decrease in pH from 6.3 to 5.6 reduced methanotrophy by almost half. In

practice, limed soils are generally drained agricultural soils and therefore either slight consumers or low sources of CH₄ due to their aerobic condition.

A final issue is the impact of liming on the production of DOC (dissolved organic carbon). An increase in the export of DOC in surface waters has been observed across much of Europe and North America over the past twenty years and the causes of this are still a subject of debate, but it has been suggested that declining acid deposition may be a major factor (Evans *et al.*, 2006) because DOC solubility is suppressed by high soil water acidity. Experiments have shown that liming can increase the concentrations of organic matter, DOC and DON in soil water (Andersson *et al.*, 1994, 1999; Curtin and Smillie, 1983). Although loss of organic matter in surface water is an issue nationwide, the management of upland organic soils may play a disproportionately large role in determining the future scale of this C loss. This is because the impact of a change in pH on DOC solubility is greatest in the pH 4 to 5 range (Thurman 1985; Peterson 1990), which is typical of upland organic soils in Britain, and titration experiments have indicated that an increase in soil water pH of 0.5 units could cause a 50% increase in DOC (Tipping and Woof 1990). This has important implications for Wales and Scotland where liming to mitigate historic acidification is underway and may have serious knock-on effects on water quality.

Good Practice Guidance for liming of organic soils

- Further lime additions should be avoided if possible, especially on wetter organic sites.
- Reducing liming in catchments which drain into water supply sources may reduce problems with 'colour' due to particulate organic matter.
- Soils in the pH 4-5 range may be most sensitive to a change in pH in terms of subsequent loss of carbon and therefore liming them, or liming more acidic soils up to this pH range, should be discouraged.

Tillage: The impacts of tillage on soil organic matter (SOM) vary depending on cropping system, residue management, and climate, but in general, it promotes decomposition through crop residue incorporation into soil, physical breakdown of residues, and disruption of aggregates protecting SOM (Baldock and Skjemstad, 2000; Paustian *et al.*, 2000; Six *et al.*, 2000). Microbial metabolic activity has been shown to increase in response to residue incorporation and improved aeration; Balota *et al* (2004) reported that, averaged across crop rotations and depth, conventional ploughing increased metabolic quotient (soil respiration per unit of microbial C) by 32 % compared with no ploughing. Impacts are however, generally confined to the surface layers, which are physically disturbed. Several workers have reported that tillage reduces soil organic carbon and nitrogen in the surface layers. Salinas-Garcia *et al* (2002) found that SOC in the top 5 cm was between 36 and 57 % lower under conventional tillage (disk ploughing

to 30 cm) compared to conservation tillage (no ploughing or disking to 10 cm), which retained crop residues at the surface, reducing contact with the decomposer community, and Wright *et al* (2005) also reported that tillage reduced SOC and SON at the surface, although the effect was not significant under all the cropping regimes.

Impacts are also dependent on the type of ploughing carried out. Moldboard ploughing is generally reported to have the most significant impacts, followed by conventional or disk ploughing, and with conservation ploughing or disking being most similar to no tillage regimes (Wright *et al.*, 2005). Deeper disturbance generally corresponds to greater overall impact, although shallower mixing may result in greater impacts at the surface: Friedal *et al* (1996) compared the effects of ploughing to a depth of 25 cm with rotary cultivation to a depth of 10-12 cm, and reported that in the upper soil layer (0-10 cm) rotary cultivation caused a greater increase in C and N mineralization potential than ploughing.

Stimulating microbial activity also leads to an increase in N₂O emissions. Although it generally improves aeration, Jackson *et al* (2003) found that tillage significantly increased denitrification from day 2 to a week after tillage, and cumulative N₂O fluxes have been reported to increase by 5 to almost 12 times, depending on crop type, in response to ploughing to 15-20 cm depth (van der Weerden *et al.*, 1999). Again, shallow tillage practices such as rotovation may lead to lower N₂O emissions than traditional ploughing, but both stimulate emissions to some extent.

Soils which are under agricultural management and suitable for ploughing are generally net consumers of CH₄ due to their aerobic surface layers, and therefore tillage does not affect CH₄ emission levels. Agricultural practices may affect their ability to oxidise atmospheric CH₄ (Le Mer and Roger, 2001) but the size of this flux is insignificant in terms of total GHG emissions.

As previously discussed (see burning section above), organic soils are more prone to erosion when bare in winter, and freeze-thaw cycles also increase N₂O emissions (Kaiser *et al.*, 1998; van Bochove *et al.*, 2000), and therefore timing of ploughing is also an important issue, as leaving stubble over winter will partially mitigate these effects.

Good Practice Guidance for tillage on organic soils

- Conservation or zero-till regimes should be encouraged
- Deep ploughing should not be allowed on soils with high carbon contents
- Winter ploughing should be avoided to reduce erosion risk and effects of freeze-thaw cycles on bare soil, and instead where necessary, ploughing should be carried out as close to new crop sowing as possible to minimise soil exposure.

Conversion to grassland: Conversion to grassland, comprising drainage to a depth of around 0.4-0.8 m below the surface (Joosten *et al.*, 2002), removing natural vegetation and replacing it with productive grass species, and moderate to high levels of fertiliser addition, has been assessed as causing fast loss of stored carbon (Byrne *et al.*, 2004). Conversion of a peat bog to pasture for dairy farming has been reported to reduce carbon storage by an average of 3.7 t ha⁻¹ yr⁻¹ over 40 years (Schipper and McLeod, 2002). Studies of boreal organic soils used for grassland in Finland have reported annual net carbon losses of 750 g CO₂-C m⁻² (Maljanen *et al.*, 2001), and 3.3-4.6 t CO₂-C ha⁻¹ yr⁻¹ (Maljanen, 2003), while in the Netherlands, carbon loss of 300 g CO₂-C m⁻² has been reported for pasture (Langeveld *et al.*, 1997).

Based on flux measurements from European sites, GHG emissions from grassland on nutrient-poor, ombrotrophic bog soils are estimated to have a GWP of 2367 CO₂-C equivalents kg⁻¹ ha⁻¹ yr⁻¹, based on CO₂ emissions of 1.5-3.5 t C ha⁻¹ yr⁻¹, CH₄ emissions of 2 kg C ha⁻¹ yr⁻¹ and N₂O emissions of 0-0.2 kg N ha⁻¹ yr⁻¹. For more nutrient-rich, minerotrophic fen soils, the values are GWP of 4794 CO₂-C equivalents kg ha⁻¹ yr⁻¹, consisting of CO₂ emissions of 0.82-6.58 t C ha⁻¹ yr⁻¹, CH₄ emissions of -1.04-105 kg C ha⁻¹ yr⁻¹, and N₂O emissions of 5.05 kg N ha⁻¹ yr⁻¹ (Byrne *et al.*, 2004). Boreal peat soils drained and used for pasture may also have higher nitrate availability than those drained for afforestation, and have been reported to emit 6-10 times more N₂O at a rate of 8-9 kg N ha⁻¹ yr⁻¹ (Nykänen *et al.*, 1995), while a range of 1.7-11 kg N₂O-N ha⁻¹ yr⁻¹ has been reported for organic soils under grass across Scandinavia (Maljanen, 2003).

Good Practice Guidance for grassland on organic soils

- Conversion to grassland causes loss of stored carbon and increased greenhouse gas emissions and therefore should be discouraged on organic soils
- Where possible, existing grassland should have drains blocked to restore a high water table, and fertiliser and lime additions stopped.
- If blocking drains completely is not feasible, maintaining as shallow a water table as possible will reduce carbon losses.
- Greenhouse gas emissions can be partially mitigated by minimising disturbance to the soil – using a permanent crop so no tillage is required and minimising fertiliser inputs.

Conversion to cropland: Conversion to arable land use is the worst case scenario for organic soils, with the combined effects of drainage, tillage and fertiliser input causing very fast loss of stored carbon (Byrne et al., 2004). Not only does this land use generally involve the greater disturbance of the soil, it also requires the most drainage, as the water table needs to be lowered to around 1.0-1.2 m below the surface (Joosten et al., 2002). Carbon losses of 400-830 g CO₂-C m⁻² yr⁻¹ have been reported for boreal organic soils used to grow barley in Finland (Maljanen, 2003). These losses are also exaggerated by the presence of bare soil between crops, which may emit more CO₂ than under crops, for example up to 11 t ha⁻¹ yr⁻¹ for Finnish organic agricultural soils (Maljanen, 2003), and may also be subject to physical erosion. Therefore, emissions can be partially mitigated by changing to more permanent crops, and also by avoiding root crops such as potatoes and sugar beet, as these involve frequent and intensive soil disturbance (Freibauer et al., 2004).

As well as losing stored carbon rapidly, organic soils used for agriculture can be a significant source of N₂O. In Finland, such soils may account for as much as 25 % of national N₂O emissions, despite making up less than 10 % of agricultural soils (Maljanen, 2003), and in general, organic soils emit more N₂O than mineral agricultural soils (Duxbury *et al.*, 1982; Martikainen *et al.*, 2002). A range of 5.4-24.1 kg N₂O-N ha⁻¹ yr⁻¹ has been given for Finnish organic soils used to grow barley (Maljanen, 2003).

Flux measurements from sites across Europe indicate that an average of 4400 CO₂-C kg ha⁻¹ yr⁻¹ is emitted from nutrient-poor bog soils converted to arable usage, with no net flux of either CH₄ or N₂O, while nutrient-rich fen soils have an average GWP of 5634 kg CO₂-C equivalents ha⁻¹ yr⁻¹, of which 1.09-10.6 t C ha⁻¹ yr⁻¹ is emitted as CO₂, 4.0-56.4 kg N ha⁻¹ yr⁻¹ is N₂O and there is a net sink of 0.2 kg C ha⁻¹ yr⁻¹ for CH₄ (Byrne *et al.*, 2004).

Good Practice Guidance for using organic soils for growing crops

- Conversion to arable land use is the most damaging practice for reducing C stocks and enhancing GHG emissions, and should therefore be strongly discouraged on soils with high carbon contents.
- Ideally, existing arable land on organic and organo-mineral soils should be restored to its natural water level and vegetation.
- Where restoration is not possible, the following measures will mitigate C and N losses
 - Maintain as shallow a water table as possible
 - Stop deep ploughing and if possible change to zero or conservation tillage practices
 - Avoid root crops such as potatoes and sugar beet which require more soil disturbance
 - Minimise time in which soil is without crop coverage and change to permanent crops if possible

6.2 Good practice guidance for the agricultural management of organic soils

6.2.1 Summary

Greater disturbance of natural conditions equates to greater impact in terms of both loss of C stocks and increased greenhouse gas emissions. Therefore, conversion to arable cropland is the worst-case scenario for organic and organo-mineral soils, as this involves drainage, tillage, fertilisation, liming, and bare soil between crops which increases erosion risk. The best-case scenario is restoring natural vegetation and high water table conditions, and managing with light grazing.

6.2.2 Good practice guidance for arable land use on organic soils

- Organic soils converted to arable land use lose stored carbon very rapidly, typically at rates of around 4-8 C t ha⁻¹ yr⁻¹, emit more N₂O than mineral agricultural soils, and have average Global Warming Potentials of 4.4-5.6 t CO₂-C equivalents ha⁻¹ yr⁻¹, making this the worst-case scenario for land use change.
- Any further conversion to arable usage on these soils should therefore be strongly discouraged.
- Ideally, existing arable land should be restored to heathland, although conversion to managed permanent grassland is less damaging than continued usage for arable crops, provided any grazing animals are stocked at low density.
- Where arable land use continues, the following measures will partially mitigate carbon losses and greenhouse gas emissions:
 - Fertiliser and lime additions should be kept to a minimum.
 - Fertiliser should be applied when crops are in growth phases and can make best use of the applied nutrients.
 - Fertiliser and lime should not be applied in wet conditions.
 - Application of solid farmyard manure is preferable to sewage sludge or liquid manure/slurry, as it has been shown to produce lower N₂O emissions.
 - Conservation or zero-till regimes should be encouraged.
 - Deep ploughing should be discouraged.
 - Winter ploughing should be avoided to reduce erosion risk and effects of freeze-thaw cycles on bare soil, and instead where necessary, ploughing should be carried out as close to new crop sowing as possible to minimise soil exposure.
 - Avoid root crops such as potatoes and sugar beet which require more soil disturbance.
 - Minimise time in which soil is without crop coverage and change to permanent crops if possible.
 - Maintain as shallow a water table as possible.

6.2.3 Good practice guidance for improved grassland on organic soils

- Organic soils converted to managed grassland lose carbon rapidly, typically at rates of around 3-5 C t ha⁻¹ yr⁻¹, and have average Global Warming Potentials of 2.4-4.8 t CO₂-C equivalents ha⁻¹ yr⁻¹.
- Therefore, further conversion of heathland by draining, fertilising and liming should be strongly discouraged.
- Ideally managed grassland should be restored to a more natural state by blocking drains and stopping applications of fertiliser and lime. Restoring high water tables could reduce their Global Warming Potential to 0.4-0.5 t CO₂-C equivalents ha⁻¹ yr⁻¹.
- Resources for blocking drains should be focused on slopes where the drainage is oldest as this will have the greatest impact for reducing carbon losses through erosion.
- Blocking drains may also help reduce problems with colour in water due to particulate organic matter in catchments where drainage water goes into the water supply, as will stopping liming in these areas.
- Drainage should not be used to mitigate N₂O emissions on non-mineral soils, instead options such as reducing N inputs and grazing intensity should be explored.
- Where grassland management continues, the following measures will reduce its impact:
 - Maintain as shallow a water table as possible
 - Minimise fertiliser and lime inputs, particularly on wetter sites
 - Permanent grassland is preferable to reduce soil disturbance
 - Where grassland is grazed, low stocking densities will reduce impacts, as will using sheep or other smaller animals, rather than cattle, and reducing numbers in winter and on wetter sites.
 - Nutrients deposited on pastures should be taken into account to reduce fertiliser requirements
 - Fertiliser inputs should be timed so that vegetation is in growth phase and can make maximum use of the nutrients, and to avoid wet conditions.
 - Application of solid manure is preferable to sewage sludge or liquid manure/slurry.

6.2.4 Good practice guidance for rough grazing on organic soils

- Organic soils store most carbon and emit less greenhouse gas when underlying natural vegetation and managed with minimal disturbance and light grazing to prevent succession to scrubland and ultimately woodland.
- Drainage, especially of water logged areas, should be prevented.
- No fertiliser or lime should be applied.
- Overgrazing causes significant damage to soil organic layers, reduces carbon storage and increases greenhouse gas emissions so stocking densities need to be

carefully managed to minimise this, while limiting succession to protect vegetation communities.

- Heavier animals such as cattle should only be used in very limited numbers, where their less selective feeding will aid vegetation management, and not on wetter sites where they are more likely to cause significant damage.
- Stocking densities should be reduced in winter or animals removed completely, particularly on wetter sites.
- Stocking densities must take into account grazing pressure from wild animals such as deer and a reduced number of domestic livestock should be used on sites where deer are numerous.
- Deer control measures should be concentrated in areas with organic soils which are at risk of overgrazing.
- Serious consideration should be given to halting the practice of burning on soils with significant stores of carbon, and the use of grazing alone to control community succession and encourage vegetation renewal, although this may require reductions in grazing density.
- Burning should not be allowed in areas with a high risk of erosion such as exposed areas or areas with extensive drainage gullies or ditches.
- Fires should be as small and controlled as possible so that only vegetation is consumed and litter and root mats remain to protect the soil from exposure
- The burning season should be confined to after wet periods in early spring, towards the end of the current legal season, as winter freeze-thaw cycles increase the risk of erosion when soil is no longer protected by vegetation.

Module 7

Suggestions For Guidance On The Likely Effects Of Changing Land Use From Grazing Or Semi-Natural Vegetation To Forestry On Soil C And N In Organic Soils

7.1 An assessment of the likely effects of changing land use from grazed or semi-natural vegetation to forestry on carbon stores and fluxes in upland organo-mineral soils in the UK

The following report is a summary of an extensive review (Reynolds, 2007) exploring the likely effects on organic carbon stores and fluxes of converting semi-natural and extensively grazed natural vegetation to forestry on upland organo-mineral soils, and does not include deep peats.

As a party to the United Nations Framework Convention on climate change, the UK is required to protect and enhance the sinks and reservoirs for greenhouse gases. Within the UK, approximately 30% (1357 Tg to a depth of 1 m) of the soil organic carbon (SOC) stock is held in peat soils, with a further 22% in organo-mineral soils (Bradley, *et al.*, 2005). In relation to land use, only 9% of the UK SOC stock resides in forest and woodland soils, although the carbon density of these soils is relatively high (25 kg m⁻² or 250 tC ha⁻¹) compared to pasture and arable soils (16 and 12 kg m⁻² or 160 and 120 tC ha⁻¹ respectively; Bradley *et al.*, 2005). Organo-mineral soils may well become increasingly important for the future expansion of forestry in the uplands if the extensive grazing agriculture they currently support becomes economically marginal because of CAP reform. An assessment of the likely effects on SOC stocks of this potentially major change in land use is therefore required.

Assessments of the effects of afforestation on soil organic carbon stocks in the UK have relied heavily on modelling approaches (Dewar & Cannell, 1992; Cannell *et al.*, 1993; Cannell and Dewar 1995) which have subsequently contributed to the national carbon inventories for the UK (Cannell *et al.*, 1999; Milne *et al.*, 2000). As far as SOC is concerned, Cannell *et al.*, (1999) consider three types of soil; peats, upland organo-mineral soils and mineral soils. For the purposes of the carbon inventory, the latter are assumed to be planted with broadleaves and are predicted to gain SOC as the trees add substantial quantities of litter to soils with relatively low organic matter content, including former agricultural land (Cannell and Dewar 1995). For peats, information from field measurements of net ecosystem carbon exchange suggest that these will be a small net source of carbon resulting from the effects of drainage and soil drying by the trees (Hargreaves *et al.*, 2003). Upland organo-mineral soils under plantation conifer forest are considered to be 'carbon neutral' having gained as much carbon from forest litter as has been lost through accelerated decomposition during site preparation and drainage. This assumption has been evaluated in this module.

Planting trees can affect the ecosystem carbon balance in two opposing ways. Drainage and disturbance during site preparation and planting may lead to higher SOC losses as

increased microbial respiration rates, aeration and disruption of soil aggregates lead to higher rates of organic matter decomposition. Conversely, carbon will be sequestered by tree growth while leaf and needle fall will contribute to the accumulating litter layer at a faster rate than for the pre-existing ground vegetation. Tree roots and root litter will also contribute to the below ground carbon stock.

Two main approaches were taken to assess the likely effects of changing land-use from grazed acid grassland or semi-natural vegetation to forestry on SOC in organo-mineral soils of the UK. These were:

- 1) Direct measurement of change in soil organic carbon
- 2) Measurement of biogeochemical fluxes to produce a carbon budget

As there are relatively few directly relevant studies, the review has drawn on studies of related systems to draw inferences and conclusions about the likely effects of afforestation on the SOC balance of organo-mineral soils

7.1.1 Evidence from direct measurement of SOC stocks

In the context of organo-mineral soils in the UK uplands, the evidence from direct measurements of SOC was inconclusive about a major effect on SOC stocks following a change in land use from semi-natural grassland / moorland to forestry. International studies suggest that planting broadleaf trees will have little effect on SOC stocks, whereas conifer planting, especially in high rainfall areas may lead to depletion of SOC stocks (Guo and Gifford 2002) However, it is important to note that the meta-analysis conducted by Guo and Gifford (2002) was dominated by studies from New Zealand and the conifer species considered were mainly pines. Few relevant UK data sets exist which address the issue and the experimental data presented in Module 2 of this report appear to contradict the findings of the international meta-analysis. The inherent problems of soil heterogeneity, vertical gradients in organic matter content and bulk density, contribute uncertainty to making these types of comparison. Furthermore, the extent to which results from other systems, for example peatlands and lowland agricultural soils can be extended to upland organo-mineral soils is also very uncertain. Rooting patterns and tree growth rates will differ considerably, and the effects of tree species must also be taken into account.

There is evidence that organic matter accumulates at the surface of organo-mineral forest soils, but the long-term fate of this material is uncertain and depends on the extent to which it is incorporated into the long-term, stable organic carbon pool within the soil profile (Pataki *et al.*, 2003). This in turn is likely to depend on the tree species, soil type, site nutrient status, site hydrology and climate. Some studies reported in a review by Post and Kwon, (2000) recorded a loss of organic carbon at depth in the profile, whilst observing accumulation at the surface. In these cases, organic carbon inputs during early growth of the forest were insufficient to replenish decomposition losses lower in the profile. A decrease in the recalcitrant soil organic carbon pool was predicted by the MERLIN model in a simulation of planting and growth of a 30 year old Sitka spruce plantation which replaced moorland vegetation growing on acid peaty podzols in north

Wales (Emmett *et al.*, 1997). The data used to parameterise the model, which came from a chronosequence of Sitka spruce plantations on similar soils, showed, however, a net accumulation of organic carbon by the ecosystem, both as new wood and in the labile soil organic matter pool mainly in the forest floor. Evidence of a loss of 'old' soil organic carbon during forest development also comes from a study of the ^{14}C signal in soil solution dissolved organic carbon (DOC) by Karlton *et al.* (2005). Samples of DOC were collected from the transition zone between the A and B soil horizons of a forest chronosequence comprising twelve sites in southwest Sweden, ranging from agricultural land to 89 year old, first generation Norway spruce. In order to explain the observed changes in ^{14}C signal along the chronosequence, the authors proposed that two processes were occurring simultaneously; namely changes in litter input and increased mobilisation of soil organic carbon formed before afforestation. The implication is that any assessment of SOC stock change in response to afforestation must extend down to the base of the rooting zone in order to account for potential losses of SOC from the subsoil as well accumulation at the soil surface.

