

Chapter 1

1.1 Introduction

1.1.1 The role of soils in the global C cycle

Carbon is present in the atmosphere mainly as CO₂, with minor amounts present as CH₄, CO and other gases. During the last glacial epoch (20,000 - 50,000 years ago) CO₂ ranged from 180 to 200 μmol mol⁻¹ in the atmosphere. CO₂ rose dramatically at the end of the last glacial epoch (10,000 years ago) and was relatively stable at 280 μmol mol⁻¹ until the Industrial Revolution (Schlesinger, 1997). In the last two centuries, the concentration of atmospheric CO₂ has been increased from 280 μmol mol⁻¹ in 1800 to about 315 μmol mol⁻¹ in 1957 and to about 356 μmol mol⁻¹ in 1993 (Schimel, 1995). CO₂ does not react with other gases in the atmosphere and its concentration is affected principally by interactions with the earth's surface by the annual cycles of photosynthesis and respiration, gas exchange with seawater following Henry's law (the concentration of a dissolved gas in a solution is directly proportional to the partial pressure of that gas above the solution), and the reactions of the carbonate-silicate cycle (Schlesinger, 1997).

Understanding the role of soils as a sink or source for C on a global scale has become vital in assessing the changes in atmospheric CO₂ concentrations (Johnson and Curtis, 2001); because soils are the largest reservoir of C in terrestrial ecosystems, containing 75% of total terrestrial C (Eswaran *et al.*, 1993). Any disruption of soils due to land use or climate change can affect the soil C pools and this can have significant impacts on the C budget of the atmosphere. On the other hand, increased storage of carbon in the world's soils could help offset further anthropogenic emissions of CO₂ (Rustad *et al.*, 2000). The increase in atmospheric CO₂ concentrations, and the likely increase in global temperatures, is expected to alter the

distribution of carbon between atmosphere, vegetation and soils (Watson *et al.*, 2000). However, appropriate land use and soil management may help offset the greenhouse effect by increasing C within the terrestrial ecosystems (Lal *et al.*, 1997).

There are various estimates of global soil carbon in the first top metre, with values between 700 and 2,946 Pg C (1 Pg = 10^{15} g) (Post *et al.*, 1982, Jobbagy and Jackson, 2000). Bouwman and Leemans (1995) estimated that global soil organic matter could range from 1,000-3,000 Pg C. Schimel (1995) estimated that soils contain approximately 1,580 Pg of C, twice as much as the atmosphere (750 Pg) and more than twice as much as the vegetation (610 Pg).

The amount of C to be stored in soils is the net balance of photosynthesis and total respiration in terrestrial ecosystems. The C storage in forest soils is affected by forest type (conifer or broadleaf species) (Lal *et al.*, 1997), with conifer and broadleaf trees having a different strategy in allocating assimilated C belowground (Guo and Gifford, 2002), and also by the site quality (Lal *et al.*, 1997) and management practices, such as fire, clearfelling etc. (Ballard, 2000). Lal *et al.* (1997) estimated the C contained in the soil for different ecosystems and the order was: swamps and marsh > boreal forest > tundra and alpine meadow > temperate grassland, pastures > temperate evergreen forest > temperate deciduous forest > tropical evergreen forest > tropical seasonal forest. But they also suggested the above trend should be further investigated, as many studies have shown no differences in soil organic C between tropical and temperate regions.

Dixon *et al.* (1994) estimated that forest ecosystems contain approximately 1146 Pg C, which is approximately 60% of the C in the terrestrial biosphere (Dixon and Krankina, 1995). Approximately 37% of this carbon is in low-latitude forests (0° to 25°), 14% in mid-latitude (25° to 50°) and 49% in high-latitude forests (50° to 75°) (Dixon *et al.*, 1994). The allocation of C between vegetation and soils differs by latitude. A large part of vegetation (25%) and especially soil C (59%) are located in

the high-latitude forests. Mid-latitude forests account for 16% of the vegetation and 13% of the soil C, while the low-latitude tropical forests contain 59% of the global forest vegetation C and 27% of soil C (Dixon *et al.*, 1994). There is an important global relationship between soil C density and climate, with soil C density generally increasing with increasing precipitation and decreasing temperature (for any particular level of precipitation) (Post *et al.*, 1982), although the storage of soil carbon is not related to latitude (Fisher and Binkley, 2000).

Globally, soil CO₂ flux to the atmosphere is estimated to range between 50 and 76.5 Pg C y⁻¹ (Raich and Schlesinger 1992; Kicklighter *et al.*, 1994; Raich and Potter, 1995); close to the GPP (Gross Primary Productivity), which is estimated to range from 100 to 120 Pg C y⁻¹ (Houghton and Woodwell, 1989). Soil CO₂ flux is a combination of biological, chemical and physical processes and also a sensitive indicator of essential processes in ecosystems such as: metabolic activity of roots, mycorrhizae and other soil organisms, decomposition of plant residues in soil and conversion of soil organic carbon to atmospheric CO₂ (Ewel *et al.*, 1987a; Rochette *et al.*, 1997).

The total soil CO₂ efflux is commonly referred to as soil respiration and the two main sources are the heterotrophic processes (decomposition of organic compounds by soil microorganisms) and the autotrophic (respiration of plant roots and rhizomes) from different soil depths (Buchmann, 2000). The relative contributions of autotrophic and heterotrophic respiration to total soil CO₂ flux are still poorly known. There are different estimates of autotrophic respiration for different forests and in temperate regions there are estimates ranging from less than 30% up to 70% (Ewel *et al.*, 1987b; Bowden *et al.* 1993, Boone *et al.* 1998; Striegl and Wickland, 1998; Buchmann, 2000). The respective contributions to total soil CO₂ flux and the factors controlling the relative contributions in different forests are not well understood (Bowden *et al.*, 1993). Soil “respiration” may also derive from abiotic reactions among carbonate species (Rochette *et al.* 1997).

Several factors can affect soil CO₂ flux rate. The most important ones are soil temperature (Lloyd and Taylor, 1994; Kätterer *et al.*, 1998) and soil water content (Singh and Gupta, 1977; Schlentner and Van Cleve, 1985; Carlyle and Than, 1988). The effects of soil temperature and water content on soil CO₂ efflux are described in more details in a subsequent section. Other important factors can be the seasonal changes in soil microclimate (Raich and Potter, 1995), soil microbial population levels (Klopatek, 2002), availability and quality of soil C substrates (Orchard and Cook, 1983), rates of C inputs to soil (Trumbore *et al.*, 1996), gas diffusivity (Davidson and Trumbore, 1995), plant root activities (Bowden *et al.*, 1993; Raich and Tufekgcioglu, 2000) and plant root production. For instance, Lee and Jose (2003) found a positive correlation between soil CO₂ flux and fine root production in cottonwood and loblolly pine plantations in Florida. Also, soil physical properties such as bulk density and chemical properties (Borken *et al.* 2002) affect soil CO₂ efflux, with fine-textured soils supporting more microbial biomass than coarse-textured ones (Zak *et al.*, 1994). Soil pH can also directly affect microbial biomass. Lee and Jose (2003) found that soil microbial biomass in cottonwood stands in Florida was negatively correlated with soil pH, while soil microbial biomass in loblolly pine stands was positively correlated with soil pH. The type of vegetation also significantly affects soil respiration, with soil respiration rates varying significantly among major biomes (Raich and Schlesinger, 1992), with broadleaves generally having higher respiration rates than conifers on the same soil type (Raich and Tufekgcioglu, 2000). Tree species and nutrient availability strongly affect litter decomposition and thus influence soil respiration (Borken *et al.* 2002). At the global scale there is a highly significant positive correlation between measured rates of soil respiration and aboveground litter production in forest ecosystems (Raich and Nadelhoffer, 1989). Finally, land use and disturbance regimes, such as tree harvesting, fire etc. (Ewel *et al.*, 1987a; Weber, 1990) have been shown to affect strongly soil CO₂ efflux.

1.1.2 Effects of afforestation and forest management on soil C storage

The role of forest plantations has received considerable attention recently, as their management could lead to C sequestration from the atmosphere to the trees and soil. Afforestation leads to an accumulation of carbon in vegetation, but little is known about changes in soil C storage with establishment of plantation forests (Scott *et al.*, 1999) and subsequent management, although soil organic matter is most likely to decrease under intensive management (Harmon *et al.*, 1990).

