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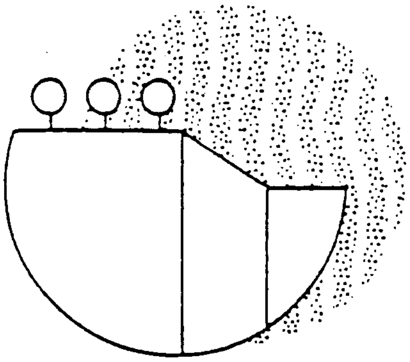

THE ROLE OF SOILS AND LAND USE IN THE GREENHOUSE EFFECT

Background paper
International Conference
"Soils and the Greenhouse Effect"

Wageningen, The Netherlands
August 14-18, 1989

A.F. Bouwman
July 1989

INTERNATIONAL SOIL REFERENCE AND INFORMATION CENTRE

The role of soils and land use in the greenhouse effect

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Background paper
International Conference on:
"Soils and the Greenhouse Effect"

The present status and future trends concerning the effect of soils and their cover on the fluxes of greenhouse gases, the surface energy balance and the water balance

Wageningen, The Netherlands
14-18 August 1989

organized by:
the International Soil Reference and Information Centre (ISRIC)

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PREFACE

The likelihood of significant changes in world climatic conditions taking place through increased atmospheric concentrations of so-called "greenhouse gases" have attracted considerable attention in recent years. Greenhouse gases can absorb thermal radiation and thus contribute to the warming of the atmosphere. In addition, changing patterns of evapotranspiration and overall reflectance (albedo) have received much attention in this respect. The world soils and their land use are important sources of a number of greenhouse gases such as carbon dioxide, methane and nitrous and nitric oxide.

The sources and sinks of most of the greenhouse gases are poorly identified. As for their contribution to the global gas budgets many knowledge gaps exist. Firstly, there is a need to study the fundamental processes underlying the emissions from soils and ecosystems. Secondly a great number of ecosystems are still undersampled. Thirdly methods for stratification of ecosystems need to be developed. On the basis of such classification systems the existing soil, vegetation and climate geo-referenced data bases need to be improved to enable a proper extrapolation of measurement data to regional and global scales. The need to assess the sources of greenhouse gases is driven by the need to parameterize global and regional models of climate. The coupling of ecosystem emissions with atmospheric chemistry is an important factor in the control of climate.

The International Geosphere-Biosphere Programme (IGBP-Global Change) of the International Council of Scientific Unions (ICSU) is starting multidisciplinary studies on the interactions between changes in biogeochemical cycles and processes in Solid Earth, Terrestrial Ecosystems, Marine Ecosystems and the Atmosphere. Recently ICSU's Scientific Committee on Problems of the Environment (SCOPE) organized a Dahlem workshop on "Exchange of trace gases between terrestrial ecosystems and the atmosphere". This workshop dealt with ecosystem processes and methods of extrapolation mainly. An important conference on this subject will be organized by SCOPE in Stockholm in 1990.

The present conference on "Soils and the Greenhouse Effect", 14-18 August 1989 in Wageningen, The Netherlands, is organized by the International Soil Reference and Information Centre (ISRIC) on behalf of the Netherlands Ministry of Housing Physical Planning and Environment (VROM). Financial support was received for the conference publications (including this background paper) from the Commission of the European Community (CEC) and the United Nations Environment Programme (UNEP). General financial support was received from the Netherlands Royal Academy of Arts and Sciences (KNAW). The Organization was made possible through cooperation with IGBP-Global Change, the International Society of Soil Science (ISSS) and the Unesco Man and the Biosphere Programme (MAB).

The Conference is one of soil scientists and researchers from many other fields. The aim is identify research gaps in the field of the geographic distribution of the world soils and land cover types (natural vegetation, cropland or grazing land) and trends in land use, on the one hand, and of greenhouse gas fluxes, evapotranspiration and albedo, on the other hand, it may form an important contribution of soil science to the study of climatic change.

This background paper is the result of 3 years of literature research and consultation with a great number of international scientists working in this field. It gives an overview of the present knowledge on the geographic distribution of soils and vegetation types, on regional and global land use changes, the role of soils and land cover in the emission of greenhouse gases, the water balance and the surface energy balance. It will be used as a reference for the discussions during the Conference. This background paper, the invited papers and papers of oral and poster presentations will be published after the Conference in the form of a book, to be published by Wiley and Sons publishing company.

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Lex Bouwman
July 1989

SUMMARY

1. Soil data bases

A global data base of soils is an important tool in trace gas studies. In many cases soil fertility, soil chemical and physical parameters play an important role in the production and emission of trace gases. Extrapolation of point measurement data for fluxes can only be done on the basis of a reliable soil data base. The best data base available is the FAO/Unesco Soil Map of the World which is available as a map or as a digital data base. The information on the map is however, not reliable in many parts of the world. Much new information has become available since its compilation. A major problem with the map is that it represents codes for soil associations, and information on texture of the topsoil for the major component of the association only. It is difficult to use this information for understanding trace gas fluxes. Only experienced soil scientists are capable to translate information included in the descriptions of the map units into terms of controlling factors of trace gas fluxes. A new soil data base, the Soil and Terrain Digital Data Base of the World at 1:1,000,000 scale, is being developed. This data base will contain, where avaialable, all the basic information needed for trace gas strudies.

2. Land cover and use

Many maps and data sets present the theoretical climax vegetation and not actual land cover or land use. For most trace gases the utilization of land determines the flux rate and pattern. For instance, rice paddies emit methane. The area of land used for rice is relatively well known. However, the number of crops per year and soil, crop and water management are still unknown factors. Shifting cultivation is a major cultivation system in the tropics. There is only little known on the present status and the changes in the area cleared annually and the number of people who are active shifting cultivators. At present, there is no good system for classifying the different land use types at a global scale at present. Some attempts to develop such a scheme have been made, but most classifications have local or regional value only.

There are many maps and digital data sets of vegetation available. Differences in the classification system used cause definitional problems which make intercomparisons difficult. Many maps and data sets present the theoretical climax vegetation and not actual land cover or land use.

There is great controversy concerning the global annual deforestation rate of tropical forests. Estimates of global forest destruction for permanent agriculture range between 10 and $20x10^{10}$ m^2 y^{-1}, much of it in the Amazonian region. There now remains only about $10x10^{12}$ m^2 of tropical forest. The figures reported for shifting cultivation show even greater variability than those for permanent clearing. Most of the data base is unreliable. Careful comparison learns that there is great controversy between the various estimates concerning the nature of changes (permanent clearing versus partial destruction or shifting cultivation; the latter process could account for an even greater extent of forest loss than the permanent clearing). There are also definitional difficulties with regard to the type of vegetation which is being cleared (primary forest, secondary forest; fallow forest, non-fallow forest; wet forest, moist forest, seasonal forest, dry forest; open forest, closed forest). Thirdly there are differences in the type of land use after clearing. This is important for making an estimate of the carbon flows and changes in fluxes of other trace gases. These will have a different pattern when forest is cleared for grassland or for cropland. Even when focussing on a specific area, the deforestation figures given in the literature disagree. Possibly the interpretation of time series of NOAA-AVHRR data (in conjunction with vegetation classifications) can improve the estimates of changes.

3. Global budgets of greenhouse gases

Carbon dioxide (CO_2)

The emissions of CO_2 from the terrestrial biota, including soil emissions and clearing and burning of forests, have contributed significantly to the present atmospheric CO_2 concentration. Presently however, the main anthropogenic source of CO_2 is the combustion of carbon based fuels. Future emissions of CO_2 resulting from fossil fuel combustion are highly uncertain. The estimated CO_2 emission for 1984 is $5.3x10^{15}$ g C y^{-1}. In projections for the year 2050 Keepin et al (1986) give estimates of CO_2 emission ranging from $2x10^{15}$ to $20x10^{15}$ g C y^{-1} depending on the application of various techniques of improving the efficiency of fuel use. In addition, biotic

sources of CO_2 (mainly deforestation) would contribute $1x10^{15}$ to $3x10^{15}$ g C y^{-1}, of which the soils contribution is estimated at $0.2x10^{15}$ to $0.9x10^{15}$ g y^{-1}.
Major problems in estimating the biotic CO_2 source are the extent of forest clearing on the one hand and the biomass of forests cleared on the other hand. Estimates of forest biomass show differences of over 200 %. This makes global C balances very unreliable. Sinks of CO_2 are the atmosphere (55%), oceans (30%) and terrestrial biota (15%). The annual increase of the atmospheric concentration is about 0.5% or $3.5x10^{15}$ g C y^{-1}. Possible effects of increased atmospheric CO_2 concentrations on plant production are not considered in this paper.

Methane (CH_4)
There is strong evidence, that the total CH_4 source has increased during the last decades. The increase of the human world population, the annual increase of the area of paddy rice cultivation (1.2%) and the increase of methane emission by termites due to shifts in land use correlate well to the atmospheric increase. This indicates, that the rise in atmospheric CH_4 is most likely related to anthropogenic activities.
The present annual release rate is $334x10^{12}$ to $714x10^{12}$ g y^{-1} of methane according to various authors. Individual sources are rice paddies (60-140), wetlands (40-160), landfill sites (30-70), oceans/lakes/other biogenic (15-35), ruminants (66-90), termites (2-42), exploitation of natural gas and coal mining (65-75), biomass burning (55-100). The global budget is well known. The allocation of the total budget among the sources is still not well known. The increase of methane over the past 200 years is probably due to the increase of emissions (70%) and a lesser amount is due to a possible depletion of OH. radicals (30%). The OH. depletion is caused primarily by the ever larger CO emission from various anthropogenic sources.
Major methane sinks are reaction with OH. radicals in the troposphere ($260x1^{12}$ g y^{-1}), transport to the stratosphere ($60x10^{12}$ g y^{-1}) and oxidation in arid soils ($60x10^{12}$ g y^{-1}). Clearly sources and sinks are not balanced in this budget.

Carbon monoxide (CO)
The major sources of CO are known, but their magnitudes are still uncertain. The background concentration of CO is increasing at a rate of 2 to 6 % per year, but estimates are uncertain due to fluctuations of sources and sinks and the relatively short residence time of CO in the atmosphere.

Estimates of global emission range between 1270 and $5700x10^{12}$ g CO yr^{-1} with an average of $2920x10^{12}$, the major sources being biomass burning (800), fossil fuel burning (450), oxidation of hydrocarbons incl. methane (960-1370). The primary sinks are oxidation to CO_2 (3000) and soil uptake (450). Estimates for the global sink strength range between $1960x10^{12}$ and $4750x10^{12}$ g CO yr^{-1} averaging $3600x10^{12}$. The model is not completely balanced, indicating the uncertainty in the estimates.

Nitrogenous greenhouse gases (N_2O and NO_x)
With the available data an attempt was made in this review to estimate the global soil flux of nitrous oxide. The global N_2O emission from cultivated soils (3 T g N_2O-N y^{-1}) and from natural soils (6 T g N_2O-N y^{-1}) form the major sources in the N_2O budget. In this estimate it is assumed that the emission induced by N-fertilization (1.5 to 3.2 T g N_2O-N y^{-1}) is included in the flux rate given for cultivated fields. Many of these estimates are highly uncertain. Particularly natural ecosystems are undersampled. One of the major sources could be the tropical forests of the earth, where in places extremely high fluxes have been measured. All N_2O is eventually tranferred to the stratosphere where it reacts with ozone. The size of this sink is 10.5 T g N_2O-N y^{-1}, while the atmospheric increase is about 2.8 T g N_2O-N y^{-1}. This gives a total sink strength of 13.3 T g.

The importance of NO and its relation to N_2O production was recognized in recent years. The production ration of NO:N_2O is a highly uncertain factor in all such calculations as it is very sensitive to abiotic controls such as oxygen pressure. Recent research indicates that NO production during nitrification of NH_3 in aerobic soils can exceed that of N_2O. NO catalyzes various atmospheric reactions in which CH_4 and CO are involved. It is very difficult to give a mass balance of these cyclic reactions and therefore a reliable estimate of the sink strength cannot be given.

4. The influence of land cover changes on the energy balance.
Land cover changes in general cause an increase in the ratio of reflected to incident radiation (albedo) of the earth's surface. Denudation or total deforestation causes the greatest albedo increase. As a consequence net radiation and the ratio of sensible heat flux to latent heat flux will increase. This implies that areas which in their original state are sources of latent heat, become foci for the generation of large amounts of sensible heat. This phenomenon has also been observed after partial deforestation. Especially in areas with tropical rainforest the albedo effect of denudation is expected to have a great impact on local, regional and possibly global climate. However, inclusion of this feature in global climate sensitivity models has received little attention so far. The attempts that have been made have not produced satisfactory results.

5. The influence of land cover changes on the hydrologic cycle.
Denudation of once forested land causes the ratio of sensible heat flux to latent heat flux to increase. The ultimate result of this change in the energy balance is that water originally lost through evapotranspiration now has to drain superficially. Apart from the erosion caused by this surface runoff a consequence is that less vapour will be emitted into the atmosphere. Less vapour will be available for cloud formation. Added to the above described shift in the energy balance this means that a decline in the degree of coverage provided by vegetation will provide a positive feedback to any tendency to aridity, to imbalance of local rainfall and also to the global heat balance.
The rates of dry canopy evaporation or transpiration are relatively well known. The rate of wet canopy evaporation and hence total evapotranspiration in relation to the vegetation type and rainfall distribution are so variable that no reliable regional or global figures can be given as yet.

5. Remote sensing
Thanks to the repeat cycle of 1 day of the NOAA-AVHRR will give a cloud free image of any geographic area during any period of time. Analysis of time series of such data enable study of spatial changes and dynamics of vegetation and land use. Therefore, for global monitoring of land cover the NOAA-AVHRR sensor data are highly suited. New AVHRR applications are being researched. The low spatial resolution AVHRR data can be supplemented with higher resolution MSS data or radar imagery when and where available.
Remote sensing of surface temperatures and correlation with meteorlogical measurements to estimate evapotranspiration has proved to be very accurate for agricultural crops on a large scale. However, since processes in the planetary boundary layer appear to be more determining for evapotranspiration than surface temperature and net radiation, remote sensing techniques for measuring evapotranspiration on a regional scale and for natural ecosystems are less suitable.

RESEARCH NEEDS

1. Soil data sets

The Soil and Terrain Digital Data Set (SOTER) is being developed. However, its completion will take another 15-20 years. Meanwhile, the old FAO/Unesco data base is a tool which shows gaps and inaccuracies in many parts of the world, particularly in the USSR, the Amazon Basin and Africa. In these areas new information is available now. This should be included in a new data base couled be developed for use in regional and global studies of climate, land use and trace gas fluxes. A workable scale for such studies would be 1:5,000,000, or in terms of resolution 1/4 x 1/4 degree. A soil data base that can be applied by users form various disciplines should include the following information: soil texture, soil depth, stoniness, slope, and some parameters important for soil fertility.

2. Land cover changes

- The question of how much of tropical forests has been cleared is very difficult to resolve. Careful definition of terminology is essential to land use studies. On a global scale the use of one single classification system for potential vegetation would improve comparison between studies. Proposed classification schemes are the physiognomic or 'landscape' classification systems by Unesco and by Gaussen. Bioclimatic schemes are those of Holdridge and Walter. The combination of the Unesco vegetation classification system with a system of land use classification as proposed by Dent or Matthews would offer a good basis.
- Coupling of a land use classification with time series analysis of data from the growing satellite data set (e.g. LANDSAT, NOAA-AVHRR) and with ground observations of actual land use should be given more attention.

3. Global budgets of greenhouse gases

Carbon dioxide (CO_2)

- There are number of areas of uncertainty in the estimates of CO_2 release from terrestrial biota. The largest biotic source of carbon dioxide is the transformation of forested lands to agriculture. This source is greatest in the tropics. Yet one of the greatest sources of uncertainty is the rate of clearing of tropical forests and to a lesser degree the total extent of each ecosystem. The use of one single vegetation classification system would improve estimates and mutual comparison.
- A second uncertainty is the volume and carbon density of forests. Field studies in combination with remote sensing techniques will prove helpful in improving knowledge of forest volumes and biomass density. However, the effect of density on the <u>net</u> flux of CO_2 is small compared to the effect of the rate of forest clearing.
- A third uncertainty is the soil carbon loss due to land cover changes. Modelling in combination with ^{14}C techniques are tools necessary to single out the pool size and historical and present carbon release from soils.
- Monitoring tropical regions with satellite imagery or with radar techniques of remote sensing may prove effective methods. The largest changes in the storage of carbon, i.e. those associated with the transformation of forest to nonforest, are the changes most easily detectable with remote sensing.

Methane (CH_4)
- As for the nitrogenous trace gases, the temporal and spatial variability of methane production is extremely high. Point measurements are helpful in uderstanding the process of methane production, but extrapolation of box measurement data to regional or global scale may be cause of serious errors.
- For the wetlands, one of the major sources, the total extent and the area of the various types of wetlands in relation to methane production are not known with sufficient accuracy to serve as a basis for global estimates.
- The extent of wet rice cultivation is known, but regarding soil properties, such as organic matter contents and temperature regimes, and land and water management in rice cultivation very few data are available on local and regional scales.
- Methane emission by landfill sites needs much more research with regard to quantities of organic wastes and methane release in relation to the types of waste.
- Methane production by termites and the increase of the termite population as a consequence of shifts in land use is still a subject of debate.
- Besides process oriented data, improved measurement techniques are essential. Source measurements for larger areas would provide a basis for global estimates of the methane sources.