Leaf and needle litter accumulation is an important process for transferring newly fixed carbon to the the surface of forest soils. Another significant but much less well understood and quantified pathway is the exchange of carbon between plant roots and the soil (Jones *et al.*, 2004). This has the potential to directly transfer newly fixed carbon into the soil at depth where it may be subsequently incorporated into the stable, long-term organic carbon pool. Recent work suggests that carbon sequestration via this pathway may decline as atmospheric CO_2 concentrations rise, with obvious implications for the terrestrial carbon sink (Heath *et al.*, 2005).

7.1.2 Evidence from flux studies

The questions addressed by flux measurements essentially come down to how much of the carbon exchanged with the atmosphere finds its way into the stable, long-term SOC store, and how much do the other flux pathways offset the carbon sink.

In order to summarise the review findings, a simple matrix was constructed which provides a qualitative indication of the likely effects of forestry development on individual fluxes (Table 1). The summary suggests that overall, there is likely to be little net effect of forest development on many of the carbon fluxes, particularly for DOC and Dissolved Inorganic Carbon (DIC), although the latter is uncertain. Particulate Organic Carbon (POC) fluxes are likely to be enhanced at all stages of the forest cycle, but changes in forest management practice in line with current Guidelines should constrain these losses (Nisbet *et al.*, 1997; Forestry Commission, 1998; HMSO, 2003).

The effect of forest harvesting on net CO_2 exchange may be neutral depending on whether carbon lost to the atmosphere from increased decomposition rates and microbial activity is subsequently balanced by uptake of carbon in re-growing vegetation. This is likely to be very site specific and will be determined to a great extent by harvest residue management and site environmental conditions. However some measurement campaigns

similar to those conducted during the afforestation of peatland sites by Hargreaves *et al.* (2003) would provide useful data.

Table 7.1. Suggested effects of forest development on carbon fluxes in organo-mineral soils.

	Preparatory phase (approx. duration of effect: up to 10 years)	Forest growth (approx. duration of effect: upto 5 decades)	Harvesting (approx. duration of effect: up to 5 years)	Comments
Net CO ₂ exchange	+ / +	- / +	+ / +	harvesting effects depend on residue management and re-vegetation
Methane emissions	- / +	- / +	+ / 0	relatively minor flux term in relation to overall carbon balance
Dissolved Organic Carbon	+ / +	+ / 0	+ / 0	
Particulate Organic Carbon	+ / +	+ / +	+ / +	can be controlled by good site management?
Dissolved Inorganic Carbon / CO ₂ evasion	(+ / 0)	(+ / 0)	(+ / 0)	uncertain with few measurements

Symbols:

First symbol represents direction of flux, where negative sign represents depletion of atmospheric carbon. Second symbol represents the change in the magnitude of the flux with respect to semi-natural or extensively grazed moorland / acid grassland. '0' means that there is no change as a result of changing land use; '+' means an increase in flux with respect to that measured in semi-natural moorland / grassland.

In conclusion, the overall effects on soil organic carbon stocks of land use change from semi-natural, extensively grazed vegetation to forestry on organo-mineral soils can be summarised as follows:

- There have been few directly relevant studies of the effects on soil organic carbon stocks of afforestation of organo-mineral soils, so all conclusions about the likely effects of land use change to forestry are inferential from related studies on UK peatlands and studies from abroad.
- It is not possible yet to identify likely effects on soil organic carbon stocks down to individual soil types or even broad categories of freely and non-freely drained soils.
- The overall conclusion (as assumed by UK carbon balance models) is that afforestation probably has little net effect on soil organic carbon stores in organo-mineral soils, but this is a very uncertain statement.

The main areas of uncertainty for organo-mineral soils are:

- What is the net effect on the soil organic carbon store of planting and disturbance versus subsequent carbon capture by trees and inputs to soil as litter?
- What effects do forest harvesting have on soil organic carbon stores?
- To what extent is the gain in soil organic carbon from litter deposited at the soil surface offset by any loss of pre-existing soil organic carbon from the subsoil?

- How much of the carbon captured by trees and deposited as litter finds its way into the “stable” soil organic carbon store?
- What are the effects of different species and forest management regimes (eg continuous cover forestry vs. conventional patch clearfell) on soil organic carbon storage?

7.2 Provisional management guidance

Some provisional guidance for managing the effects of afforestation on SOC stocks in organo-mineral soils can be suggested from the outcome of the review. These are summarised in Table 7.2. Currently this guidance must be regarded as very preliminary it is based on limited data and has not been tested against modelled scenario outcomes. However the broad principles behind the guidance are to:

- Minimise disturbance during any site operations so as to limit particulate organic carbon losses and to reduce the opportunities for mineralising the pre-existing soil organic carbon store, for example by limiting soil drying and aeration
- Take steps to maintain the net CO₂ sink strength at all stages of the forestry cycle by maximising vegetative cover

Table 7.2. Guidance for managing the effects of land use change from semi-natural moorland / acid grassland to forestry on upland organo-mineral soils.

Measure	Guidance	Purpose
Site preparation	Minimise disturbance	Reduce Particulate Organic Carbon losses, minimise losses of pre-existing Soil Organic Carbon stock
	Maintain as much vegetation cover as possible	Maintain net CO ₂ sink strength
Forest management	Minimise disturbance	Reduce Particulate Organic Carbon losses
Harvesting	Use low impact technologies	Minimise Particulate Organic Carbon losses & site disturbance
	Encourage rapid re-vegetation of felled areas	Maintain net CO ₂ sink strength; retain nutrients on site

Module 8

Use Of The Land-Use Change Data Derived From The Countryside Surveys Of Scotland And Wales To Provide Predictive Estimates For Changes To C And N Balance In Organic Soils Over Time

8.1 Introduction

Changes in land use have shaped the landscape of the UK over many centuries and there have been significant changes in land use even in recent decades. The rates and types of land use change over time are produced by the interaction between the physical environment (for example, climate and soil type limit certain types of land use) and social/economic forces governing the relative value of land under different usages (for example, tax concessions encouraged afforestation during the 1970s and 1980s on land previously used for extensive grazing). Assessments of national land use change are of interest for development and agricultural policy, for habitat conservation and, more recently, for the estimation of greenhouse gas emissions and carbon storage. Land use change is responsible for both emissions and removals of greenhouse gases, due to the associated changes in vegetation inputs and turnover and especially in soil carbon stocks. Land use change is one of the main contributors to the Land Use, Land Use Change and Forestry sector in the UK National Greenhouse Gas Inventory, the only sector that is a net sink for greenhouse gases (Baggott *et al.*, 2006).

8.1.1 The matrix approach for assessing land use change

At any point of time, an area of land can be classified as being of a particular land use type. Many different classifications are possible but for Greenhouse Gas Inventory (GHGI) purposes the IPCC Good Practice Guidance for Land Use, Land-Use Change and Forestry (IPCC, 2003) defines five types useful in the UK: Forest land, Grassland, Cropland, Settlements and Other Land (the IPCC Wetlands category is included in either Grassland or Other land for the UK). For the ECOSSE project the Grassland category is split into Natural Grasslands and Managed Pasture. Between points in time, areas of land may remain in the same category of use, or need to be reclassified because of a change in use. The total areas changing can be summarised in a matrix (Table 8.1). The diagonal cells of the matrix represent no change in categorisation (i.e. use) of land. The matrix gives the areas and categories of land that have undergone land use change during a period: needed for estimating changes in soil carbon stocks, and hence GHG emissions.

Table 8.1 Example of a land use change matrix (Scotland 1990-1999). Data are annual changes (000 ha) in land use from the 1990 and 1998 Countryside Surveys.

Change to ...	Change from...				
	Forest	Cropland	Grassland	Settlement	Other
Forest		0.6	11.1	0.2	0.0
Cropland	0.1		21.4	0.3	0.0
Grassland	5.0	16.8		0.7	0.0
Settlement	0.3	0.1	2.2		0.0
Other	0.0	0.0	0.0	0.0	

8.1.2 Sources of information on national land use change

There are three national datasets on land use change that cover the period from 1950 to the present.

- The Monitoring Landscape Change (MLC) project (Hunting Technical Services Ltd, 1986) which assessed land use change in England and Wales between 1947, 1969 and 1980 using aerial photography.
- The National Countryside Monitoring Scheme (NCMS) (Mackey *et al.*, 1998) which assessed land cover change in Scotland between 1947, 1973 and 1988 using aerial photography.
- The Countryside Surveys (Barr *et al.*, 1993; Firbank, 2003), which are national (GB) field surveys managed by the Centre for Ecology and Hydrology. These are available for 1984, 1990 and 1999.

Each of these data sources uses a different land classification system, and the original classes must be grouped into the land use categories used for the GHGI to allow comparison between datasets and over time. The groupings for each dataset are given in Tables A3.1-A3.3 in Appendix 3.

Previously, only country-level (England, Scotland, Wales and Northern Ireland) land use change matrices for the GHGI have been developed. This approach is unsatisfactory as land use change varies greatly within these countries. The available data sources are all sample-based surveys so although it is possible to develop matrices at finer scales from these sources, there is an issue of statistical validity: at very fine scales the matrices will be too uncertain to be useful. The target scale chosen here is 20km x 20km, as this is the scale found to achieve an acceptable balance between detail and accuracy in other components of the GHGI.

For Scotland and Wales (England also) measured land area and change data are available from surveys taken in 1947, 1969/1973, 1980 (England and Wales only), 1984, 1990 and 1998. Measured land use change data over the different periods were used to estimate annual changes by assuming that these were uniform across the measurement period, e.g. the period 1980-84 was filled using data from the Countryside Survey (CS) assuming the same annual rate of change as seen for 1984-90. The period 1999-2003 was extrapolated forward from the CS assuming the same annual rate of change as seen for 1990-98. Another CS is planned for 2007-2008, which will allow the land use change estimates to be updated in the future.

8.2 Methods and results

8.2.1 Development of land use matrices 1950-1980 for Wales.

Matrices of change at the national (Welsh) level and areas of different land use types (in 1947, 1969 and 1980) at the Welsh county level are available in the MLC reports (volumes 1, 3-6 and 10) (Hunting Technical Services Ltd, 1986). By combining these sources of information it is possible to produce estimated land use change matrices at the county level (Appendix 3; Figure A3.1a). The MLC data were only available in the paper reports, so the areas and standard deviations of the MLC land use classes in each county

in 1947, 1969 and 1980 were typed into spreadsheets to enable further analysis. A new spreadsheet was constructed that compiled the land use area information for all counties and for Wales as a whole. National matrices of land use change (1947-69 and 1969-80) using the inventory land classification had been constructed in a previous MSc project (Hodgson, 2004). These tables were used to construct national matrices of percentage change for each time period, i.e. the percentage contribution from each land class at $t=i$ to the area in that land class at $t=i+1$ (Table 8.2).

Table 8.2 Percentage contribution matrices of change for Wales for (a) 1969, and (b) 1980.

(a)		Land class in 1947					
		<i>Forestland</i>	<i>Natural</i>	<i>Cropland</i>	<i>Pasture</i>	<i>Settlement</i>	<i>Other</i>
Land class in 1969	<i>Forestland</i>	69.8%	19.5%	0.2%	9.8%	0.2%	0.3%
	<i>Natural</i>	0.9%	96.7%	0.0%	2.1%	0.2%	0.1%
	<i>Cropland</i>	0.8%	0.2%	32.7%	66.1%	0.2%	0.0%
	<i>Pasture</i>	2.4%	6.5%	3.8%	86.8%	0.4%	0.0%
	<i>Settlement</i>	1.7%	3.6%	2.7%	25.8%	66.0%	0.2%
	<i>Other</i>	0.0%	3.9%	0.0%	0.8%	0.0%	95.3%
(b)		Land class in 1969					
		<i>Forestland</i>	<i>Natural</i>	<i>Cropland</i>	<i>Pasture</i>	<i>Settlement</i>	<i>Other</i>
Land class in 1980	<i>Forestland</i>	74.3%	13.4%	0.2%	11.9%	0.1%	0.0%
	<i>Natural</i>	0.5%	97.0%	0.0%	2.4%	0.0%	0.1%
	<i>Cropland</i>	1.3%	0.1%	49.4%	48.8%	0.3%	0.0%
	<i>Pasture</i>	1.8%	3.6%	1.4%	92.7%	0.5%	0.0%
	<i>Settlement</i>	1.2%	1.9%	1.0%	19.1%	76.8%	0.0%
	<i>Other</i>	6.4%	3.6%	0.0%	2.1%	0.0%	87.9%

The county area data and the national matrices of change were used to estimate land use change at the county level in the two time periods (Equation 8.1). The data for the year at either end of the time period can be used for the estimation of matrices of land use change during that period, e.g. either the 1947 areas or the 1969 areas can be used to estimate change in 1947-1969. After an error comparison, the end year was chosen as the baseline for calculation as this generated slightly smaller errors overall.

$$L_{mjk} = \frac{\left(A_{mk} \frac{W_{jk}}{\sum_{k=1}^n W_{jk}} \right)}{\sum_{j=1}^n W_{jk}} A_{mj}$$

Equation 8.1

Where L is land use change in county m in time period $t=i \rightarrow i+1$, j is the land class in $t=i$, k is the land class in $t=i+1$, A is area and W is land use change in Wales in time period $t=i \rightarrow i+1$.

Land use data are disaggregated to the 20km scale by splitting the land use in each county in proportion among all the 1km cells in that county and then aggregating the 1km cells into the appropriate 20km squares. The different spatial patterns of land use change can be seen in Figure 8.1. The increased spatial and temporal precision in the estimation of land use change results in larger estimates of change *to* land classes at the national scale and different rates of change between decades (Table 8.3). Some of these differences are due to interchange between land classes during the longer time period. The greatest rates of change ($>2\text{kha a}^{-1}$), in order of size, are: N→P, P→C, C→P in the 1950s; P→C, N→P and C→P in the 1960s; and P→C, N→P, N→F, P→F and P→S in the 1970s. (N=Natural, P=Pasture, C= Cropland, F= Forest land and S= Settlement). The rates of change increase over time for all land classes except for Pasture, where rates decline over time.

Table 8.3 Comparison of annual land use change per decade in Wales 1950s-1970s, 000 ha a⁻¹

Change to Land Class	20km sum estimate, 1950s	20km sum estimate, 1960s	20km sum estimate, 1970s	National estimate, 1950s-1970s
	000 ha a ⁻¹			
<i>Woods</i>	2.50	2.74	4.87	2.58
<i>Natural</i>	0.69	0.74	1.15	0.42
<i>Farm (Pasture)</i>	7.11	7.09	6.90	5.17
<i>Farm (Arable)</i>	3.05	3.59	8.45	4.10
<i>Urban</i>	1.54	1.64	2.60	1.64
<i>Zero</i>	0.06	0.08	0.32	0.00

8.2.2 Development of land use matrices 1950-1980 for Scotland

Land use matrices for Scotland for 1950 to 1980 were developed using the database of the National Countryside Monitoring Scheme (NCMS) project (Mackey *et al.*, 1998). The NCMS database can be analysed using user-defined geographic regions using an extension (the NCMS Visualisation and Analysis System) in the ArcView 3.2 geographic information system. Analysis regions were developed from groups of LAU1 administrative areas (Appendix 3; Figure A3.1b), 22 regions in total, which are similar in size to the county regions used for Wales.

Each region was analysed for the two time periods 1947-1973 and 1973-1988, using a 95% confidence interval and minimising dependence on squares outside the region. The analysis outputs are area estimates for each land class at the beginning and end of the time period and interchange area estimates between land classes, i.e. land use change matrices. The individual regional outputs were collected into a single spreadsheet. The resulting matrices were assigned to 20km cells in the same way as described for Wales.

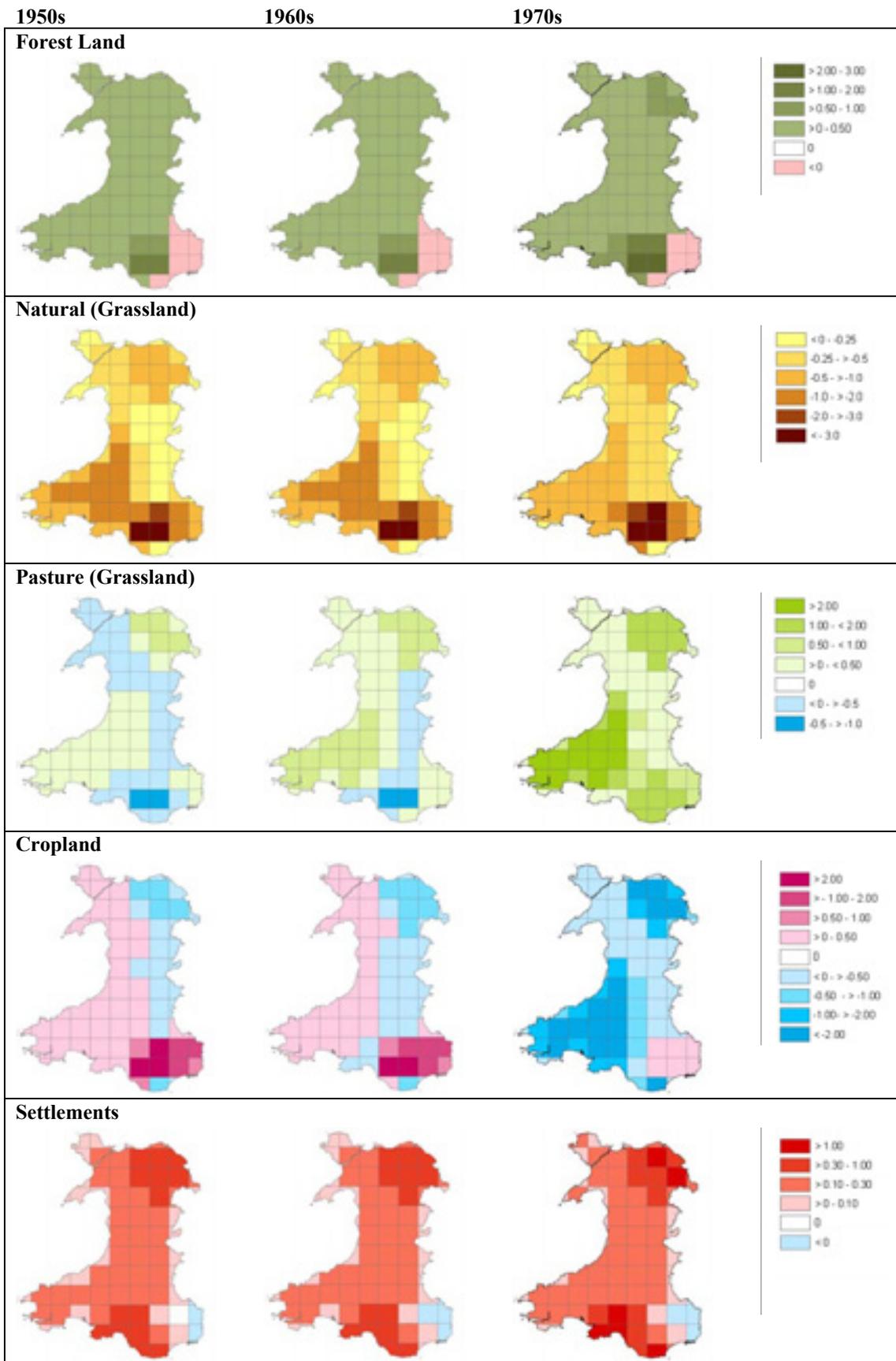


Figure 8.2 Pattern of NET land use change in Wales in the 1950s, 1960s and 1970s, kha/decade/20km
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The new matrices for Scotland have been derived from a different dataset (NCMS) than previously used matrices (MLC). There is broad agreement in the total annual change (Table 8.4), although there is much variation in change between land classes. There is more change to Forest land in the NCMS matrices, largely due to increased change from the Natural class, although there is also less change from the Pasture class. There is less change to Pasture in the NCMS matrices, due to less change from Forest land and Natural classes. The annual change in the Cropland class is similar in both datasets in the 1950s and 1970s, but much greater in the NCMS matrices in the 1970s due to greatly increased change from the Pasture class. The differences between the other classes in the matrices are smaller. As in Wales, there are increased rates of change in all classes in the 1970s when compared to the 1950s and 1960s. The different spatial patterns of land use change can be seen in Figure 8.2.