Guo and Gifford (2002) conducted a meta-analysis of the literature on the effects of land use changes on soil C stocks. They concluded that changing from pasture (including natural grassland) to conifer plantation causes an decrease of 12% in soil C stocks. Jobbagy and Jackson (2000) found that the relative distribution of soil organic C in the top metre of soil was deeper in grasslands than in forests (42% and 50%, respectively, in the uppermost 20 cm, where relative distribution is estimated by dividing the soil organic C content of the 20 cm by the total organic C content in the first metre of the profile). That indicates that more organic matter is available for decomposition in the upper part of the soil in a forest than in grassland. It also means that lowering the water table by afforestation brings more C into the aerated zone where it can be decomposed.

Special attention needs to be given to plantations created on peaty soils, as cultivation and drainage of peat can lead to increased decomposition rates and thus release of C to the atmosphere. Peatlands in temperate and boreal regions have been extensively used for forestry (Laine and Minkkinen, 1996). Peatlands in their original state accumulate C, but drainage to lower the water table for the planting of trees can lead to increased decomposition of organic matter that results in higher CO₂ fluxes and a decrease of soil C (Cannell *et al.*, 1993). Laine and Minkkinen (1996)

estimated a loss of $1.4 \text{ t C ha}^{-1} \text{ y}^{-1}$ from the drainage of a mire peatland and planting with Scots pine in Finland. However, the loss of C from the peat can be compensated by a C increase in standing biomass. Laine and Vasander (1991) found an overall ecosystem C increase of 9% due to increases in tree, litter and peat C, in a drained and afforested peat bog in Finland, which compensated for any loss of peat C due to increased decomposition rates. Cannell *et al.* (1993) argued that the benefit of afforestation with conifers on peaty soils depends on the rate of peat oxidation. In the short-term (50 to 200 years), net carbon sequestration will increase if peat oxidation rates are low ($<1 \text{ t C ha}^{-1} \text{ y}^{-1}$) but if rates of oxidation are more than $3 \text{ t C ha}^{-1} \text{ y}^{-1}$, there might be no net benefit from afforestation.

Practices such as site preparation, forest harvesting and afforestation have different effects on soil C input, accretion and loss (Bashkin and Binkley, 1998) and have many similarities on forest soils across different climatic zones (Ballard, 2000). Clearfelling of trees also causes changes in soil physical properties, such as an increase in bulk density, because of soil compaction due to harvesting machine operations that can affect soil C. Schilling *et al.* (1999) found that clearfelling of a bottomland hardwood forest in Mississippi increased soil bulk density. Tree harvesting usually results in higher soil moisture and temperature that in turn result in increased decomposition. Johnson *et al.* (1991) studied the effects of clearfelling on the soil horizons and the organic matter pools in a northern hardwood forest in New Hampshire. They sampled soil immediately before and three years after clearfelling. They concluded that although the thickness of the O horizon decreased by 20% (from 6.9 cm to 5.5 cm) the O organic matter content increased (from 60 to 64 t ha^{-1} for the 0 to 10 cm depth and 54 to 58 t ha^{-1} for the 10 to 20 cm depth). However, the organic matter content summed over forest floor and mineral soil layers remained essentially unchanged following harvesting (334 and 337 t C ha^{-1} before and after clearfelling respectively), although the organic matter was redistributed within the soil profile. Johnson *et al.* (2002) studied the effect of tree harvesting and type of harvesting on soil C by re-sampling mixed deciduous and

conifer forests stands, in various locations in USA, 15 to 16 years after treatment. They concluded that harvesting had little lasting effects on soil C and the effect occurred at early stages. They also found that soil C varied considerably over time, and time elapsed since tree harvesting was an important factor, thus frequent sampling should be conducted in order to assess accurately the harvest effects on soil C.

1.1.3 The forest chronosequences approach for the study of changes soil C stock

The chronosequence approach has been commonly used (Covington, 1981; Wallace and Freedman, 1986; Johnson *et al.*, 1991; Knoepp and Swan, 1997; Klopatek, 2002) in order to determine the long-term changes in soil C stocks due to afforestation and clearfelling, since direct investigation of the forest succession is virtually impossible because of the long period of time required. The major assumption of the chronosequence approach, is that the pattern observed among stands of different age is the expected pattern of temporal development that would be exhibited by a particular stand (Wallace and Freedman, 1986). In order to implement the chronosequence approach the sites to be chosen must be directly comparable with each other and that means they must have similar vegetation on similar soils, as well as a similar water regime, topography, and management history (i.e. site preparation, clearfelling method, etc). However, in reality these assumption are rarely all met in chronosequence studies.

Wallace and Freedman (1986) measured forest floor C dynamics in a hardwood chronosequence after clearfelling in Nova Scotia. They found that the intra- and inter-site variation in their study sites was greater than any effects resulting from stand age. Reasons for this variation included the presence of stumps, logs and slash piles and differences in decomposition rates due to microclimate and

microtopography. They also argued that the high intra- and inter stand variation could mask an existing trend. Yanai *et al.* (2000) tested the validity of the patterns of organic matter loss and accumulation predicted by Covington (1981). Covington (1981) predicted a loss of 50% of forest floor organic matter in the first 20 years after disturbance. Federer (1984) added 13 more stands in the same region in order to test Covington's curve. He found that forest floor C did not show such a sharp minimum as in Covington's study. Yanai *et al.* (2000) re-sampled the 13 stands that had been previously measured by Federer (1984). They observed that changes in the stands did not agree with the pattern in the original chronosequence of Covington (1981) and they also found very high spatial variation in forest floor organic matter. This made it impossible to detect small rates of change in a 15 years period, by using reasonable sampling schemes. Thus, they could not determine if forest floor was gaining or losing carbon. Although young stands had less forest floor than old stands, this pattern could equally be explained by the year of harvesting, which is related to the method of logging, rather than the age of the stand. Tractor logging presumably causes more disturbance to the soil than horse logging in earlier times and rubber-tired skidders (Yanai *et al.*, 2000). Also the amount of slash left behind after tree logging affects the amount of organic matter input to the soil and this amount differs with the method used for tree harvesting. Additionally, in Covington's (1981) study, a bias in the true pattern of the chronosequence might have arisen from the restrictive criteria he used in soil sampling, as he excluded sample points, e.g. points with abnormal horizon layers (other than L, F, H) and on windthrow mounds or depressions. Federer (1984) adopted less strict criteria, however, he still excluded sampling of poorly drained soil, on places with extreme rockiness and steep slopes. Knoepp and Swan (1997) used a chronosequence approach to measure changes in soil C after clearfelling in forests in North Carolina. They also noticed high inter-annual variability in the soil C data, possibly indicating that the sample year may affect the observed value.

Despite all these caveats, the chronosequence approach maintains an important role in the study of the long-term changes of soil C stocks, as it allows to obtain the expected patterns of change over long time periods. Ideally, it is coupled to permanent sample plots, which will be subsequently re-surveyed to confirm or disprove the observed patterns.

1.1.4 Effects of forest management on soil CO₂ efflux

Forests comprise more than one third of the world's land area (Fisher and Binkley, 2000), with temperate forests representing 11% of the land area (Ewel *et al.*, 1987a; Kicklighter *et al.*, 1994). Tans *et al.* (1990) suggested that the terrestrial biosphere at temperate latitudes can be a major sink for atmospheric CO₂, sequestering between 2 and 3.4 Pg C y⁻¹. While changes in land use and land use intensity in the world's forests may account for the net release of as much as 2.5 Pg C y⁻¹ as CO₂, (mostly in the tropics) (Schlesinger, 1997; Houghton, 2003), 61.4 Pg C y⁻¹ are entering the soil as detritus and 60-75 Pg C y⁻¹ are leaving by respiration (Schimel, 1995).