Nitrogenous greenhouse gases (N_2O, NO and NO_x)
- The production of nitrogenous oxides under different soil moisture and soil temperature regimes, and in relation to other soil conditions needs further study. In this respect the relation between morphometric soil properties and soil processes and measured gas fluxes should be given more attention.
- The production of all nitrogenous gases is extremely sensitive to environmental factors. This is cause of the observed high temporal and spatial variability. Currently estimates are made using point source measurements. Extrapolation of these estimates are fraught with potential errors. Currently methods to measure these gas emissions directly over large land or water surfaces are in developmental stages (lidar measurements, eddy-correlation techniques). These measurements over large ecosystems should be made complementary to point measurements. The continued basic study of cycles of C and N is a prerequisite for the understanding of these gas fluxes.

4. Surface energy balance
Much is known about the complex relations as presented by the energy balance. A number of global coarse resolution data sets are available. However, no large scale data set is available as yet for characterization of land cover types in terms of albedo in relation to zenith angle and season. Such a data set would provide the basis for a global geographic quantification of the heat and energy fluxes.

5. Hydrologic cycle
- All the presented models and equations have only a local or temporally limited validity. Most of the evaporation models give reliable results under conditions of plentiful water supply; most of them work only for summer conditions. The prediction of wet canopy evaporation, evaporation from arid and semi arid lands and water loss during winter is still in its infancy.
- Water recycling or the vegetation feedback onto climate requires further study especially where the density of a vegetation cover is changed as in shifting cultivation.

6. Remote sensing
- Methods for using satellite imagery of the NOAA-AVHRR sensor to measure current rates of change in areas of forests globally are being developed. Such data combined with high resolution data such as radar, Thematic Mapper, SPOT or other sensors, should be correlated with existing vegetation and land use classification systems and field data.
- Narrower bands will be available in the near future. A proper selection of bands will enhance the capability of identifying land cover changes.
- Further research on lidar measurement of gas concentrations and eddy correlation techniques should be stimulated.
- Synthetic aperture radar provides all weather sensing. Operational spacecraft is still in future. Radar is especially useful for measurement of stand density, species differentiation, area delineation and tropical forest inventory. However, much more research and correlation with field observations and other sensors is required.

1. INTRODUCTION

1.1 GENERAL

The steadily increasing atmospheric concentration of carbon dioxide has provided an unequivocal signal of the global impact of human activities. Moreover, a number of other gases of biotic or industrial origin, or both, are increasing in concentration. The sources, sinks and dynamics of these gases are important for several reasons. First, many gases affect the chemistry and physics of the atmosphere. They alter characteristics as diverse as the energy budget of the earth, concentrations of oxidants in the atmosphere and absorption of ultraviolet radiation in the stratosphere. Second, trace gases or their reaction products affect terrestrial biota directly in ways that can be more or less species specific and that range from enhancing productivity or competitive ability to causing substantial mortality. Third, the production and consumption of trace gases in terrestrial ecosystems indicate the presence and magnitude of particular physiological processes or ecosystem fluxes or both. Measurement and understanding of trace gas fluxes will thus be useful for understanding terrestrial ecosystem dynamics as well as the atmosphere.

Climatologists are also concerned about the role of forests in recycling water vapour and the effects of massive deforestation on the rainfall regime and the hydrology of large river systems. Water vapour is one of the major greenhouse gases. Besides its direct role as absorber of thermal radiation in the atmospheric water vapour is also of eminent importance in cloud formation and rainfall. Moreover water vapour indirectly affects the radiation balance through cloud formation. Vapour also plays an important role in the energy transport between equatorial and temperate regions. Human interference in the land cover also brings about changes in the earth's albedo. The albedo plays a role in the surface energy balance and therefore has a large (mostly local and regional) climatic impacts. Since the latent heat flux also is a term in the energy balance, the subjects of albedo and evapotranspiration are closely related.

Present knowledge regarding the sources of disturbances to atmospheric chemistry and the impact of atmospheric chemicals on valued atmospheric components is presented in table 1.1. The convention used indicates only direct effects. The gases relevant to the subject chosen for this background paper are those gases originating from soils and which directly affect the thermal radiative budget or the concentrations of photochemical oxidants. These gases have asterixes in the source columns 3, 5 or 7 and in effect column B or C. The resulting gases are: carbon dioxide (CO_2), methane (CH_4), nitrogen dioxide and nitric oxide (as a group denoted by NO_x) and nitrous oxide (N_2O).

To assemble a complete picture of the role of soils in the greenhouse effect, it is necessary to attend to indirect effects as well. Indirect effects of a source change (i.e. soil change or land cover change) are induced changes in chemical species A which affect a given valued atmospheric component through an intermediate influence on chemical species B.

Atmospheric compounds having an indirect effect are carbon monoxide (CO) and ammonia (NH_3/NH_4^+). CO is oxidized to CO_2 whereby either ozone may be formed or consumed (see 1.2.2). 10% of atmospheric ammonia is oxidized to nitrogenous oxides thereby influencing the concentration of NO_x and in cases of ozone (see 1.2.3).

Thus, the complete list of gases with relevance to this paper is:

$$H_2O,\ CO_2,\ CO,\ CH_4,\ N_2O,\ NO_x,\ NH_3$$

These gases will be referred to as "greenhouse gases".

In appendix I the 1968 composition of the atmosphere is compared with the present (1985) concentrations of gases. From these data the annual increase rates are calculated. Possible causes of the increase in the concentrations of CO_2, CH_4, CO, N_2O, O_3, NO_x and NH_3 are the increasing emissions by the various sources and in some cases a reduced sink strength.

Table 1.1 Sources of major disturbances to atmospheric chemistry and the impacts of atmospheric chemistry on valued atmospheric components. The o's and *'es indicate that the listed chemical is expected to have a significant <u>direct</u> effect on the indicated atmospheric property or that the expected source is expected to exert a significant <u>direct</u> effect on the listed chemical (data from Clark ,1986; Crutzen and Graedel ,1986; Mooney et al., 1987). *'s are items relevant to this paper while o's indicate sources and effects not relevant. Gases for which both the source of disturbance and the impact are relevant to the subject of this paper are printed in italics.

chemical constituent	sources of disturbances											Impact					
	1	2	3	4	5	6	7	8	9	10	11	A	B	C	D	E	F
C (soot)						o			o	o			*			o	
CO_2	o	o	*			o	*		o	o			*				
CO	o		*			o			o	o	o						
CH_4			*	o	*	o	*	o					*				
C_xH_y	o	o				o	*							*		o	
NO_x	o		*			o	*		o	o	o			*	o	o	
N_2O	o		*			o	*		o	o			*				
NH_3/NH_4		o	*	o	*	o	*	o		o					o		
SO_x									o	o	o		*		o	o	o
H_2S	o		*		*		*										o
COS	o		*	o													
Org.S	o		*		*												o
Halo-carbons											o		*				
Other halogens	o								o	o	o						o
trace elements	o					o			o	o	o						
O_3												o	*	*			

Sources of disturbances:
1. Oceans and estuaries
2. Vegetation
3. Soils
4. Wild animals
5. Wetlands
6. Biomass burning
7. Crop production
8. Domestic animals
9. Petroleum combustion
10.Coal combustion
11.Industrial processes

Impacts:
A. Ultraviolet energy absorption
B. Thermal infrared budget alteration
C. Photochemical oxidant formation
D. Precipitation acidification
E. Visibility degradation
F. Material corrosion

[1] NMHC = Non-Methane hydrocarbons (isoprene, C_5H_8, and terpenes).

The topics which will be addressed in the Conference "Soils and the Greenhouse Effect" are (see Conference circular):

-quantification of the global spatial distribution of soils and their land cover;
-quantification of the fluxes rates of greenhouse gases, evapotranspiration and albedo for the world soils and their land cover types;
-mechanisms and modelling of gas fluxes, evapotranspiration and albedo for the major soils and land cover types;
-global trends in land use and changes taking place in soils and land cover under human influence;

-quantification of the global effect of soil and land use changes on the rates of emission of greenhouse gases, evapotranspiration and albedo;
-remote sensing techniques for monitoring land use.

To address the impact of changes in global land cover and in soils, it is attempted to evaluate the present state of knowledge (and lacunae) of greenhouse gas fluxes, evapotranspiration and the surface energy balance in chapters 2 to 7.
In chapter 2 the extent of the various soils and land cover types will be estimated using various sources. In chapter 3 a number of studies concerning the present global spatial land cover distribution and its changes are compared. Section 4.1 evaluates the fluxes of CO, CO_2 and CH_4. In section 4.2 the fluxes of nitrogen compounds (N_2O, NO_2, NO and NH_3) are discussed. In chapter 5 the effect of land cover changes on the water balance will be discussed. In chapter 6 the reflection characterisitics of soils and vegetation will be assessed. The remote sensing techniques for the monitoring of vegetation will be discussed in chapter 7. In the remainder of chapter 1 some aspects of the atmospheric chemistry of carbon and nitrogen compounds will be discussed.

1.2 ATMOSPHERIC CHEMISTRY OF CARBON AND NITROGEN COMPOUNDS

1.2.1 General. Tropospheric oxidizers.

Data on the atmospheric composition and properties of the various components are presented in appendix II. Regarding the most important greenhouse gases the following table 1.2 gives data on the annual rise in concentration, heat absorbing capacity and the contribution to the global temperature rise.

Table 1.2 Atmospheric concentrations of the major greenhouse gases, their rise, residence time and contribution to the global warming. Reproduced from: Bouwman (1989a).

type	residence time (y)	annual rise (%)	1985 concentration	radiative absorption potential[8]	contribution to global warming (%)[9]
CO_2	2[1]	0.5[2]	345ppm[2]	1	45
CO	2-3[3]	2-6[3]	90ppb[1]	n/a	?
CH_4	7-8[4]	1.3[4]	1.65ppm[6]	+	10
N_2O	100-200[5]	0.25[5]	300ppb[5]	++	5
O_3[10]	0.1-0.3[1]	2.0[1]	n/a[7]	+++	5
CFC's[11]	65-110[1]	3.0[1]	0.18-0.28ppbl	++++	25

[1] Ramanathan et al. (1985) (data for 1980); [2] Bolin (1986); [3] Khalil & Rasmussen (1984); [4] Khalil & Rasmussen (1985); [5] Crutzen & Graedel (1986); [6] Bolle et al. (1986); [7] O_3 varies from 25ppb at surface to 70 ppb at 9 km (Ramanathan et al., 1985); [8] Swart (1988; pers. comm.); CO_2=1; + = 10-100; ++ = 100-1000; +++ = 1000-10000; ++++ = > 10000; [9] calculated for period 1980-2030 with data from Ramanathan et.al. (1985); [10] O_3 = ozone; [11] Chlorofluorocarbons; data presented are for the two major CFC's.

The troposphere is the part of the atmosphere nearest to the earth's surface and extends to 10 km in polar regions to 15-20 km in the tropics. The tropopause forms an abrupt change to the stratosphere. The stratosphere extends to about 55 km.

The principle oxidizing reagents in the lower atmosphere are ozone (O_3) and hydroxyl-radicals (OH.). Ozone is produced in the troposphere by the peroxyl radical oxidation of NO:

$$NO + RO_2 \xrightarrow{\quad} NO_2 + RO.$$
$$NO_2 \xrightarrow{hv} NO + O.$$
$$O. + O_2 \xrightarrow{m} O_3$$

Ozone is also formed in the stratosphere through dissociation of molecular O_2 and transported towards the troposphere. Tropospheric destruction of ozone also occurs and this process constitutes the primary source of OH.:

$$O_3 \xrightarrow{hv} O_2 + O.$$
$$O. + H_2O \xrightarrow{\quad} 2\ OH.$$

A second source of OH. are oxygenated organic compounds,

e.g.:
$$HCHO \xrightarrow{hv} H. + CHO.$$
$$H. + O_2 \xrightarrow{m} HO_2.$$
$$CHO. + O_2 \xrightarrow{\quad} HO_2. + CO$$

followed by either:
$$HO_2.+HO_2. \xrightarrow{\quad} H_2O_2 + O_2$$
$$H_2O_2 \xrightarrow{hv} OH. + OH.$$

or:
$$HO_2. + NO \longrightarrow NO_2 + OH.$$

1.2.2 Carbon compounds

The carbon components, which are involved in the atmospheric carbon cycle are CO, CH_4, CO_2 and NMHC (non methane hydrocarbons). Since only the soils- and land cover related sources are considered the compounds CO, CH_4 and CO_2 are of interest. However, CO is not interacting in the atmospheric radiation balance and is oxidized to CO_2 relatively quickly, thereby influencing the mixing ratios of other greenhouse gases. CO_2 is chemically not reactive in the atmosphere.

Methane (CH_4)

Most of the methane present in the troposphere is oxidized to CO. All reaction paths proceed via the intermediate product formaldehyde CH_2O:

CH_4 + OH.	-->	CH_3. + H_2O
CH_3. + O_2	-->	CH_3O_2.

Possible subsequent reaction paths are:

CH_3O_2. + HO_2.	-->	CH_3O_2H + O_2
CH_3O_2H + OH.	-->	CH_3O_2. + H_2O

Reaction path with NO>10pptv:

CH_3O_2. + NO	---->	CH3O. + NO_2
CH_3O. + O_2	---->	HCHO. + HO_2.
HO_2. + NO	---->	OH. + NO_2
2[NO_2	-hv->	NO + O.]
2[O. + O_2	--m->	O_3]
net:		
CH_4 + HO_2.	-->	HCHO + 2 O_3 + H_2O

Reaction path with NO<10pptv:

CH_3O_2.+ HO_2.	---->	CH_3O_2H + O_2
CH_3O_2H	-hv->	CH_3O. + OH.
CH_3O. + O	---->	HCHO + HO_2.
net:		
CH_4 + O_2	---->	HCHO + H_2O

After both paths (with and without sufficient NO present) the next step is equal for both:

HCHO	-hv->	H. + HCO.
H. + O_2	--m->	HO_2
HCO. + O_2	---->	CO + HO_2.
CO + OH.	---->	H. + CO_2

H. + O_2	--m->	HO_2.
3[HO_2. + NO	---->	NO_2 + OH.]
3[NO_2	-hv->	NO + O.]
3[O. + O_2	--m->	O_3]
HCHO + 6O_2	-->	CO_2 + 3O_3 + 2OH.

2[H. + O_2	-->	HO_2.]
3[HO_2. + O_3	-->	OH. + 2O_2]
HCHO + 3O_3	-->	CO_2+ 3O_2+2OH.

The important implications are, that in the presence of sufficient NO there is a net production of 3.7 O_3 molecules and 0.5 OH. radicals for each CH_4 molecule oxidized. In the absence of sufficient NO a net loss of 1.7 O_3 and 3.5 OH. may occur.

For other, more complex hydrocarbons (represented by RH) the reaction path is similar to that of methane in the first two reactions:

RH + OH.	-->	R. + H_2O
R. + O_2	-->	ROO.

Two possible consecutive paths exist:

$ROO. + R'OO. \text{-->} ROOR' + O_2$
$ROOR' \text{-hv->} RO. + R'O.$

$ROO. + NO \text{-->} RO. + NO_2$
$RO. + O_2 \text{-->} R_{-1}CHO + HO_2.$

Carbon monoxide (CO)

The oxidation of carbon monoxide proceeds as follows (Logan et al., 1981):

with> 4-20 pptv NO:

$$
\begin{array}{lll}
CO + OH. & \text{-->} & H. + CO_2 \\
H. + O_2 & \text{-->} & HO_2. \\
HO_2. + NO & \text{-->} & OH. + NO_2 \\
NO_2 & \text{-hv->} & NO + O. \\
O. + O_2 & \text{-m->} & O_3 \\
\hline
\text{net:} & & \\
CO + 2O_2 & \text{-->} & CO_2 + O_3
\end{array}
$$

with < 4-20 pptv NO

$$
\begin{array}{lll}
CO + OH. & \text{-->} & H. + CO_2 \\
H. + O_2 & \text{-->} & HO_2. \\
HO_2. + O_3 & \text{-->} & OH. + 2O_2 \\
\hline
\text{net:} & & \\
CO + O_3 & \text{-->} & CO_2 + O_2
\end{array}
$$

The ratio of the CO concentration in the Northern hemisphere to that in the Southern hemisphere is 1.5:1 to 2:1 due to higher emissions. There is also a seasonal fluctuation.

1.2.3 Nitrogen compounds

Nitrogen dioxide and nitric oxide (NO_x)

As can be deducted from the above reactions in which methane and carbon monoxide are involved, NO plays an essential role in the oxidation of CH_4 and CO. The reactions of NO and NO_2 (together denoted NO_x) in the atmosphere are diverse. NO_x molecules play an important catalytic role in many photochemical reactions. In the troposphere NO_x enhances the formation of O_3, while in the stratosphere the opposite is the case.

During daytime HNO_3 is formed according to:

$$NO_2 + OH. \quad \text{-m->} \quad HNO_3$$

during nighttime a different path is followed:

$$
\begin{array}{lll}
NO_2 + O_3 & \text{-->} & NO_3 + O_2 \\
NO_3 + NO_2 & \text{-->} & N_2O_5 \\
N_2O_5 + H_2O_{aq} & \text{-->} & 2\ HNO_3
\end{array}
$$

During the photochemical breakdown of many non-methane hydrocarbons various organic nitrates are formed. The most important of these, PAN (peroxy acetylnitrate: $CH_3C[O]O_2NO_2$) is an important reservoir of NO_x in urban areas, but may occur in the middle and upper troposphere as well (Levine et al., 1984). PAN is formed as an intermediate, and it decomposes according to the reaction:

$$CH_3C[O]O_2NO_2 \text{-->} \quad CH_3C[O]O_2. + NO_2$$

thereby liberating NO_x radicals.