Table 8.4 Matrices of annual land use change per decade in Scotland 1950s-1970s

Change to Land Class	NCMS 20km sum estimate, 1950s-1960s	NCMS 20km sum estimate, 1970s	NCMS national estimate, 1950s-1960s	NCMS national estimate, 1970s	MLC national estimate, 1950s-1970s
	000 ha a ⁻¹				
<i>Forestland</i>	18.86	26.67	19.06	26.42	17.27
<i>Natural</i>	4.73	16.36	4.75	16.34	7.94
<i>Pasture</i>	17.60	24.30	17.67	24.33	23.79
<i>Cropland</i>	11.00	19.82	11.09	20.31	10.74
<i>Settlement</i>	2.61	3.72	2.67	3.77	3.63
<i>Other</i>	0.81	2.59	0.84	2.64	0.00
<i>Total</i>	55.60	93.47	56.07	93.82	63.37

8.2.3 Development of matrices 1980-present

Countryside Survey data were used to estimate land use change in Scotland and Wales from 1980 to the present. These data are based on detailed field observations, collected in ~ 500 samples of 1 km grid squares across Great Britain (Barr *et al.*, 1993; Firbank, 2003). The sampling is stratified using a system of classifying geographical locations depending on many mapped attributes ('ITE Landclass') so the ecological and land use data from the samples can be scaled up to the national or other level.

In order to calculate the land use change matrix for each 20x20 km grid cell it was first necessary to generate the numbers of 1x1 km cells of each Landclass within each 20x20 km cell. A weighted average for each 20x20km cell of areas of land use change of the field results for sites of known and appropriate Landclass then provided the land use change matrices. These matrices were calculated separately for the two periods of 1984 to 1990 and 1990 to 1998. Figure 8.3 shows examples of the mapped land use changes at 20km scale.

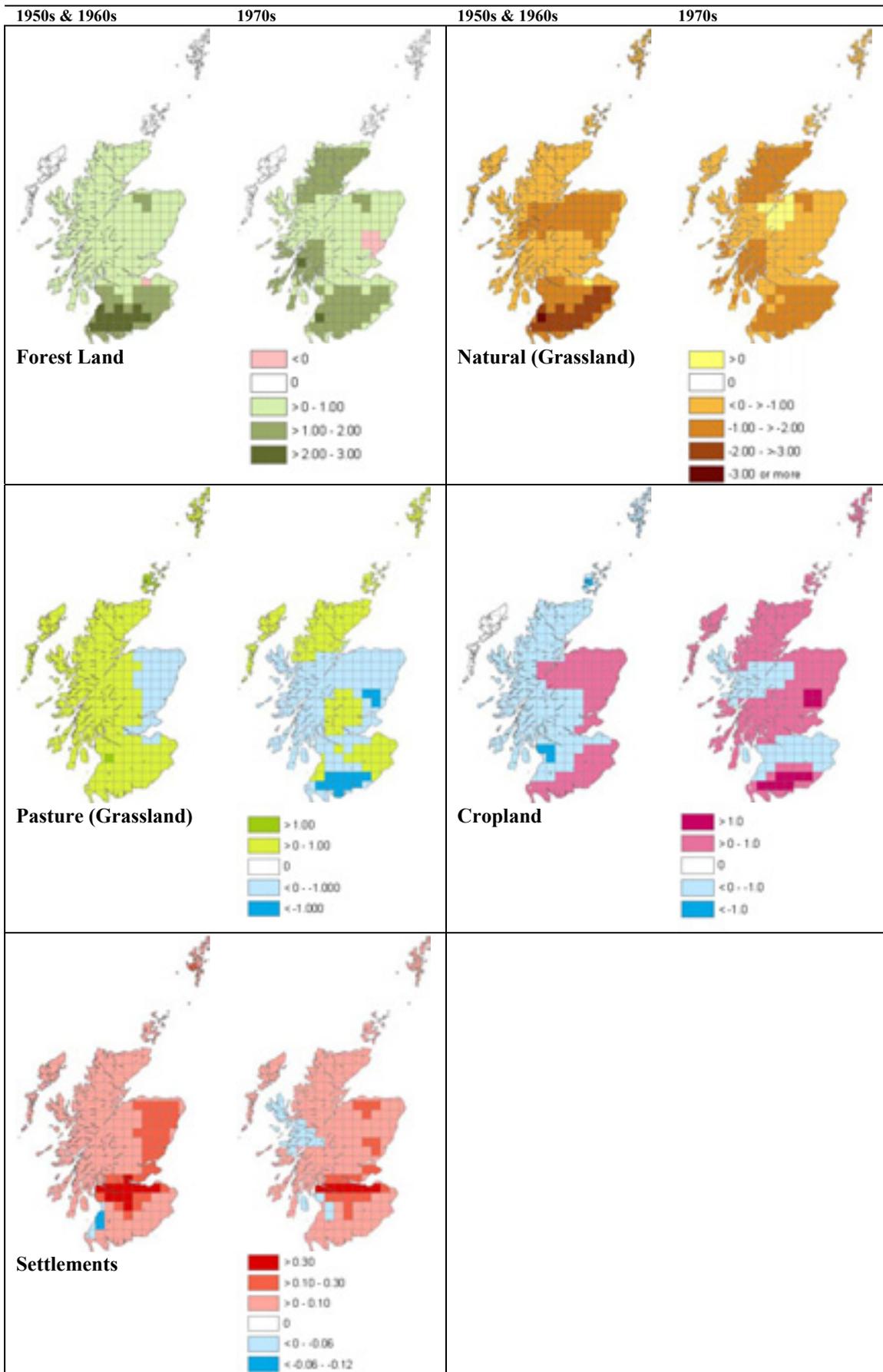


Figure 8.3 Pattern of NET land use changes in Scotland in the 1950s-60s and 1970s, units in kha/decade/20km square

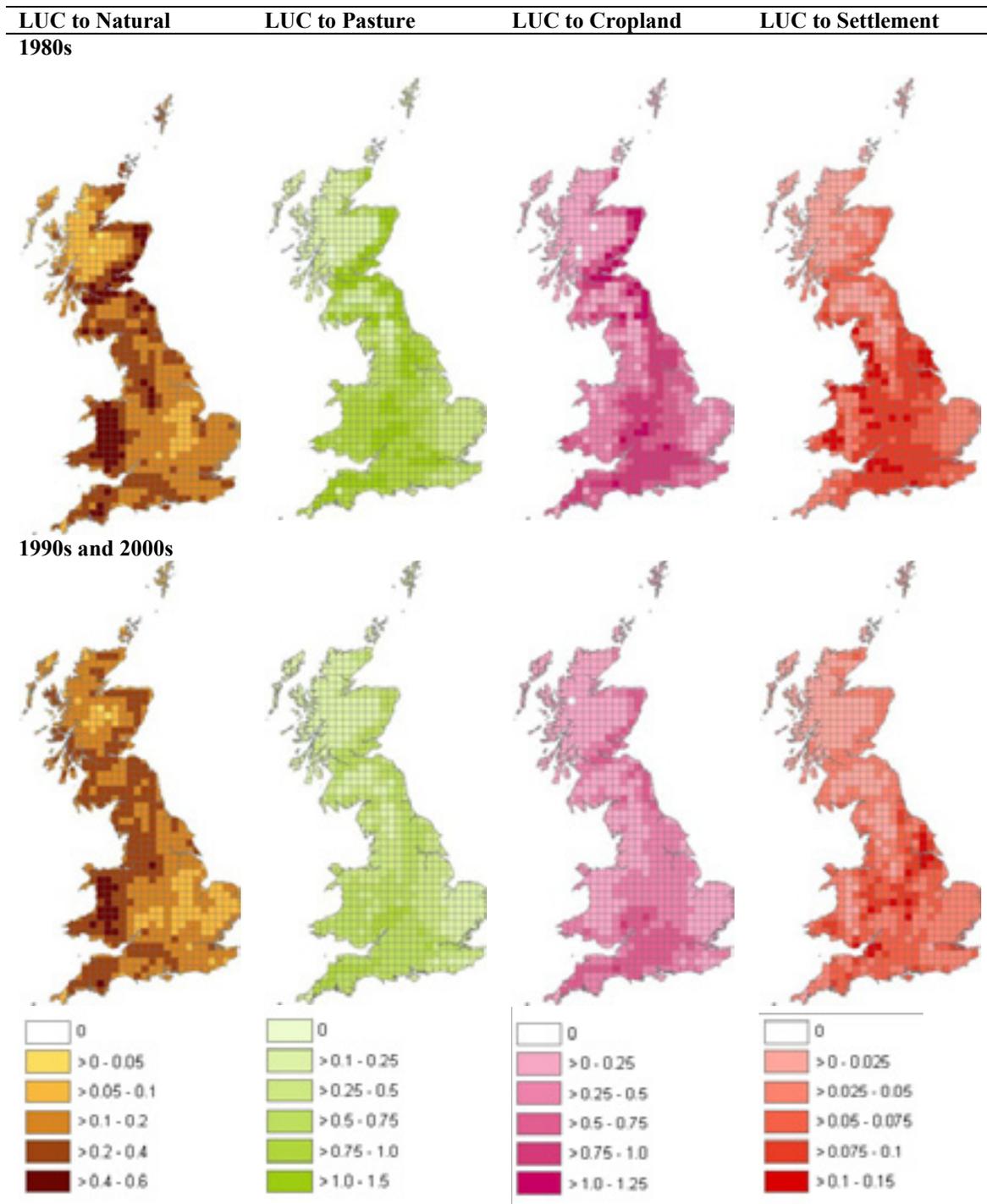


Figure 8.4 Annual NET land use change in the 1980s, 1990s and 2000s in the UK, units in kha/year/20km square

8.2.4 Relationship to soil type

In the Defra Soil Carbon Database (Bradley *et al.*, 2005) the area of the each of the five most common soil series has been recorded for 1km grid-cell in GB. The soil series have also been classified into four broad types: Peats (here termed organic), Organo-mineral, Mineral and Other (Figure 8.4). These data have been co-registered with the land use

change maps using the unique 20km cell identifier codes. When the organic/organo-mineral soil distribution map is used to mask the land use change maps the comparative distribution of land use change on these soil types becomes apparent (Figure 8.5). As the soils dataset is at 1km scale and the land use change dataset is at 20km scale, comparative analysis must be careful, as land use change at the 20km scale may not be representative of land use change occurring specifically on organic soils at the 1km scale.

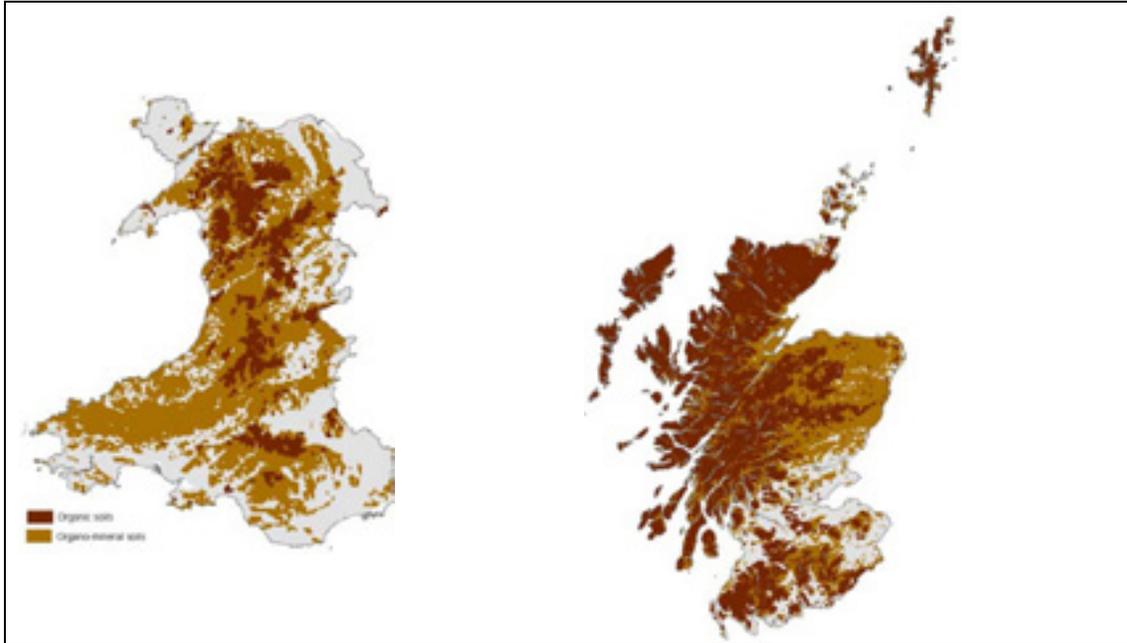


Figure 8.5 Distribution of peats (here termed organic) and organo-mineral soils in Scotland and Wales

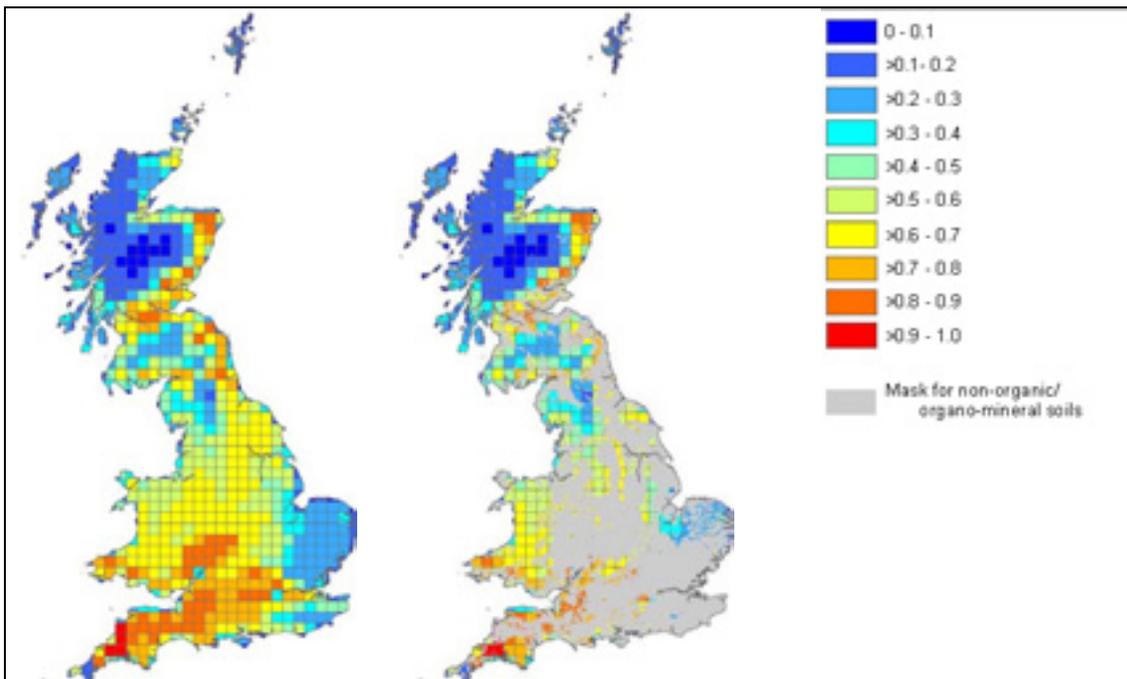


Figure 8.6 Total land use change masked by the peats (here termed organic) and organo-mineral soil distribution maps, units in kha/decade/20km square

The new land use change matrices and maps produced for Module 8 are a great improvement on the previous estimations of land use change at the national scale. The increased spatial and temporal detail results in better estimation of where and when land use change has taken place in the past. This work also has implications for the calculation of carbon fluxes resulting from changes in soil carbon stocks, and the improved matrices will be incorporated into existing models for calculating land use change fluxes for the UK national greenhouse gas inventory.

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Appendix 1

Additional information for Module 1: Regression equation to predict dry bulk density of organic surface horizons in Scotland.

Organic carbon values are held within the Scottish soils database as a percentage of dry weight, however, in order to determine the overall carbon store, it is necessary to convert this value to a volumetric basis. This is achieved by multiplying the gravimetric measurement by the dry bulk density. However, there is insufficient data on the dry bulk density of Scottish soils in order to give meaningful estimates of mean values or any other ‘average’ values for individual soil types. Thus, regression equation-based pedotransfer functions were developed to predict bulk density for soil horizons grouped according to their mode of deposition and subsequent pedogenesis. In order to develop these equations all data on organic soil horizons (those where the organic carbon content exceeds 18%) were grouped.

There are dry bulk density measurements for 39 soil horizons. The soil horizons come from 15 profiles that were sampled for various research projects and 3 profiles sampled specifically to derive dry bulk density for organic horizons. The profiles were located in Galloway, Loch Bradan catchment, Mourne Mountains (Northern Ireland), Mharcaidh catchment and Shetland.

The organic carbon content of each horizon was determined by C and N elemental analyser and bulk densities were determined by oven drying a known volume of peat for 48 hrs at a temperature of 105° C and expressing the results as grams dry weight per cubic centimetre.

A number of regression equations were tested: simple linear regression between organic carbon content (org_C) and dry bulk density (Db), between log transformed organic carbon (logn_org_C) and dry bulk density and a multiple linear regression with organic carbon content and depth as independent variables.

Results: Initial simple linear regression between dry bulk density and organic carbon content explained 63.5% of the variation in dry bulk density.

$$Db = 0.6654 - 0.00999 * org_C \quad (1)$$

A log normal transformation of the organic carbon value improved this simple linear relationship by explaining 70.4% of the variation.

$$Db = 1.772 - 0.4127 * logn_org_C \quad (2)$$

As the degree of peat decomposition generally increases with depth, the uppermost depth that the sample was taken from was used as a predictor along with org_C and logn_org_C in a multiple linear regression. There was no improvement in the percentage variation explained (63.4% and 70.1% respectively).

Plots of dry bulk density against depth showed a sharp decline up to approximately 15 cm depth but then leveled out until depths of around 100cm where there was a slight increase in dry bulk density. The inclusion of depth as a predictor means that this slight increase can be modeled though this has to be balanced by the decline in the percentage variance explained. As this increase in bulk density with depth is only evident in a few samples, it is perhaps best that further data be collected to substantiate this before applying such a model to the data in this project.

There is a little to be gained by using the log transformed organic carbon contents as a predictor of dry bulk density though this should be matched against the additional effort involved in converting all organic carbon values and the small increase in percentage variance explained. It is more than likely that the bulk density values of fibrous 'litter' layers will be over predicted using these transfer functions.

The recommended equation to predict dry bulk density is equation 2. However, it would be prudent to test this equation against an independent dataset when available. This has significant implications for soil carbon stock estimates for Scotland.

Appendix 2

Development and description of the ECOSSE model

A2.1 Modularisation of SUNDIAL/MAGEC

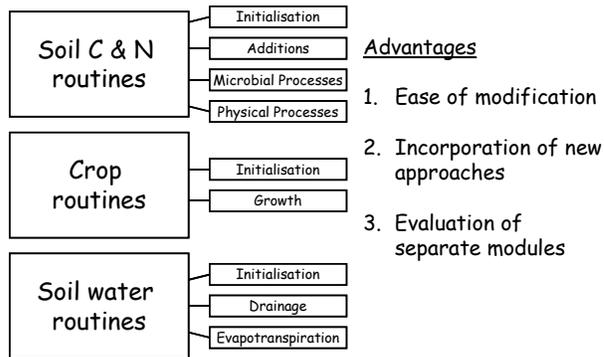


Figure A2.1 Modularisation of SUNDIAL-MAGEC

The model was restructured into modules that contain routines describing soil C and N turnover (initialisation, additions, microbial and physical processes), crop processes (growth, N demand and return of C and N in debris), and the soil water processes (drainage and evapotranspiration). This allows easy modification of individual processes, replacement of modules with new approaches, and independent evaluation of individual modules (Figure A2.1).

A2.2 Improved description of N₂O production

In SUNDIAL-MAGEC, N loss by denitrification is modelled using a simple linear relationship. Losses are not partitioned into N₂, N₂O and NO gas, so emissions of greenhouse gases cannot be estimated. In ECOSSE, emissions of N are simulated due to denitrification and partial nitrification (Figure A2.2). The emissions are then partitioned into emissions of N₂, N₂O and NO.

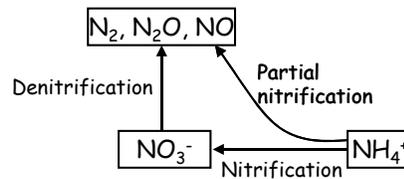


Figure A2.2 Emissions of N₂O

Denitrification: Denitrification is a process that responds to changes in the system on a shorter timescale than many of the other C and N turnover processes. As a result, denitrification is difficult to measure accurately, and difficult to simulate without short-term and detailed input data. As a result, many models of denitrification have high data requirements. The aim in ECOSSE is to develop a model that will simulate processes to the required accuracy but without the need for detailed data that will not be available at the large scale.

The denitrification sub-model of DNDC calculates N₂O and NO production, consumption and diffusion during rainfall, irrigation and flooding events. DNDC simulates relative growth rates of nitrate, nitrite, NO, and N₂O denitrifiers based on soil heterotrophic respiration, and concentration of DOC and N oxides. The soil matrix is divided into aerobic and anaerobic zones using an "anaerobic balloon", which swells and shrinks in response to oxygen diffusion and consumption in the soil profile. Substrates allocated to the anaerobic zone of the profile are then used to determine the denitrifier growth rates based on a simple function of multinutrient-dependent growth (based on Bader, 1978). The death rate is a constant fraction of total denitrifier biomass. Relative growth rates for denitrifiers with different substrates are assumed to be independent, and competition takes place via the common DOC substrate (based on Leffelaar & Wessel, 1988). Substrate consumption rates are calculated using growth rates and biomass. NO, N₂O and N₂ fluxes are calculated using the basic laws of kinetics, as denitrification is a typical sequential reaction. As intermediates of the reactions, NO and N₂O fluxes are determined by rates of production, consumption and escape from the reacting system. Escape is controlled by diffusion rate, a function of soil porosity, moisture, temperature and clay content (Li, 2000).