Forest clearfelling is one of the management practices used most frequently for the harvesting of timber. It can have a significant effect on soil CO₂ efflux by altering the biogeochemical cycles of a forest ecosystem. Clearfelling alters environmental factors which have an effect on soil CO₂ efflux, such as soil temperature, soil water content and depth to the water table, and consequently it will alter soil CO₂ efflux. Adams *et al.* (1991), in a long-term study in a Douglas-fir (*Pseudotsuga menziesii* (Mirb.) Franco) forest in Oregon, observed significantly higher soil water contents in the clearfelled site than in the control stand, during the first 2 years after clearfelling. Frazer *et al.* (1990) measured higher temperatures in fall, spring and summer in a clearfelled than in an uncut mixed conifer forest in Sierra Nevada, but not in winter. They also observed that the soil in the clearfelled site was drier during the summer and autumn and wetter during the winter. Tree harvesting also results in a rise of the

water table because of the reduced evapotranspiration. Pyatt *et al.* (1985) measured a significant rise in the water table after clearfelling of a Sitka spruce stand in England. The mechanical disturbance from machines used for tree harvesting and site preparation causes changes in the physical properties of the soil. Furthermore, compaction of the soil by heavy equipment decreases the soil macroporosity and causes reduction in air diffusion and water infiltration rates (Pritchett, 1979), thus causing increases in the soil water content and consequently causing the soil environment to become more anaerobic.

A number of studies have compared soil CO₂ efflux under different forest management practices, including clearfelling (Edwards and Ross-Todd, 1983; Weber, 1985; Ewel *et al.*, 1987a; Gordon *et al.*, 1987; Hendrickson *et al.*, 1989; Weber, 1990; Fernandez *et al.*, 1993). The pattern of soil CO₂ efflux reported after clearfelling is not the same in all studies. Ewel *et al.* (1987), Gordon *et al.* (1987), and Hendrickson and Robinson (1984) reported that soil CO₂ efflux increased after clearfelling compared to an uncut “control” forest stand, while Weber (1990) and Nakane *et al.* (1986) found that clearfelling caused a decrease in soil CO₂ efflux. Fernandez *et al.* (1993) found that clearfelling did not significantly affect soil CO₂ efflux in a red spruce (*Picea rubens* Sarg.)-balsam fir (*Abies balsamea* (L.) Miller) forest in Maine (USA). It is difficult to identify any characteristic pattern of soil CO₂ efflux, following clearfelling of a forest. The magnitude of change in soil CO₂ efflux depends on whether or not litter and organic layers are removed, roots are disturbed and mineral soil horizons are exposed or mixed (Buchmann, 2000), and on the rates of C input to soil from the logging residues and the response of soil microbial biomass to microclimatic conditions created after the removal of trees. All these responses can differ among ecosystems. Also, the different types of harvesting machines used in each occasion can cause different degrees of disturbance to the soil. Hence, the response of soil CO₂ efflux to clearfelling may vary among different locations.

Additionally, in different studies soil CO₂ efflux has been measured for different time periods since clearfelling. For example, Nakane *et al.* (1986) started measuring soil CO₂ flux in a red pine (*Pinus densiflora*) plantation just three months after clearfelling, while Gordon *et al.* (1987) measured soil CO₂ efflux in a white spruce (*Picea glauca*) forest in Alaska three and four years after clearfelling of the stand. Depending on the time since clearfelling, the contributions from the amount of decomposing roots and litter, the invasion of grasses on the site or the re-planting of trees will differ.

1.1.5 Spatial variability of soil CO₂ efflux

Singh and Gupta (1977) indicate that temperate habitats show large variability in soil CO₂ efflux, with values in temperate forests ranging from 2.4 to 12 g CO₂ m⁻² d⁻¹, and with occasional maximum values up to 56.16 g CO₂ m⁻² d⁻¹. Shibistova *et al.* (2002) found that CO₂ fluxes in areas around trees were about double the fluxes measured in open areas in between trees of a Scots pine (*Pinus sylvestris* L.) forest in Siberia. This variability is due to the several factors affecting soil CO₂ flux, such as soil temperature and water content (Raich and Potter, 1995; Davidson *et al.*, 1998), soil microbial population levels (Klopatek, 2002), availability and quality of soil C substrates (Orchard and Cook, 1983), levels of plant root activity (Ewel *et al.*, 1987, Bowden *et al.*, 1992), plant root density (Shibitova *et al.*, 2002), and vegetation type (Raich and Schlesinger 1992). Longdoz *et al.* (2000) found larger spatial variability in a beech stand (*Fagus sylvatica* L.) (3.6 to 22.2 g m⁻² d⁻¹) than in a Douglas fir (*Pseudotsuga menziesii* (Mirb.) Franco) stand (3.5 to 10 g m⁻² d⁻¹) in Belgium.

Land use changes, like the conversion of grassland to forest, may have considerable effects on the spatial variation of soil CO₂ flux, due to the mechanical preparation of the site for the planting of trees. Forest management practices such as clearfelling may result conversely in changes in spatial variation of soil CO₂ flux, since the

removal of the trees alters dramatically the microclimate at the forest floor, while mechanical disturbance caused by the harvesting machines causes compaction of the soil and changes its physical properties (Pritchett, 1979).

Spatial variability of soil CO₂ flux has not been studied extensively, although very high spatial variability of soil CO₂ flux has been reported in some ecosystems (Raich *et al.*, 1990; Hanson *et al.*, 1993). It is important to estimate soil CO₂ flux accurately in order to evaluate the carbon budget of an ecosystem. However, this may require a large number of sampling points for the estimation of a representative value of soil CO₂ flux within an ecosystem. This may be practically difficult to implement in the field due to time and high cost limitations.

Geostatistics provide powerful analytical tools to capture the horizontal variability of a property and have received increasing interest from soil biologists in recent years. In its simplest form geostatistics define the degree of autocorrelation among the measured data points and interpolate values between measured points based on the degree of autocorrelation encountered (Robertson, 1987). Spatial patterns are usually described using the experimental semivariogram $\gamma(\mathbf{h})$ which measures the average dissimilarity (or inverse autocorrelation) between data separated by a vector \mathbf{h} . Variograms can be derived from as little as a few to as many as several thousand data points, but those based on fewer than 50 data points are often erratic sequences of experimental values with little or no evident structure. The form of the variogram becomes clearer as the size of the sample increases (Webster and Oliver, 2001).

Another useful indicator of spatial variation of soil CO₂ efflux is the coefficient of variation (CV), resulting from the variability among different sampling locations and it is useful for comparing the variation of different sets of observations of the same property (Webster and Oliver, 2001), e.g. in this case the variation of soil CO₂ efflux across the different dates of the study period.

1.1.6 Importance of soil temperature and water content on soil CO₂ efflux

Temperature is the main environmental factor affecting soil CO₂ efflux, although not all studies concur. Giardina and Ryan (2000) concluded that the decomposition (respiration) of organic carbon contained in forest mineral soil can not be stimulated by increased temperature alone. However, their data are in conflict with the soil CO₂ flux data reviewed by Loyd and Taylor (1994). Liski *et al.* (1999) also concluded that the decomposition of old soil organic matter is more tolerant of temperature change than is the decomposition of young litter. The same idea was supported by Thornley and Cannell (2001) who argued that the temperature sensitivity soil of CO₂ efflux has been over-estimated. They concluded that, although in the short-term warming increases soil CO₂ efflux, in the long-term the carbon lost by microbial respiration is being offset due to acceleration of soil physico-chemical “stabilisation” reactions and inputs in the soil from net production. Furthermore, they criticised soil CO₂ efflux data from short-term measurements (incubation), as short-term experiments are heavily affected by the transient loss of readily metabolizable carbon (same argument as in Giardina and Ryan, 2000). Additionally they questioned the soil temperature response functions derived from the seasonal changes of soil CO₂ efflux in the field. Despite all these arguments, many studies on different individual sites and at different time scales of the order of hours to months, have stated that soil CO₂ efflux is strongly temperature-dependent (Howard and Howard, 1993; Kirschbaum 1995; Davidson *et al.*, 1998).

Although many studies have shown a positive correlation between soil CO₂ efflux and temperature, there is not an agreement on the exact form of the relationship. Also, most of these relationships tend to be site-specific and so far no agreement has emerged on one common model. Some researchers have used a linear relationship (Froment, 1972; Raich and Schlesinger, 1992; Howard and Howard, 1993), especially when soil water content was included in the analysis (Gordon *et al.*, 1987;

Rey *et al.*, 2002). Soil CO₂ efflux usually increases exponentially with increasing temperature and this relationship can be described with exponential or Arrhenius-type equations, which have different theoretical bases (Lloyd and Taylor, 1994). It has been suggested that the use of Arrhenius type functions effectively calculates the influence of temperature on rate constants for most chemical reactions and many biological reactions, although in biological systems the functions can only be applied over a limited range of temperatures (Winkler *et al.*, 1996). It is unclear if it would be useful to look for a theoretical description of such a complex system as the community of soil organisms, where the total activity is determined by the combined activity of a whole range of different organisms that may have different individual temperature sensitivities (Kirschbaum, 1995).