NO_x catalyses the destruction of ozone:

$$
\begin{array}{lll}
O_3 & \text{-hv->} & O. + O_2 \\
O. + NO_2 & \text{---->} & NO + O_2 \\
NO + O_3 & \text{---->} & NO_2 + O_2 \\
\hline
2O_3 & \text{---->} & 3\ O_2
\end{array}
$$

Below 40 km ozone is formed via the reaction path:

$$
\begin{array}{lll}
O_2 & \text{-hv->} & 2O. \\
2[O.+O_2 & \text{--m->} & O_3] \\
\hline
3O_2 & \text{---->} & 2\ O_3
\end{array}
$$

Below 25 km ozone is formed as follows:

$$
\begin{array}{lll}
HO_2. + NO & \text{---->} & OH. + NO_2 \\
NO_2 & \text{-hv->} & NO + O. \\
O. + O_2 & \text{--m->} & O_3 \\
\hline
HO_2. + O_2 & \text{---->} & OH. + O_3
\end{array}
$$

Above about 25 km the effect of NO_x additions would be to lower the ozone concentration, below 25 km NO_x protects ozone from destruction.

Nitrous Oxide (N_2O)

In the stratosphere N_2O is oxidized to NO via:

$$
\begin{array}{lll}
O_3 & \text{-hv->} & O. + O_2 \\
O. + N_2O & \text{---->} & 2\ NO
\end{array}
$$

Ammonia (NH_3)

Ammonia itself is not capable of absorbing thermal radiation. It is lost from the atmosphere through wet and dry depostion, but 10% (Crutzen, 1983) to 20% (Levine et al., 1984) of all atmospheric NH_3 will be oxidized by OH.. The reaction shown below is rather slow compared to the estimated residence time of NH_3 in the atmosphere.

$$OH. + NH_3 \quad \text{-->} \quad NH_2 + H_2O$$

The subsequent chemistry of NH_2 is uncertain:

$$
\begin{array}{llll}
 & NH_2 + O_2 & \text{-->} & NH_2O_2 \\
\text{or:} & NH_2 + NO & \text{-->} & \text{products } (N_2, N_2O) \\
 & NH_2 + NO_2 & \text{-->} & \text{products } (N_2, N_2O) \\
 & NH_2 + O_3 & \text{-->} & \text{products } (NH., HNO, NO)
\end{array}
$$

At NO_x concentrations below 60 ppt the oxidation of ammonia leads to enrichment of oxides, while with NO_x concentations exceeding 60 ppt, the oxidation process could provide a sink for NO_x.

Ammonia may also react with gaseous HNO_3 to form aerosol nitrate:

$$NH_{3(g)} + HNO_{3(g)} \text{<-->} NH_4NO_{3(s\ or\ aq)}$$

2. DISTRIBUTION OF THE MAJOR SOILS OF THE WORLD

2.1 INTRODUCTION

In this chapter different soil classification systems for global scale maps and global maps and data bases are evaluated.

Reviews of various systems of soil classification are presented in most books on soil science (e.g. Buringh, 1979; Fitzpatrick, 1983). Two systems are of particular interest in the context of global ecological studies. These are the legend of the FAO/Unesco soil map of the world (1971-1978) and the USDA Soil Taxonomy (USDA, 1975). Both systems will be discussed. Furthermore, a number of global soil maps and global data bases will be compared.

2.2 SOIL CLASSIFICATION SYSTEMS

FAO Soil Map of the World

The FAO/Unesco (1971-1978) Soil Map of the World is a compilation of all soil maps available. All the maps had to be translated in order to achieve uniformity. Soils that were geographically related were combined in soil associations called 'major soils'. These major soils, 106 in total, are shown separately or in combination. Some special properties are indicated by extra map symbols. At the time of compilation, soil maps were available for limited areas of many countries. The classification used in Volume I (legend) is not a systematic classification of all soils of the world. It is a classifcation of soil mapping units. The maps presents more than 5000 map units or delineations. The original systems distinguishes 28 groups of major soils, and 106 soil units. In the codes of map units the textural class of the major component of the association and the map unit slope class is indicated. On the map phases such as stony, lithic, petric, petrogypsic, saline, are indicated. The FAO legend is especially designed for use on a very small scale. The map units are sufficiently broad to have general validity and contain sufficient elements to reflect as precisely as possible the soil pattern of a large region. A modification of the traditional genetic classification system combined with the concept of the diagnostic horizon was used. The only separations based on climate are the Yermosols and Xerosols, otherwise soil temperature and moisture are only used where they correlate with other properties such as in Gleysols and the separation of areas of permafrost. It is easily transcribed to other classifications, and the legend is used in many countries at a national level too. Very recently a new version of the classification system was proposed (FAO, 1989). The new version distinguishes 28 major soils and 153 soil units. The properties of the major soils is discussed in FAO (1989).

USDA Soil Taxonomy

The USDA (1975) soil taxonomy is a rather complicated systems which requires thorough training to handle it. It is a morphometric system. This means that all properties used to characterize soils can be measured in the field or in the laboratory. Soil are classified according to the absence or presence of such properties. In many situations this is a disadvantage, in particular when the laboratory analysis data are not available. Soil taxonomy distinguishes 10 soil orders, which are further subdivided into suborders, great groups, subgroups, families and series. The orders in Soil Taxonomy are in alphabetical order: Alfisols, Aridisols, Entisols, Histosols, Inceptisols, Mollisols, Oxisols, Spodosols, Ultisols and Vertisols. Some of these names are similar to those used on the FAO/Unesco soil map of the world, such as Histosols and Vertisols. Alfisols are similar to Luvisols, Aridisols are almost similar to Xerosols and Yermosols; Oxisols to Ferralsols, and Ultisols to Acrisols. However, some major soils of FAO/Unesco soil map of the world are not in the first category of Soil taxonomy, e.g. Solonchaks, Andosols, Gleysols, Chernozems.
The second category is the suborder level, according to diagnostic soil horizons and properties. Many of them are similar to FAO/Unesco diagnostic horizons. For example, a mollic epipedon (USDA, 1975) is a mollic A horizon (FAO/Unesco, 1971-1978). There are also umbric, histic and ochric epipedons (epipedon = surface layer). For the subsurface horizons (B horizons of FAO/Unesco) there is also similarity, e.g. the argillic, natric, cambic and oxic horizons. However, there are also some differences not discussed here. For global and continental studies

however, only the orders, suborders and possibly great groups are of relevance. Finer subdivisions are too detailed.

2.2 GLOBAL SOIL MAPS AND DATA SETS

A great number of global soil maps are available. Each of them offers specific advantages, which will be outlined shortly in this section. In recent years digital geographically referenced data sets are gaining popularity beacause of their flexibility of use, and their capability of combining and overlaying with other types of maps, such as vegetation maps.

Table 2.1 List of global soil maps and digital data sets

reference	scale	classification used	No.of types
maps			
FAO/Unesco (1971-1981)	1:5,000,000	FAO	126
USDA (1972)	1:50,000,000	USDA (1975)	117
FAO (1983)	1:25,000,000	FAO	18
Kovda et al. (1975)	1:10,000,000	Kovda et al. (1967)	280
CEC (19..)	1:1,000,000	FAO	126
digital data sets			
Gildea and Moore (1985)	½x½ degree	FAO	106
Zobler (1986)	½x½ degree	FAO	106
UNEP/GRID	1:5,000,000	FAO	106
Matthews & Rossow (1987)	1x1 degree	USDA (1975)	43
Bouwman (1989c)	1x1 degree	FAO (based on FAO,1983)	18

Most of the digital data bases shown in Table 2.1 are based on FAO/Unesco (1971-1981). A disadvantage of that map is that it shows only codes for soil associations, with special symbols for topsoil texture of the major component of the association, slope class. Furthermore there are phases for e.g. stoniness indicated with a map symbol. However, most information is in the soil reports and not on the map, in a form difficult to translate into parameters such as soil available water capacity and soil fertility. Only experienced pedologists are able to read the report and to use the map. A further disadvantage of the FAO/Unesco map is that in many areas it is based knowledge available in the 1960's. In the mean time much information has become available for regions such as the Amazon basin, the USSR, parts of Eastern Africa, Western Europe.

A very recent development is the compilation of a 1:1,000,000 Global Digital Data Base (SOTER) (Shields and Coote, 1988) which has started in pilot areas in South America and North America. Further pilot areas are projected in areas such as West Africa and the mediterranean coastal zone (southern Europe). SOTER uses no soil classification in order to be universally applicable and to avoid problems of interpretation. Furthermore it includes attribute files containing all available soil data necessary to derive information on e.g. available water capacity, rooting depth, drainage condition.

Maps of the agro-ecological zones of the devloping world were published by FAO (1978-1981) in the Agro-Ecological Zones project. Soil climatic data can be derived from the small maplets in the reports of the Soil map of the World (FAO/Unesco, 1971-1981). A map of soil moisture regimes was compiled by USDA (1971). A map of wetlands and soils with hydromorphic properties was produced by Van Dam and Van Diepen (1982). This map shows all the flat wetlands and in the accompanying report the suitability for rice cultivation is assessed. Aselmann and Crutzen (1989) compiled a data set of the world's wetlands using a 2.5x2.5 degree grid and Matthews and Fung (1987) also compiled a data set of the wetlands of the world. The latter data set is not geo-referenced, however.

2.3 DESCRIPTION OF THE WORLD'S MAJOR SOIL GROUPINGS

In this section short descriptions will be given of the major soil groupings occurring on the global map compiled by FAO (1983). Much of the information is from Fitzpatrick (1983), Buringh (1979) and FAO/Unesco (1971-1981). At the end of section 2.3 the areal extent of these major soil groupings is given. In table 2.2a a matrix is presented of major soil groupings and land cover types. Table 2.2b shows a matrix of the same soil groupings and the Holdridge Bioclimatic Zones (Holdridge, 1967). For this compilation the following digital data bases were used: for soils the data base compiled by Bouwman (1989c), for land cover types Matthews (1983) and for bio-climates the data base compiled by Leemans (1989). The bioclimatic zones defined by Holdridge (1967) are listed in Appendix V. In Appendices IV matrices of ecosystems for the 6 continents presenting bioclimatic zones, soil groupings. For each ecosystem the % distribution of land cover is also given.

1. Acrisols (A) (USDA:Ultisols)
Characteristics of Acrisols:
- strong acidity;
- high aluminium saturation, low base saturation;
- very low availability of nutrients;
- topsoil organic matter easily lost;
- weak physical structure, high susceptibility to rainfall erosion;
- water may stagnate on the argillic B-horizon, causing impeded internal drainage;
- low in trace elements.

Acrisols are formed in Wet Equatorial to Tropical wet-dry climates. The thick, acid and strongly leached upper horizons create many problems for utilization. Even after liming and heavy fertilizer dressings their productivity is low. Deficiencies of micronutrients is common. Annual cultivation leads to rapid decomposition of the soil organic matter, and also to rapid structural deterioration, deficiencies of minor and major nutrients. Continued cultivation may lead to compaction and increased danger of water erosion. The required fallow periods are long under low management, with 1-2 years of cultivation followed by 20 or more years of rest. Under high management levels, shorter rest periods under grass or green manure crops are required.

2. Chernozems (C), Phaeozems (H) and Greyzems (M) (USDA:Mollisols)
* Characteristics of Chernozems:
- inherent fertility good;
- excellent physical structure;
- high available water capacity;
- rich in organic matter;
- calcareous layer within 125 cm;
- moderate to high cation retention capacity.

Chernozems are developed almost exclusively in loess, but they also occur in other sediments. Chernozems are confined largely to continental conditions from Humid Continental to Mid-Latitude steppe. Annual cultivation of these soils leads to rapid loss of soil organic matter. Required fallow periods are short.

* Characteristics of Phaeozems:
- dark organic rich topsoil;
- good structure;
- high available water capacity;
- moderate to high cation retention capacity.

Phaeozems are developed under continental conditions with dry summers; evapotranspiration in summer exceeds precipitation. Therefore Phaeozems are susceptible to droughtiness.

* Greyzems are integrades between Chernozems and Luvisols (see Soil Group No. 8). They are formed in warm continental areas under grassland cover. Topography in general is gently sloping.

3. Podzoluvisols (D) and albic Luvisols (La)
(USDA: Alfisols and Eutroboralfs respectively)
* Characteristics of Podzoluvisols:
- poor drainage;
- moderately acid soil reaction.

Podzoluvisols occur in cool humid continental regions. They are confined to flat or gently sloping areas where moisture can accumulate in the upper part of the soil.

* Characteristics of albic Luvisols
The subgroup of albic Luvisols are Luvisols (see 8 and 9) with a bleached horizon in between the humus rich topsoil and a heavier subsoil. This bleached horizon may exceed 20 cm in thickness. These soils occur in humid climates with a marked dry season. Albic Luvisols are moderately suitable to highly suitable for cropping, depending on the thickness of the above discussed bleached horizon. With high levels of management these soils are always good.

4. Ferralsols (F) (USDA: Oxisols)
Characteristics of Ferralsols:
- strong acidity;
- very low availability of nutrients;
- high aluminium saturation and low base saturation;
- low cation retention;
- no reserves of weatherable minerals;
- organic matter predominantly in topsoil, is easily lost after cultivation;
- good physical structure and low inherent susceptibility to rainfall erosion.

Ferralsols are formed in Wet Tropical, Trade wind Littoral and Tropical Wet-Dry climates, with mean annual temperatures of generally over 25°C and mean annual precipitation of over 1000 to 1200 mm. Annual cultivation of Ferralsols leads to rapid loss of organic matter. Since organic matter is responsible for a great part of the cation retention, the latter will also decrease. Under a high level of management limited rest periods are required to control pests and diseases. Although inherent erodibility of Ferralsols is low, their usually sloping to rolling topography may induce high run off and soil loss rates.

5. Histosols (O) and Gleysols (G) (USDA: Histosols and Aquepts/Aquents, respectively)
* Histosols are soils with an organic layer of at least 40 cm thickness. A more common name is organic soils peat soils. These soils frequently occur in association with Gleysols and usually have a flat topography. Drainage is a problem common to these soils.

* Gleysols (G) occur on level land, in many cases with a high water table. Their properties, such as texture, physical and chemical properties, vary widely. The Gleysols (and also Fluvisols) formed in deposits of the tributaries of the Amazon for example are formed in very poor material deposited by the rivers carrying material derived from very poor eroded Ferralsols, Acrisols and acid rocks. In other regions however, Gleysols and Fluvisols may be among the best soils available if their drainage is well managed. With intermediate and high input levels there is no need for rest periods.

6. Lithosols (I) (in new FAO legend: Leptosols) (USDA: lithic subgroups)
The group of lithosols comprises all the soils which are shallow (less than 30 cm deep) independent of the parent rock, climate or physiographic position.

7. Kastanozems (K) (USDA: Mollisols)
Characteristics of Kastanozems:
- high organic matter content in the topsoil;
- good structure;
- high available water capacity;
- high inherent fertility, high cation retention capacity.

Kastanozems are found predominantly in semi arid climates in the middle latitude steppe conditions. Grassland of medium height is their natural vegetation, their overall topography is gently sloping. There is a fairly wide range of variability within this soil group as determined by differences in parent material, age and topographic position. Usually these soils are found on elevated situations, but they may form a continuum to Solonetz or Solonchaks (Soil group 13) in depression and also to Chernozems (Soil Group No. 2).

8. orthic Luvisols (L) and Cambisols (B) (USDA: Alfisols and Inceptisols, respectively)
* Characteristics of Luvisols:
- inherent fertility moderate;
- moderate to high cation retention capacity;
- organic matter content low to moderate;
- physical structure weak in the topsoils, moderate but unstable in the argillic B-horizon;
- moderate to high available water capacity.

These soils are formed under humid conditions with a marked dry season. The precise temperature regime and total precipitation are variable. Luvisols can be found in Tropical Wet-Dry, Humid Subtropical, Mediterranean and Humid Continental climates. In addition they are found in semi arid climates where they were fotrmed under more humid conditions and are now fossiled to drier climatic conditions.

They constitute one of the major soils for food production, particularly maize, sorghum, groundnuts, and other foodcrops. At a low input level, a fallow period of about 2 years in 3 is required to keep the soils in a good condition. At high levels of management, the required fallow period is shorter (Young and Wright, 1980).

* Characteristics of Cambisols:
This group of soils comprises a variety of soils in the tropics and temperate regions. The soils range from shallow to moderately shallow soils occurring in cool upland regions (South America) to deep soils in old alluvium in the Ganges flood plain. Generally the organic matter status is moderate to good. With high inputs almost continuous cultivation is possible.

9. Nitosols (N) and ferric Luvisols (L) (USDA: Alfisols/Ultisols)
* Characteristics of Nitosols:
- moderate to high cation retention;
- nutrient availability high in eutric nitosols, low in dystric nitosols;
- dystric nitosols have a problem of acidity;
- moderately high reserves of weatherable minerals;
- generally higher organic matter levels than in other freely drained soils;
- high available water capacity;
- moderately high erosion hazard.

Nitosols are strongly weathered kaolinitic soils, the principal feature being the steady increase in clay with depth to a maximum in the middle layer and deeper remaining uniform for some depth. Nitosols are among the most fertile tropical soils. At low input levels eutric nitosols may well be cultivated for at least 1 out of 2 years, while dystric nitosols need a longer rest period. Nitosols respond well to inputs and under high levels of management they can be continuously cropped.

* Characteristics of Ferric Luvisols: these soils are the tropical lateritic podzolic soils with high base saturation. They are widespread in the tropical savanna zone. They have an horizon with clay illuviation with a moderate but unstable structure. The physical structure is commonly weak in the topsoils. Organic matter content and inherent fertility are moderate.

10. Podzols (P) (USDA: Spodosols)
Characteristics of Podzols:
- chemically poor;
- good drainage;
- acid soil reaction;
- low cation retention capacity;
- generally low available water capacity.