The CENTURY denitrification model in NGAS is based on data from Weier *et al.* (1993). Total denitrification N gas fluxes (N₂ + N₂O) are a function of soil heterotrophic respiration rate (index of carbon availability), soil NO₃⁻ levels, and soil water filled pore space. Soil NO₃⁻ levels and respiration rate determine the maximum total N gas flux according to s-shaped curves fitted to experimental data. This maximum is then reduced by a fraction determined by water filled pore space, as a function of soil texture. Maximum denitrification occurs where water filled pore space is 0.9 or above. Below this value, denitrification is reduced exponentially, with fine soils declining most rapidly, followed by medium and then coarse soils. Denitrification drops to zero at water filled pore space of 0.6 for fine soils, 0.5 for medium and 0.4 for coarse. Total denitrification N gas flux is then partitioned into N₂ and N₂O as a function of soil NO₃⁻, WFPS and soil respiration (Parton *et al.*, 1996).

In ECOSSE, a simpler approach is used in which total denitrification is simulated as a proportion of the nitrate content of the layer, modified according to water content and biological activity of the soil (as described by CO₂ release during decomposition). The denitrified N is then divided into N₂, N₂O and NO according to the water and nitrate content of the soil. The total loss of N due to denitrification is given by

$$N_d = m_{NO_3} \times m_{water} \times m_{bio} \times NO_3,$$

where N_d is the amount of nitrogen emitted from the soil during denitrification (kg N ha^{-1}); m_{NO_3} modifies this amount depending on the nitrate level, m_{water} depending on the water content, and m_{bio} depending on the biological activity of the soil; and NO_3 is the amount of nitrate in the soil (kg N ha^{-1}).

The nitrate modifier, m_{NO_3} , is based on the model developed by Henault & Germon (2000), i.e.

$$m_{NO_3} = \frac{NO_3}{((3.3 \times 5) + NO_3)}$$

The response of the denitrification process to the amount of nitrate in the soil is shown in Figure A2.3 (A).

The water modifier, m_{water} , is based on the fitted equation of Grundmann & Rolston (1987), i.e.

$$m_{water} = \left(\frac{\left(\left(\frac{W}{W_{max}} \right) - 0.62 \right)}{0.38} \right)^{1.74},$$

where W is the water content and W_{max} is the maximum water content of the soil (mm/layer). The response of the denitrification process to the soil water content is shown in Figure A2.3 (B).

The biological activity modifier, m_{bio} , is based on the relationship developed by Bradbury *et al.* (1987), i.e.

$$m_{bio} = CO_2 \times 0.005,$$

where CO_2 is the amount of CO_2 produced during mineralization (kg C ha^{-1}). This acts as a surrogate measure of the soil biological activity. The response of the denitrification process to the soil biological activity is shown in Figure A2.3 (C).

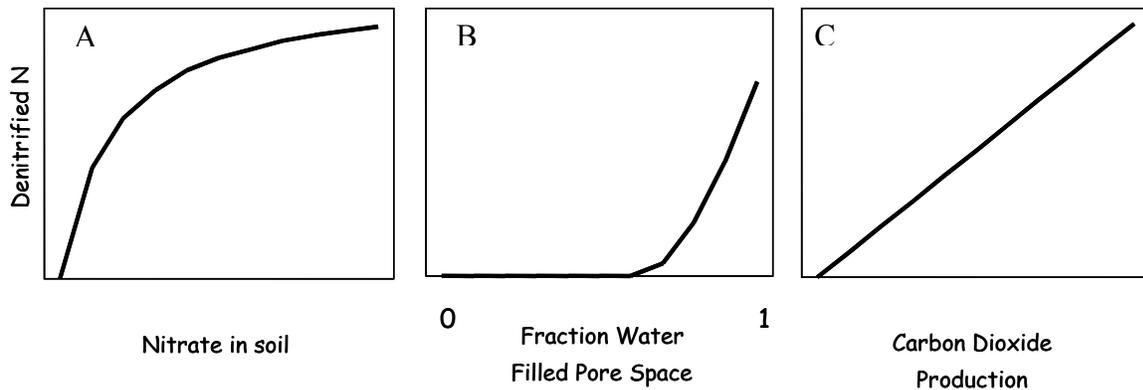


Figure A2.3 Response in amount of N emitted due to denitrification to (A) nitrate in soil; (B) soil water content; (C) biological activity

The N lost by denitrification, N_d , is partitioned into N_2 and N_2O using simple linear relationships shown in Figure A2.4.

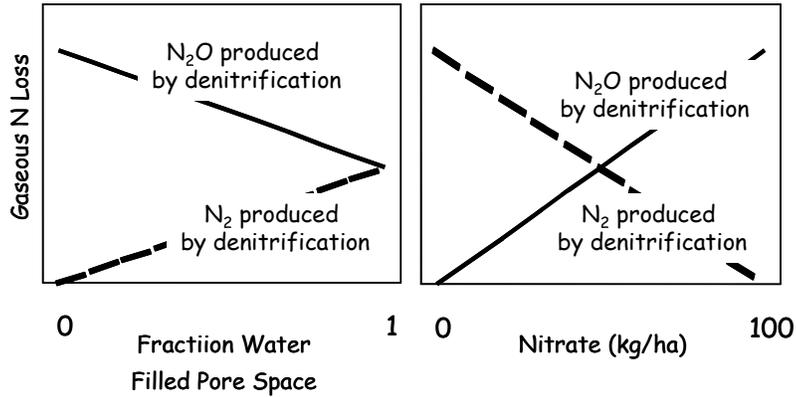


Figure A2.4 Partitioning of N lost by denitrification into N_2 and N_2O with respect to soil water and nitrate content

The amount of N_2 gas lost by denitrification, N_{d,N_2} , is given (in $kg\ N\ ha^{-1}$) by

$$N_{d,N_2} = p_{water} \times p_{NO_3} \times N_d,$$

where N_{d,N_2} is amount of N_2 gas lost by denitrification ($kg\ N\ ha^{-1}$), p_{water} proportions denitrification into N_2 according to the water content of the soil, ($p_{water} = d_{N_2,FC} \times \frac{W}{W_{max}}$

where $d_{N_2,FC} = 0.5$ is the proportion of denitrified N lost as N_2 at field capacity), and p_{NO_3} proportions denitrification into N_2 according to the nitrate content of the soil,

$$(p_{NO_3} = 1 - \left(\frac{NO_3}{N_d \times d_{zero}} \right)) \text{ where } d_{zero} = 0.1 \text{ is the proportion } \frac{NO_3}{N_d} \text{ at which } N_2 \text{ emission}$$

falls to zero and all denitrified N is lost as nitrate.

The amount of N_2O gas lost by denitrification, N_{d,N_2O} is similarly calculated (in $kg\ N\ ha^{-1}$) by the following expression,

$$N_{d,N_2O} = (1 - (p_{water} \times p_{NO_3})) \times N_d.$$

Nitrification: DNDC predicts nitrification rates by tracking nitrifier activity and NH_4^+ concentration. Growth and death rates of nitrifying bacteria are calculated as a function of DOC concentration, temperature and moisture, based on Blagodatsky & Richter (1998) and Blagodatsky *et al.* (1998). Nitrification rates are then predicted as a function of the nitrifier biomass, NH_4^+ concentration and pH. Nitrification-induced NO and N_2O production are calculated as a function of nitrification rate and temperature (Li, 2000).

NGAS takes a much simpler approach, calculating nitrification directly as a function of water filled pore space (WFPS), pH, temperature and soil NH_4^+ levels. Each of the four variables controls the fraction of nitrification that occurs under the given conditions. The effect of WFPS is based on Doran *et al.* (1988) and is a function of soil texture. It

consists of two bell-shaped curves where maximum nitrification occurs at of 0.55 for the sandy soils and 0.61 for medium and fine-textured soils. The effect of temperature is an exponential function with a Q_{10} of 2, based on the work of Sabey *et al.* (1959):

$$T'_{eff} = -0.06 + 0.13 \exp(0.07 \times T)$$

where T'_{eff} describes the effect of temperature on nitrification, and T is the air temperature at the soil surface ($^{\circ}\text{C}$).

The effect of pH on nitrification is an S-shaped curve based on data presented by Gilmore (1984) and Motavalli *et al.* (1996), where the optimum pH is taken to be pH 7. Finally, NH_4^+ affects nitrification based on Malhi & McGill (1982):

$$NH_{4\text{eff}} = 1 - \exp(-0.0105 \times NH_4)$$

where $NH_{4\text{eff}}$ is the effect of NH_4^+ on the nitrification fraction and NH_4 is the concentration of ammonium in the soil.

Maximum nitrification is set as a constant based on field data from Mosier *et al.* (1991), although (Parton *et al.* 1996) suggests it varies as a function of soil texture. None of these factors are parameterised for highly organic or very acidic soils. Frohling *et al.* (1998) modified these relationships when incorporating NGAS into CENTURY, changing the temperature curve according to Malhi & McGill (1982), specifying a maximum nitrification rate of 10 % of soil NH_4^+ per day, and setting N_2O emissions as 2 % of nitrification.

Using a similar but simpler approach, nitrification is simulated in ECOSSE according to the amount of ammonium in the soil layer, and is modified according to the temperature, moisture content and soil pH. The amount of N nitrified, N_n , is calculated (in kg N ha^{-1}) using the expression for nitrification developed by Bradbury *et al.* (1987), i.e.

$$N_n = NH_4 \times \exp(-0.6 \times m'_{temp} \times m'_{water} \times m'_{pH}),$$

where NH_4 is the amount of ammonium in the soil (kg N ha^{-1}), m'_{temp} is a rate modifier due to soil temperature, m'_{water} is a rate modifier due to soil water, and m'_{pH} is a rate modifier due to soil pH.

The expression for the water rate modifier is adapted from that used by Bradbury *et al.* (1987) to include reduced nitrification in anaerobic soils, i.e.

$$m'_{water} = \left(1 - \frac{(0.8 \times (W_{\max} - W - d_{(-1\text{bar})}))}{(W_{\max} - d_{(-1\text{bar})})} \right) \times m'_{anaer},$$

where $d_{(-1\text{bar})}$ is the water deficit on the soil layer at -1bar (mm), and m'_{anaer} reduces nitrification in anaerobic soils. If the $W < 0.9W_{\max}$, $m'_{anaer} = 1$, otherwise m'_{anaer} is given

$$\text{by } m'_{anaer} = \left(1 - W + \frac{(0.9 \times W_{\max})}{W_{\max} - (0.9 \times W_{\max})} \right).$$

The response of nitrification to the soil water content is shown in Figure A2.5 (A).

The expression for the temperature rate modifier, m'_{temp} , is given by the expression used by Bradbury *et al.* (1987) for both mineralization and nitrification, i.e.

$$m'_{temp} = \frac{47.9}{\left(1 - \exp\left(\frac{106}{T+18.3}\right)\right)},$$

where T is the air temperature ($^{\circ}\text{C}$).

The response to air temperature is shown in Figure A2.5 (B), and the response to ammonium is shown in Figure A2.5 (C).

The pH response function suggested by Parton *et al.* (1996) was used. In this model, rate is about 1 (unmodified) at about neutral pH, but decreases below neutral:

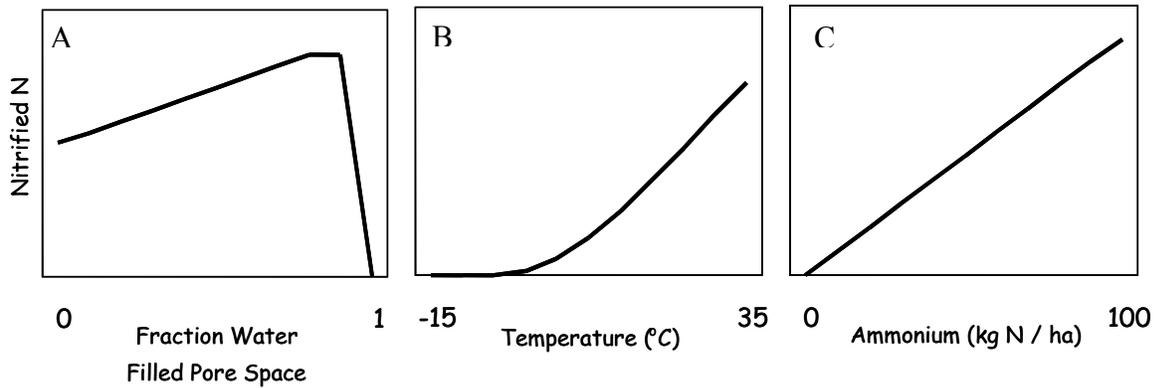


Figure A2.5 Response in amount of N nitrified to (A) soil water content; (B) temperature; (C) ammonium content

$$m'_{pH} = 0.56 + \frac{(\text{atan}(3.14 \times 0.45 \times (-5 + pH)))}{3.14}.$$

Conversion of ammonium to nitrate by nitrification is accompanied by gaseous losses of N due to complete and partial nitrification. The gaseous losses as NO and N₂O are calculated using linear relationships shown in Figure A2.6.

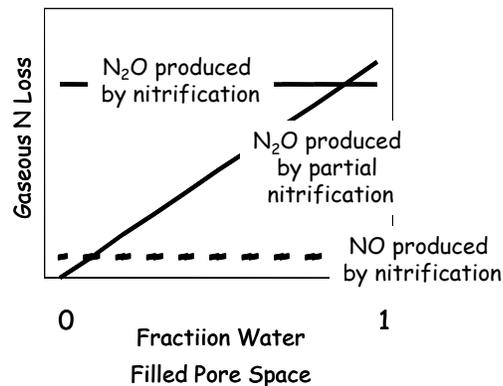


Figure A2.6 Partitioning of N lost by denitrification into N₂ and N₂O with respect to soil water and nitrate content

The amount of gas emitted as N₂O during nitrification, N_{n,N_2O} , is given (in kg N ha⁻¹) by

$$N_{n,N_2O} = \left(\left(n_{FC} \times \frac{W}{W_{\max}} \right) + (n_{gas} \times (1 - n_{NO})) \right) \times N_n,$$

where $n_{FC} = 0.2$ is the proportion of N₂O produced due to partial nitrification at field capacity, $n_{gas} = 0.02$ is the proportion of full nitrification lost as gas, and $n_{NO} = 0.1$ is the proportion of full nitrification gaseous loss that is NO.

Similarly, the amount of gas emitted as NO during nitrification, $N_{n,NO}$ is given (in kg N ha⁻¹) by

$$N_{n,NO} = n_{gas} \times n_{NO} \times N_n.$$

A2.3 Description of methane production and oxidation

Methane is an important contributor to global warming, which is produced by methanogenic bacteria in soil when decomposition occurs under anaerobic, reducing conditions. Wetlands represent the most important natural source of methane emissions to the environment. As the rate of methane emission is often reported to increase with temperature, there is potential for a positive feedback due to climate change. This emphasises the need to understand the processes that control methane emission from wetlands and how they react to both environmental and land use changes.

DNDC calculates CH₄ production as a function of DOC concentration and temperature, under anaerobic conditions where soil Eh is predicted to be 150 mV or less. CH₄ oxidation is calculated as a function of soil CH₄ and Eh. CH₄ moves from anaerobic production zones to aerobic oxidation zones via diffusion, which is modelled using concentration gradients between soil layers, temperature and soil porosity. CH₄ flux via plant transport is a function of CH₄ concentration and plant aerenchyma. Plant aerenchyma is a function of the plant growth index, which is calculated using plant age and season days. If plant aerenchyma is not well developed, or soil is unvegetated, the efflux is determined by ebullition (bubbling). In DNDC, this is assumed to occur only at the surface level, and is regulated by CH₄ concentration, temperature, porosity, and any existing plant aerenchyma.

By contrast, to the very complex approach used in DNDC, Christensen *et al.* (1996) describe methane emissions very simply as a proportion of the total heterotrophic respiration. In ECOSSE we simulate methane emissions using a process-based but simple approach, as the difference between methane production and methane oxidation, the oxidation process adding to emissions of carbon dioxide (Figure A2.7).

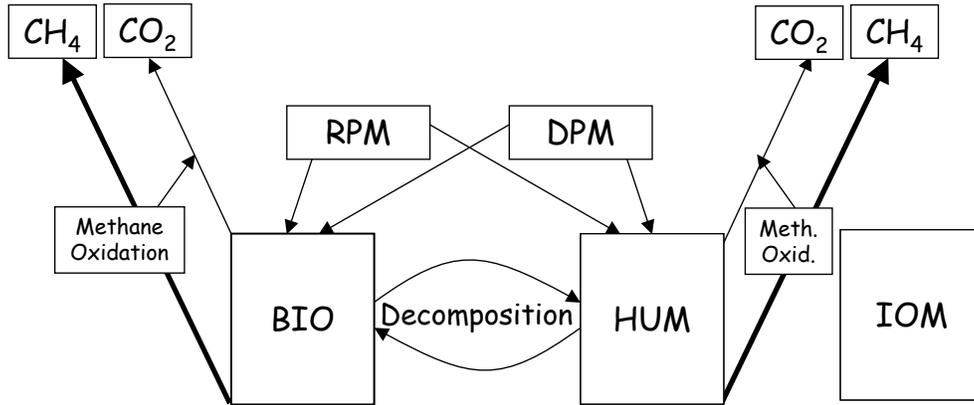


Figure A2.7 Structure of methane model in ECOSSE

Methane production during anaerobic decomposition is simulated using a similar pool approach as is used for aerobic decomposition. The amount of anaerobic decomposition (D_{an}) is calculated using the following expression:

$$D_{an} = C_{pool} \exp(k_{pool} m_{water}''' m_{temp}''' m_{pH}'''),$$

where C_{pool} is the amount of carbon in the biomass, humus, decomposable plant material or resistant plant material pool; k_{pool} is the rate constant for decomposition (initially assumed be the same as in aerobic decomposition: for decomposable plant material $k_{DPM} = 10 \text{ yr}^{-1}$; for resistant plant material $k_{RPM} = 0.3 \text{ yr}^{-1}$; for soil biomass $k_{BIO} = 0.66 \text{ yr}^{-1}$; and for humus $k_{HUM} = 0.02 \text{ yr}^{-1}$); m_{water}''' is the rate modifier for moisture; m_{temp}''' is the rate modifier for soil temperature; and m_{pH}''' is the rate modifier for pH. The difference between the rates of aerobic and anaerobic decomposition is simulated through the different functions describing these rate modifiers. Derivation of the rate modifying factors is described in the following sections.

The production of methane (C_{CH_4}) is then given by

$$C_{CH_4} = (1 - \alpha - \beta) D_{an},$$

where α is the proportion of decomposing materials partitioned to biomass, and β is the proportion partitioned to humus. The values of α and β are calculated from the efficiency of decomposition (f_{eff}) as $\alpha = (1 - f_{eff}) \left(1 - \frac{1}{1 + 1.1}\right)$ and $\beta = 1 - \alpha$. The efficiency of decomposition is set for each given soil type, and is currently assumed to be the same as for aerobic decomposition.

The oxidation of methane ($C_{CH_4 \rightarrow CO_2}$) is calculated from methane production as

$$C_{CH_4 \rightarrow CO_2} = (1 - \tau) \nu d \times C_{CH_4},$$

where τ is a transport factor (for non-transporters $\tau = 0$; for transporting non-sedges $\tau = 0.25$; for transporting sedges $\tau = 1$ (Kettunen, 2002); ν is a soil dependent factor that

accounts for different rates of diffusion and oxidation (derivation is described in the following sections); and d is the depth (cm).

Soil Temperature Rate Modifier, m'''_{temp}

Field measurements of changes in methane emissions with temperature can be difficult to unravel, as many confounding factors can contribute to the observed emissions. Christensen *et al.* (2002) report that mean seasonal temperature is the best predictor of methane fluxes on a large scale. Hargreaves *et al.* (2001) observed an exponential relationship between surface temperature (0-10 cm) and methane flux for measurements without a thaw period, with a Q_{10} of 4, while Rask *et al.* (2002) found a linear relationship between the same factors in a shallow bay area of a fen. Hargreaves & Fowler (1998) reported a linear relationship with temperature between 7 and 11°C, with a slope of $5 \mu\text{mol CH}_4 \text{ m}^{-2} \text{ h}^{-1} \text{ } ^\circ\text{C}^{-1}$ for peat wetlands in Caithness, Scotland. Other workers have also reported significant relationships between soil temperature in the surface layer and methane flux, in wet tundra sites (Christensen *et al.* 1995) and subalpine wetlands (Wickland *et al.* 1999, 2001). Although these field observations describe different responses, a positive relationship between methane emissions and temperature is generally observed (Figure A2.8).