In order to describe the temperature sensitivity of soil CO₂ efflux, the Q₁₀ is commonly used as a convenient indicator to summarise the observed responses (Kirschbaum, 1995). In the most commonly used model, the Q₁₀ value is conceptually constant. In the other models Q₁₀ is a variable with temperature. Lloyd and Taylor (1994) pointed that Q₁₀ decreases with increasing temperature. Consequently, for biomes in areas where soil temperatures are low, soil CO₂ efflux is more sensitive to changes in soil temperature. Winkler *et al.* (1996) in incubation experiments with soil from the A horizon, calculated a Q₁₀ from 1.7 to 1.9, over the temperature range from 4 °C to 28 °C. Kirschbaum (1995) found over a wide range of studies that Q₁₀ decreases linearly with increasing temperature. In his review, the Q₁₀ value ranged from about 8 at 0 °C to about 2 at 30 °C. However, different Q₁₀ values, may be obtained with the same data set. Fang and Moncrieff (2001) incubated soils from farmland and forest and derived different Q₁₀ values at 10 °C and 30 °C, from different fitted relationships in their data. For example, when they used the first order exponential function, the Q₁₀ was constant at both 10 °C and 30 °C (3.1), while when used the Arrhenius equation, Q₁₀ decreased from 3.4 at 10 °C to

2.9 at 30 °C. However, when they used the equation from Schlentner and Van Cleve (1985) the Q_{10} increased from 3.5 at 10 °C to 8.1 at 30 °C.

Q_{10} varies among ecosystems and temperature ranges because the various components of soil respiration (e.g. root respiration, oxidation of plant detritus and organic matter by heterotrophs) exhibit different sensitivities to temperature (Kirschbaum, 1995; Trumbore *et al.*, 1996). It has also been found that Q_{10} varies with the depth of the active soil layer and the depth at which temperature is measured (Kirschbaum, 1995; Swanson and Flanagan, 2001).

Water content is another important abiotic factor affecting soil CO₂ efflux and often it interacts with soil temperature to affect soil respiration, making it difficult to separate the two factors. Generally, dry soils exhibit lower CO₂ efflux than wet soils (Reichstein *et al.* 2002; Davidson *et al.* 2000, Mielenick and Dugas, 2000). Davidson *et al.* (2000) found decreasing soil CO₂ efflux in forests and pastures in eastern Amazonia with decreasing water content. However, at high levels of soil water content limitations on the diffusion of oxygen suppress soil CO₂ efflux. The results from published studies do not reveal a consistent relationship of soil water content with soil CO₂ efflux and different functions have been used to describe the relationship. Gordon *et al.* (1987) used a linear regression to best fit the relationship between soil CO₂ efflux and soil water content for the mineral horizon. The exponential function has also been used (Davidson *et al.*, 1998), while many researchers have used quadratic functions (Bunnell *et al.*, 1977; Linn and Doran, 1984; Howard and Howard, 1993; Mielenick and Dugas, 2000). Finally, hyperbolic functions have been also often used (Bunnell *et al.*, 1977; Schlentner and van Cleve 1985; Hanson *et al.*, 1993). All these functions are empirical and site-specific. Howard and Howard (1993) tried to establish a regression model for the relationship between CO₂ efflux and soil water content for different soil types. They found a range of curves with different shape for each soil type, with the water content where maximum CO₂ efflux occurred differing between soil types. They concluded that the

shape of the curve is related to the structural stability of the soils at different soil water contents and the type of the soil organic matter, rather than the amount. Soil water content has been expressed in different forms such as, volumetric soil water content (Hanson *et al.*, 1993; Mielnick and Dugas, 2000), matric potential (Orchard and Cook, 1993; Davidson *et al.*, 2000), gravimetric water content (Schentner and Van Cleve, 1985), precipitation (Raich and Potter, 1995), water table depth (Kim and Verma, 1992).

Soil CO₂ efflux is not very sensitive to temperature at very low soil water content (<75%), but it is responsive at higher water content (100 to 250%). Also, soil CO₂ flux is not sensitive to water content at low temperatures (<5 °C), but it is responsive at higher temperatures (10-20 °C) (Schlenter and Van Cleve, 1985; Carlyle and Than, 1988; Howard and Howard, 1993). The optimum water content is usually somewhere near field capacity, where the macropore spaces are mostly air filled, thus facilitating O₂ diffusion, and the micropore spaces are mostly water filled, thus facilitating diffusion of soluble substrates (Davidson *et al.*, 2000). When soils are too wet, aerobic respiration is inhibited by oxygen deficiency (Skopp *et al.*, 1990). Water content also affects the rate of diffusion of CO₂ within the soil profile (Davidson and Trumbore, 1995).

1.1.7 Methods for measuring soil CO₂ efflux

The study of soil respiration dates back more than 100 years (cf., Schwartzkopf, 1978). Soil respiration is difficult to measure because of its large temporal and spatial variability and dependence on many environmental and substrate nutrients characteristics (Norman *et al.*, 1997; Lund *et al.*, 1999).

There are different techniques for measuring CO₂ efflux with different advantages and disadvantages, resulting in different accuracy in the measurements and different

spatial and temporal resolution and thus making the comparison of measured rates problematic (Raich and Nadelhoffer, 1989). So the choice of the method to be used is often a compromise between accuracy, applicability and cost. The methods that have been used to measure soil CO₂ efflux include open dynamic chambers (Rayment and Jarvis, 1997), closed dynamic chambers (Desjardins, 1985; Rochette *et al.*, 1991; Kim *et al.*, 1992; Norman *et al.*, 1997), closed static chambers (Grahammer *et al.*, 1991), soil CO₂ profile and micrometeorological techniques (e.g., eddy covariance). The latter two methods have the advantage of avoiding the confounding chamber effects, however their cost and methodological requirements (e.g. the estimation of soil transport coefficients in the first and requirement of specific atmospheric conditions and the estimation of aboveground plant respiration in the second) (de Jong *et al.*, 1979; Baldochhi and Meyers, 1991) make their application more limited compared to the use of chamber methods. The closed static chamber method has sometimes been criticised for underestimating the soil CO₂ efflux at low flux rates and severely underestimating at high flux rates, compared to the closed dynamic chamber method (Healy *et al.*, 1996; Jensen *et al.*, 1996; Janssens *et al.*, 2000). However, Rochette *et al.* (1997) found little differences in fluxes measured by either the closed dynamic or the static chamber method. Also, the closed static chamber method is currently the only accepted standard for the measurements of other trace gases such as nitrous oxide (N₂O) and methane (CH₄).

All the methods that have been compared *in situ* can only give relative answers. Some researchers have attempted to evaluate the absolute accuracy of systems by using a simulated soil and known CO₂ fluxes. Nay *et al.* (1994) compared a static closed chamber with soda lime as a CO₂ absorbent and a dynamic closed chamber (LI-COR 6200, Lincoln, Nebraska, USA). They found that the static closed chamber greatly overestimated the zero flux and by about 25%, while the dynamic closed chamber consistently underestimated all fluxes above zero by 15%. They attributed the overestimation and underestimation of the static closed chamber to the use of the soda lime as a CO₂ absorbent, that caused changes in the diffusion of CO₂ from the

foam (simulated soil) into the chamber. Widén and Lindroth (2003) used a calibration system where CO₂ was allowed to diffuse through a layer of sand on top of a box of known volume and calculated the exact CO₂ efflux by measuring the decrease in CO₂ concentration inside the box. Then they compared this flux with the fluxes measured from a dynamic closed chamber (LI-COR 6200) and a dynamic open chamber. For the dynamic closed chamber they found that any errors were proportional to the flux. CO₂ efflux was overestimated by 21% when the sand water content was 0.06 cm cm⁻³, compared to the flux when the sand was dry. CO₂ efflux was underestimated by 19% in sand with high porosity and low water content (dry), and that may have been due to the increased air volume in the soil that was not accounted for in the total system volume. Bekku *et al.* (1997) tested the fluxes given by a static closed chamber and a dynamic closed chamber (LI-COR 6400) by using an artificial soil medium inoculated with *Trichoderma* sp., in which the absolute CO₂ efflux could be estimated by measuring the weight loss of the medium. They found that the rates of CO₂ efflux measured by the static closed chamber were slightly lower than those measured by the dynamic closed system; however, they were not significantly different and the accuracy of the two systems was 94 and 95% respectively of the absolute estimated CO₂ efflux. However, the chamber of both systems covered the whole surface from where CO₂ was emitted and thus any chamber effects were avoided; therefore the conclusion may not be completely reliable.