Podzols are usually formed in coarse to medium textured unconsolidated deposits, often containing a high proportion of boulders and stones. They generally occur in any topographic situation where aerobic conditions prevail and water is allowed to percolate freely through at least the upper and middle parts of the soil profile. The range of climatic conditions under which Podzols are formed is quite wide. They are most widespread under a tundra or marine climate with rainfall variation from 450 to 1250 mm per annum. They also occur under Humid Comtinental cool summer climate and Humid Tropical climate. In both of these environments the parent material is highly siliceous producing conditions which favour Podzol formation. Under high levels of management these soils may be improved a lot. Tropical Podzols occur either in the cold tropics or in the lowland rainforest areas. The latter Podzols are formed in sands and their suitability for cultivation is probably low.

11. Arenosols (Q) and sandy Regosols (R) (USDA: Psamments)
Characteristics of both Arenosols and sandy Regosols:
- high sand content;
- low organic matter levels;
- low cation retention and nutrient availability;
- few or no reserves of weatherable minerals;
- rapid permeability, high susceptibility to leaching;
- low available water capacity.

The soils of this group are formed in coarse textured materials, exclusive of recent alluvial deposits of aeolian, colluvial or alluvial origin. In many cases they seem to have formed from a previous soil or by the leaching of clay formed in material with high contents of quartz such as granite. The greatest extent of these soils seem to be in the Tropical Wet-Dry and tropical Desert and Steppe climates. Their topography varies from flat to moderately sloping.

Annual cultivation of these poor soils leads to rapid depletion of the organic matter content. This causes a rapid decline of the cation retention. In humid climates leaching is severe, in semi-arid and arid zones drought hazard is serious. Due to the rapid deterioration the period of sustainable cultivation is short and required fallow periods are long.

12. Chromic Luvisols (Lc) and Cambisols (B) (USDA: Alfisols and Inceptisols, respectively)
* Characteristics of chromic Luvisols:
- good physical structure;
- moderate organic matter content;
- moderate to high cation retention and inherent fertility;
- moderate to high available water capacity.

Chromic Luvisols are soils also known under the name Terra Rossa. They are usually soils found in areas with mediterranean conditions. Their potential for agriculture is moderate to high. Generally their susceptibilty to erosion is high. Topography is gently sloping to rolling.

* For a description of Cambisols see Soil Grouping No.8.

13. Solonchaks (Z) and Solonetz (S) (USDA: Salorthids and Natr-Great Groups)

* Solonchaks are grouped because of their high salinity. They are soils with very little profile development. Due to their salinity they are not suitable for cultivation.
* Solonetz have a natric horizon in common, which is a subsoil with a clay illuviation from the topsoil, having a sodium (Na) saturation of over 15% at the cation exchange complex. The structure in the subsoil is columnar. These soils are considered virtually not suitable for cultivation.

14. Andosols (T) (USDA: Andepts)

Characteristics of Andosols:

- good physical structure;
- good drainage;
- high available water capacity;
- high cation retention;
- problems of phosphorous fixation.

Andosols are developed from volcanic ash. They are generally very good soils for cropping when they are well managed. In places Andosols may have fertility problems, usually due to phosphorous fixation. They frequently occur on steep slopes and this feature makes them highly susceptible to erosion. A further property is the phenomenon of thixothropy. Andosols occur in a wide variety of climates and under various vegetation types.

15. Vertisols (V) (USDA: Vertisols)

Characteristics of Vertisols:

- high nutrient availability;
- relatively high organic matter content;
- high cation retention;
- low water available capacity;
- slow permeability when wet, leading to low infiltration and high run off.
- soil is very hard when dry, causing problems of cultivation and seedbed preparation.

Vertisols in general have favourable chemical properties, but are problem soils in their physical qualities. Under high levels of management however, some of the physical problems can be overcome and high intensities of cultivation may become possible. Vertisols may be highly suitable for paddy rice cultivation, but lack of drainage may induce salinization.

16 Planosols (W) (USDA: Alfisols, Ultisols, Aridisols)

Planosols have a slowly permeable subsoils, which may cause problems of drainage. Hydromorphic properties are general, and their topsoils may be of poor physical and chemical properties. Planosols with a leached topsoil are moderately suitable for cultivation, when their topsoil is richer, these soils are good.

17/18 Xerosols (X) and Yermosols (Y) and shifting sands (USDA: Aridisols, Psamments)

In the new FAO legend (FAO, 1988) Xerosols and Yermosols have been combined in the Xerosols). Xerosols and Yermosols are soils of semi arid (Xerosols) to arid (Yermosols) climatic conditions. Xerosols and Yermosols may be very fertile soils. Due to the lack of rainfall these soils are of little or no value for agriculture. Xerosols usually occur in areas with a growing period of less than 75 days. With irrigation these soils may be classified among the best soils.

2.4 CONCLUSIONS

The best soil map with global coverage available at present is the FAO/Unesco (1971-1981) Soil Map of the World. This map is also available in a digital form. However, in a number of areas the information on this map is not accurate. New information has become available for many regions of the world. A global Soil and Terrain Digital Data Base at 1:1,000,000 scale is in development. The data base will include all information available on the soil, in separate attribute files, so as to enable scientists from other fields to use it in e.g. climate studies and trace gas studies. This work will be completed in 15-20 years. In the mean time an updated version of the FAO/Unesco Soil Map of the World would be very useful in studies on climate and trace gas fluxes. This map could be compiled as a digital data base with a ½ x ½ degree longitude-latitude resolution (which is comparable to the 1:5,000,000 scale of the FAO/Unesco soil map) and with attribute files for the most important information on the soil profile, such as soil depth, stoniness, texture, and some soil chemical parameters.

Table 2.2a Matrix of Major Soil Groupings according to FAO (1983) and the major land cover types according to Matthews (1983). Indicated are the areas in 1,000,000 ha. Description of the Major Soil Groupings are in section 2.3.

Land cover types	Major Soil Groups (FAO, 1983)																		
(Matthews, 1983)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	totals
1 Tropical evergreen rainforest, mangrove forest	576			301	121	42		79	51	11	30			1	5		6	1	1223
2 Tropical/Subtrop.evergreen seasonal broadleaved forest	164			42	13	22		54	19			10		1	5			1	331
3 Subtropical evergreen rainforest		4		1	3							1	1	2		6		1	19
4 Temperate/subpolar evergreen rainforest						14		11				3		11					39
5 Temperate evergreen seasonal broadleaved forest summer rain	42				7	8		18				5			2				81
6 Evergreen broadleaved sclerophyllous forest winter rain					1	6		9				22	1			6	2		47
7 Tropical/subtropical evergreen needleleaved forest	17					22		10											49
8 Temperate/subpolar evergeen needleleaved forest		20	232		131	144	16	57		276		13	3	14			1	5	909
9 Tropical/subtrop.drought deciduous forest	57			70	20	6		40	26		12	17		6	29	7	2	0	292
10 Cold-deciduous forest with evergreens	45	11	66		78	102	0	109		53		32	1	6	5	1	1		510
11 Cold-deciduous forest without evergreens	16	33	5		123	147		56		5		2	3	1	2				393
12 Xeromorphic forest/woodland	22	2		34	17	9	39	17	2		12	46	6	3	11	18	21	12	270
13 Evergreen broadleaved sclerophyllous woodland	18			2	3	38	1	20	1		9	18	5		19	3	15	17	170
14 Evergreen needleleaved woodland	8		8		24	51	16	68		50		6		9	2		4	2	247
15 Tropical/subtrop. drought-deciduous woodland	19	0		107	27	15	2	6	86		47	2	6		6		10	37	371
16 Cold-deciduous woodland			1		121	117				5									244
17 Evergreen broadleaved shrubland/thicket, dwarf-shrubland	3			1	3	21		2	5		2	10	6		0	7	9	60	130
18 Evergreen needleleaved/microphyllous shrubland/thicket					7	20	6					2		9	1		5	16	66
19 Drought-deciduous shrubland/thicket				1		8	3	12	6		9	2			1			41	84
20 (Dwarf) Shrubland, cold-deciduous subalpine/subpolar		1			26	3				6				9					45
21 Xeromorphic shrubland/dwarf shrubland		2		9	3	90	38	13	19		39	19	34	7	28	1	173	409	883
22 Arctic/alpine tundra, mossy bog		1	34		379	214		33		42			1	10					715
23 Grassland, 10-40% tree cover	39	40	1	212	23	78	23	9	64		73	15	19		10	15	6	17	642
24 Grassland, <10% tree cover	53	3	2	76	15	31	1	11	56	1	29	3	3		54	1	14	4	359
25 Grassland, shrub cover	8	14		16	16	115	128	38	52		205	25	54		34	1	101	122	929
26 Tall grassland, no woody cover	6	11		16	5	2	1		4		1		9		26				81
27 Medium grassland, no woody cover	1					1	16	10	9		18	1	2	1	3	1	4	11	79
28 Meadow, short grassland, no woody cover	2	41		5	7	240	84	30		2	11	19	17	42	6	2	28	67	604
29 Forb formations					16	10												1	27
30 Desert					47	393	1	3			159		44		2		49	845	1544
31 Ice																			1640
32 Cultivated area	149	196	18	95	128	133	156	273	116	26	56	127	14	31	86	18	54	51	1726
totals	1245	378	368	988	1363	2101	531	986	515	478	713	398	233	163	339	87	503	1721	14748

Table 2.2b Matrix of extents of Major Soil Groupings of the World according to FAO (1983) and Bio-Climates according to Holdridge (1967) using the data base compiled by Leemans (1989). Areas in 1,000,000 ha. Descriptions of the major Soil Groupings are presented in section 2.3. The Holdridge bioclimatic zones are described in Appendix V.

	Major Soil Groupings (FAO, 1983)																		
Bioclimates (Holdridge, 1967)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	totals
1 Polar desert					216	47				0		0		0				2	265
2 subpolar dry tundra					17	19		0											36
3 subpolar moist tundra			0		154	57		7		2				1				1	221
4 subpolar wet tundra			3		96	67		15		28				3				1	213
5 subpolar rain tundra						10				3		0		2					15
6 boreal desert			1		3	18		5									0	2	30
7 boreal dry scrub		3	6		65	99	31	10		1		1		1			13	8	237
8 boreal moist forest		98	251		231	341	29	62		112		2	9	8			1	5	1148
9 boreal wet forest		3	27		56	92	0	14		191		2		25		0		0	412
10 boreal rain forest					1	23		3		7		1		5					40
11 cool temperate desert			0			38	4	2				1	16	1			8	93	164
12 cool temperate desert scrub		0	3		1	109	36	5		1		3	18	5			47	119	345
13 cool temperate steppe		89	4		5	152	269	55		4		34	9	7	1	1	83	74	786
14 cool temperate moist forest	7	132	75		47	136	26	311		82		44	0	18	0	8	3	1	891
15 cool temperate wet forest	4	0			3	32	1	34	0	29		5		13				2	122
16 cool temperate rain forest					0	12		3		7		1		1					24
17 warm temperate forest						32		2			1	1	14	2			1	40	93
18 warm temeprate desert scrub					1	85	4	2			13	7	12	2	1		15	69	211
19 warm temperate thorn steppe					3	39	17	8			2	30	16	7	7	5	38	56	228
20 warm temperate dry forest	7	22		0	39	61	29	40	4	0	5	73	28	12	7	19	28	10	384
21 warm temperate moist forest	134	12		9	12	49		53	11	0	1	9	2	4	11		0	1	308
22 warm temperate wet forest	2			1	1	2		8	0					1					15
23 warm temperate rain forest														0					0
24 subtropical desert					1	148		0			7	2	25	5			24	437	648
25 subtropical desert scrub	0				1	80	11	4	0		90	20	46	1	15	1	75	198	543
26 subtropical thorn woodland				2	8	25	22	22	8		154	22	14	3	55	6	63	105	511
27 subtropical dry forest	26	9		135	48	38	44	60	122	0	84	66	5	6	73	8	48	41	812
28 subtropical moist forest	411	11		530	125	47	1	80	88	1	97	19	3	10	31	33	5	2	1493
29 subtropical wet forest	97			26	11	9		35	9					2	0				190
30 subtropical rain forest	1				2	3		1	1					1					8
31 tropical desert						140		2			13		1		1		15	330	502
32 tropical desert scrub	1			0	1	23		0			59	0	6		5		10	74	180
33 tropical thorn woodland	3			2	9	9		2	11		79	4	3	1	22		4	27	178
34 tropical very dry forest	9			8	13	27	4	13	106		59	24	9	1	67		3	14	356
35 tropical dry forest	159			130	42	30	3	53	141	4	37	26	2	4	42	9	7	1	689
36 tropical moist forest	310			129	75	6		34	13	6	11	1		3					588
37 tropical wet forest	22			1		1		8											32
38 tropical rain forest																			0
Totals	1193	379	370	975	1286	2106	531	953	515	478	712	398	237	155	339	90	491	1712	12920

3. PRESENT STATUS AND TRENDS OF THE GLOBAL SPATIAL LAND COVER DISTRIBUTION

3.1 INTRODUCTION

The type of land cover forms an important factor in the radiation balance of the earth and in numerous biogeochemical cycles. The land cover is subject to modifications by natural cycles and through human activities. Direct effects of human activities are the clearing of forests for agriculture, selective clearing during shifting cultivation in the tropics, afforestation, etc. One indirect effect of human activities is the forest degradation due to acid rain occurring at an ever larger scale particularly in industrialized parts of the world.

Historically the study of the spatial distribution of plant communities has been approached in terms of vegetation mapping. Small scale vegetation maps were compiled using field observations and other small and large scale thematic maps. The final product is generally a subjective composite of information reflecting the unrecorded choices of the compiler.

In chapter 2 an attempt has been made to compile a data base of the world soils and their cover types based on the FAO (1983) Resource Base of the world at scale 1:25,000,000, the Matthews (1983) 1x1° longitudexlatitude land cover data base and the 0.5x0.5° longitude x latitude Holdridge life zone data base (Leemans, 1989). The published literature deals with vegetation and land use only and do not consider the soils distribution. Therefore, in this chapter the subject of the spatial land cover distribution will be discussed separately.

3.2 REVIEW OF VEGETATION/LAND USE CLASSIFICATION SYSTEMS AND MAPS

3.2.1 Vegetation classification systems

The vegetation cover of the earth has basically two important aspects: variation in time and variation in space. One of the important tools for determining vegetation's change in space is mapping. Determining its variation in time is by repeated mapping. The latter approach has not been used as widely as it could have been, but it is gaining importance through the development of remote sensing techniques.

Vegetation is classified according to a number of criteria based on the vegetation itself, its surrounding environment or a combination. A large number of classification methods have been used. In most systems the potential vegetation or climax vegetation is distinguished. A number of classifications and maps with relevance to global and regional scales are listed in Table 3.1.

Potential vegetation is commonly mapped using bioclimatic parameters. This is based on the idea that typical communities develop in response to climate, particularly temperature and precipitation. Examples of such classification the system are those proposed by Holdridge (1967) and Schroeder (1983). Seasonal effects which may influence ecosystem structure are not taken into account, and landscape and edaphic factors are also omitted. Classifications based on bioclimatic factors can be used to predict primary production. Mueller-Dombois (1984) gives a good overview of the value of the Holdridge classification system. In Appendix V the classes of the Holdridge system are listed. This bioclimatic classification has been used in combination with soil groupings and land cover in Appendices IV for defining ecosystems. For this purpose the digital data base of the Holdridge bioclimatic zones compiled by Leemans (1989) has been used.

Existing vegetation is mapped usually on the basis of physiognomy or floristics. Physiognomy provides a better estimate of phytomass since it includes variations due to succession and habitat (Mueller-Dombois, 1984). The most widely known physiognomic classification is the system of Unesco (1973) which was designed for mapping of vegetation on a 1:1,000,000 scale. Unesco (1973) gives guidelines for representing cultivated areas graphically. Dent (1978) complemented this with a classification (coding) system for cultivated areas, for soil survey and land evaluation purposes. The Faculty of Geography of Moscow State University (Rjabehakov, ?) produced a map of actual land use of the world at a 1:15,000,000 scale. Apart from the Unesco (1973) method, the Institut de la Carte Internationale de la Vegetation (1988) uses the physiognomic or 'landscape' classification method proposed by Gaussen (1954) for mapping at a 1:1,000,000 scale. A number of small scale maps of vegetation which give a good global overview are listed in Table 3.2. Apart from these global maps a number of large scale vegetation maps were prepared, such as Hueck and Seibert (1972) of South America at a 1:8,000,000 scale; Institut de la Carte Internationale du Tapis Vegetal (1982, South America 1:5,000,000, Africa 1:5,000,000, based on Unesco); White (1981; Africa 1:5,000,000); Küchler (1965, North America 1:7,500,000); the Mediterranean Region (Unesco, 1:5,000,000). Apart from the above continental maps the maps produced by the Institut de la Carte Internationale de la Vegetation (1988) are useful. They are all based on the system proposed by Gaussen (1954) on a 1:1,000,000 scale for India, Madagascar, Sri Lanka, Cambodia, Mexico. The same institute also produced a number of bioclimatic maps.

Table 3.1 A number of vegetation classification schemes for small scale mapping.

Main criteria used	units distinguished	classification
1. Properties of the vegetation		
- physiognomic properties	dominant life form	Unesco (1973)
- floristic properties	dominant species/associations of species	
2. Properties outside the vegetation		
- environment	climate mainly; topography, soil, landform, combinations	Köppen (1936) Thorntwaite (1948) Gaussen (1954) Holdridge (1967) Walter (1973) Schroeder (1983)
- geographical location		Hueck and Seibert (1972)
- successional stage		Clements (1916, 1928)
3. Combination of vegetation and environmental properties		
- overlay of vegetation and environmental properties		Küchler (1965)
- ecosystem classification		Ellenberg (1973)

Ecosystems are self contained units in terms of primary production and nutrient cycling (Burton, 1987). Most descriptions of the composition of ecosystems are based on the dominant species, but deciding which characterisitics should be used to determine the ecosystems boundaries presents problems. To some extent the problem is one of scale as shown in Table 3.2 and discussed in the review by Mueller-Dombois (1984).