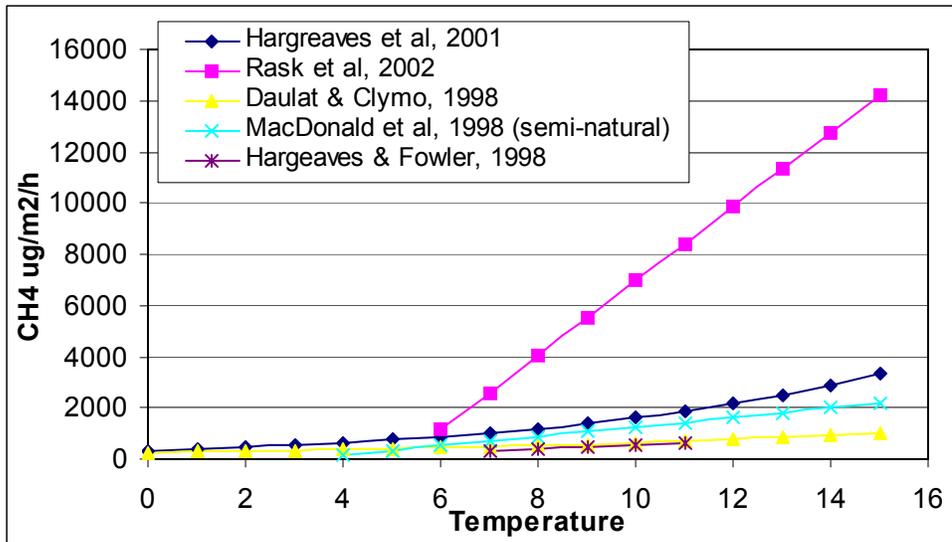


Figure A2.8 Reported responses of methane emissions to temperature

Micro- and mesocosm measurements provide a less complex picture, allowing confounding factors to be removed from the experimental setup. Daulat & Clymo (1998) reported an exponential relationship between soil temperature at 5 cm depth and mean methane flux in peat cores from Scotland. Lloyd *et al.* (1998) commented on the high sensitivity of methane fluxes from Scottish peat cores to temperature, reporting a Q_{10} of 3 in the dark. MacDonald *et al.* (1998) also report Q_{10} values of around 3 (between 5 and 15°C) for relationships between peat temperature and methane flux from peat cores from Northern Scotland, which are linear under semi-natural conditions, and exponential under controlled conditions of constant humidity and light.

Following the model of Kettunen (2002), the temperature rate modifier is given by the equation:

$$m'''_{temp} = \frac{47.91}{\exp\left(\frac{125}{(T_{soil} + 18.27)}\right) + 1},$$

where T_{soil} is the temperature of the soil layer ($^{\circ}\text{C}$)

In agreement with the observations of Daulat & Clymo (1998), Lloyd *et al.* (1998) and MacDonald *et al.* (1998), this relationship shows a Q_{10} value close to 3 between 5 and 15 $^{\circ}\text{C}$ (see Figure A2.9).

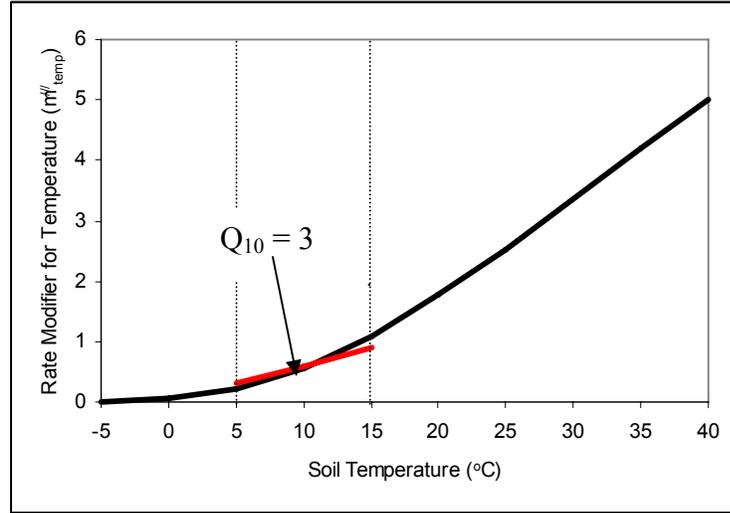


Figure A2.9 Response of anaerobic decomposition to soil temperature

Soil Water Rate Modifier, m'''_{water}

Methane emissions only occur in strongly anoxic soils (Le Mer & Roger, 2001), therefore the rate modifier is assumed to be non-zero only at water contents over field capacity.

Following the approach used wetlands DNDC (Zhang *et al.*, 2002), the rate of methane production is assumed to increase exponentially above field capacity (see Figure A2.10) and is calculated as

$$m'''_{water} = c_2 \exp((c_1 \times \vartheta_{5cm}) - 1),$$

where ϑ_{5cm} is the available water in a 5cm layer (mm); and c_1 and c_2 are fitted constants

($c_1=0.5$ and $c_2 = \frac{1}{\exp((0.5 \times \vartheta_{sat}) - 1)}$ where ϑ_{sat} is the available water at saturation (mm)).

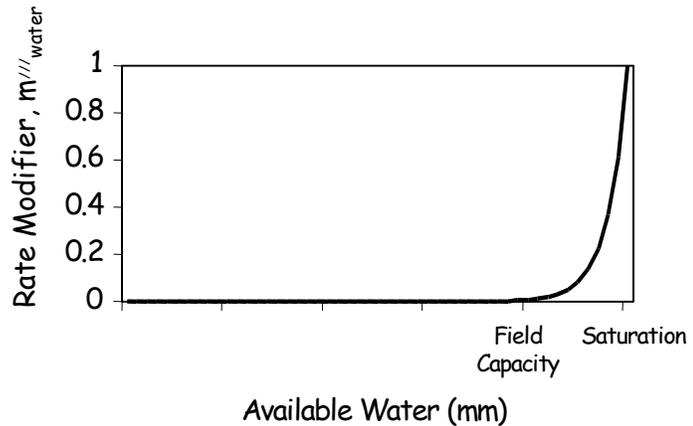


Figure A2.10 Response of anaerobic decomposition to available water in the soil

Soil pH Rate Modifier, m'''_{pH}

Methanogenic bacteria are generally reported to exhibit maximum activity under neutral or slightly higher pH conditions (Garcia *et al.* 2000) and to be very sensitive to variations in soil pH (Wang *et al.*, 1993). Garcia *et al.* (2000) reported that 68 species of methanogenic bacteria could not grow at a pH lower than 5.6.

However, methane producers can adapt to more acidic environments, as many studies have recorded methanogenic activity in soils with a lower pH. Williams and Crawford (1985) reported that a mixed bacterial culture from a Minnesota peatland produced methane at pH values between 3 and 4. Dunfield *et al.* (1993) investigated methane production in peat soil samples from temperate and subarctic areas (pH 3.5–6.3) and reported an optimum production rates at pH of 5.5 to 7.0. Inubushi *et al.* (2005) reported a positive correlation between methane production activity and soil pH ($r^2 = 0.802$, $P < 0.01$) for peat soil samples from a temperate Japanese wetland, which had a pH range of 5–7. However, more acidic Indonesian peat soils, which ranged from pH 3.9–5, showed no correlation with pH (Inubushi *et al.* 2005).

Depth can also affect the influence of soil pH on methane production. Williams & Crawford (1984) found that a pH increase from 3.2 to 5.8 increased the methane production of an incubated peat from a Minnesota peatland by 1.5 fold, for samples from 10 cm depth, and 2.2 fold for samples from 60 cm depth.

To complicate things further, some studies have reported negative relationships. Bergman *et al.* (1999) reported a negative effect of pH on methane production in incubations of peat soil from a Swedish mire, but it was only statistically significant for one of two years data. Valentine *et al.* (1994) also reported a decrease in methane production rates in peat with increasing pH. Bergman *et al.* (1998) reported both positive and negative relationships with pH for peat originating from different plant communities within the same mixed mire site in Sweden, and suggested that conflicting results may be due to competition for hydrogen between methanogens and homoacetogens, increasing pH favouring the later since it increases the degree of dissociation of acetate, or due to differences in cation exchange capacity (CEC).

It is clear that the response of methane production to pH is very complex. Some of these conflicting results may have been confounded by methane oxidation. However, from the above results it can be stated that the rate of methane production is usually at an optimum ($m'''_{pH} = 1$) at around pH 7 (Garcia *et al.*, 2000; Wang *et al.*, 1993), and is close to the optimum between pH 5.5–7 (Dunfield *et al.*, 1993), decreasing to close to zero at around pH 3 (Williams & Crawford, 1984, 1985; Dunfield *et al.*, 1993).

This response can be simulated using a sigmoid relationship, as

$$m'''_{pH} = \left(1.0^{1/c_4} + \exp(c_3 \times pH) \right)^{c_4}$$

where pH is the measured pH of the soil layer, and c_3 and c_4 are constants ($c_3 = -1$, and $c_4 = -50$). The relationship between m'''_{pH} and pH is shown in Figure A2.13.

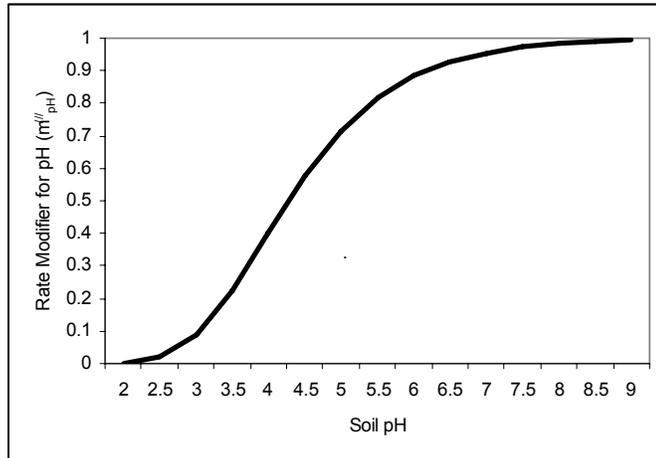


Figure A2.11 Response of anaerobic decomposition to soil pH

Soil factor for diffusion and oxidation of methane

Water table depth is one of the main factors controlling methane emissions as it determines the position of the boundary between the anaerobic and aerobic zones. When the water table is below the soil surface, oxidation of methane becomes a major controlling variable for methane efflux (Christensen *et al.*, 2000). As a result, a lower water table decreases methane emission (Blodau *et al.*, 2004) and draining peats may even convert the soil to a net methane sink (Blodau & Moore, 2003; Huttunen *et al.*, 2003; Maljanen *et al.*, 2002). Hargreaves & Fowler (1998) measured methane fluxes over a peat wetland in Caithness and related them to water table depths in different areas of the bog.

They found a negative linear relationship between depths of around 8-17 cm. Daulat & Clymo (1998) also report a linear relationship between water table depth and methane emission, with depths of more than 15-20 cm below the soil surface stopping peat cores from being a net emitter of methane (Figure A2.12). MacDonald *et al.* (1998) found a reduction in methane emissions from

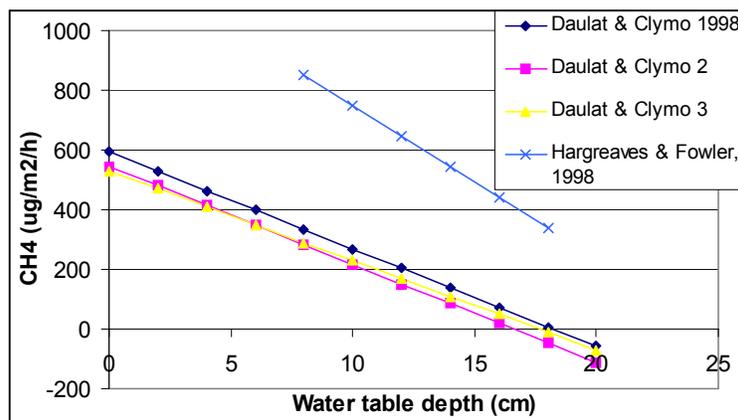


Figure A2.12 Reported responses of methane emissions to water table depth

Scottish peat cores with increasing water table depth, but in this case, the observed relationship is a decay curve and the cores do not become net sinks for methane even when the water table is 40 cm below the surface. Moore & Dalva (1993) found a negative logarithmic correlation with water table depth, down to 60 cm, in cores of peatland. Where the water table is above the surface, oxidation can also reduce methane emissions,

particularly if light availability allows benthic photosynthetic activity (Le Mer & Roger, 2001).

If it can be assumed that methane oxidation is close to zero when the water table is at the surface, the soil factor accounting for diffusion and oxidation of methane (ν) can be calculated from the depth (in cm) at which methane emissions cease and the soil becomes a net methane sink (d_{sink}) as follows:

$$\nu = \frac{1}{d_{sink}} \cdot$$

This results in a value of $\nu = 0.056 \text{ cm}^{-1}$ for the soils of Daulat & Clymo (1998), and $\nu = 0.04 \text{ cm}^{-1}$ for the soils of Hargreaves and Fowler (1998). The methane emissions calculated in this way are shown in figure A2.13. Further work is required to determine the relationship between soil type and the depth at which the soil becomes a net methane sink.

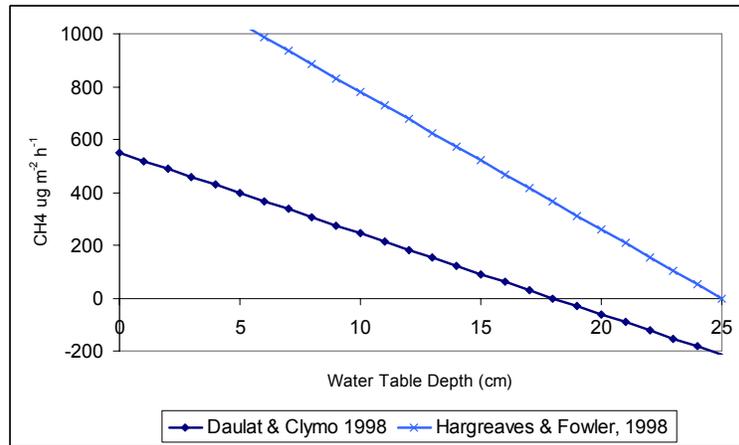


Figure A2.13 Reduction in methane emissions with depth of water table, as calculated by ECOSSE

Methane oxidizing bacteria (methanotrophs) are more tolerant to pH variations than methanogens, with a reported optimum pH of 5.0 to 6.5 in temperate and subarctic peats (Dunfield *et al.*, 1993), and have been discovered in peat soils below pH 4.7 using molecular ecological methods (McDonald *et al.*, 1996). However, Hutsch *et al.* (1994) reported that, in a non-fertilised permanent grassland at the Rothamsted experimental station, a decrease in pH from 6.3 to 5.6 reduced methanotrophy by almost half. As a first approximation, it is assumed that methane oxidation is not influenced by soil pH.

A2.4 Description of dissolved organic matter (C and N): turnover and losses

Dissolved Organic Carbon (DOC) is a potentially important component of carbon export from the soil system under highly organic conditions. Dissolved Organic Nitrogen (DON) is closely linked to the production of DOC, and Jörgensen & Richter (1992) demonstrated that the two soil characteristics must be considered together in order to avoid unrealistic simulations of carbon or nitrogen concentration in the soil. Evans *et al.* (2005) and Evans *et al.* (in press) give a convincing argument that DOC production in organic soils is increasing, not due to climate change-induced oxidation, but instead as a result of soils recovering from acid deposition, which suppresses DOC solubility. They demonstrate a link between increased temperature, declining sulphur deposition and sea-salt loading. Strong links are already known to exist between soil water pH and DOC solubility, with

an inverse relationship between mineral and organic acid export from soils (Krug & Frink, 1983). Palmer *et al.* (2001) used a combination of stable isotope and ^{14}C -dating to identify the main sources and processes controlling DOC production in a temperate non-forested watershed underlain by mostly organic and podzolic soils. They found that the main source of DOC was the readily-decomposable carbon, such as leaf litter, rather than the longer term storage of SOM. They also showed that wetter soils export more ‘recent’ carbon than dry soils: DOC from dry soils contains more of the older carbon than does DOC from wet soils.

DyDOC is a model of DOC production that utilises a layered soil description (Michalzik *et al.*, 2003), and contains different organic fractions. Three processes are represented in DyDOC: hydrology, metabolism and sorption. Hydrology is modelled in terms of macropores and micropores, between which dissolved organic carbon diffuses. Metabolism is modelled using relatively simple relationships to define the transformations between the three pools of organic carbon, with only temperature being a variable. Sorption occurs to and from soil solids from the different pools, and is controlled by DOC type, soil solution pH and cation content, and the nature of the soil solids (texture etc.). In ECOSSE, we follow a similar approach, simulating the hydrological, metabolological and sorptive processes using the 5 soil organic matter pools: decomposable and resistant plant material, soil biomass, humus, and inert organic matter. A further organic matter pool is introduced, containing the DOC and DON that is in solution (see figure A2.14).

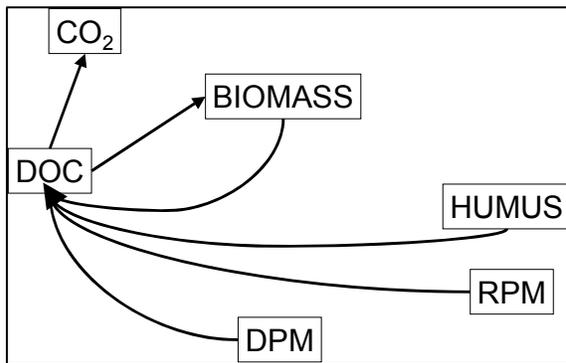


Figure A2.14 Pool structure for DOC production and uptake

This dissolved component is then susceptible to leaching by the same processes as already exist in the model to describe leaching losses. According to the model of Aguilar & Thibodeaux (2005), there are two fractions of DOC, one fraction that is readily available in the soil solution at all times, and another that is created from slower decomposition processes. This observation is consistent with the transfer into the DOC pool during decomposition of material from the rapidly turning over biomass and plant material pools, as well as from the more slowly turning over humus pool.

In DyDOC production by metabolism is calculated using a Q_{10} relationship with a rate dependent only on temperature:

$$\Delta C_{pool} = -k_{pool} Q_{10}^{(T/10)} C_{pool} \Delta t$$

where k_{pool} is the rate constant for DOC production specific to the pool; Q_{10} is a constant, usually set to 2.0; T is the temperature ($^{\circ}\text{C}$); C_{pool} is the amount of carbon in the pool (kg C ha^{-1}) and t is the size of the timestep.

In ECOSSE we differentiate the equation, and include rate modifiers for moisture (m''_{water}), temperature (m''_{temp}), crop cover (m''_{crop}) and pH (m''_{pH}) as used in the existing

calculation of SOM decomposition. The amount of DOC produced ($C_{pool \rightarrow DOC}$) by a given pool is then calculated as:

$$C_{pool \rightarrow DOC} = C_{pool} \exp(k_{pool} m''_{water} m''_{temp} m''_{crop} m''_{pH})$$

where k_{pool} is set to 0.0001 day^{-1} for decomposable plant material; $0.000005 \text{ day}^{-1}$ for resistant plant material; 0.00005 day^{-1} for soil biomass; and $0.000002 \text{ day}^{-1}$ for humus.

The moisture modifier (m''_{water}) is calculated from the available water at saturation (ϑ_{sat}), the available water at field capacity (ϑ_{FC}) and the actual available water in the 5cm layer (ϑ_{5cm}) (all in mm) as follows:

$$\begin{aligned} \text{if } (\vartheta_{FC} - \vartheta_{5cm}) > 4\text{mm then } m''_{water} &= (1 - 0.2) \times \frac{(\vartheta_{FC} - \vartheta_{5cm} - 4)}{\vartheta_{FC} - 4}; \\ \text{if } (\vartheta_{FC} - \vartheta_{5cm}) \leq 4\text{mm then } m''_{water} &= 1.0; \text{ and} \\ \text{if } (\vartheta_{5cm} > \vartheta_{FC}) \text{ then } m''_{water} &= \frac{(\vartheta_{sat} - \vartheta_{5cm})}{\vartheta_{sat} - \vartheta_{FC}}. \end{aligned}$$

The temperature rate modifier is calculated from the temperature of the soil layer in $^{\circ}\text{C}$ (T_C) as follows:

$$\begin{aligned} \text{if } T_C < -10^{\circ}\text{C}, m''_{temp} &= 0; \text{ and} \\ \text{if } T_C \geq -10^{\circ}\text{C}, m''_{temp} &= \frac{47.91}{\left(1 + \exp\left(\frac{106.06}{T_C + 18.27}\right)\right)}. \end{aligned}$$

The crop rate modifier is set as follows:

$$\begin{aligned} \text{if no crop, } m''_{crop} &= 1.0; \\ \text{if crop present, } m''_{crop} &= 0.6. \end{aligned}$$

Finally, the pH rate modifier is calculated as discussed in section 2.3.7 as

$$m''_{pH} = \frac{pH - pH'_{min}}{pH'_{max} - pH'_{min}},$$

where pH is the pH of the soil layer; and pH'_{min} and pH'_{max} are the pH values at which the minimum and maximum rate of decomposition occur respectively.