The use of static closed chambers can generate subtle pressure fluctuation associated with the turbulence of air over the soil surface (Hutchinson and Livingston, 1993), but this problem does not occur with the use of dynamic closed chamber systems. Hutchinson and Mosier (1981) proposed that a properly vented closed chamber causes no perturbation in mean air pressure. However, Conen and Smith (1998), after testing vented and non-vented chamber in field experiments, recommended the use of non-venting chambers for the measurements of trace gases. They concluded that the potentially beneficial effect of venting a chamber is outweighed by the

disadvantages, such as the depressurisation of the chamber by the Venturi effect and subsequent flushing of soil air into the chamber that can lead to significant overestimation of fluxes on permeable soils. Also, diffusion losses through the vent tube can lead to underestimation of fluxes in soils with small air permeability.

1.1.8 Effects of afforestation and forest management on soil carbon balance

When estimating the ability of forests to sequester carbon from the atmosphere for long-term storage, the net soil CO₂ efflux from the forest floor must be taken into account (Schlesinger, 1990). Soil carbon storage is an important factor in long-term ecosystem stability but slow changes in a large pool are difficult to detect, although even small changes in the soil carbon pool can result in relatively large changes in fluxes of CO₂ to the atmosphere (Keith *et al.*, 1997). Because of the difficulty in determining small changes in large pools, the changes in soil CO₂ fluxes may be more informative.

Soil CO₂ efflux is also a good indicator of total belowground carbon allocation and ecosystem productivity (Raich and Nadelhoffer, 1989). For instance, Davidson *et al.* (2000) found higher respiration rates in a primary forest in Amazonia than in a pasture. They also found that a secondary forest (which was once a degraded pasture) had regained most of its productivity, but the rates of respiration were not yet as high as the primary forest.

Roots play an important role in soil C cycling; stand age and disturbances due to forest management can affect the number and mass of roots present in the soil (Grier *et al.*, 1981; Vogt *et al.*, 1983). A significant amount of C is allocated to fine roots annually, even though they comprise only a small fraction of the C content of forest trees (Gower *et al.*, 1996). Accurate measurements of turnover of fine roots and

mycorrhizae are essential for understanding and quantifying belowground carbon allocation (Nadelhoffer and Raich, 1992; Ryan *et al.*, 1997, McDowell *et al.*, 2001).

Raich and Nadelhoffer (1989) used the conservation of mass approach to estimate the belowground C allocation for forest ecosystems globally. Total belowground carbon allocation (*TBCA*) is the sum of mycorrhizae and root (fine and coarse) production, respiration and exudates. All carbon that enters the soil must either leave the soil or increase soil carbon stocks. Raich and Nadelhoffer (1989) assumed that if the forest is near steady-state, the change in soil carbon stock will be near zero and thus the main regulators of *TBCA* can be estimated from measurements of annual rates of soil respiration and aboveground litterfall. Many have followed their example (e.g. Ryan, 1991; Smith and Resh, 1999); however, this simple approach cannot be applied to aggrading or disturbed forests (Gower *et al.*, 1996). A few studies have examined the *TBCA* patterns with age and the findings are not consistent. Ryan *et al.* (1997) concluded that *TBCA* may increase with stand age, while Smith and Resh (1999) estimated that *TBCA* declines with stand age in a age sequence of Lodgepole pine (*Pinus contorta* L.) stands and Giardina and Ryan (2002) also concluded similarly for *Eucalyptus saligna* stands. Although *TBCA* may be a large fraction of GPP (more than 30%, Gower *et al.*, 1996) and it is the primary source of detrital C to the soil (Giardina and Ryan, 2002), the mechanisms that control the allocation of carbon belowground are still poorly understood.

Hargreaves *et al.* (2003) used the eddy covariance method roving between different sites to investigate the carbon balance of afforested peatlands with Sitka spruce, in Scotland. They identified four stages in the state of the peatland: 1) the undisturbed peatland was a C sink of about $0.25 \text{ t C ha}^{-1} \text{ y}^{-1}$. 2) The freshly drained and ploughed peatland was a carbon source of about $2 \text{ to } 4 \text{ t C ha}^{-1} \text{ y}^{-1}$. 3) The four to eight years old plantation was a C sink of up to $3 \text{ t C ha}^{-1} \text{ y}^{-1}$, due to C fixation by re-colonising ground vegetation. 4) The growing trees dominated the C budget with an uptake of 5

t C ha⁻¹ y⁻¹; however, peat decomposition contributed an emission of about 1 t C ha⁻¹ y⁻¹.

1.1.9 The role of soils as a source of N₂O flux

Nitrous oxide (N₂O) is a powerful greenhouse gas, which destroys ozone in the stratosphere. The concentration of N₂O in the atmosphere has increased from about 0.275 μmol mol⁻¹ in pre-industrial times to about 0.320 μmol mol⁻¹ at present times (Smith and Conen, 2004). The atmospheric N₂O concentration increase started in the 1940s (Khalil and Rasmussen, 1992) and the current rate of increase is about 0.3% y⁻¹ (Schlesinger, 1997). Schlesinger (1997) estimated a global pool of about 2.4 × 10¹⁵ g N₂O (1.5 × 10¹⁵ g N) in the atmosphere.

Soils are the dominant source of N₂O emissions to the atmosphere, accounting for about 9 Tg N y⁻¹ (Batjes, 1996), which is about 50% of the total annual global emissions of about 17.7 Tg N y⁻¹ (Houghton *et al.*, 2001). The most significant natural sources of N₂O in the atmosphere are tropical soils, contributing between 2.2 and 3.7 Tg N y⁻¹, while temperate forest soils are estimated to contribute between 0.1 and 2 Tg N y⁻¹ (Houghton *et al.*, 2001). Agricultural soils also significantly contribute, releasing 4.2 Tg N y⁻¹ to the atmosphere (Houghton *et al.*, 2001). The oceans also appear to be a source of N₂O of about 4 (1.2-6.8) Tg N y⁻¹, as a result of nitrification in the deep sea (Nevison *et al.*, 1995). Fossil fuel combustion and biomass burning were considered to be significant sources of N₂O, but Cofer *et al.* (1991) concluded that biomass burning contributes only 7%, while Berges *et al.* (1993) estimated that the fossil fuel combustion by cars contributes 3-16% of the global emissions of N₂O.

The emissions from the soil have increased in the last few decades due to N inputs by fertilisers to agricultural soils and N deposition onto natural soils (Skiba and Smith,

2000). The increase of N₂O in the atmosphere is expected to have an impact on climate change over the next century, as each kg of N₂O has the potential to contribute about 300 times as much to the greenhouse effect as each kg of CO₂, because of its high absorption of infrared radiation and its long (~120 years) residence time in the atmosphere.

Evaluation of N₂O fluxes has been one of the most problematic topics in environmental biogeochemistry over the last 10-15 years (Groffman *et al.*, 2000) as N₂O fluxes exhibit extraordinary spatial and temporal variability (Ball *et al.*, 1997; Brumme *et al.*, 1999). The global budget of N₂O is not perfectly understood. There are many factors contributing to this high degree of uncertainty, including the following: 1) the estimates are still based on a limited number of data sets from field measurements, which comprise mostly sporadic measurements and frequently do not cover complete seasons. 2) The estimates do not differentiate between the contribution of different forest types (e.g. temperate coniferous and deciduous forests) to global N₂O emissions (Papen and Butterbach-Bahl, 1999).