The physiognomic vegetation types shown on maps as listed in Table 3.2 are also called <u>biomes</u>. A biome, as defined by Whittaker (1975) is a group of terrestrial ecosystems on the same continent, which are similar in physiognomy, in the main features of their environment and in some characterisitics of their animal communities. Biomes in similar environments common to more than one continent are <u>biome types</u> or, in the terminology of Walter (1985), <u>zonobiomes</u>.

It remains virtually impossible to draw a sharp distinction between any of these ecosystem types or their subdivisions. Each division is arbitrary. Most boundaries are gradual. Only tundra and boreal forest have some continuity throughout the northern hemisphere. Other ecosystems such as tropical rainforest and temperate grassland are isolated in different biogeographical regions and therefore may be expected to have ecologically equivalent, but often taxonomically unrelated species (Ajtay et al., 1979). The terms 'forest', 'woodland', 'grassland' and 'savanna' are difficult to specify, as each author has his own concept of what they consitute.

Table 3.2 List of global vegetation maps and digital data sets

reference	map scale	classification used	Number of types
maps			
Schmithüsen (1968)	1: 25,000,000	physiognomic-environmental	144
Brockmann-Jerosch (1918)	1:100,000,000	physiognomic-environmental	10
Whittaker (1970)	1: 50,000,000	physiognomic-environmental	25
Odum (1959)	?	physiognomic-environmental	12
Duvigneaud (1972)	?	physiognomic-environmental	17
Whittaker and Likens (1975)	?	ecosystems	13
Schroeder (1983)	?	bioclimatic	42
Rjabehakov (?)	1:15,000,000	Actual land use	
digital data sets			
Olson et al., 1983)	0.5x0.5° grid (1:30,000,000)	physiognomic-environmental	43
Matthews (1983)	1°x1° grid	physiognomic (Unesco, 1973)	32
Emmanuel (1985)	0.5°x0.5° grid	bioclimatic(Holdridge, 1967)	37
Leemans (1989)	0.5°x0.5° grid	bioclimatic(Holdridge, 1967)	
Henderson-Sellers (1985)	1°x1° grid	land cover	?

The Unesco (1973) classification has been used in its modified form by Whittaker and Likens (1975), Ajtay et al. (1979), Matthews (1983), Goudriaan and Ketner (1984), and many other authors. Furthermore it has been applied in the map unit descriptions of the FAO/Unesco soil map of the world and this makes linking to soil geography an easy task. Comparisons between the various studies are therefore possible without much problems of data interpretation and conversion.

3.2.2 Global land cover data sets

A feature of all the above vegetation classifications is that they represent rather theoretical potential vegetation or climax vegetation. A general aspect omitted is the perturbation of the earth's vegetative cover by man. Clearing of vegetation for permanent cropping or grazing, shifting cultivation or clearing for other than agricultural purposes (or without any purpose) have changed the theoretical vegetation. A number of authors have attempted to compile vegetation and land use data sets for the world. A selection of these studies will be discussed here.

Hummel and Reck (1979) produced digital data files with resolution ranging from 0.4° to 0.9° latitude and longitude for albedo studies. The results of Olson et al. (1983) were based on this study. Their map has a 1:30,000,000 scale and a 0.5°x0.5° resolution on digital format. An older study is that of Whittaker and Likens (1975) and Ajtay et al. (1979). The most recent and possibly best documented study is the analysis by Matthews (1983).

In Table 3.3 a number of global land cover distributions are presented. The data by Matthews (1983) concerning the pre-agricultural and present land cover distribution are shown in Table 3.4. Matthews (1983) used the Unesco (1973) vegetation classification scheme, Olson et al. (1983) used a land systems grouping while in the other studies the scheme proposed by Whittaker and Likens (1975) was used. The latter system is also based on the Unesco (1973) classification. Comparison is possible but difficult. A number of land cover types show a great similarity between studies. Houghton (1983) and World Resources Institute (1987) (Table 18.3, on page 272, prepared by Houghton) calculated land use changes between 1700 and 1980 using the FAO Production Yearbooks (FAO, 1949-1985) complemented with a model for the increase in agricultural land based on population growth.

There is great controversy about the areas of the land area occupied by each ecosystem. Estimates of the coverages in the tropics may differ by a factor of 3.7 (highest and lowest literature estimate as presented in Brown and Lugo, 1982, Table 8 page 173). In many cases the map does not include documentation of the sources used in compilation or of the relative weight given to various sources which may conflict. Where source information is provided, it is not uncommon to find, that several maps have borrowed heavily from a single source (Matthews, 1983).

The forest area calculated by Whittaker and Likens (1975) is about 25% higher than for all other studies. However, tropical rainforest areas show a considerable spread. Rainforest areas are highest in Whittaker and Likens (1975) and lowest in the analysis made for this paper. It is clear that the various authors have used different concepts for woodland and deserts. The area of cultivated land shows a good similarity between all the studies. Matthews based her land use data on the cultivation intensities of land cover areas. Her figure for cultivated areas corresponds well with the FAO (1985) reports from local and national sources.

Table 3.3. Area coverage for the major land cover types according to: A.Whittaker and Likens (1975); B. Ajtay et al. (1979); C. Olson et al. (1983); D. Matthews (1983); E. Houghton et al. (1983).

Land cover type	area (10^{12} m^2)				
	A	B	C	D	E
1. tropical rainforest	17.0	10.3	{	12.3	11.6
2. trop. seasonal forest	7.5	4.5	{12.1	6.7	7.7
3. temperate forest	12.0	7.0	12.0	{20.3	10.0
4. boreal forest	12.0	9.5	11.5	{	11.4
5. woodland, shrubland and interrupted woods	8.5	4.5	4.7	25.2	14.8
6. savanna(+trop.grassl.)	15.0	22.5	24.6[3]	{	12.5
7. temperate grassland	9.0	12.5	6.6	{27.4[4]	17.7
8. tundra, alpine	8.0	9.5	13.6	7.3	7.0
9. desert, semidesert	18.0	21.0	8.5	{	{21.3
10.extreme desert	24.0	24.5	16.4	{15.6	{
11.cultivated land	14.0	16.0	15.9	17.6	15.0
12.swamps, marshes and coastal land	2.0	2.0	2.5	____	____
13.bogs and peatland	____	1.5	____	____	____
14.other	2.0	4.0	____	2.4	5.2[5]
total	149.0[2]	149.3[2]	129.7	134.8	133.6

[1] data presented here are calculated from the data presented by Houghton et al. (1983);
[2] includes areas of ice, lakes and rivers;
[3] includes "warm or hot shrub and grassland" and "tropical woods" (remnants with fields and grazing).[4] denoted as "grassland";[5] pasture.

The above simple comparison highlights the need for critical evaluation of vegetation definitions and classifications. Apart from the areal extent of land cover types, the description in terms of forest volumes and densities are a subject of deep controversy (Brown and Lugo, 1984).

Table 3.4 illustrates the changes in the land cover distribution since pre-agricultural times. The figures presented are not consistent with those of other authors. For example, Postel (1984) reported that up till now 13.3% of all the tropical forests have been logged, while Matthews reported a decline of 3.75% for tropical rainforests since pre-agricultural times. According to the data presented by Houghton (1983) the decline of the extent of tropical rainforest and tropical seasonal forests since 1700 is 11.4%.

Table 3.4 Comparison of areal estimates (in 10^{12} m^2) of pre-agricultural and present land cover distribution.

cover type	pre-agricultural	present	%reduction
total forest	46.28	39.27	15.2
tropical rainforest	12.77	12.29	3.8
other forest	33.51	26.98	19.5
Woodland	15.23	13.10	13.8
Shrubland	12.99	12.12	6.7
Grassland	33.90	27.43	19.1
Tundra	7.34	7.34	0.0
Desert	15.82	15.57	1.6
Cultivation	0.93*	17.56	____

*limited areas with long use histories and for which reliable vegetation data could not be acquired have been designated as cultivated land. From: Matthews (1983)

3.2.3 Tropical forest resources and distribution

Table 3.5. The extent of tropical forests by various authors.

reference	extent (10^{10} m^2)			
	Asia	Latin America	Africa	total
Persson (1974)				
closed forest	290.4	576.8	195.9	1067.8
total forest	406.0	734.1	754.9	1895.0
Myers (1980)				
moist forest	271.4	641.6	151.4	1064.4
Postel (1984)				
moist forest	305.0	629.0	217.0	1201.0
total forest	445.0	1212.0	1312.0	2969.0
Lanly (1982)*				
closed forest	305.5	678.7	216.6	1200.8
open forest	30.9	217.0	486.4	734.4
total forest	336.5	895.7	703.1	1935.2
World Resources Institute (1988)*				
closed forest	269.4	679.4	217.6	1166.4
open forest	26.8	207.2	481.1	715.1
total	296.2	886.6	698.7	1888.5

* data for the same 16 countries (Asia), 23 countries (Latin America) and 37 countries (Africa); see Lanly (1982) for the list used of temperate and tropical countries.

The figures reported by Postel (1984) and Myers (1980) for moist forest correspond well with the figures presented by Matthews (1983) for tropical rainforests (see Table 3.3 and 3.4). Myers defines tropical moist forest as "evergreen or partly evergreen forests in areas receiving not less than 100 mm of precipitation in one month for 2 out of 3 years with mean annual temperatures of greater than 24°C and essentially frostfree" while Matthews (1983) used the Unesco (1973) defined tropical evergreen rainforest and mangrove forests (codes 1A1 and 1A5 resp.).

There are few reliable estimates of the actual area covered by tropical rainforest. Moreover, the available estimates vary widely because of the differences in criteria used to define this ecosystem. The statistics used are often unreliable. The great mass of existing information has not been "organized" at a national level in many tropical countries. Even when national syntheses exist it is not possible to regroup them together because the classifications and concepts used differ from one country to another (Lanly, 1982). A selection of literature data on the areal extent of tropical forests is presented in Table 3.5.

3.2.4 Temperate forest resources and distribution

Table 3.6 The extent of temperate forests (10^{10} m^2)

	Area						
	temp Africa	USA/ Canada	temp. Latin America	temp. Asia	USSR	Europe	temp. Oceania
closed forest	3.7	459.4	121.9	161.7	791.6	137.0	52.1
open forest	2.8	275.1	nd	38.3	137.0	21.9	67.4
total	6.5	734.5	nd	200.0	928.6	158.9	119.5

Source: World Resources Instite (1988); nd = no data available.

Data for the temperate forest resources were taken from World Resources Institute (1988). The distribution is presented in Table 3.6. The extent of temperate forests is greater than that of tropical forests. Great part of temperate forests are formed by the boreal forest in the USSR and North America.

3.3 CAUSES OF DEFORESTATION

Most of the present agricultural and urban land was once under forest cover. Traditionally there has been a need for development of land resources for economic benefit. Deforestation has been a necessary evil. An important issue to be considered is whether additional deforestation is necessary.

A list of principal causes of deforestation is given in Table 3.7. A cause of forest degradation (not conversion) which is playing a role of growing importance is the forest degradation due to acid precipitation. The observed consequences of this phenomenon will be discussed briefly in section 3.4.3.

Historically forest dwellers have sought their livelihood and basic necessities of life by harvesting forest products. In addition to this harvesting of minor products in the humid tropical forest zones the system of shifting cultivation is widespread. This system involves incomplete clearing and cultivation in such a way that the forest is allowed to regenerate during a fallow period. If the shifting cultivators maintain a fallow period long enough, the forest is allowed to regenerate. This is often not the case. Jackson (1983) estimated that there are 200 million people who rely solely on forest lands for their living. Some reports indicate that perhaps 50% of all present tropical deforestation is caused by permanent clearing of land that was previously used for shifting cultivation (e.g. Myers, 1980; Houghton, 1987). Deforestation in developing tropical countries is also occurring for providing fuelwood to meet household energy needs.

Table 3.7. Causes of deforestation and forest degradation.

"natural"	fires acid precipitation
tradition	shifting cultivation fuel wood harvesting forest products
economic	agriculture ranching plantation crops timber and commercial wood infrastructure and urbanization
socio-political	strategic population migration speculation

National resources planning often involves deforestation for alternative land utilization such as agriculture, ranching, plantation crops, urbanization. Postel (1984) reported that about 4.4 million hectares of land per year of tropical moist forest are being logged.
Population migration from densely populated areas to forest land is a major socio-political reason for the present large scale deforestation in some regions, e.g. Sumatra and Amazonia. Often the forest is cleared to establish propietary claims (Lal, 1986). The result is a rapid depletion of the existing forest reserves in favour of new land development and colonization schemes.

Fearnside (1987) lists causes of deforestation in the Amazon. He distinguishes underlying and proximal causes. A number of proximal causes of deforestation in the Amazon basin mentioned by Fearnside are: tax incentives, tax penalties, interest incentives, subsidies and loans.

3.4 LAND COVER CHANGES

3.4.1 Deforestation in tropical regions

Present deforestation rates for the tropics are difficult to estimate. Lack of infrastructure, accessibility and scarcity of trained manpower in many countries in the tropics contribute to the scarcity of reliable surveys of the existing forest resources and conversion rates. Inadequate communication is also a major hindrance in updating the field records particularly in those countries where fast deforestation occurs. Therefore, the presently available estimates for forest conversion rates in the tropics are unreliable and obsolete.

A thorough study of tropical deforestation was carried out by FAO (1981a,1981b and 1981c) for tropical Africa, tropical Asia and tropical America. The results of these analyses are summarized in Lanly (1982). The data used are local and country land use statistics for 63 countries complemented with 1972-1978 LANDSAT imagery for 13 additional countries. The deforestation rates as presented by Lanly are summarized in Table 3.8. The deforestation rates estimated by Lanly are 0.623% for tropical America, 0.615% for tropical Africa and 0.596% for tropical Asia. The total rate for the tropics is 0.58% or 11.3×10^{10} m^2 y^{-1}. Lanly (1982) has not included partial or selective felling of trees or managed productive forests. Other authors have accounted for fallow lands used for shifting cultivation. A selection of estimated deforestation rates is given in Table 3.9.

Houghton et al. (1983) used three data sets in their calculations of the annual carbon release from the biota. In one they used a conversion rate of $10.5 \times 10^{10} m^2 y^{-1}$ of forest to agriculture and

$1.4 \times 10^{10} m^2 y^{-1}$ of forest to grazing land (based on the population growth) and an afforestation rate of $3.9 \times 10^{10} m^2 y^{-1}$. In the second data set they use a deforestation rate of $11.8 \times 10^{10} m^2 y^{-1}$ (based on Myers, 1980) and the third deforestation rate is $3.26 \times 10^{10} m^2 y^{-1}$ (based on FAO,1949-1975).

Table 3.8. Average annual deforestation and reafforestation (10^7 m^2) in tropical zones in the periods 1976-1980 and 1981-1985. n.a.=no data available.

	area (10^7 $m^2 y^{-1}$)						
	trop.America		Trop.Africa		Trop.Asia		Total
	76-80	81-85	76-80	81-85	76-80	81-85	81-85
closed broadleaved forests	3807	4006	1319	1318	1767	1782	7106
closed coniferous forests	312	309	8	7	35	30	346
closed bamboo forests	n.a.	n.a.	6	6	13	14	20
total closed forests	4119	4339	1333	1331	1815	1826	7496
open forests	____	1272	____	2345	____	190	3805
total open + closed forests	____	5611	____	3676	____	2016	11303
reafforestation**	---	63	---	85	---	416	563

*refer to Lanly (1982) for a list of countries.
**from World Resource Institute (1988) for the 1980's.

Table 3.9. Rates of tropical deforestation reported by selected authors.

reference	deforestation rate($10^{10} m^2 y^{-1}$)
Seiler and Crutzen (1980)	9 - 15[1]
Myers (1980)	24.5[2]
Dickinson (1982)	6
Lanly (1982)	11.3
Houghton (1983)	3.3-11.8[2]
Houghton (1987)	22.3-28.0[2]

[1] incl. clearing for cattle grazing;
[2] incl. clearing as a consequence of shifting cultivation.

Houghton et al. (1987) used two data sets with data for shifting cultivation and arrived at a conversion of forest to agriculture of $5.8 \times 10^{10} m^2 y^{-1}$ and a deforestation rate of $16.5 \times 10^{10} m^2 y^{-1}$ as a consequence of shifting cultivation. Their second data set had rates of 2.9×10^{10} and $25.1 \times 10^{10} m^2 y^{-1}$ for conversion to agriculture and permanent deforestation due to shifting cultivation respectively.

The data proposed by Seiler and Crutzen (1980) and Lanly (1982) show a good similarity. Both use strictly deforestation rates without accounting for permanent clearing of fallow forest lands. As was shown in Table 3.3 there is no agreement about the definition and classification of forests. The conversion rates will thus disagree as well, even when focussing on a specific area. Henderson-Sellers (1987) concluded that the question of how much tropical forests have been cleared is very difficult to resolve. The interpretation of LANDSAT and other satellite is unlikely to provide a complete answer.

3.4.2 Deforestation in temperate regions

The average deforestation rates in temperate zones as presented in the World Resources Institute (1988) are in Table 3.10. The World Resources Institute has consulted a great number of sources to compile their forest resources tables.

Apparently, data for deforestation in temperate regions are not available. However, the reafforestation rate for all temperate countries for which data are available is $13.5 \times 10^{10} m^2 y^{-1}$. This figure represents both afforestation of previously not forested land and re-afforestation of land which was under forest cover within the previous 50 years. Therefore the above estimate of afforestation cannot be compared with deforestation data.