The carbon in the DOC pool (C_{DOC}) can further decompose to produce additional biomass ($C_{DOC \rightarrow BIO}$) following a similar expression to the above:

$$C_{DOC \rightarrow BIO} = C_{DOC} \exp(k_{DOC} m''_{water} m''_{temp} m''_{crop} m''_{pH}),$$

where k_{DOC} is set to 0.005 day^{-1} .

The decomposition to biomass produces a fixed proportion of carbon dioxide ($C_{DOC \rightarrow CO_2}$), i.e.

$$C_{DOC \rightarrow CO_2} = f_{eff} C_{DOC \rightarrow BIO},$$

where f_{eff} is a fraction representing the efficiency of decomposition, and is set according to soil type to the same value as used in the decomposition routines. Having calculated the changes in carbon, the transformations in DON are calculated using the C:N ratio of the decomposing pool. Leaching of DOC takes place only for the portion that is not sorbed onto soil solids. The proportional availability (DOC_{SOLV}) of DOC is dependent on pH using the following relationship

$$DOC_{SOLV} = C_{DOC} (pH - 3) / 4,$$

giving a maximum availability at $pH = 7$, and a minimum at $pH = 3$. If $pH < 3$ then $DOC_{SOLV} = 0$, while if $pH > 7$ then $DOC_{SOLV} = C_{DOC}$.

A2.5 Improved layer structure in the soil profile

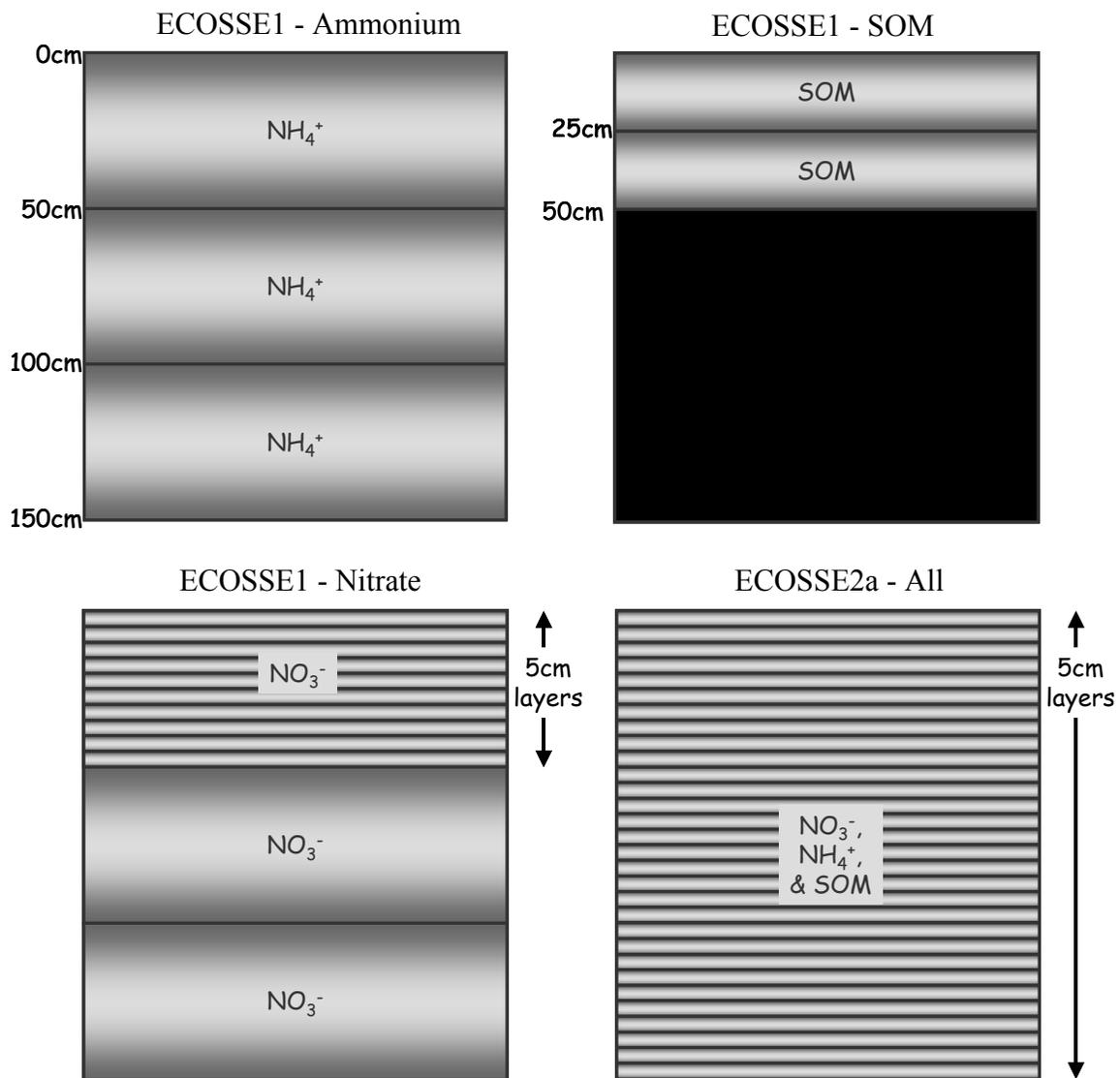


Figure A2.15 Distribution of ammonium, nitrate and soil organic matter in original model (ECOSSE1) and model with improved layers (ECOSSE2a)

In SUNDIAL-MAGEC, the soil is divided into different layers for the different components of the soil: ammonium is simulated in 50cm layers down the soil profile; SOM is simulated in 25cm layers to a maximum depth of 50cm; and nitrate is simulated in 5cm layers for the top 50cm, followed by 50cm layers for the remainder of the profile. This structure does not allow C and N turnover in deep organic soils to be simulated. In ECOSSE, the layers are unified into 5cm layers throughout (Figure A2.15), so allowing C and N turnover to be simulated to depth. The soil parameters are also described in 5cm layers, so allowing more detailed descriptions of soil profile characteristics to be included. This is important in highly organic soils where the organic layer can be very deep.

A2.6. Initialisation of the size and characteristics of the soil organic matter pools

In SUNDIAL-MAGEC, the status of the SOM at the start of the simulation is entered as a soil parameter. These parameters describe not only the total amount of organic matter in the soil, but also its activity, which is an important driver for the amount of C and N released from the organic matter during the simulation. For mineral soils, these parameters have been derived from measurements of soil C in long-term experiments, for which average climate and long-term land use can be estimated. The standard parameters derived for mineral soils are not transferable to organic soils. Furthermore, organic soils contain a higher percentage of SOM, so the influence of measured C content on C and N release is more pronounced. It is not likely that fixed soil parameters describing amount and activity of SOM will provide an accurate simulation of C and N release in organic soils: these factors should instead be entered as input variables that change for each simulation.

The developments contained in ECOSSE allow the amount and activity of SOM to be estimated using two additional input variables: the dominant land use over the last 10,000 years, and the measured soil C content at the start of the simulation. If it can be assumed that the soil is close to its equilibrium SOM content, the activity of the SOM can be estimated using an equilibrium run of ROTH-C (Coleman & Jenkinson, 1996, Figure A2.16). The equilibrium run has been built into the initialisation routines of ECOSSE.

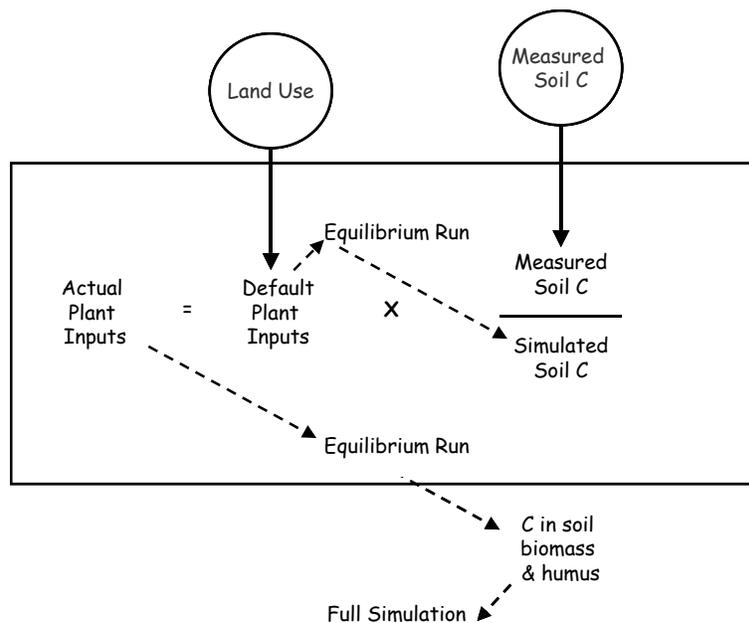


Figure A2.16 Initialisation of amount and activity of soil organic matter

The dominant land use determines the pattern of plant inputs used in the ROTH-C equilibrium run. The distributions used, obtained from Smith *et al.* (2005), are shown in Figure 2.20. These distributions are used as the default monthly plant input of C in a first equilibrium run. The equilibrium run calculates the amount of C in each SOM pool. The total C at equilibrium can be obtained by summing the C content of these pools. The annual input of plant C needed to achieve the measured soil C at equilibrium can then be estimated from the ratio of the measured and simulated soil C and the default plant input used in the run (Figure A2.17). The estimated annual plant C input is then redistributed as specified by the dominant land use (Figure A2.17), and used in a further equilibrium run. This run, calculates the amount of C in each SOM pool, given the new estimate of input plant C. The annual input of plant C needed to achieve the measured soil C at equilibrium is then re-estimated from the measured and new simulated soil C. This procedure is continued until the measured and simulated values of soil C differ by less than $0.00001 \text{ t C ha}^{-1}$. The final estimates of monthly inputs of plant C and the amount of C in each SOM pool are then used in the full simulation.

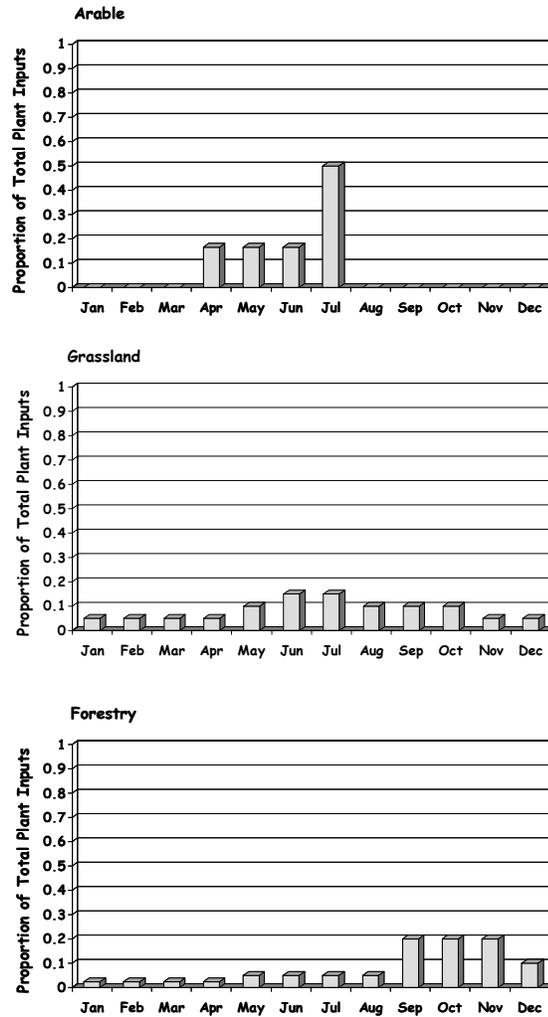


Figure A2.17 Distribution of plant inputs for 3 major land uses

A2.7 Incorporation of the effect of pH on soil processes

pH effect on rate constant: A significant effect of soil pH on the rate of decomposition has been observed in many studies (e.g. Andersson and Nilsson, 2001; Hall *et al.* 1997; Situala *et al.* 1995). High soil acidity is generally considered to limit the activity of decomposers but studies that have manipulated pH in the field or laboratory have found conflicting results. This can be explained by the interaction between soil pH, organic matter decomposition and nutrient cycling (Binkley and Richter, 1987).

Motavalli *et al.* (1995) found a positive correlation between biological activity and soil pH in soils amended with ^{14}C labelled plant residues. Similarly, Situala *et al.* (1995) used

acid irrigation to examine the effect of low pH and reported that pH 3 produced CO₂ fluxes 20 % lower than pH 4 and 5.5, between which there was no significant difference. Persson and Wiren (1989) reported increasing the acidity of forest soil from pH 3.8 to 3.4 reduced CO₂ evolution by 83 % and from pH 4.8 to 4 by 78 %.

However, these effects differ with redox conditions. Increases in pH have been reported to increase CO₂ production 1.4-fold under anaerobic conditions but decrease it by 53 % under aerobic conditions (Bridgham and Richardson 1992). Bergman *et al.* (1999) compared CO₂ production rates at pH 4.3 and 6.2, and found that under anaerobic conditions rates were 21 (at 7°C) and 29 (at 17°C) times greater at the more neutral pH, while under aerobic conditions rates were 3 times greater at 7°C but pH had no significant effect at 17°C.

In a simple regression model, Reth *et al.* (2005) included the effect of soil pH in terms of the deviation of the soil pH (pH) from the optimum pH for decomposition (pH_{opt}), and the sensitivity of the decomposition processes in this soil to pH (pH_{sens}):

$$W_{pH} = \exp \left[- \left(\frac{pH - pH_{opt}}{pH_{sens}} \right)^2 \right].$$

The plot in Figure A2.18 shows how Reth's model can be used to simulate the aerobic observations described above.

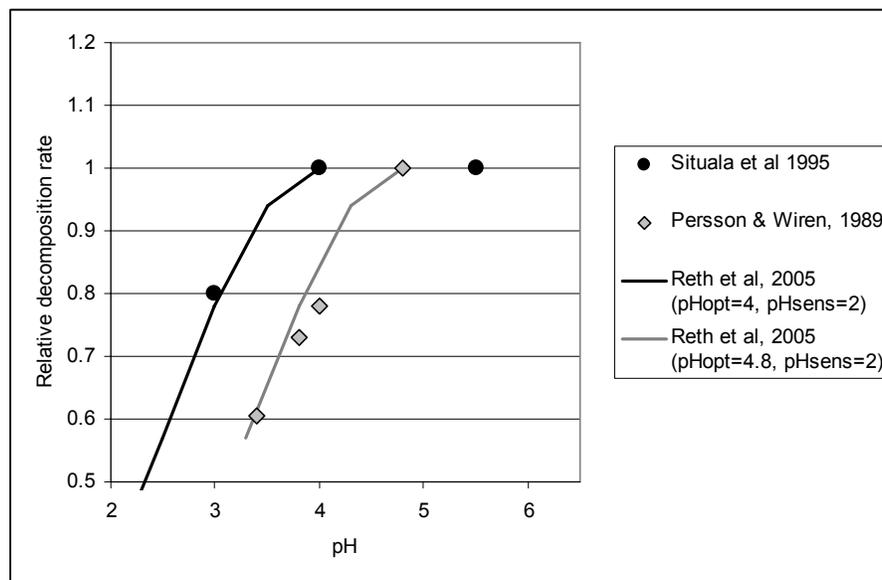


Figure A2.18 Fitted equation by Reth *et al.* (2005) to the observations of changes in aerobic decomposition with pH made by Situala *et al.* (1995) and Persson & Wiren (1989).

The plots show that Reth's approach can be used to accurately describe the response of aerobic decomposition to pH, if the optimum pH for decomposition and the sensitivity to pH are known, and the response is assumed to flatten off above the optimum pH and

below a minimum pH. The difficulty in implementing this in a functional model is in knowing what these values should be.

The influence of soil pH on decomposition is not implemented in the SUNDIAL model as it was originally designed to work in well-managed arable soils, and so it could be assumed that the pH was close to neutral (Bradbury *et al.*, 1993). This assumption breaks down in natural and managed highly organic soils, where the pH is more variable. The above discussion suggests that the implementation of decomposition in ECOSSE should include a different description of the effects of pH on aerobic and anaerobic decomposition. In an approach that follows that of Reth *et al.*, 2005, but with a simplified formula that uses more explicit terms, aerobic decomposition is described as proceeding at an optimum rate (rate modifier $W_{pH} = 1$) until the pH falls below a critical threshold (pH_{max}), after which the rate of decomposition falls to a minimum rate (rate modifier $W_{pH} = W_{pH,min}$) at pH_{min} (see Figure A2.19a):

$$W_{pH} = W_{pH,min} + (1 - W_{pH,min}) \left(\frac{pH - pH_{min}}{pH_{max} - pH_{min}} \right).$$

Similarly, anaerobic decomposition should proceed at an optimum rate (rate modifier $W_{pH}' = 1$) until the pH falls below a critical threshold (pH_{max}'), after which the rate of decomposition falls to a minimum rate (rate modifier $W_{pH}' = W_{pH,min}'$) at pH_{min}' (see Figure A2.19b):

$$W_{pH}' = W_{pH,min}' + (1 - W_{pH,min}') \left(\frac{pH - pH_{min}'}{pH_{max}' - pH_{min}'} \right).$$

That these much simpler equations can equally be used to describe the observations of aerobic and anaerobic decomposition is demonstrated in figure A2.20. In order to use

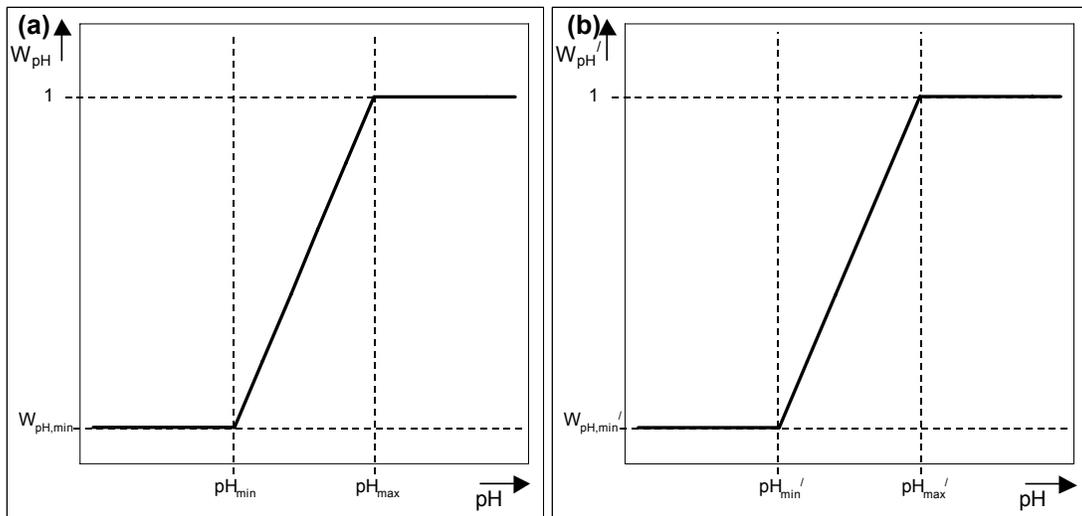


Figure A2.19 Decomposition rate modifiers for pH; (a) W_{pH} = aerobic rate modifier; (b) W_{pH}' = anaerobic rate modifier.

these equations in a range of soil types, experimental data is needed to obtain generic parameters for the soils according to texture and organic matter content (as required in

large scale simulations). Alternatively, observations of changes in CO₂ production with pH could be used to set the parameters at a specific site.