N₂O is the intermediate product of several processes in soil: nitrification (the oxidation of ammonium (NH₄) to nitrite (NO₂⁻) and nitrate (NO₃⁻), and denitrification (the reduction of nitrate and nitrite to nitric oxide (NO), nitrous oxide and dinitrogen (N₂), depending on contrasting requirements for oxygen. N₂O can also form by chemical reactions when NO₂⁻ or NH₂OH are decomposed in acid soils, but the amount of N₂O produced by these processes is relatively little compared with the amount produced by nitrification and denitrification (Sahrawat and Keeney, 1986). These two processes contribute two thirds of all emissions and they vary with climate, soil conditions and soil management (Skiba and Smith, 2000). Nitrification is promoted by low rainfall, good drainage, and coarse soil texture while denitrification is promoted by high rainfall, poor drainage, fine soil texture and high organic carbon (Groffman, 2000). The production of N₂O through nitrification or denitrification is not static and it can switch rapidly between the two processes, as the

soil aeration state (within the biologically active sites) changes because of changes in the rainfall or increased O₂ demand caused by the presence of easily mineralizable organic matter (Skiba and Smith, 2000). Nitrification is an aerobic process, but when the supply of O₂ is limited by diffusional constraints the nitrifying bacteria can use nitrite as an electron acceptor and reduce it to NO and N₂O. The necessary conditions for denitrification occur when respiratory consumption of O₂ in the soil by plant roots and soil microorganisms exceeds the rate of replenishment by diffusion from the atmosphere and anaerobic microsites within the soil profile are created (Smith *et al.*, 2003). The rate of denitrification is controlled by the rate of nitrification, which supplies NO₃ as a substrate (Schlesinger, 1997). Generally, rates of N₂O production during nitrification are smaller than the ones produced during denitrification (Sahrawat and Keeney, 1986).

1.1.9.1 Effects of environmental factors on soil N₂O flux

While much is known about process level control on N₂O production by nitrification and denitrification, knowledge on the environmental controls responsible for site variation in annual N₂O fluxes at an ecosystem level is low. Brumme *et al.* (1999), established a new classification for N₂O emissions from forest ecosystems, based on differences in the emission patterns of existing whole-year measurements. For temperate and tropical forest regions three emission types could be separated: seasonal, background and event emission types. The seasonal type in temperate regions is characterized by a period of elevated rates in summer. The background emission type has low annual N₂O emissions (range from 0.17 to 0.80 kg N₂O-N ha⁻¹ y⁻¹, mean of 0.39 ± 0.27 kg N ha⁻¹ y⁻¹) and has been identified in most forests (n=21). Background types have lower annual rates than the seasonal type. Emission of N₂O based on events could be observed during frost-thaw periods and after rewetting and are combined with the seasonal or background emission type.

It has been very difficult to establish strong predictive relationships between N₂O fluxes and parameters such temperature, soil water content and soil inorganic N concentrations (Groffman *et al.*, 2000). In many soils the water content has been shown to be the dominant variable controlling the N₂O emission rate. Goodroad and Keeney (1984a) in incubation experiments found that N₂O emissions increased when soil water content increased from 0.1 to 0.3 cm³ cm⁻³. Dobbie *et al.* (1999) suggested that maximum N₂O emissions occur at a WFPS (Water Filled Pore Space) of 80-85% in agricultural soils. But in forests there is often no apparent relationship between soil N₂O fluxes and water content. Bowden *et al.* (1990) found no relationship between soil N₂O fluxes and soil water content in red pine (*Pinus resinosa* Ait.) and hardwood forests in Massachusetts. Henrich and Haselwandter (1997) also found no relationship between N₂O fluxes and soil water content in a Norway spruce stand in Austria, while Papen and Butterbach-Bahl (1999) found very weak coefficients of determination between soil N₂O fluxes and WFPS in a 89-years-old Norway spruce forest in Germany. But even if there is no apparent relationship, soil water content is still an important factor regulating soil N₂O fluxes, although its importance being more apparent in rapidly N cycling ecosystems (Verchot *et al.*, 1999).

The depth to the water table has also been shown to influence N₂O fluxes. Lowering the water table has been found to increase N mineralization in peat and to enhance N₂O fluxes in both field and laboratory experiments (Martikainen *et al.* 1993; Kliewer and Gilliam, 1995). Kliewer and Gilliam (1995) measured N₂O evolution under different water table depths, in undisturbed soil cores in fallow agricultural soils in North Carolina and measured higher N₂O evolution when the water table was 15 cm below the soil surface, than when it was at 30 or 40 cm. Martikainen *et al.* (1993) found that a lower water table in drained peatlands in Finland increased N₂O emissions, but emissions were poorly correlated with the actual level of water table. Regina *et al.* (1999), in laboratory experiments with soil monoliths of previously drained and forested sites (with birch, *Betula pendula*, and pine, *Pinus sylvestris*)

measured N₂O emissions of 12 mg N m⁻² d⁻¹ at a water table depth of -25 cm below the soil surface, and emissions of 2.4 mg N m⁻² d⁻¹ at a depth of -2 cm.

Peterjohn *et al.* (1994) found that an increase in soil temperature could only explain 14% of the variability in soil N₂O fluxes in experimentally heated soils in a mixed deciduous forest in Massachusetts; soil water content explained 23% of the variability. Brumme *et al.* (1999) conducted an analysis of N₂O in 11 German forest sites and found very poor relationships between annual N₂O fluxes and a series of long-term ecosystem-scale state variables such as soil organic C, annual temperature, annual precipitation, pH, N leaching, N deposition, base saturation, and bulk density. Groffman *et al.* (2000) combined the data from Brumme *et al.* (1999) with data from several other forest studies and still did not manage to establish strong predictive relationships between soil N₂O flux and soil pH, annual precipitation or annual temperature, while they found that soil bulk density was the only statistically significant predictor of annual N₂O flux. They also questioned the robustness of annual estimates of soil N₂O based on a single year of data, as it cannot be known if this is a representative year for a given ecosystem type. Papen and Butterbach-Bahl (1999) measured soil N₂O fluxes from a Norway spruce and a beech forest ecosystem in Germany for 3 years (1994-1996). They found very high inter-annual variability, which was controlled by the annual climate (temperature).

1.1.9.2 Effects of forest management on soil N₂O flux

The rate of N₂O production and emission depends primarily on the availability of mineral N in the soil, soil temperature, water content and organic matter content, and the availability of labile organic compounds (Sahrawat and Keeney, 1986; Skiba and Smith, 2000). Therefore clearfelling of a forest can have an effect on soil N₂O fluxes, by altering these factors. As was mentioned before, after tree harvesting soil temperatures increase because direct solar radiation reaches the ground. Soil water

content increases and water table rises due to ceased transpiration, while logging residues and dying roots add labile organic matter. All these can favour nitrification and denitrification. Frazer *et al.* (1990) measured 18% higher N concentration, 343% higher specific rates of N mineralisation (the specific rate is calculated as the amount of N mineralised per unit total N) and higher nitrification rates in a forest soil five years after clearfelling than in an uncut mixed conifer forest, in Sierra Nevada of California. Dutch and Ineson (1990) measured increased denitrification during the first two years after clearfelling of a Sitka spruce forest on peaty gley soil, in Northern England; with a total loss of N between 10 and 40 kg N ha⁻¹ y⁻¹ from the clearfelled site compared to the 1.7 kg N ha⁻¹ y⁻¹ from the standing stand. These values included also losses as N₂, but from measurements in the same plots later, it was estimated that losses through N₂O accounted for about 80% of the total flux. Smolander *et al.* (1998) measured nitrification and denitrification before and after clearfelling of a Norway spruce stand in Finland. Before clearfelling, nitrification and denitrification occurred only in the plots that had been fertilised, while after clearfelling nitrification started in all plots indicating that clearfelling had increased nitrifier populations by increasing pH and net formation of mineral N. In incubation measurements they found that N₂O was produced only from the clearfelled plots, which was in accordance with their field measurements. Paavolainen and Smolander (1998) conducted further investigations in the same forest the third summer after clearfelling and they confirmed the same trend as Smolander *et al.* (1998).

1.1.10 The role of soils as sources and sinks for CH₄

The amount and concentration of methane (CH₄) in the atmosphere (10¹²–10¹⁴ g C y⁻¹ and 1.75 μmol mol⁻¹, respectively) (Schlesinger, 1997) are much lower than the ones for CO₂. However CH₄ has a global warming potential 23 times that of CO₂ (Houghton *et al.*, 2001) and it increases at an average rate of 1% y⁻¹, which is much faster than the rate of increase of CO₂ (Schlesinger, 1997); thus it can have a large

effect on climate change. The atmospheric CH₄ concentration varies with climate, as does that of CO₂, and the historic record of atmospheric CH₄ obtained from ice cores which extends to 420,000 years before present, indicates that at no time during this record has atmospheric CH₄ approached today's values (Houghton *et al.*, 2001). CH₄ has increased from a concentration of 0.75 μmol mol⁻¹ during the pre-industrial period to 1.75 μmol mol⁻¹ today, while Lelieveld *et al.* (1998) estimated that mean global CH₄ surface concentration will further increase to 2.55 μmol mol⁻¹ by 2050. Mean CH₄ levels in the northern hemisphere are 5 to 6% higher than in the southern hemisphere (Lelieveld *et al.*, 1998).