Table 3.10. Average annual deforestation and reafforestation (10^7 m^2) in temperate zones.[1] (n.d. = no data available).

	temp. Africa	temp. North America	temp. South America	temp. Asia	USSR	Europe	temp. Oceania
deforestation	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
reafforestation[2]	111[3]	2495	40	5233[4]	4540	1031[5]	112

[1] countries for each zone are listed in FAO (1949-1985) Production Yearbooks; [2] Reafforestation includes reafforestation of previously unforested lands and reafforestation of land which was under a forest cover within the previous 50 years; [3] only data available for Algeria, Egypt, Libya, Morocco and Tunisia; [4] only data available for China, Iraq, Israel, Japan, Jordan, Korea Dem. People's Rep., Korea Rep., Saudi Arabia, Syrian Arab. Rep., Turkey and Un. Arab. Emirates; [5] No data available for Albania, Denmark, German Dem. Rep., Greece, Iceland, Luxembourg, Maltha and Romania.

3.4.3 Forest decline due to air pollution

Almost all the work on so called acid precipitation related to atmospheric pollutants has so far been carried out in Europe and North America. The results of a SCOPE conference on acidification of tropical ecosystems (Rodhe and Herrera, 1988) point out that with the exception of China, little evidence exists of serious regional acidification effects of anthropogenic emissions of sulphur and nitrogen.

There are four main hypotheses concerning the cause of the forest decline due to acid precipitation:

- acid rain, causing increased leaching of cations from soils and consequently soil acidification, increased aluminium concentration which may at high levels become toxic to plant roots;
- ozone, direct leaf injury due to high ozone concentrations in the troposphere;
- nitrogen excess; an excess of nitrogen may cause a disturbed nutrient balance in the soil. The best known symptom is potassium and magnesium deficiency, decreasing photosynthesis capacity, weakening rooting systems, greater frost susceptibility, greater susceptibility to pests and diseases;
- stress hypothesis; during long periods forests have been influenced by air pollutants; the combined effect of all pollutants is greater than the sum of the individual effects.

Most data published on forest decline and effects of air pollution on forest growth have mainly been descriptive or hypothesizing with a significant lack of experimental support (Andersen and Moseholm, 1987).

Novel forest decline in Europe was first observed in silver fir (Abies albas) in rural areas in the southern part of the Federal Republic of Germany (FRG) in the beginning of the 1970's.

Subsequently the decline was described in the early 1980's for Norway spruce (Picea abies) and recently for deciduous trees, i.e. European beech (Fagus sylvatica).

In the Federal Republic of Germany (FRG) the forest inventory for 1986 showed no further decline between 1985 and 1986. In 1986 19% of the forested area was moderately to severely damaged. In Austria and Norway about 5% of the forested area is reported to show moderate to severe loss of vitality. In Denmark 3% of the forested area shows a decreased vitality. In The Netherlands an investigation showed that 21% of all forests has a severe loss of vitality (Staatsbosbeheer, 1986). For the United States, Eastern Europe and the USSR similar data on forest decline are known although no exact decline figures can be given.

Damage to cultivated areas is less pronounced. Cultivated soils usually have a higher buffering capacity. The relative contribution of atmospheric deposition to the total soil acidification in cultivated soils is 5 to 35 % in calcareous soils, while its contribution is 7 to 55 % in noncalcareous mineral soils. The atmospheric contribution in forest soils on calcareous soils is 5 to 45 % and 56 to 99 % in noncalcareous soils (Staatsbosbeheer, 1987).

There is recent evidence of a direct effect of soil pollution on plant leaves. At the present atmospheric concentrations of O_3, SO_2 and HF a 5% yield reduction of agricultural crops can be expected (Staatsbosbeheer, 1987).

3.5 CONCLUSIONS

Tropical deforestation is the present major global land cover change. The data pertaining to land use change are difficult to obtain and exceedingly difficult to verify, particularly data from developing countries. Estimates of forest destruction range between 10 and $20 \times 10^{10}\ m^2\ y^{-1}$, much of it in the Amazonian region. Most of the data base is unreliable. Great controversy exists concerning the nature of changes (permanent clearing versus partial destruction or shifting cultivation; the latter process could account for an even greater extent of forest loss than the permanent clearing), while definitional differences add to the difficulty of comparing estimates.

The quantification of the geographic distribution of vegetation and land use involves the use of some sort of classification system. A great number of classifications have been proposed, each having its own specific criteria for separating vegetation types. Of the most important classifications, i.e. the physiognomic-environmental and the bioclimatic systems, the first is most suited for prediction of phytomass and is most widely used for description of units covering a large area of land. The latter is most suited for potential production figures. If maps on a global or continental scale are compared, one aspect becomes clear immediately: it is very hard to compare maps produced on the basis of different criteria. With regard to the vegetation mapping can be said that the Unesco (1973) system is one of the most widely accepted system. The global vegetation data set made by Matthews (1983) is probably the best documented and has the advantage that it is based on Unesco. Concerning the bioclimatic systems there are a number of good classifications, i.e. Holdridge, Gaussen and Schroeder. The Holdridge system has a number of disadvantages (Mueller-Dombois, 1984) but is as practical as the other classifications. The maps produced by the Instut Internationale de la Carte Vegetal (on the basis of Gaussen) are very useful, but their global coverage is very incomplete.

For classifying actual vegetation and land use a system such as the one proposed by Dent (1978) will prove very useful since it has been widely used in soil and land use surveys. Comparing the studies on land use changes the same problem of classification becomes clear. Apart from that the lack of data, especially in the third world where most changes occur, is enormous. Intensive satellite monitoring, e.g. with the NOAA-AVHRR sensor, will improve the capacity of measuring changes regionally and globally. However, the digital sensor data will have to be linked to one of the vegetation/land use classification systems in the near future.

4. EXCHANGE OF GREENHOUSE GASES BETWEEN SOILS AND THE ATMOSPHERE

4.1 CARBON DIOXIDE

4.1.1 Introduction

CO_2 is the most abundant so called greenhouse gas. Its current atmospheric concentration is approximately 345 ppm and its annual increase is 0.5%. The total size of the global biomass carbon pool is estimated at: 835 Gt C (Whittaker and Likens, 1975), 560 Gt C (Ajtay et al., 1979), 594 Gt C (Goudriaan and ketner, 1984). The other carbon pools are: atmosphere (700 Gt C), oceans (38000 Gt C), fossil reserves (6000 Gt C) (Goudriaan and Ketner, 1984) and caliche (petrocalcic horizons in arid and semi-arid regions; 780-930 Gt C; Schlesinger, 1982, 1985). Net primary production by terrestrial biota is 60 Gt C y^{-1}, litterfall 40-50 Gt y^{-1}, the total stock of litter 60 Gt (Ajtay et al., 1979). Annual decomposition releases about 40-50 Gt C y^{-1} to the atmosphere.

The estimated total CO_2 emission from fossil fuel combustion was 5.3 Gt C y^{-1} in 1984 (Rotty, 1987) while the annual release caused by deforestation is estimated to be 0.3-1.7 Gt C y^{-1} (Detwiler and Hall, 1988) to 1 to 2.6 Gt C y^{-1} (Houghton et al., 1985). Projections for the year 2050 for the emission from fossil fuel combustion range from 2 to 20×10^{15} yr^{-1} (Keepin et al., 1986). According to the different scenarios the atmospheric CO_2 concentration will rise to 440 to 660 ppm in the year 2050. In equilibrium the sinks for atmospheric carbon dioxide are the atmosphere (71%), oceans (18%) and the terrestrial biota (11%) (Goudriaan, 1988; pers. comm.) Due to lack of time for redistribution at present oceans absorb only 40% of the annual carbon injection into the atmosphere. The resulting increase of the atmospheric concentration is 0.5% or 3.5 Gt C y^{-1}.

Table 4.1. Estimates of the pool of organic carbon in world soils (all figures in 10^{15} g C).

approach	reference	carbon pool
vegetation types	Bolin (1977)	700
	Schlesinger (1977)	1456
	Bolin et al. (1979)	1672
	Ajtay et al. (1979)	1635
Soil types	Bohn (1976)	3000
	Ajtay et al. (1979)	2070
	Post et al. (1982)	1395
	Schlesinger (1984)	1515
	Buringh (1984)	1477
	This study	1700
Modelling	Meentemeyer et al. (1981)	1457
	Goudriaan and Ketner (1984)	1400
	Goudriaan (1988)	2000

On a world wide basis the amount of carbon in decaying plant litter and soil organic matter may exceed the amount of carbon in living vegetation by a factor 2 to 3. There have been several attempts to estimate the storage of soil organic matter in world ecosystems with basically four different approaches: the vegetation groups, the soils, the life zone groups and through modeling. Some of these estimates are listed in table 4.1. Bohn (1976) presented a figure for the carbon pool which does not fit within the range of estimates by other authors. His estimate is partly based on the world soil map by Gaussen and Hädrich (1965) since the FAO soil map of the world was not yet completed. The estimation made by Post et al. (1982) is based on the world life zone groups according to Olson (1983) and analyses of about 2700 soil profiles representing virtually every major ecosystem. Samples from each soil horizon or at standard depths were analyzed for carbon content (not incl. litter) and bulk density. The global estimates of soil carbon are much more accurate than estimates of the area assigned to the different ecosystems as was discussed in chapter 3.

The emission of CO_2 from the world soils will be evaluated in 4.1.2 - 4.1.8. In 4.1.9 estimates of the global emission of CO_2 from soils will be evaluated and in 4.1.10 different global C-models will be discussed to illustrate the total CO_2 emission from vegetation and soils.

4.1.2 The soil organic matter fractions

Organic matter in soils is represented by plant debris or litter in various stages of decomposition through to humus and also includes microbial biomass. Soil organic matter exists in many forms. Kononova (1975) separates the soil organic matter into fresh and incompletely decomposed plant residues and a stable component (humus). Humus in turn can be subdivided into strictly humus substances (humic and fulvic acids) and other products of advanced decomposition of organic residues and products synthesized by microbes. Since a large portion of the soil carbon is found above a depth of 50 cm, Schlesinger (1984) suggests, that nearly half of the carbon in a typical profile is relatively labile. The remainder is found in the lower sections of the soil profile and is more stable. Spycher et al. (1983) found that the light fraction (density <1.6 g cm^{-3}) of the soil organic material in a temperate forest was 34% and they suggest that the light fraction is equivalent to the labile components of the organic matter. Kortleven (1963) also divided soil carbon into humus and a fraction of relatively stable carbon, that contains recalcitrant humus, charcoal and other forms of elementary carbon. Janssen (1984) divides soil organic matter into 'young' and 'old' soil organic matter. 'Young' soil organic matter represents material younger than 1 year. The lifespans of humus and charcoal according to Goudriaan and Ketner (1984) are 10 to 50 years and 500 years respectively. Paul and Van Veen (1978) use the terms physically protected and non-protected soil organic matter, the former having the lowest decomposition rate.

The stability and related age of soil organic matter is closely related to the pedogenic factors under which the soil is formed. Martel and Paul (1974) found that the mean residence time of organic carbon in a chernozemic soil to be 350 years. The mean residence time of the $ZnBr_2$ residue (= organic material excluding the light material) was 500 years. The oldest fraction found was the NaOH extract with 1900 years. Jenkinson and Rayner (1977) used a soil sampled in 1881 with an equivalent age of 1450 years between 0 and 23 cm, 2000 years between 46 and 46 cm and 3700 years between 46 and 67 cm. Scharpenzeel and Schiffmann (1977) found a linear relationship of mean residence time with depth, the average increase per cm being 46 years for Chernozems. Soils formed under different conditions showed different patterns of residence time with depth.

4.1.3 Models of litter decomposition and soil organic matter accumulation

A variety of different models have been developed for calculating plant residue decomposition rates and soil organic matter levels:

* models with decompostion rates which are constant in time. Basic formulae are:

$$dC/dt = h^*A - k^*C \quad (4.1)$$
$$C_E = h \times A/k \times (1 - e^{-kt}) \quad (4.2)$$

where: C = quantity of organic carbon;
C_E= steady state level of organic C;
k = decomposition contstant;
h = humification constant;
A = addition of organic material.

The decomposition of soil organic matter often shows an initial period of rapid loss of labile constituents such as soluble carbohydrates, followed by a longer period of years of ever decreasing decomposition rates during which more recalcitrant components such as lignin are oxidized. Examples of models based on this first order decomposition process are Henin and Dupuis (1945), Jenny et al. (1949), Greenland and Nye (1959), Kortleven (1963) and Olson (1963).

* Models with decomposition constants which are variable in time. An example of this approach is Janssen (1984), who proposed:

$$k = 2.82 \times (a + t)^{-1.6} \quad (4.3)$$

where: a = apparent age of the added material.

* Models with different fractions of soil organic matter, each having a different decomposition rate. As was reported by Minderman (1968) it is necessary to know the decomposition rate of the separate chemical constituents of litter and soil organic matter to calculate the decomposition of the total mass of soil organic matter. Models of this type were developed after the introduction of computer simulation models, which enable repetitive execution of complex calculations. Representative models of this type are Jenkinson and Rayner (1977), Van Veen and Paul (1981) and Parton et al. (1987).

4.1.4 Factors affecting the stability of organic matter in soils

- <u>mineralogy</u>

Base status and clay content are often related. Soils derived from basic rocks are usually more fertile than their acidic counterparts, which gives rise to higher annual inputs of organic matter to the soil. A basic environment accellerates the decomposition of litter (short term effect), but the mixing with soil components, which retards organic matter turnover, increases the retention and leads to higher organic matter contents (Oades, 1988). In acidic soils, the initial decomposition of debris is retarded, but subsequent oxidation of organic matter proceeds relatively quickly because of lack of stabilizing mechanisms. Jones (1972) indicated, that in savanna soils parent material has an effect on soil carbon contents, but this effect may be attributed to the clay content. The clay content is related to the parent material.

There is little evidence, that organic matter intercalates the plates of montmorillonite or vermiculite, which would offer protection against attack by enzymes. In such clays organic matter may penetrate the intertactoid pores, which are generally less than 1 um in diameter and would not be accessible to microorganisms (Oades, 1988).

In tropical soils the carbon in the lower profile may be fixed to allophane or other amorphic minerals. The organic matter in tropical soils is often red or colourless. In other soils crystalline clay minerals are important in complexing soil organic matter.

- soil texture

Clay soils generally show lower decomposition rates. This can be explained by the adsorption of organic matter on clay surfaces in combination with the spatial arrangements of substrates and organisms with the soil pore system. In heavy textured soils the movement of meso and micro life may be restricted. In heavy soils the protection from predation by larger organisms (amoeba, eelworms) because of restricted movement of these organisms (Van Veen et al., 1985). The movement of enzymes may also be restricted. Many authors (e.g. Kortleven, 1963; Jenkinson, 1977; Schimel et al., 1985a, 1985b) reported that finer, clayey soils tend to be more preservative than the coarser, sandy soils. Jenkinson (1977) indicated that this texture effect however would be small under both near neutral and acidic conditions. Jones (1972) found a good positive correlation between the clay content and soil carbon for well drained soils. For poorly drained soils the correlation was less. For Vertisols (USDA, 1975; FAO/Unesco, 1972) Jones indicates a linear relationship between soil carbon and clay content between 35 and 80% clay.

Clays and organic molecules are both negatively charged. Cations may build bridges between clay particles and organic molecules. A substantial proportion of this protected organic matter is found in stable aggregates. Not only clay bridging, but also interactions between charged and uncharged polymers with clay surfaces are involved in determining the stability of microaggregates.

- soil structure

Organic matter is undoubtedly stabilized by physical processes. The protective action of clay is an illustration (Jenkinson and Rayner, 1977; Jenkinson, 1977; Van Veen et al., 1987). Probably soil organic matter is a continuum, with physically protected but decomposable materials at one end and free organic matter, that is highly resistant to biological attack at the other (Jenkinson and Rayner, 1977). Tiessen and Stewart (1983) showed, that during decades of cultivation particle size fractions of 0.5 to 10 um have lost the least organic matter and organic matter associated with fine clay (<0.5 um) was rapidly depleted. Apparently new biomass is separated mainly in the clay and fine clay fractions. Cultivation of virgin grasslands enhances the accessibility of organic matter to microbial attack bybreaking soil structure. This explains considerable losses of soil organic matter afeter cultivation of such soils (Voroney et al., 1981).

- soil fertility

Although the effect of calcium on the formation of stable humus compounds was known, in the numerous organic matter studies where the influence of soil texture was assessed (Jenny et al., 1968; Jones, 1973; many others) the base status was usually not taken into account.

The soil's base saturation may be an important factor in stabilizing organic matter in clay soils through clay bridging (Oades, 1988). The higher organic matter content of calcareous soils compared with non calcareous soils was in part due to Ca-humates. Removal of Ca from a soil stimulates the decomposition of organic matter. Addition of Ca inhibits the release of CO_2 and stabilizes soil structure. The above discussed relations (Oades, 1988) may not be valid in acid soils receiving Ca-fertilization, where organic matter decomposition is stimulated.

The nutrient status of a soil, which is inluenced by fertilizer applications and nutrient uptake by plant roots, may also have an effect on the quality of the microbial population. At higher mineral nutrient levels probably bacteria are preferentially stimulated. At lower levels a shift towards a more fungi dominated population might occur (Van Veen et al., 1987).

- vegetation-soils effect

Although forests produce more organic matter and more organic debris reaches the soil surface, the decomposition is faster in forest soils than in grassland soils and resulting organic matter contents are lower. Most organic matter in forests is added as litter and is decomposed before it is incorporated in the soil. In grasslands additions are in the form of root system which is well distributed through the complete soil profile. It is thoroughly mixed before decompostion. In evergreen forests the turnover time is 2 times that in deciduous forests. Oades (1988) concludes, that the physical accessibility of soil organic matter is of prime importance in determining the retention time in soil.