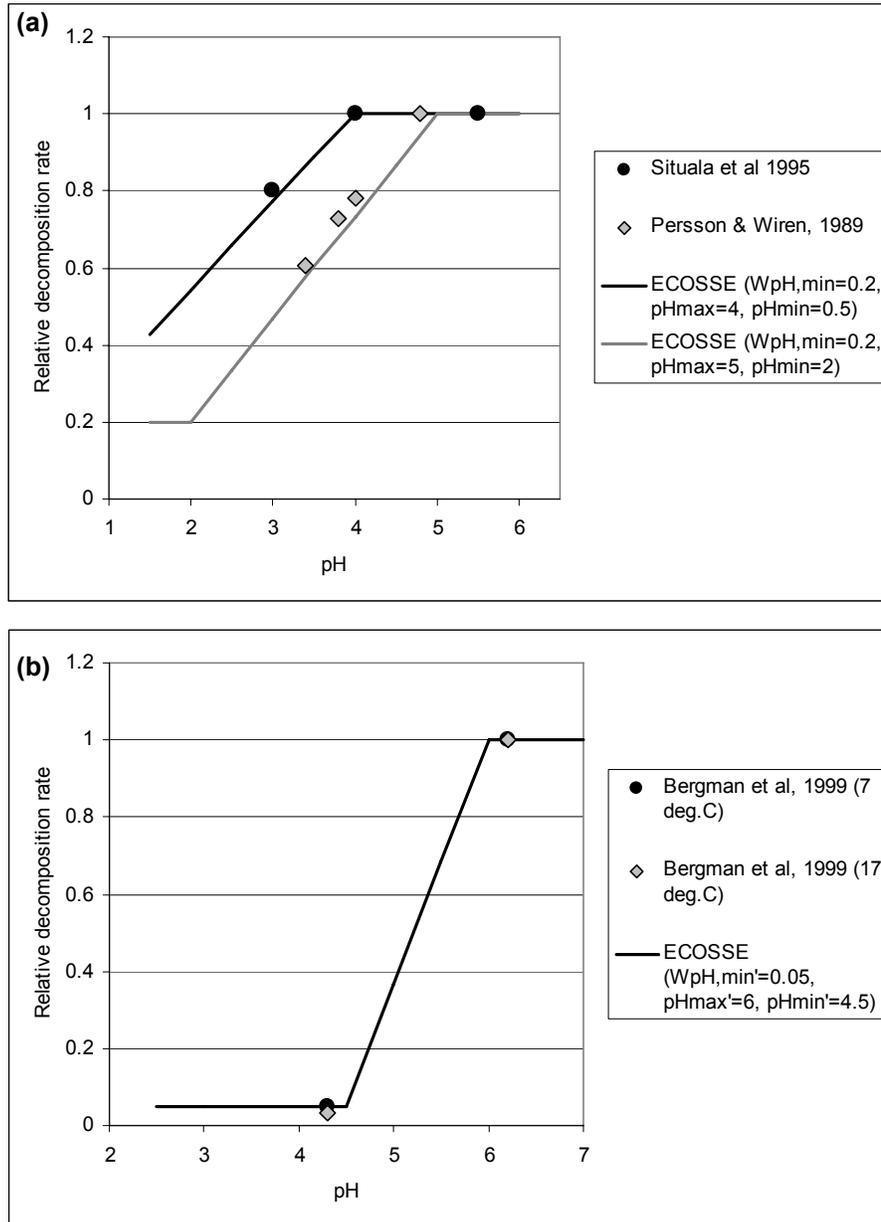


Figure A2.20 ECOSSE equation for effect of soil pH on rate constant, fitted to (a) changes in aerobic decomposition with pH (Situala *et al.*, 1995; Persson & Wiren, 1989); (b) changes in anaerobic decomposition with pH made (Bergman *et al.*, 1999)

pH effect on stable C:N ratio: Change in pH has also been observed to result in a change in the C:N ratio of the organic matter in the soil. The stable C:N ratio is an important driver of mineralization or immobilisation of N during the decomposition of organic matter. In SUNDIAL-MAGEC, the stable C:N ratio of SOM is assumed to be 8 (Bradbury *et al.*, 1987). If organic matter has a C:N ratio greater than 8, then the organic matter is deficient in C, and N will be mineralised to ammonium during decomposition. If organic matter has a C:N ratio less than 8, then the organic matter is deficient in N, and N will be immobilised during decomposition, first from the ammonium, and then from the nitrate pool.

The choice of a stable C:N ratio of 8 is a result of the weighted average of the C:N ratios of the decomposing fungi and bacteria in the soil. However, as the soil pH falls, the ratio of fungi to bacteria increases because fungi are more tolerant of acidic conditions. Because the C:N ratio of fungi is higher than the C:N ratio of bacteria, this change in the population is accompanied by an increase in the stable C:N ratio of the SOM.

This is due to the decomposer community in the soil becoming dominated by fungi with a higher C:N ratio than bacteria. This suggests that the stable C:N ratio of the organic matter pools in ECOSSE should change with changing pH. Again, in the absence of more detailed experimental data, a simple linear approach is used (see Figure A2.21).

The proportion of bacteria in the soil, p_{bac} , is calculated as

$$p_{bac} = b_{min} + \left((b_{max} - b_{min}) \times \frac{pH - pH_{min}}{pH_{max} - pH_{min}} \right),$$

where $b_{min} = 0.2$ is the minimum proportion of bacteria found in the soil, $b_{max} = 0.5$ is the maximum proportion of bacteria found in the soil, pH is the pH of the soil, pH_{min} is the pH at which minimum soil bacteria occurs, and pH_{max} is the pH at which maximum soil bacteria occurs.

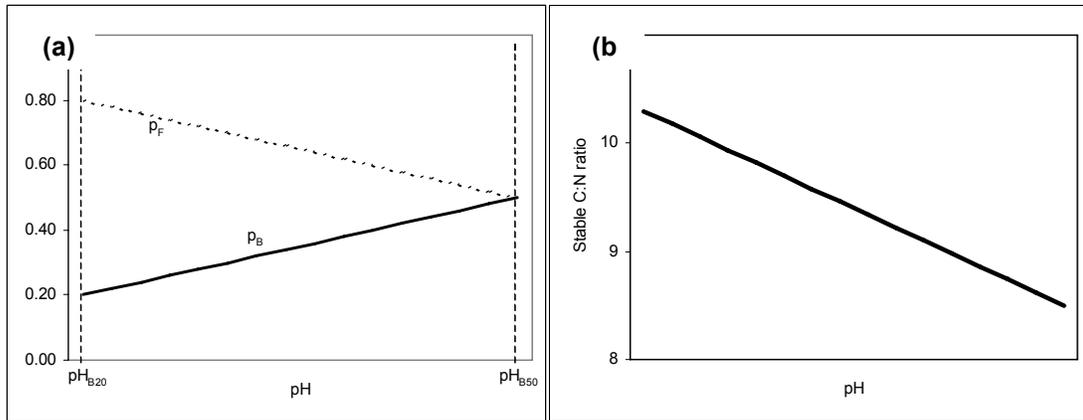


Figure A2.21 Calculation of change in stable C:N ratio with pH; (a) proportion of fungi (p_F) and bacteria (p_B); (b) stable C:N ratio.

Similarly, the proportion of fungi in the soil, p_{fun} , is calculated as

$$p_{fun} = 1 - \left(b_{min} + \left((b_{max} - b_{min}) \times \frac{pH - pH_{min}}{pH_{max} - pH_{min}} \right) \right).$$

The stable C:N ratio of the SOM, $(C : N)_{stable}$, can then be calculated from the proportion of bacteria and fungi in the soil, and the typical C:N ratios of bacteria and fungi, $(C : N)_{bac} = 5.5$ and $(C : N)_{fun} = 11.5$ respectively, i.e.

$$(C : N)_{stable} = (p_{bac} \times (C : N)_{bac}) + (p_{fun} \times (C : N)_{fun}).$$

The change in stable C:N ratio of SOM between pH 5.5 and 4 is shown in Figure A2.22.

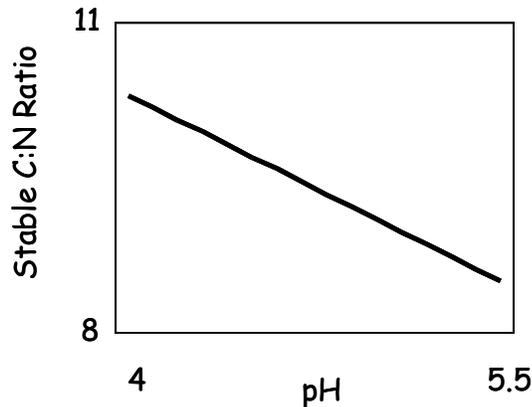


Figure A2.22 Change in stable C:N ratio with pH

A2.8 Incorporation of the effect of saturated conditions on soil processes

There are a number of different ways to express soil water content. Volumetric water content (% or mm/mm) gives a good indication on how much water is there, but to assess the effect on soil processes also soil specific parameters must be taken into account. There are 3 descriptors of soil water behaviour that are needed: permanent wilting point, field capacity and saturation. Permanent wilting point is the limit below which plants can extract no more water from the soil. Field capacity is the maximum water that the soil can hold against gravity, and therefore the maximum soil water content that can occur in a freely drained soil. As agricultural soils are usually freely drained, agricultural models usually ignore water contents above field capacity. The real maximum water content of a soil is called saturation. That is the maximum amount of water that can fit in the soil before it overflows, when all pores are filled with water.

Field capacity is generally assumed to be the optimum water content for biological processes because both water and oxygen are easily available. Below field capacity water availability may be limiting; above field capacity oxygen availability may be limiting. This can lead to reduction of the rate of some processes, but also to whole new processes taking place (such as methane and nitrous oxide formation) as electron acceptors other than oxygen are used.

In models where the purpose is to express possible water limitation, water content is usually expressed as available water, as is the case in ECOSSE.

When water content < permanent wilting point, available water = 0.

When water content > permanent wilting point, available water = water content – permanent wilting point.

In models where the purpose is to predict methane emissions and denitrification, water contents above field capacity must also be considered. Whilst methane emissions occur in strongly anoxic soil, denitrification occurs at water contents below field capacity and increases with water content. The water content is more usually expressed as the percentage of soil pores filled with water.

To use equations from other models for the effect of soil water content on nitrous oxide emissions it is necessary to calculate water filled pore space from water content and vice versa. Water filled pore space can be expressed as:

$$w = 100 \cdot \frac{vw}{wp}$$

where w is water filled pore space, vw is volume of water and wp is total pore volume. The total pore volume can be calculated as: pore volume = total volume – volume of solid material. For soils, that simplifies to:

$$S_t = 1 - b / p$$

where S_t = total soil porosity, b = bulk density and p = particle density (Danielson & Sutherland, 1986). As most mineral soils are from similar material (“rock”), particle density does not vary much between mineral soils, and is usually close to 2.65 g cm^{-3} . The measured values for peat are somewhat lower: Pihlatie *et al.* (2004) measured a soil porosity for peat of 1.9 g cm^{-3} . If saturation water content is known, water filled pore space will be the percentage of that water content, as saturation water content equals the total pore volume filled. To calculate saturation water content, bulk density and particle density must be known.

Another way to express oxygen limitation is redox potential (Li, 2000). The advantage of this measure is that the effect of excess water can be integrated over time: When the soil becomes saturated, first stored oxygen will be used up. After that, other electron acceptors are used in a sequence depending on their redox potential. One of the effects of this is that methane emissions usually only starts after the soil has been saturated for weeks. This is difficult to simulate if only the water content in the individual time step is input. A simplification is to introduce a time lag in the onset of methane emissions.

Most of the models that do incorporate an effect of water saturation on decomposition simply decrease decomposition rate of all pools equally (Parton *et al.*, 1994). Whilst this can be tuned to fit most data, there is experimental evidence that that a stronger effect on resistant pool may be more mechanistically correct. Lignin, the most resistant material does not decompose at all under anaerobic conditions, and cellulose decomposes only slowly (Kögel-Knabner, 2002). Litter quality is also the main determinant of decomposition rate under anaerobic conditions (Miyajima *et al.*, 1997). There are also reports that sugar and easily decomposable material decomposes at about the same rate

under anaerobic as under aerobic condition (Bergman *et al.*, 1999), but with a larger portion of carbon is released as CH₄ rather than CO₂. Therefore, the reduction in decomposition rate under anaerobic conditions should be implemented most strongly in the RPM pool. The implication of this will be that RPM will accumulate under anaerobic conditions, as is expected during peat accumulation. That means that if conditions become aerobic, this organic matter will decompose quickly. Again, this agrees with observations. Figure A2.23 shows the structure of the model.

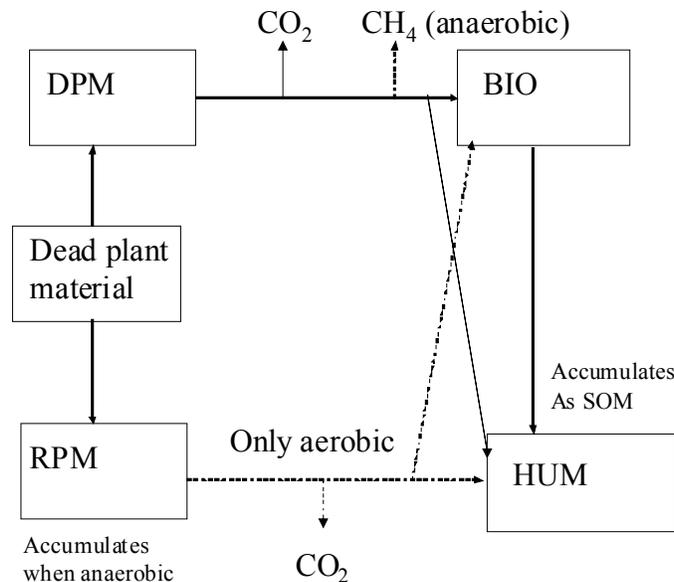


Figure A2.23 Diagram of how anaerobic decomposition can be integrated into ECOSSE soil organic matter model. The upper pathway shows decomposition of easily decomposable parts of the litter (Shirato & Yokozawa, 2006). This is envisaged to occur both at anaerobic and aerobic conditions, but under anaerobic conditions some of the carbon is released as methane. When the microbes die, parts of the dead cells are decomposed, but parts are left in the soil. This forms the majority of soil organic matter under aerobic conditions. The lower pathway shows decomposition of more resistant material (Shirato & Yokozawa, 2006). The microbes dealing with these processes are severely limited under anaerobic conditions, therefore recalcitrant material is left un-decomposed or partly decomposed. This will form the bulk of soil organic matter under anaerobic conditions.

To obtain parameters for the model, data were used that had been collected from incubation experiments where flooded and non-flooded decomposition was compared (Figure A2.24). An incubation experiment with rice straw indicated that the mass loss over 160 days under flooded conditions was about 80% of that at 50% WHC (Devevre & Horwarth, 2000). This experiment was used to fit the water modifier to the decomposition rate. Measurements of respiration in peat soils, which can be expected to contain more DPM, were more limited by anaerobic condition, 34-62% of aerobic

(Bridgeham & Richardson, 1992), so the modifier for RPM was set to a lower level (0.2), and the modifier for DPM was then fitted. It was assumed that the modifier for biomass and DPM were the same, as well as the modifier for RPM and humus. The fitting was done not only to optimize fit, but also to make any discrepancies in temperature response similar at flooded and non-flooded conditions. It is difficult to say at what water-filled pore space decomposition rate should start to decline except that it should be at water contents above field capacity. We therefore assume a linear decline from field capacity to saturation.

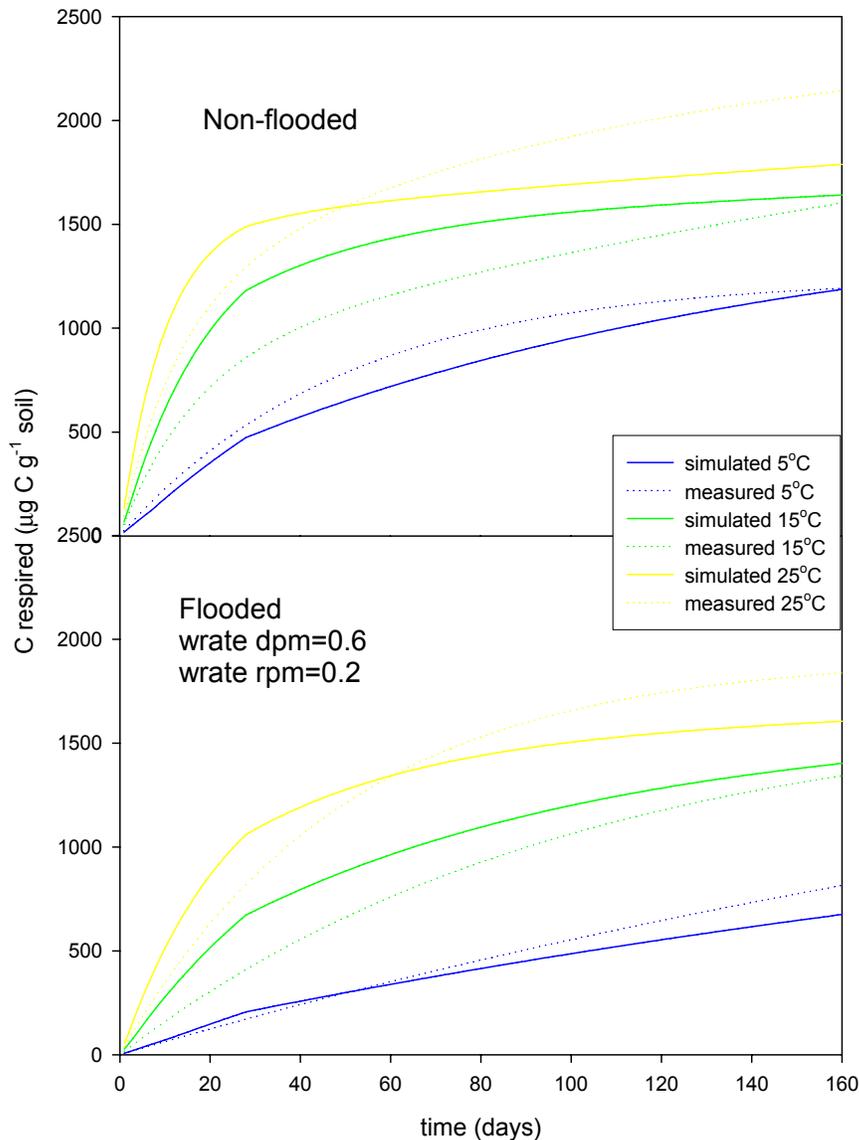


Figure A2.24 Measured and simulated graphs for mass loss during incubation of rice straw under flooded and non-flooded conditions (Devevre & Horwarth, 2000). The water modifier is set at 1 for the non-flooded treatments and fitted for the flooded treatment.

Appendix 3

Additional tables and figure for module 8

Table A3.1: Relationship between land categories of Countryside Surveys (CS) and IPCC LULUCF Guidance. For ECOSSE Grassland is split into Pasture (CS Improved grassland) and Natural (other CS grasslands).

FOREST	CROPLAND	GRASSLAND	SETTLEMENTS	OTHER
Broadleaved/ mixed	Arable	Improved grassland	Built up areas	Inland rock
Coniferous	Horticulture	Neutral grassland	Gardens	Supra littoral rock
		Calcareous grassland		Littoral rock
		Acid grassland		Standing waters
		Bracken		Rivers
		Dwarf shrub heath		Sea
		Fen, marsh, swamp		
		Bogs		
		Montane		
		Supra littoral sediment		
		Littoral sediment		

Table A3.2: Relationship between land categories of MLC data and IPCC LULUCF Guidance. For ECOSSE Grassland is split into Pasture and Natural (see Table A3.1)

FOREST	CROPLAND	GRASSLAND	SETTLEMENTS	OTHER
Broadleaved wood	Crops	Upland heath	Built up	Bare rock
Conifer wood	Market Garden	Upland smooth grass	Urban open	Sand/shingle
Mixed wood		Upland coarse grass	Transport	Inland water
Orchards		Blanket bog	Mineral workings	Coastal water
		Bracken	Derelict	
		Lowland rough grass		
		Lowland heather		
		Gorse		
		Peat bog		
		Fresh Marsh		
		Salt Marsh		
		Improved grassland		
		Rough pasture		
		Neglected grassland		

Table A3.3 Correspondence of NCMS land cover classes to GHG inventory land classes

NCMS Group	NCMS feature	Inventory land class
Grassland	Rough grassland	Natural
Grassland	Intermediate grassland	Pasture
Grassland	Smooth grassland	Pasture
Mire	Blanket mire - heather	Natural
Mire	Blanket mire - grass	Natural
Mire	Lowland mire	Natural
Heather moorland	Heather moorland	Natural
Arable	Arable	Arable
Woodland	Broadleaved woodland	Woodland
Woodland	Mixed woodland	Woodland
Woodland	Broadleaved plantation	Woodland
Woodland	Parkland	Woodland
Woodland	Coniferous woodland	Woodland
Woodland	Young plantation	Woodland
Woodland	Coniferous plantation	Woodland
Woodland	Felled woodland	Woodland
Fresh water	Lochs	Zero
Fresh water	Reservoirs	Zero
Fresh water	Rivers	Zero
Fresh water	Canals	Zero
Fresh water	Marginal inundation	Natural
Fresh water	Wet ground	Natural
Built and bare ground	Built land	Urban
Built and bare ground	Recreation	Urban
Built and bare ground	Transport corridor	Urban
Built and bare ground	Quarry	Zero
Built and bare ground	Rock	Zero
Built and bare ground	Bare ground	Zero
Bracken and scrub	Bracken	Natural
Bracken and scrub	Low scrub	Natural
Bracken and scrub	Tall scrub	Natural