Methane emission to the atmosphere mainly derives from natural wetlands, rice paddies, enteric fermentation in ruminant livestock, landfills, fossil fuel production and biomass burning, which result in total emissions of about 600 Tg y⁻¹, of which about 60% are related to human activities (i.e. agriculture, fossil fuel use, etc.). Natural wetlands contribute about 145 ± 30 Tg CH₄ y⁻¹, with 60% of global emissions deriving from tropical wetlands (mostly due to the relatively high soil temperatures and high plant production). However, assessment of individual CH₄ source strength is very difficult and it is associated with uncertainties up to a factor of two (Lelieveld *et al.*, 1998).

There are also two significant sinks for atmospheric CH₄: 1) the reaction with OH radicals in the troposphere (506 Tg y⁻¹) (Houghton *et al.*, 2001); about 90% of the removal of CH₄ from the atmosphere occurs through this reaction (Lelieveld *et al.*, 1998). 2) The oxidation of CH₄ in aerobic soils, which is estimated to be between 29 and 38 Tg CH₄ y⁻¹ (Ridgwell *et al.*, 1999; Smith *et al.*, 2000). Temperate and boreal forest soils are considered to be a major sink for atmospheric CH₄ with a consumption of up to 9.3 Tg CH₄-C y⁻¹ (Stuedler *et al.*, 1989). Yavitt *et al.* (1990) estimated that forest soils in the temperate region consume 2.4 Tg CH₄ y⁻¹. However, the net soil consumption of CH₄ can be reduced by land use change (Hütsch *et al.*, 2001). Dobbie and Smith (1996) measured three times higher CH₄ oxidation rates in

a mixed deciduous forest soil than in adjacent agricultural land in Scotland. Maljanen *et al.* (2001) studied the effect of afforestation on previous agricultural peatlands on soil CH₄ fluxes. They measured CH₄ fluxes in a 1-year-old birch stand and two pine stands, 6 and 32 years old. They found that the oldest stand was a small sink or source of CH₄, while the younger stands were sources of CH₄, indicating that when the land use changes from agriculture to forest CH₄ oxidation might increase but very slowly.

1.1.10.1 Effect of environmental factors on soil CH₄ flux

Soil water content and water table depth are the major factors controlling CH₄ production and oxidation. Also, soil temperature is important, with soil CH₄ oxidation being less temperature dependent than CH₄ production (Dunfield *et al.*, 1993). CH₄ production by methanogenic bacteria is limited to strictly anaerobic environments, and thus the process is normally assumed to play a minor role for CH₄ flux in well-drained soils, although CH₄ production may be also significant from anaerobic microsites of such soils during periods with high soil water content (Sitaula *et al.*, 1995).

Liblik *et al.* (1997) found that mean water table depth accounted for 62% of variation in soil CH₄ emissions in peatlands in Canada, while incorporating both mean water table and mean temperature at the water table depth into the regression analysis, did not improve the R² (water table depth accounted for 61.2% of the variation in CH₄ flux and temperature at the water table added only another 1.7%). Castro *et al.* (1995) observed that temperature was a strong controller of soil CH₄ consumption at low temperatures (–5 °C to 10 °C) but it became less important between 10 and 20 °C in red pine (*Pinus resinosa* Ait.) and mixed hardwood stands in Massachusetts. Dunfield *et al.* (1993) in incubation experiments observed that CH₄ production rates showed a marked dependence on temperature with optima in the region of 25 – 30

°C and extremely low activities in the low temperature range 0-15 °C. Bowden *et al.* (1998) in incubation experiments with soil from mixed hardwood stands in Massachusetts, found that CH₄ uptake rates were strongly related to soil water content in the forest floor, with a maximum uptake at approximately 70% of field capacity, and lower values as soil water content approached 100% of capacity. Uptake rates were highest under the highest temperature conditions but the influence of temperature was not nearly as strong as that of water content.

However, Yavitt *et al.* (1990) found no significant correlation between CH₄ fluxes and soil temperature or water table depth in temperate wetlands and forests in the Appalachian Mountains. Also, Sjögersten and Wookey (2002) found no significant relationship between CH₄ fluxes and temperature in forest (*Betula pubescens*)-tundra ecotones in Sweden and Norway.

1.1.10.2 Effect of clearfelling on soil CH₄ flux

The effect of clearfelling on soil CH₄ fluxes has not been extensively studied. As was previously mentioned in the context of N₂O emissions, clearfelling alters many factors and these may also influence CH₄ fluxes. These factors include soil N dynamics (Vitousek and Matson, 1985; Smolander *et al.*, 1998), decomposition of organic matter (Binkley, 1986; Hendrickson *et al.*, 1989) and availability of substrate, soil water content, soil water table depth (Smethurst and Nambiar, 1990; Keenan and Kimmins, 1993) and soil temperature (Chen *et al.*, 1993).

It has been generally reported that clearfelling of temperate forests causes a reduction in CH₄ uptake. Bradford *et al.* (2000) observed reduced oxidation of CH₄ in the soil of clearfelled temperate forests (beech, Japanese larch and oak) in the UK. Steudler *et al.* (1989) reported inhibitions of the CH₄ oxidation by elevated soil N, and increases in soil mineral N concentrations have been observed after clearfelling

(Vitousek and Matson, 1985; Steudler *et al.*, 1991). Castro *et al.* (2000) measured CH₄ fluxes in two slash pine plantations in Florida before and after clearfelling and found that, although before clearfelling the soils were CH₄ sinks, after clearfelling they became a CH₄ source.

1.1.11 Importance of British soils for soil C storage and trace gas emissions

Britain has a forest area of 2.4×10^6 ha, covering 10.3% of its land area (Harrison *et al.*, 1995). A great amount of this carbon is contained in peatlands, approximately 30 times larger than the amount of carbon contained in all forests plus non-forest vegetation in Britain (Cannell *et al.*, 1993). Britain has about 2.11×10^6 ha of deep peatlands (more than 45 cm depth), excluding lowland fens and an unknown, but probably greater area of peaty gley and peaty ironpan soils classified as shallow peats (less than 45 cm depth). It is estimated that British peatlands (excluding lowland fens) contain about $3,000 \times 10^6$ t of carbon, 76% of which is in deep peats (Cannell *et al.*, 1993).

Commercial forestry represents a large land use category in Britain and during the period 1950s to 1980s about 190,000 ha of deep peatlands and about 315,000 ha of shallow peatlands (peaty gley soils), were drained, ploughed and planted with coniferous forests, mostly Sitka spruce (Hargreaves *et al.*, 2003).

The vital role of soils as a sink or source for C at the global scale and in offsetting atmospheric CO₂ concentrations makes it important to evaluate accurately the effects of forest management on soil C storage and trace gas emissions to the atmosphere. Furthermore, it is important to understand the soil carbon dynamics under intensively managed forest plantations (primarily Sitka spruce) and to evaluate the impact of

forest management practices on soil C storage and on emissions of the most important greenhouse gases: CO₂, N₂O and CH₄.

1.2 Site description

The study area is located in Harwood Forest, Northumberland, North East England (55° 10' N, 2° 3' W), 30 km inland from the North Sea coast. The establishment of the forest started in the 1950s and it currently consists of even aged stands of two main species, Sitka spruce (*Picea sitchensis*) and Lodgepole pine (*Pinus contorta*), both individually and in mixture. The size of the forest is approximately 4000 ha and its elevation varies from 200 to 400 m a.s.l.. The average annual rainfall is about 950 mm.

The dominant soil type in the area is Peaty gley (Pyatt and Suarez, 1997). It is a slowly permeable, seasonally waterlogged, fine loamy upland soil with a peaty surface horizon and frequently a clayey subsoil. The upper parts of Harwood are characterised by a coarse, loamy, very acid upland soil over rock with a wet peaty surface horizon and a thin ironpan. Basin depressions within the forest are characterised by a thick, very acid, raised bog peat soil. In the lower parts of the forest there are slowly permeable, seasonally waterlogged, fine loamy soils, usually with a clayey subsoil. Also, very small areas of acid brown earth occur within the forest (Wilson, 2000).