For temperate conditions a rule of thumb is, that soil carbon decreases in the order: natural grassland (prairie)- forest- cultivated land. The total biomass of a forest is greater than the biomass of prairies, but the recycling of organic matter in prairies is faster. The quantity of organic material added to the soil in prairies is 2-4 times greater than in forest areas. Additionally, less leaching occurs in prairies and leguminous species present in the plant population in prairies can make nitrogen available for the formation of stable humus compounds. The organic carbon content is higher in cultivated soils than in soils under forest or grass, since:

- the biomass produced on arable land is harvested;
- the organic substances of crops are less resistant against decomposition. Substrates are less well decomposable and possibly more re-immobilization of N occurs in rangeland soils than in cultivated soils (Schimel et al., 1985a). Decomposition rates are higher in cultivated soils. Possibly the microbial species composition was changed with the importance of fungi declining (see also Voroney et al., 1981).
- the average temperature on the soil surface in a forest is lower than in cultivated fields;
- losses due to leaching are smaller in forests.

- climate

Temperate and tropical soils should be treated separately in the assessment of the soil carbon content under natural conditions. Considering that temperate summers are as warm as or even warmer than tropical (rainy) periods, the data for soil carbon loss under cultivation may be similar in the tropics. Due to the fact that in tropical areas with a udic (moist) soil moisture regime (USDA, 1975) the biological activity continues during most of the year, the equilibrium after clear cutting of forests or after cultivating grassland, will be reached after a shorter period than in temperate regions. In the 22% of the tropics with udic soil moisture regimes (USDA, 1975) neither temperature nor moisture limits the biomass production and decomposition of organic matter. Forests in these areas produce about 5 times as much biomass and soil organic matter per year as comparable temperate forests, but the rate of decomposition of organic matter is also 4 to 5 times greater than in temperate regions (Sanchez, 1976; Jenkinson and Ayanaba, 1977).

In dry tropical areas, the biological activity is reduced during parts of the year, thereby extending the period needed to reach equilibrium. In the temperate regions the biological activity is greatly reduced during the winter . In the 78% of the tropics, that has an ustic (alternating moist and dry season) or aridic soil moisture regime (USDA, 1975), the lack of moisture during parts of the year has a similar effect. Temperatures during tropical rainy seasons are similar to -but seldom as high as- the corresponding summer temperatures in temperate regions (Sanchez, 1976). The carbon content in arid and semi-arid soils may even increase after agricultural use with irrigation. Schlesinger (1982) reports, that arid soils lose organic matter upon cultivation but may gain in carbonate carbon in calcrete or caliche.

The effect of temperature and moisture stress on metabolism and composition of bacteria, fungi and yeasts isolated from soil have provided evidence, that these two factors may act independently (Van Veen, 1987). Jenkinson and Ayanaba (1977) in comparing Rothamsted data with experiments carried out in Nigeria, concluded that the decomposition coefficient in Nigeria is 4 times that in temperate zones. They found no effect of soil type. If the decompostion process is 4 times faster in the tropics, then the annual input of organic matter must be 4 times as high to maintain a soil at the same C-content once steady state conditions have been achieved. Oades (1988) reports that in mediterranean Australian soils decomposition proceeds 2 times faster than in England.

- <u>the soil's preservation capacity</u>

Though the microbial biomass is a relatively small and labile fraction of the soil organic matter complex, its turnover is a rate determining process in the cycling of organic matter and nutrients. Soil texture and structure affect microbial biomass turnover and the related cycling of carbon and nitrogen. Van Veen (1987) related the soil conditions to the soil's preservation capacity. Each set of soil physical, chemical, hydrological and climatic conditions gives the soil characteristic capacities to preserve both organic matter and microorganisms. Preservation of organic matter could be the result of protection of microbial biomass and organic matter against predation or amelioration of harsh environmental conditions. Organic matter additions in excess of the soil's preservation capacity will be decomposed at an accelerated rate (Jenkinson and Rayner, 1977; Van Veen and Paul, 1981). Biomass formed in excess of a soil's preservation capacity for microorganisms is assumed to die at a relatively high rate (Van Veen et al., 1984; Van Veen, 1987). Biomass and its immediate products of decay are considered to form a fairly tightly closed system from which only small proportions of the products leak out as stabilized materials.

4.1.5 Soil carbon loss due to mineralization following interference in natural conditions

Modern agricultural practices, i.e. monocultures with limited return of crop residues to soil and use of chemical fertilizers, have been reported to cause a serious decline of soil organic matter levels (Allison, 1973; Diez and Bachtaler 1978; many others). However, cultivation generally increases the soil's bulk density and thus affects comparison of virgin and cultivated soils (Voroney et al., 1981, Mann, 1986).

Agricultural conversion disrupts the steady state conditions that exist in many natural communities. Annual organic carbon additions are drastically reduced when forests are brought under cultivation. Annual decomposition rates increase with higher temperature and aeration. The depletion of organic carbon and nitrogen can be attributed to changes in the magnitude of biological and physical processes in the soil. The rate of decomposition increases during cultivation due to increased microbial activity under a more favourable soil moisture and temperature regime. Incorporation of plant residues and pulverization of soil structure caused by tillage operations would increase the rate of mobilization of microorganisms (Voroney et al., 1981). The disruption of soil aggregates increases the accessibility of soil organic matter and enhances its mineralization.

Table 4.2. Mean loss of soil carbon after agricultural conversion for different ecosystems.

type	No. of studies	mean loss (%)	range(%)
temp.forest	5	34.0	3.0-56.5
temp. grassland	24	28.6	-2.5-47.5
tropical forest	19	21.0	1.7-69.2
tropical savanna	1	46.0	______

Source: Schlesinger (1986)

The data listed in table 4.2 are from a literature review by Schlesinger (1986) of long term paired plot experiments. The soils were sampled to various depths. Steady state soil carbon levels have been reached presumably in most experiments. The losses of soil humus listed in table 4.2 are likely to be equivalent to the 'light' or labile fraction of the soil organic matter. Mann (1986) calculated the carbon loss from 650 studies from temperate and tropical zones. The carbon loss from the top 30 cm varied from 0 to 20%. The greatest average loss was from the taxonomic class Borolls (USDA, 1975).

The loss of carbon from soils with a high initial carbon content tended to be higher than from soils poor in carbon.

From long term experiments on the behaviour of soil carbon in different soils under cultivation in Denmark (Dam Kofoed, 1982) can be concluded that soil carbon contents decrease by about 25% on fertilized (NPK) and on unmanured and unfertilized soils. The decrease was slightly lower (15 to 25%) on heavily manured soils. After prolonged periods of cultivation a new steady state level was reached. Jenkinson and Rayner (1977) indicated, that soil carbon remained equal in the Rothamsted classical experiments if no manure is applied and annual additions of organic matter are 1.1 t C y^{-1}, exclusively from crop residues. In their experiments the addition of farm yard manure enhanced the build up of soil organic matter substantially.

Organic matter levels have dropped to less than 50% after cultivating virgin grasslands on Chernozems (Van Veen and Paul, 1981; Voroney et al., 1981). The loss of soil carbon from the plow layer (Ap horizon) may be 90% of the total loss (Voroney et al., 1981). The degree of physical protection of soil organic matter decreased from 50% in virgin grasslands to 20% (in the top 15 cm) and 40% (below 15 cm). Schimel et al. (1985a) found a decrease in soil carbon after 44 years of cultivation of grasslands on Haploborolls and Argiborolls (USDA, 1975) of 24 to 56% depending on the soil's texture and the rate of erosion. The lowest loss rates were found in places were a deposition of soil material was recorded. The amounts of organic C increased and proportional losses decreased downslope.

Brams (1971) showed the general effects of cultivation on organic carbon contents in two West African soils. Cultivated soils had 40-60% of the soil carbon present in ferralitic soils under secundary forest. After 2 years of rice cultivation and 3 years of intertilled crops soils contained 30% of the original carbon present in alluvial soils with secundary forest cover. Jones (1972) calculated from data of the Guinea savanna, that soils under cultivation or under fallow following cultivation, have a mean carbon content slightly more than 50% of the content in virgin soils. Lugo et al. (1986) in an analysis of Central American subtropical soils report losses after clearing of wet forest of 46% after 10 years of cultivation, 70% after 100 years of cultivation of moist forest. In dry areas losses were lower. In all cases in relatively short periods of time (decades) most of the losses could be regained by abandonment of cultivation and secondary forest succession or use as pasture.

Shifting cultivation seldom results in substantial soil organic matter depletion; usually the carbon contents are maintained at 75% of natural equilibrium levels (Greenland and Nye, 1960:105), but in conditions of higher population pressure and lower rehabilitation periods soil carbon contents may drop to 50% of the original levels (Sanchez, 1976).

The variation in the above presented data is enormous, probably because the lack of systematic research on the carbon loss from pedogenetically uniform units or soils with similar textural and structural properties.

4.1.6 Soil carbon loss due to erosion

The carbon loss from cultivated soils with a soil loss through erosion of 5 to 10 tons of soil ha^{-1} and an average carbon content of 3% will amount to .15 to .30 tons C ha^{-1}. In areas of extreme water erosion these figures may be considerably higher.

Compared with the heavy losses through oxidation the above range is small (see table 4.4). However, after prolonged periods of cultivation the cumulative effect of erosion may be dramatic (Voroney et al., 1981; Schimel et al., 1985a; Schimel et al., 1985b). Moreover, organic matter is among the first constituents removed due to its weight. Eroded material is about 5 times richer in organic matter than the residual soil material (Allison, 1973). Erosional removal of organic matter following forest cutting is probably unimportant as long as vegetation is allowed to regenerate immediately (Bormann et al., 1974). Bouwman (1989b) reported that rainfall erosion of tropical soils used for cultivation may increase the carbon loss from soils considerably.

4.1.7 CO_2 evolution from soils

The soil carbon loss from a forest directly after clearing was assessed by Edwards and Ross-Todd (1983). They found an increase of the annual C evolution in the first year after cutting of only 79 kg C ha^{-1} for clear cut mixed deciduous forest with only sawlogs removed. The increase in CO_2 and only 169 kg C ha^{-1} for clear cut forest with removal of all woody material. This indicates, that a forest soil where forest is allowed to regrow after clearing and is <u>not</u> cultivated, will not produce substantial quantities of CO_2. This finding contradicts Scharpenseel (in prep.), who states in his review, that alternating drying and wetting of a bare soil with fresh decomposable organic material, such as after wood clearing, can lead to considerable 'priming effects' and organic matter destruction. The degree of pulverization or destruction of soil aggregates is the primary factor in determining the C-loss and CO_2 evolution.

Table 4.3. Total CO_2 evolution (from root + soil respiration) from different temperate soils calculated from various authors.

	C-loss	
	$g\ C\ d^{-1}\ m^{-2}$	$10^2 g\ C\ m^{-2} y^{-1}$
cropped soils/bare soil		
oats[1]	2.7	—
cabbage[1]	1.9	—
bare soil[5]	0.8	—
bare soil (winter)[2]	0.4	—
bare soil (summer)[2]	1.6-1.9	—
barley[2]	2.7-5.4	—
wheat after summer fallow[3]	3.4-3.8	23.5-25.5
wheat[4]	1.8	6.4
6-year crop rotation[8]	0.19-0.38	0.7-1.4
8-year rotation, heavily manured[8]	0.56	2.0
forest soils		
spruce forest[6]	1.3	4.9
mixed forest[7]	2.3	8.5
mixed deciduous forest[9]	2.3	8.4
mixed oak forest, summer[10]	0.55-0.95	—
mixed oak forest, winter[10]	0.14-0.22	—
oak forest[11]	2.2	7.9
cedar swamp forest[11]	2.1	7.4
fen forest[11]	1.9	7.1

[1] Lundegard(1927) quoted in: Buyanovski et al. (1986);[2] Monteith (1964) quoted in: Buyanovski et al. (1986);[3] de Jong and Schappert (1972);[4] Buyanovski et al. (1986);[5] Koepf(1953) quoted in: Minderman and Vulto (1973);[6] Hager (1975) quoted in Baumgartner and Kirchner(1980);[7] Bouten et al. (?);[8] Janssen (1984);[9] Edwards (1975);[10] Froment (1972);[11] Reiners (1968)

Only a portion of the decomposition is accounted for by measuring CO_2 output (Van Veen and Paul, 1981). If X is the amount of CO_2-C evolved during decomposition and Y is the efficiency of the use of C for biosynthesis expressed as a percentage of the total C-uptake under aerobic conditions and non-cell metabolite production is negligible, the actual amount decomposed is: $X(1+Y/(100-Y))$ (Paul and Van Veen, 1978). Microorganisms use carbon compounds for biosynthesis forming new cellular or extracellular material and as energy supply. In the latter process carbon compounds are converted to CO_2 and to a lesser extent low molecular weight compounds.

In table 4.3 CO_2 evolution given by various authors for cropped and forested soils are presented. The data shown are quite consistent, although in all the experiments with native vegetation and crops it is difficult to account for the root respiration and calculate the production of C purely from the soil organic matter decomposition. Due to litter fall and subsequent oxidation the CO_2 release from forest soils should be higher than from cultivated soils. The contribution of oxidation of crop residues to CO_2 evolution from cultivated soils however can be expected to be relatively small.

4.1.8 Estimation of the global soil C pool

Assuming a total decline of soil carbon of 75 tons C ha $^{-1}$ during a 15-year period after forest clearing (see table 4.4) this figure can be compared with experimental data on CO_2 production in cultivated soils. The period necessary to reach equilibrium and the soil humus content at equilibrium conditions depend on additions of organic material (direct effect) and fertilizers (indirect through increased dry matter production). In case of clearing a forest and cultivating the soil the initial decompostion rate will be high since first the easily oxidizable substances are decomposed. The break-down will proceed ever slower up till the equilibrium level is reached and the annual additions equal the decomposition. For simplicity an average carbon loss of 5 tons ha^{-1} (= $5x10^2$g m^{-2}y^{-1}) or 1.4 g C d^{-1} m^{-2} is assumed in this example. Table 4.3 suggests, that this loss rate fits well within the ranges for bare soils.

In the composition of the generalized soil carbon profiles data from Fitzpatrick (1983), Kononova (1975), Sanchez (1976) and Buringh (1983) have been used to to compute the carbon content at various depths for equilibrium conditions. The bulk density used in the calculation of table 4.4 is 1.5 g cm^{-3} for all soil units. The total (maximum) thickness of the profiles is taken as 1.0 m (after Houghton, 1983; Schlesinger, 1977). In the estimates for the generalized soil carbon profiles is taken into consideration, that soil carbon is not equal to total organic carbon in soils, since living roots and partly decomposed material is not included. Dead leaves and litter are not included in soil carbon.

In the calculation of the soil carbon profiles (0-1.0 m) in the different vegetation types it is assumed, that soil carbon decreases in the top 1.0 m by 40-60 % when forest or grassland is converted to cropland and by 25-35% when forest is converted to grassland. The depletion rate is taken as 25-40% for conversion of grassland to cropland. In this repect the difference in speed of decomposition between temperate and tropical regions is of no importance. Only the absolute differences are taken into account. The above assumptions are similar to those proposed by Buringh (1983). For other C depletion rates after land use changes is referred to section 4.1.9.

Table 4.4. Generalized organic carbon profiles for the major soil units of the FAO/UNESCO (1971-1978) 1:5,000,000 soil map of the world

FAO Major Soil		Generalized soil carbon profile					total soil C ($10^6 gCha^{-1}$)			
		generalized soil carbon profile in natural (typical) ecosystem					forest		grass	crop
	ecosystem	C (%) at depth (cm)					land prim.	land sec.		
		0 - 20	40	60	80	100				
Acrisols	rainforest	3	2.5	0.6	0.5	0.4	220	165	160	110
Cambisols	forest	5	3	1	0.5	0.1	290	215	210	140
Chernozems	prairie	4	3	2	1	0.5			315	235
Podzoluvisols	forest	4	0.5	0.5	0.5	0.1	160	120	120	100
Rendzinas	forest	5.8	3.0	0	0	0	130	100	90	65
Ferralsols	rainforest	3	1	0.5	0.2	0.1	116	120	100	70
Gleysols	grassland	10	5.8	0.5	0.25	0.1	350	260	350	200
Phaeozems	grassland	3	2.5	1.5	1	0.2	250	200	225	150
Lithosols	forest	2	0.5	0	0	0	75	50	50	40
Fluvisols		1	1	0.5	0.5	0.5	100	75	100	60
Kastanozems	grassland	2.9	2.3	1.5	1	0.5	300	225	250	190
Luvisols	forest	6	4	1	0.5	0.5	200	150	150	115
Greyzems	grassland	3	2.5	1.5	0.8	0.2	300	250	240	180
Nitosols	forest	3	1	1	0.8	0.5	200	150	150	115
Histosols*										
Podzols	forest	2	0.5	0.2	0.7	0.7	100	75	100	60
Arenosols	semides.	1.8	0.5	0.1	0	0			70	50
Regosols	forest	3	0.5	0.1	0.1	0	110	85	85	60
Solonetz	grassland	1.8	0.6	0.2	0.1	0.1			85	65
Andosols	forest	8	5	3	2	1	330	250	250	185
Rankers	forest	2	0.5	0	0	0	75	55	50	40
Vertisols	grassland	1.2	1.0	0.5	0.1	0.1	190	145	135	115
Planosols	forest	3	0.1	0.5	0.5	0.1	150	115	115	85
Xerosols	s.des./desere	0.8	0.4	0.1	0.1	0.1			50	30
Yermosols	s.des./desert	0.8	0.4	0.1	0.1	0.1			50	30
Solonchaks	grassland 1.8	1.2	0.5	0.1	0.1			110	80	

* Histosols as C pools are not considered here.