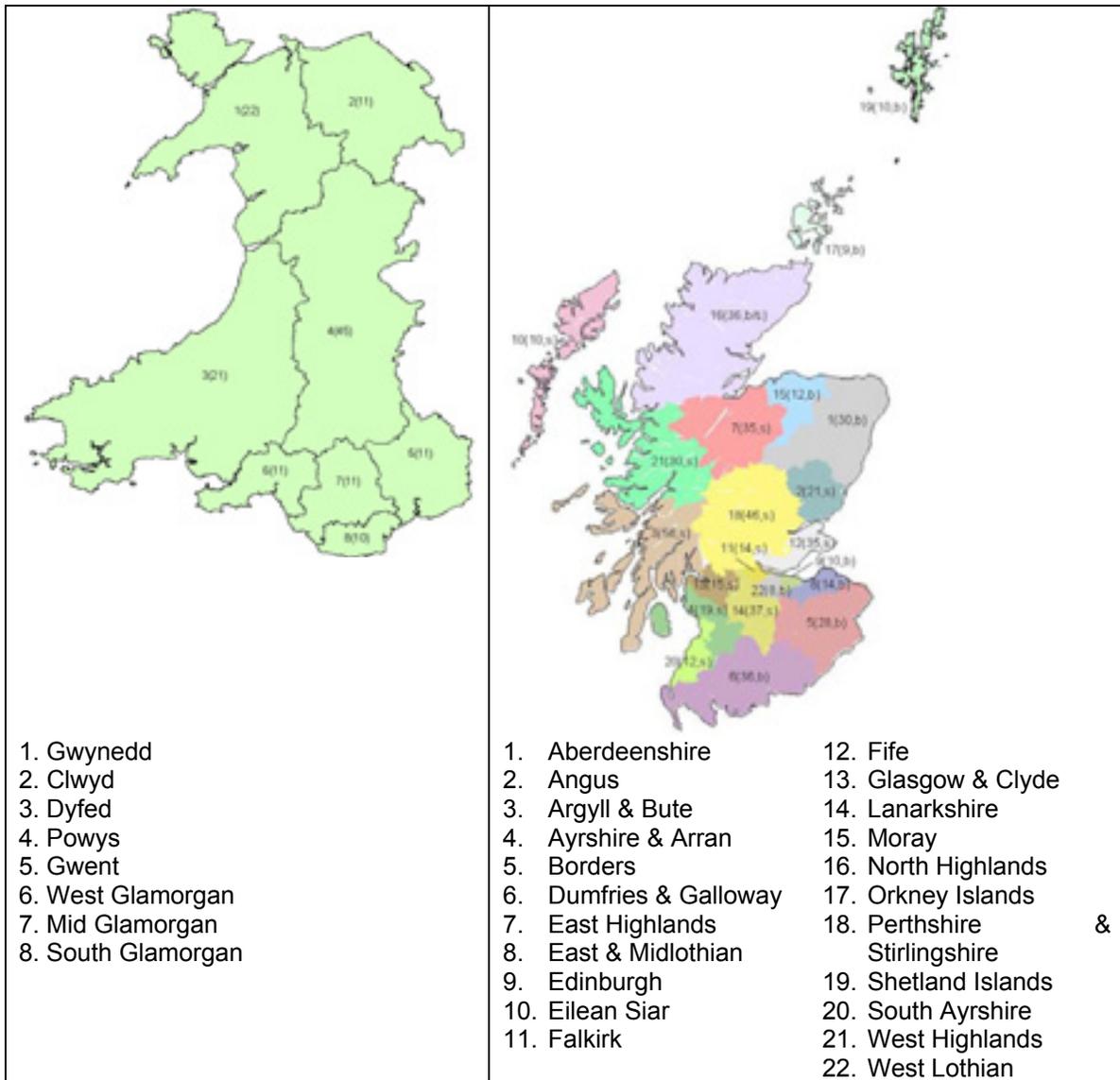


Figure A3.1 Analysis regions in (a) Wales (pre-1996 boundaries) and (b) Scotland, with the number of sample strips/squares in brackets (b=5x5km sample squares, s=2.5x2.5km sample squares)

Appendix 4 Literature consulted when building the ECOSSE model

	pH effect	Water content effect	Freeze/thaw effect	Physical mixing effect
Aerobic decomposition	<p>Murakami M, Furukawa Y, Inubushi K (2005) Methane production after liming to tropical acid peat soils. <i>Soil Science and Plant Nutrition</i> 51(5):697-699.</p> <p>Bergman I, Lundberg P, Nilsson M (1999) Microbial carbon mineralization in an acid surface peat: effects of environmental factors in laboratory conditions. <i>Soil Biology & Biochemistry</i> 31(13):1867-1877.</p> <p>Jugsujinda A <i>et al</i> (1996) Factors controlling carbon dioxide and methane production in acid sulfate soils. <i>Water air and soil pollution</i> 87(1-4):345-355.</p> <p>Sitaula BK, Bakken LR, Abrahamsen G (1995) N-fertilization and soil acidification effects on N₂O and CO₂ emission from temperate pine forest soil. <i>Soil biology & Biochemistry</i> 27(11):1401-1408.</p> <p>Zhu WX, Ehrenfeld JG (1999) Nitrogen mineralization and nitrification in suburban and undeveloped Atlantic white cedar wetlands. <i>Journal of Environmental Quality</i> 28(2):523-529.</p> <p>Khalil MI, Hossain MB, Schmidhalter U (2005) Carbon and nitrogen mineralization in different upland soils of the subtropics treated with organic materials. <i>Soil biology & Biochemistry</i> 37(8):1507-1518.</p> <p>Bridgham SD, Richardson CJ (2003) Endogenous versus</p>	<p>Bubier JL <i>et al</i> (1998) Seasonal patterns and controls on net ecosystem CO₂ exchange in a boreal peatland complex. <i>Global Biogeochemical Cycles</i> 12(4):703-714.</p> <p>Kessavalou A <i>et al</i> (1998) Greenhouse gas fluxes following tillage and wetting in a wheat-fallow cropping system. <i>Journal of Environmental Quality</i> 27(5):1105-1116.</p> <p>Best EPH, Jacobs FHH (1997) The influence of raised water table levels on carbon dioxide and methane production in ditch-dissected peat grasslands in the Netherlands. <i>Ecological Engineering</i> 8(2):129-144.</p> <p>Updegraff K <i>et al</i> (1995) Environmental and substrate controls over carbon and nitrogen mineralization in northern wetlands. <i>Ecological Applications</i> 5(1):151-163.</p> <p>Prieme A, Christensen S (2001) Natural perturbations, drying-wetting and freeze-thawing cycles, and the emission of nitrous oxide, carbon dioxide and methane from farmed organic soils. <i>Soil Biology and Biochemistry</i> 33(15):2083-2091.</p> <p>Stottlemeyer R (2001) Biogeochemistry of a treeline watershed, northwestern Alaska. <i>Journal of Environmental Quality</i> 30(6):1990-1998.</p> <p>Bridgham SD, Richardson CJ (2003) Endogenous versus</p>	<p>Bubier JL <i>et al</i> (1998) Seasonal patterns and controls on net ecosystem CO₂ exchange in a boreal peatland complex. <i>Global Biogeochemical Cycles</i> 12(4):703-714.</p> <p>Sulkava P, Huhta V (2003) Effects of hard frost and freeze-thaw cycles on decomposer communities and N mineralization in boreal forest soil. <i>Applied Soil Ecology</i> 22(3):225-239.</p> <p>Bubier J, Croll P, Mosedale A (2002) Net ecosystem CO₂ exchange measured by autochambers during the snow-covered season at a temperate peatland. <i>Hydrological Processes</i> 16(18):3667-3682.</p> <p>Herrmann A, Witter E (2002) Sources of C and N contributing to the flush in mineralization upon freeze-thaw cycles in soils. <i>Soil Biology and Biochemistry</i> 34(10):1495-1505.</p> <p>Larsen KS, Jonasson S, Michelsen A (2002) Repeated freeze-thaw cycles and their effects on biological processes in two arctic ecosystem types. <i>Applied Soil Ecology</i> 21(3):187-195.</p> <p>Prieme A, Christensen S (2001) Natural perturbations, drying-wetting and freeze-thawing cycles, and the emission of nitrous oxide, carbon dioxide and methane from farmed organic soils. <i>Soil Biology and Biochemistry</i> 33(15):2083-2091.</p>	<p>Kessavalou A <i>et al</i> (1998) Greenhouse gas fluxes following tillage and wetting in a wheat-fallow cropping system. <i>Journal of Environmental Quality</i> 27(5):1105-1116.</p> <p>Friedel JK, Munch JC, Fischer WR (1996) Soil microbial properties and the assessment of available soil organic matter in a haplic Luvisol after several years of different cultivation and crop rotation. <i>Soil Biology & Biochemistry</i> 28(4-5):479-488.</p> <p>La Scala N <i>et al</i> (2005) Soil CO₂ efflux following rotary tillage of a tropical soil. <i>Soil & tillage Research</i> 84(2):222-225.</p> <p>Reicosky DC <i>et al</i> (2005) Tillage-induced CO₂ loss across an eroded landscape. <i>Soil & Tillage Research</i> 81(2):183-194.</p> <p>Olson KR, Lang JM, Ebelhar SA (2005) Soil organic carbon changes after 12 years of no-tillage and tillage of Grantsburg soils in southern Illinois. <i>Soil & Tillage Research</i> 81(2):217-225.</p> <p>Zsembeli J <i>et al</i> (2005) CO₂ measurements in a soil tillage experiment. <i>Cereal Research Communications</i> 33(1):137-140.</p> <p>Al-Kaisi MM, Yin XH (2005) Tillage and crop residue effects on soil carbon and carbon dioxide emission in corn-soybean rotations. <i>Journal of Environmental Quality</i> 34(2):437-445.</p> <p>Lal R (2004) Carbon emissions</p>

	pH effect	Water content effect	Freeze/thaw effect	Physical mixing effect
<p>Aerobic decomposition continued...</p>	<p>exogenous nutrient control over decomposition and mineralization in North Carolina peatlands. <i>Biogeochemistry</i> 65(2):151-178. Chapin CT <i>et al</i> (2003) Nitrogen, phosphorus, and carbon mineralization in response to nutrient and lime additions in peatlands. <i>Soil Science</i> 168(6):409-420. Seneviratne G <i>et al</i> (1999) Differential effects of soil properties on leaf nitrogen release. <i>Biology and Fertility of Soils</i> 28(3):238-243. Arunachalam A <i>et al</i> (1998) Leaf litter decomposition and nutrient mineralization patterns in regrowing stands of a humid subtropical forest after tree cutting. <i>Forest Ecology & Management</i> 109(1-3):151-161. Mugendi DN, Nair PKR (1996) Predicting the decomposition patterns of tree biomass in tropical highland microregions of Kenya. <i>Agroforestry Systems</i> 35(2):187-201. Xu ZH, Ladd JN, Elliot DE (1996) Soil nitrogen availability in the cereal zone of South Australia 1. Soil organic carbon, total nitrogen, and nitrogen mineralization rates. <i>Australian Journal of Soil Research</i> 34(6):937-948. Riffaldi R, Saviozzi A, LeviMinzi R (1996) Carbon mineralization kinetics as influenced by soil properties. <i>Biology & Fertility of Soils</i> 22(4):293-298. Friedel JK, Munch JC, Fischer WR (1996) Soil microbial</p>	<p>exogenous nutrient control over decomposition and mineralization in North Carolina peatlands. <i>Biogeochemistry</i> 65(2):151-178. van Oorschot M <i>et al</i> (2000) Experimental manipulation of water levels in two French riverine grassland soils. <i>Acta Oecologica - International Journal of Ecology</i> 21(1):49-62. Arunachalam A <i>et al</i> (1998) Leaf litter decomposition and nutrient mineralization patterns in regrowing stands of a humid subtropical forest after tree cutting. <i>Forest Ecology & Management</i> 109(1-3):151-161. Mugendi DN, Nair PKR (1996) Predicting the decomposition patterns of tree biomass in tropical highland microregions of Kenya. <i>Agroforestry Systems</i> 35(2):187-201. Humphrey WD, Pluth DJ (1996) Net nitrogen mineralization in natural and drained fen peatlands in Alberta, Canada. <i>Soil Science Society of America Journal</i> 60(3):932-940. Rey <i>et al</i> (2005) Effect of temperature and moisture on rates of carbon mineralization in a Mediterranean oak forest soil under controlled and field conditions. <i>European Journal of Soil Science</i> 56(5):589-599. Emmett BA <i>et al</i> (2004) the response of soil processes to climate change: results from manipulation studies of shrublands across an environmental gradient. <i>Ecosystems</i> 7(6):625-637.</p>	<p>Panikov NS (1999) Fluxes of CO2 and CH4 in high latitude wetlands: measuring, modelling and predicting response to climate change. <i>Polar Research</i> 18(2):237-244. Hobbie SE, Chapin FS (1996) Winter regulation of tundra litter carbon and nitrogen dynamics. <i>Biogeochemistry</i> 35(2):327-338. Harris MM, Safford LO (1996) Effects of season and four tree species on soluble carbon content in fresh and decomposing litter of temperate forests. <i>Soil Science</i> 161(2):130-135. Dorsch P, Palojarvi A, Mommertz S (2004) Overwinter greenhouse gas fluxes in two contrasting agricultural habitats. <i>Nutrient Cycling in Agroecosystems</i> 70(2):117-133. Grogan P <i>et al</i> (2004) Freezing thaw regime effects on carbon and nitrogen dynamics in sub-arctic heath tundra mesocosms. <i>Soil Biology & Biochemistry</i> 36(4):641-654. Neilson CB <i>et al</i> (2001) Freezing effects on carbon and nitrogen cycling in northern hardwood forest soils. <i>Soil Science Society of America Journal</i> 65(6):1723-1730. Stottlmyer R (2001) Biogeochemistry of a treeline watershed, northwestern Alaska. <i>Journal of Environmental Quality</i> 30(6):1990-1998. Panikov NS, Dedysh SN (2000) Cold season CH4 and CO2 emission from boreal peat bogs (West Siberia): Winter fluxes and</p>	<p>from farm operations. <i>Environment International</i> 30(7):981-990. Balota EL <i>et al</i> (2004) Long-term tillage and crop rotation effects on microbial biomass and C and N mineralization in a Brazilian Oxisol. <i>Soil & Tillage Research</i> 77(2):137-145. Jackson LE <i>et al</i> (2003) Responses of soil microbial processes and community structure to tillage events and implications for soil quality. <i>Geoderma</i> 114(3-4):305-317. Wright AL, Hons FM, Matocha JE (2005) Tillage impacts on microbial biomass and soil carbon and nitrogen dynamics of corn and cotton rotations. <i>Applied Soil Ecology</i> 29(1):85-92.</p>

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<p>Aerobic decomposition continued...</p>	<p>properties and the assessment of available soil organic matter in a haplic Luvisol after several years of different cultivation and crop rotation. <i>Soil Biology & Biochemistry</i> 28(4-5):479-488. Humphrey WD, Pluth DJ (1996) Net nitrogen mineralization in natural and drained fen peatlands in Alberta, Canada. <i>Soil Science Society of America Journal</i> 60(3):932-940. Motavalli PP <i>et al</i> (1995) Soil pH and organic C dynamics in tropical forest soils: Evidence from laboratory and simulation studies. <i>Soil Biology & Biochemistry</i> 27(12):1589-1599. Baath E, Arnebrant K (1994) Growth-rate and response of bacterial communities to pH in limed and ash treated forest soils. <i>Soil Biology & Biochemistry</i> 26(8):995-1001. Włodarczyk T, Stepniewski W, Brzezinska M (2002) Dehydrogenase activity, redox potential, and emissions of carbon dioxide and nitrous oxide from Cambisols under flooding conditions. <i>Biology and Fertility of Soils</i> 36(3):200-206.</p>	<p>Glatzel S, Basiliko N, Moore T (2004) Carbon dioxide and methane production potentials of peats from natural, harvested, and restored sites, eastern Canada. <i>Wetlands</i> 24(2):261-267. Olayinka A (2003) Microbial respiration and nitrate production in cowdung amended alfisol as affected by temperature and moisture content. <i>Discovery and Innovation</i> 15(1-2):75-82. Borken W <i>et al</i> (2003) Drying and wetting effects on carbon dioxide release from organic horizons. <i>Soil Science Society of America Journal</i> 67(6):1888-1896. Paul KI <i>et al</i> (2003) Defining the relationship between soil water content and net nitrogen mineralization. <i>European Journal of Soil Science</i> 54(1):39-47. Sato A, Seto M (1999) Relationship between rate of carbon dioxide evolution, microbial biomass, and amount of dissolved organic carbon as affected by temperature and water content of a forest and an arable soil. <i>Communications in Soil Science and Plant Analysis</i> 30(19-20):2593-2605. Włodarczyk T, Stepniewski W, Brzezinska M (2002) Dehydrogenase activity, redox potential, and emissions of carbon dioxide and nitrous oxide from Cambisols under flooding conditions. <i>Biology and Fertility of Soils</i> 36(3):200-206.</p>	<p>thaw activation dynamics. <i>Global Biogeochemical Cycles</i> 14(4):1071-1080. Wang FL, Bettany JR (1993) Influence of freeze-thaw and flooding on loss of soluble organic-carbon and carbon-dioxide from soil. <i>Journal of Environmental Quality</i> 22(4):709-714.</p>	
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Anaerobic decomposition continued...	<p>peat: effects of environmental factors in laboratory conditions. <i>Soil Biology & Biochemistry</i> 31(13):1867-1877.</p> <p>Chapin CT <i>et al</i> (2003) Nitrogen, phosphorus, and carbon mineralization in response to nutrient and lime additions in peatlands. <i>Soil Science</i> 168(6):409-420.</p> <p>Hobbie SE, Miley TA, Weiss MS (2002) Carbon and nitrogen cycling in soils from acidic and nonacidic tundra with different glacial histories in Northern Alaska. <i>Ecosystems</i> 5(8):761-774.</p>	<p>mineralization in northern wetlands. <i>Ecological Applications</i> 5(1):151-163.</p> <p>Glatzel S, Basiliko N, Moore T (2004) Carbon dioxide and methane production potentials of peats from natural, harvested, and restored sites, eastern Canada. <i>Wetlands</i> 24(2):261-267.</p>		
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<p>N₂O emissions continued...</p>	<p>chinese agricultural soils by denitrification. <i>Soil Science and Plant Nutrition</i> 50(6):909-915. Cheng <i>et al</i> (2004) N₂O and NO production in various Chinese agricultural soils by nitrification. <i>Soil Biology & Biochemistry</i> 36(6):953-963. Clough TJ <i>et al</i> (2003) Emission of nitrogen oxides and ammonia from varying rates of applied synthetic urine and correlations with soil chemistry. <i>Australian Journal of Soil Research</i> 41(3):421-438. Khalil MI <i>et al</i> (2003) Nitrous oxide formation potential of various humid tropical soils of Malaysia: A laboratory study. <i>Nutrient Cycling in Agroecosystems</i> 66(1):13-21. Rochester IJ (2003) Estimating nitrous oxide emissions from flood-irrigated alkaline grey clays. <i>Australian Journal of Soil Research</i> 41(2):197-206. Włodarczyk T, Stepniewski W, Brzezinska M (2002) Dehydrogenase activity, redox potential, and emissions of carbon dioxide and nitrous oxide from Cambisols under flooding conditions. <i>Biology and Fertility of Soils</i> 36(3):200-206. Simek M, Jisova L, Hopkins DW (2002) What is the so-called optimum pH for denitrification in soil? <i>Soil Biology & Biochemistry</i> 34(9):1227-1234. Simek M, Cooper JE (2002) The influence of soil pH on denitrification: progress towards the understanding of this</p>	<p>(2003) Nitrous oxide fluxes from maize fields: relationship to yield, site-specific fertilization, and soil conditions. <i>Agricultural Ecosystems & Environment</i> 99(1-3):97-111. Hellebrand HJ, Kern J, Scholz V (2003) Long-term studies on greenhouse gas fluxes during cultivation of energy crops on sandy soils. <i>Atmospheric Environment</i> 37(12):1635-1644. Kammann C <i>et al</i> (1998) Seasonal variability and mitigation options for N₂O emissions from differently managed grasslands. <i>Environmental Pollution</i> 102(1):179-186. Reth S <i>et al</i> (2005) DenNit – Experimental analysis and modelling of soil N₂O fluxes in response on changes of soil water content, soil temperature, soil pH, nutrient availability and the time after rain event. <i>Plant and Soil</i> 272(1-2):349-363. Włodarczyk T, Stepniewski W, Brzezinska M (2002) Dehydrogenase activity, redox potential, and emissions of carbon dioxide and nitrous oxide from Cambisols under flooding conditions. <i>Biology and Fertility of Soils</i> 36(3):200-206.</p>	<p>Sehy U <i>et al</i> (2004) Adding dissolved organic carbon to simulate freeze-thaw related N₂O emissions from soil. <i>Journal of Plant Nutrition and Soil Science</i> 167(4):471-478. Koponen HT, Martikainen PJ (2004) Soil water content and freezing temperature affect freeze-thaw related N₂O production in organic soil. <i>Nutrient Cycling in Agroecosystems</i> 69(3):213-219. Syvasalo E <i>et al</i> (2004) Emissions of nitrous oxide from boreal agricultural clay and loamy sand soils. <i>Nutrient Cycling in Agroecosystems</i> 69(2):155-165. Teepe R, Ludwig B (2004) Variability of CO₂ and N₂O emissions during freeze-thaw cycles: results of model experiments on undisturbed forest-soil cores. <i>Journal of Plant Nutrition and Soil Science</i> 167(2):153-159. Grogan P <i>et al</i> (2004) Freeze-thaw regime effects on carbon and nitrogen dynamics in sub-arctic heath tundra mesocosms. <i>Soil Biology & Biochemistry</i> 36(4):641-654. Koponen HT, Flojt L, Martikainen PJ (2004) Nitrous oxide emissions from agricultural soils at low temperatures: a laboratory microcosm study. <i>Soil Biology & Biochemistry</i> 36(5):757-766. Pei ZY (2004) N₂O exchange within a soil and atmosphere profile in alpine grasslands on the Qinghai-Xizang Plateau. <i>Acta Botanica Sinica</i> 46(1):20-28.</p>	<p>Japan. <i>Global Biogeochemical Cycles</i> 18(4): Jackson LE <i>et al</i> (2003) Responses of soil microbial processes and community structure to tillage events and implications for soil quality. <i>Geoderma</i> 114(3-4):305-317.</p>

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