The forest was created by the cultivation and subsequent planting of the moorland and upland pasture. Ditches have been created to improve soil drainage and the trees were planted on ridges. This process resulted in peculiarities in the structure of the soil locally. In the furrows the organic horizon is often absent, due to depletion of the soil by running water. On the ridges there is often an inversion of the usual horizon arrangement (litter, organic, mineral), because of ploughing. Sometimes, the mineral

horizon is above the organic, while the litter horizon can appear in deeper layers. The forest is managed with rotations of about 40 years. At this age a whole stand is clearfelled and the planting of trees takes place after two or three years. Figure 1 presents the map of Harwood and the selected study areas within the forest, while Figure 2 shows a typical course of a chronosequence in the area, from an unplanted grassland to tree stands of varying age up to a clearfell.

The reasons for the choice of Harwood as a study area were: a) Harwood is typical of the Sitka spruce plantations found throughout Britain. Sitka spruce is the most common plantation tree in Great Britain (Cannell *et al.*, 1993; Adam, 1999), b) Harwood is a plantation on a fairly uniform concerning soil type and includes all ages along the desired chronosequence, c) it was also easily accessible from the place of study. Finally, Harwood was a site of the European CARBO-AGE project, which had as a central objective to explore the age-related dynamics of C exchange of European forests and to determine the role of European forests as C sinks or sources over their lifetime.

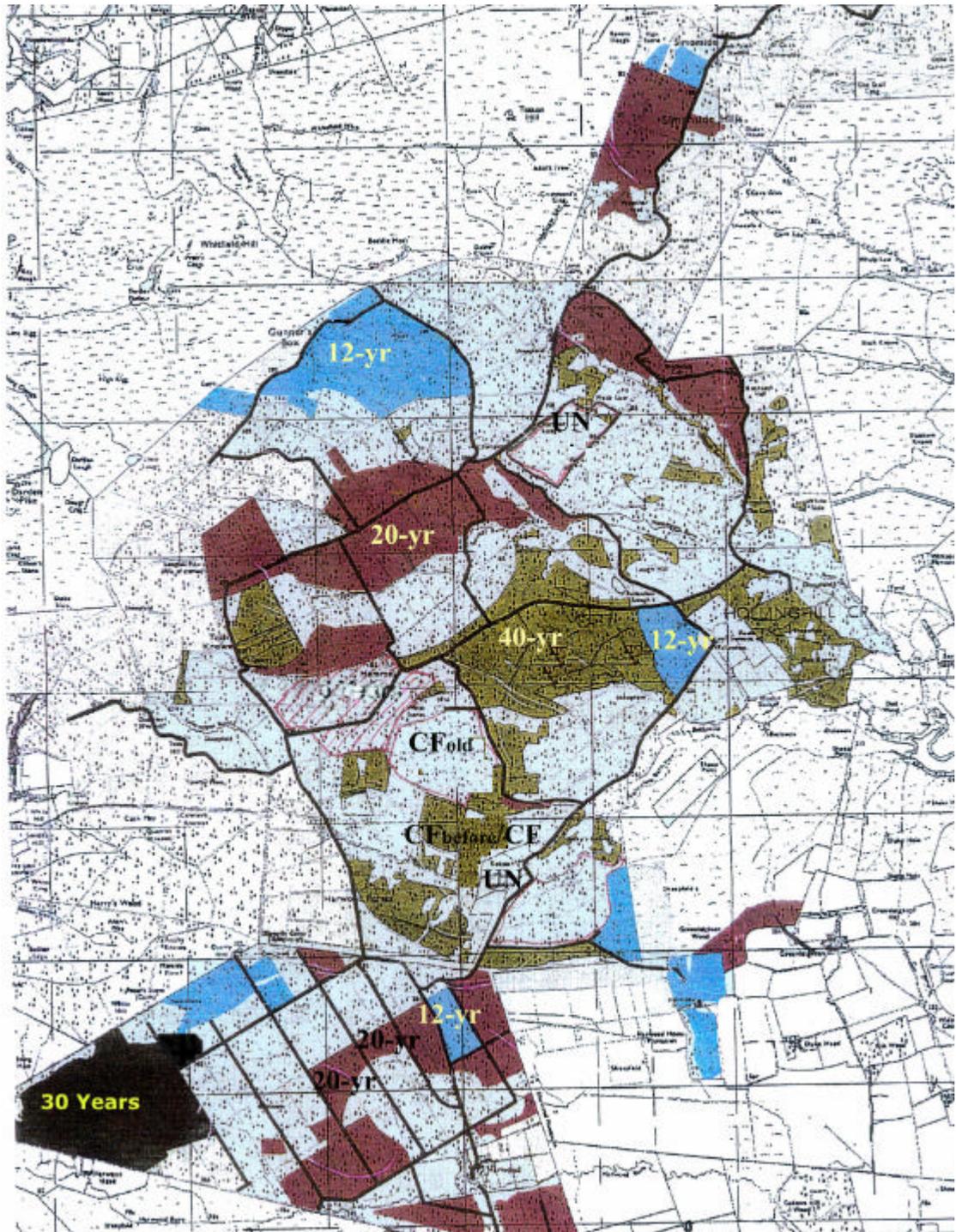


Figure 1: The map of Harwood and the study areas.



Figure 2: An example of the chronosequence studied in Harwood forest.

1.3 Aims of the study

The main aims of this study were to determine:

- i) the effects of afforestation with Sitka spruce (*Picea sitchensis* (Bong). Carr) on upland grasslands on peaty gley soil, and the effect of subsequent forest management on soil C stocks.
- ii) The effects of forest management on the soil carbon balance in the Sitka spruce chronosequence

- iii) the short-term effect of clearfelling on soil a) CO₂ b) N₂O and c) CH₄ fluxes and to examine possible relationships with the abiotic factors controlling the emissions of these gases (temperature, soil water content, water table depth).

Other aims were:

- iv) to investigate the spatial variability of soil CO₂ effluxes
v) to compare two methods used to measure soil CO₂ efflux.

1.4 Structure of the thesis

Chapter 1: This chapter introduces the importance of forest soils for C storage and fluxes of CO₂, N₂O and CH₄. It presents the method of forest chronosequence used in the studies of long-term changes on soil C storage as well as the methods used for the measurement of the trace gases. It also states the importance of environmental factors (soil temperature, soil water content and water table depth) on the fluxes of the trace gases and introduces the effect of forest clearfelling on these fluxes. It illustrates the importance of forest soils in Britain and the aims of this study. The site description of the study area is also presented here.

Chapter 2: This chapter describes the effects of afforestation with Sitka spruce on upland grassland, and the effects of forest management (i.e. clearfelling of a mature stand and re-growth of forest stands in second rotation) on the soil C stocks.

Chapter 3: The short-term effect of clearfelling on soil CO₂ fluxes and on the major environmental parameters (soil temperature, water content and depth to the water table) is investigated here. It also investigates the relationship of soil CO₂ efflux with the environmental parameters and how the relationship is modified by clearfelling, while two models are used (a multiple linear regression and a non-linear model) for

the combination of the effects of soil temperature and soil water content on soil CO₂ efflux.

Chapter 4: This chapter presents the short-term effect of clearfelling on soil N₂O and CH₄ fluxes and investigates their dependence on the environmental parameters and how they are affected by clearfelling.

Chapter 5: The C balance method is used in order to describe the effect of forest management on total belowground allocation in the whole chronosequence. The components contributing to the C balance: aboveground litterfall, root biomass, soil CO₂ efflux and change in the soil C stocks are analytically presented.

Chapter 6: The spatial variability of the measured soil CO₂ efflux is studied using a geostatistical approach. It also investigates the spatial variability of soil CO₂ efflux after disturbances, such as ploughing for tree planting and clearfelling, by using coefficients of variation of the measured fluxes.

Chapter 7: The two methods used for the measurement of soil CO₂ efflux (dynamic closed chamber and static closed chamber) are compared. It presents a series of steps taken for the explanation to explore the causes of the different estimates of CO₂ effluxes given by the two methods.

Chapter 8: This chapter gives a summary of the main conclusions derived from the integration of the entire body of work.