The resulting soil carbon profiles for the different land cover types are presented in table 4.4. The carbon profiles are not exact averages applicable to any particular country. They are based mainly on assumptions and generalizations, but in the scope of this study they are hoped to be of use.

Using the generalized soil carbon profiles and the world distribution of soils and ecosystems (table 2.2 and 4.4 respectively) the total global soil carbon pool can be estimated. The soil carbon pool estimated this way is about 1700 Gt C.

4.1.9 Fluxes of CO_2 from soils due to global land use changes

CO_2 release from mineral soils
Since pre-historic times the contribution of soil carbon to atmospheric carbon dioxide must have been considerable. Estimates range from about 537x10^{15} g C for the total flux since pre-historic times (Buringh, 1984) to 36x10^{15} g C for the loss since the mid-1800s (Schlesinger, 1984).

Table 4.5. Estimates of the carbon release from soils circa 1980. Fluxes in 10^{15} g y^{-1})

source of estimate	flux
Bolin (1977)	0.3
Schlesinger (1977)	0.85
Buringh (1984)	1.5 - 5.4
Schlesinger (1984)	0.8
Houghton (1987, pers. comm.)	0.2 - 0.52
Bouwman (1989b)	0.1 - 0.4
Detwiler & Hall (1988)	0.11- 0.25

The wide spread in estimates of the carbon release from soils due to land use changes is shown in Table 4.5.

Houghton et al. (1983) assumed losses of 35%, 50% and 15% in the top 1.0 m for agricultural conversion of tropical forests, temperate forests and boreal forests respectively. Houghton et al. (1987) applied loss rates of 25% after cultivation of forests in temperate regions and 20% after forest clearing a d regrowth. In the tropics Houghton et al. (1987) assumed loss rates of 30% for conversion to cropland and 25% for conversion to grassland. Palm et al. (1980) used losses of 30% and 20% for conversion of primary resp. secundary forest to permanent agriculture and 25% and 17% loss for conversion to primary resp. secundary forest in shifting cultivation in South east Asia. Soil carbon levels recovered to 90% of primary forest levels after forest regrowth. In Buringh's estimate (Buringh, 1984) however, a major source of CO_2 is caused by transition of forest to non agricultural land (urban land).

CO_2 release due to drainage of organic soils
Organic soils increase in mass by vertical accretion or by paludification (the lateral spreading of peat). Armentano (1980) used a global area of natural Histosols of 450 x 10^{10} m^2. With an average carbon accumulation of 300 kg ha^{-1} y^{-1} total C sequestering is 0.135 Gt C y^{-1}. Duxbury (1979, quoted in Armentano, 1980) assumed that 7 to 35x$10^{10}$$m^2$ of the global area of Histosols is being drained annually. This causes a C release by oxidation of the organic material of 10 t C ha^{-1}, yielding a global annual C release of 0.05 to 0.35 Gt C. Another approach is based on a subsidence rate of 1-3 cm, a bulk density of 0.5 g cm^{-3}, a C content of 30%, yielding a release of 0.02-0.07 GtC y^{-1}. Drainage of 10^{10} m^2 of Gleysols causes an extra release of 0.01 Gt C y^{-1}. Thus, the total global release from Gleysols and Histosols ranges between 0.03 and 0.37Gt C y^{-1}.

Armentano and Menges (1986) analysed organic soil wetlands of the temperate zones, occupying about 350x10^{10} m^2. Based on an average storage rate of 200kg C ha^{-1}, the annual storage before disturbance was 0.06-0.08 GtCy^{-1}. Armentano and Menges (1986) estimated that the total area drained annually is 8.2x10^{10} m^2 for crops, 5.5x10^{10} m^2 for pasture and 9.4x10^{10} m^2 for forests. The annual shift (loss of sink strength and gain of source strength) in the global C balance is 0.063-0.085 Gt C due to draining of Histosols in temperate regions. Including tropical Histosols, the global shift would be 0.15 to 0.184 Gt C.

4.1.10 Fluxes of CO_2 from vegetation due to land use changes; discussion of global carbon models

Recent analyses have shown, that the biota is not currently a net accumulator of atmospheric CO_2 but is releasing stored carbon into the atmosphere (Bolin, 1977; Woodwell et al., 1978; Houghton et al., 1983, 1985, 1987; Detwiler and Hall, 1988). Other authors suggest that the biosphere is a net absorber of CO_2 (Goudriaan and Ketner, 1984; Goudriaan, 1987; Esser, 1986).

There are two approaches to the assessment of the role of terrestrial biota in the CO_2 budget. First, there are bookkeeping models which account for rates of deforestation and forest volumes and yield the CO_2 release. The best known representatives of this approach are Houghton et al. (1983, 1985, 1987) and Detwiler & Hall (1985, 1988). Secondly, a number of researchers have developed dynamic models of the global carbon cycle including the CO_2 injection into the atmosphere by fossil fuel combustion and forest clearing, the oceans as a CO_2 sink and the fertilization effect of CO_2. Examples of this approach are Goudriaan and Ketner (1984) and Esser (1986).

There are a number of difficulties in such global analyses:
- The biomass of the forests which are being cleared is an uncertain factor. Estimates of the biomass of tropical forests are extremely variable. Whittaker and Likens (1975) use 202 t C ha^{-1} and 156 t C ha^{-1} for tropical rainforest and tropical seasonal forest, respectively. The C densities estimated by Brown and Lugo (1982) are 46 and 183 t C ha^{-1} and their later estimate (Brown and Lugo, 1984) amounts 55-90 for broadleaf closed forest, 31-70 for conifer closed forest and 20 for open forests. The weighted averages are 188 (Whittaker and Likens, 1975), 124 (Brown and Lugo, 1982) and 53 (Brown and Lugo, 1984).
- The area of forest cleared annually. This aspect is discussed in chapter 3.
- The stimulation of growth by inceased CO_2 partial pressure in the atmosphere. Some authors (e.g. Woodwell et al., 1978) indicate that this fertilization effect is not great enough to compensate for the vast amounts of CO_2 injected into the atmosphere by deforestation. In the discussion of the dynamic models below, this aspect will be analysed further.

Bookkeeping models
- Houghton et al. (1983) used the biomass figures presented by Whittaker and Likens (1975) and for land use changes both data in the FAO production yearbooks (1949-1978) and data presented by Myers (1981). The estimated total annual release rates from the terrestrial biota are 1.82 and 4.70 Gt C y^{-1} respectively. In a third data set Houghton et al. (1983) assumed that the rate of agricultural expansion since 1950 has been proportional to the rate of growth of populations in the tropical regions. 40% of the required increase in food production was assumed to be achieved through increasing agricultural output on existing cropland. This third data set yielded a release rate of 2.6 Gt C y^{-1}.

Houghton et al. (1983) also assessed the effect of a reduced soil response to agricultural clearing. Assuming only 20% loss of soil carbon after clearing the total annual flux was reduced by only 11%. , but the net flux from harvested and regrowing forest decreased by almost 50%. In another test 0 soil carbon loss was assumed following the clearing of forests. The effect of this assumption is only app. 10% on the total calculated flux.

The analysis was also tested for the low estimate of biomass per unit area of tropical forests made by Brown and Lugo (1982). The test shows, that with lower carbon contents of forests less carbon will be released, but net biomass accumulation will be lower during forest regrowth. So the total net release does not proportionally reflect the differences in forest biomass.

- Houghton et al. (1985, 1987) repeated their analysis with reduced soil carbon loss rates (see 4.1.8) and including data on shifting cultivation and clearing of fallow forests (forests that had been previously in agriculture) for permanent cultivation. Using the FAO/UNEP data the spatial distribution of the CO_2 emission could be calculated. The model was calculated for carbon stocks given by Brown and Lugo (1982) and Brown and Lugo (1984) respectively. The range in carbon release rates calculated by Houghton et al. (1985,1987) is 1.0 to 2.6 Gt C y^{-1}. In this figure the contribution of the tropics is 0.9 to 2.5 Gt C y^{-1} of which 0.4 to 0.8 Gt C y^{-1} is due to permanent deforestation of fallow forests. In relation to previous calculations (Houghton et al., 1983) the range is narrower, whereby the maximum flux decreased from 4.7 to 2.6 Gt C y^{-1}. The soils contribution to this emission is about 20% or 0.2 to 0.5 Gt C y^{-1} (Houghton, 1987, pers.comm.).

- Detwiler and Hall (1988) analysed the annual release due to forest clearing using all possible statistical land use data and carbon stock data. Land use data used are from FAO Production yearbooks (FAO, 1949-1985), Seiler and Crutzen (1980), and Myers (in Houghton, 1983). The carbon density data used are from Brown and Lugo (1982) and brown and Lugo (1984) respectively. The resulting annual release is 0.3 to 1.7 Gt C, most of which is from tropical regions. Soil organic matter loss causes an emission of 0.11 to 0.25 Gt.

Dynamic models
A number of dynamic carbon models have been developed. Examples are Goudriaan & Ketner (1984), who used a simple matrix of conversions of land use and a 12 layer ocean model, and Esser (1987) who used soil, climate and land use data sets and an ocean consisting of 2 layers, all regionalized on a 2.5° latitude x2.5° longitude grid. The results of both dynamic models suggest that at present the biosphere is a net sink of CO_2 (see Table 4.6). The CO_2 induced incease of net primary production is greater than the loss of CO_2 caused by deforestation. Goudriaan & Ketner (1984) suggest that charcoal formation in the process of biomass burning is another important sink of carbon. The point in time where the biosphere turned into a sink of CO_2 is around 1970 according to both dynamic models. Further deforestation however, will repress the CO_2 fertilizing effect (Esser, 1987). The ever decreasing vitality of forests caused by acid precipitation in industrialized regions may also become increasingly important.

Table 4.6. Comparison of the annual flux of carbon from terrestrial biota by various authors. All figures in GT C y^{-1}

reference	present release caused by forest clearings	present net release including increased net primary production
1. bookkeeping models		
Houghton et al. (1983)	1.8-4.7	
Houghton et al. (1987)	1.0-2.6	
Detwiler & Hall (1988)	0.3-1.7	
2. dynamic biosphere models		
Goudriaan and Ketner (1984)		0 to -0.5
Esser (1987)	2.7	- 0.1

Adapted from Bouwman (1989a).

Other sinks and sources of CO_2
Aspects of the carbon cycle, that would influence the release rate (and the kind of influence), are suppression of forest fires (reduction), expansion of paddy rice (reduction), expansion of no tillage agriculture (reduction), drainage of wetlands (increase), increase of the intensity of shifting cultivation (increase) and increase of the area of arid lands (increase). These changes will be of large local importance, but at a global scale insignificant.

Another possible source of atmospheric CO_2 is the digestion by termites. Termites process material equivalent to about 28% of the earth's net primary production and about 37% of the net primary production in areas where they occur. Thereby amounts of CO_2 are released of 13 Gt C y^{-1}

(Zimmermann et al., 1982). With all uncertainties the above emission of CO_2 by termites could range from 6.5 to 26 Gt C y^{-1}. Later studies by other authors revealed however, that the biomass comsumption rate assumed by Zimmermann et al. (1982) is much too high. Termite induced CO_2 release is only important where the population increases. Although CO_2 would probably be released as the result of any decomposition process, termites serve to accelerate carbon cycling. The ecological areas that should show the largest increase in emissions from termites are tropical wet savanna, areas that have been cleared or burned and cultivated land in developing countries (see also section 4.2.9).

4.1.11 Conclusions

There are a number of areas of uncertainty in the estimates of CO_2 release from terrestrial biota. This is expressed by the variation in the estimates produced by the various dynamic carbon models, such as Goudriaan and Ketner (1984), Esser (1987), Goudriaan (1988), and bookkeeping models, such as Houghton et al. (1983) and Houghton et al. (1987).
- The largest biotic source of carbon dioxide is the forest conversion to agriculture, most of it occurring in the tropics. Yet one of the greatest uncertainties is the rate of deforestation, the type of land use after clearing, and to a lesser degree, the areal extent of each life zone. The definitional contoversy is also responsible for differences in the estimates.
- A second important cause of unreliability of the estimates is the volume and carbon density of tropical forests. However, the effect of density on the net flux of CO_2 is small compared to the effect of the rate of forest clearing.
- A third uncertain factor is the soil carbon loss due to land cover changes and the amount of charcoal formed during biomass burning and forest burns.

The use of one single vegetation and land use classification method would improve the comparison between estimates. In this respect is referred to chapter 3, where differences and points of similarity of the various methods is discussed. With respect to the forest biomass intensified field work linked with the use of remote sensing techniques, especially radar, would improve characterization of the distinguished ecosystem types.
To improve estimates of soil carbon pools and the annual release of carbon from soils, modelling of decomposition combined with C^{14} techniques will be necessary to single out pool size and release rate.
Finally, for monitoring of tropical regions on regional and global scales a number of remote sensing techniques are available. The NOAA-AVHRR sensor data as discussed in chapter 7, combined with radar techniques and field observations, will prove a very helpful tool in future.

4.2 METHANE

4.2.1 Introduction

The presence of methane (CH_4) in the atmosphere has been known since the 1940's, when strong absorption bands in the infra-red region of the electromagnetic spectrum were discovered which were attributed to the presence of atmospheric CH_4. Concentrations (mixing ratios) of CH_4 in the troposphere vary from about 1.7 ppmv in the Northern hemisphere to about 1.6 ppmv in the Southern hemisphere (Rasmussen and Khalil, 1986; Steele et al., 1987).

The first evidence for an increase of methane in the atmosphere was given by Rasmussen and Khalil (1981). This was later confirmed by Blake et al. (1982). This rise was attributed to the existence of an upward trend of CH_4 emissions. The average temporal increase of atmospheric CH_4 during the 1978-1983 was about 18 ppbv per year (Blake and Rowland, 1986) or 1.1% per year (Bolle et al., 1986). Steele et al. (1987) observed a slowing down of the growth rate of the methane concentration to 15.6 ppb y^{-1} in 1983/1984, or globally 0.78%.

The atmospheric burden is about 4700 T g (Terragram; 1 Terragram = 10^{12} g) (Wahlen et al., 1989) to 4800 T g (Cicerone and Oremland, 1988). With a residence time of 8.1-11.8 years (Cicerone and Oremland, 1988) to 8.7-8.8 years (Wahlen et al., 1989) the total sources in quasi steady state must be 500 ± 95 T g CH_4 (Cicerone and Oremland, 1988) to 580-590 T g y^{-1} (Wahlen et al., 1989). An earlier estimate of steady state sources is 320 T g (Crutzen and Gidel, 1983) including 60 Tg for transport to the stratosphere.

Abiotic sources account for 21 ± 3 % (Wahlen et al. 1989) to about 30% (Manning et al., 1989) of total sources, being 100-140 Tg y^{-1}. Possible abiotic sources of CH_4 are coal mining, exploitation of natural gas, leaks, and CH_4 from very old peats.

The observed increase over 1970-1980 requires a yearly excess of 70 T g CH_4 of sources over sinks (Blake et al., 1982) to 50 T g CH_4 (Bolle et al., 1986) or 40-46 T g (Cicerone and Oremland, 1988). Khalil and Rasmussen (1983) showed that during summer in the Northern hemisphere CH_4 concentrations are 2% lower and that a sharp rise occurs in fall. This phenomenon suggests a fall source at Northern latitudes. These seasonal variations are generally consistent with seasonal and latitudinal variations of OH. concentrations.

The only major removal process for methane is in the atmospheric process presented by reaction (1), with the OH. production driven by photochemistry as in (2) and (3) (after Blake et al., 1982):

$$OH. + CH_4 \longrightarrow H_2O + CH_3 \qquad (1)$$
$$O_3 + h\nu \longrightarrow O + O_2 \qquad (2)$$
$$O + H_2O \longrightarrow OH. + OH. \qquad (3)$$

Khalil and Rasmussen (1985) calculated, that the increase of methane over the past 200 years is probably due to the increase of emissions (70%) and a lesser amount is due to a possible depletion of OH. radicals (30%). The OH. depletion is caused primarily by the ever larger CO emission from various anthropogenic sources. Recent estimates of the sources recognized so far are listed in Table 4.7.

The role of methane in the atmosphere is a complex one:
1. methane has absorption bands in the infrared region of the spectrum.
2. CH_4 is oxidized in the troposphere by the free radical OH., whereby products such as ozone are formed.
3. methane is a sizeable source of CO through its oxidation by OH.
4. methane is a source of water vapour in the stratosphere as a result of its oxidation to CO_2 and H_2O.

5. stratospheric methane can react with Cl. radicals, forming HCl which slows the rate at which Cl and ClO destroy stratospheric ozone.

Methane is produced during microbial decomposition of organic materials under strictly anaerobic conditions, e.g. in waterlogged soils. The processes involved in methanogenesis are discussed in 4.2.2. The factors influencing the actual emission (i.e. production minus oxidation or consumption) are highlighted in 4.2.3.
CH_4 may also be consumed by microorganisms in aerobic soils in tropical regions (see 4.2.11).

In the present study the methane emission from paddy rice soils and natural wetlands (sections 4.2.2-4.2.8) and emission by herbivorous animals (termites and ruminants, section 4.2.9) will be discussed in more detail. Methane production due to biomass burning is discussed only briefly.

Table 4.7. CH_4 emission rates from various sources.

Source	CH_4 emission ($Tg\,C\,y^{-1}$)
rice fields	60 -140[a]
natural wetlands	40 -160[a]
landfill sites	30 - 70[f]
oceans/lakes/other biogenic	15 - 35
intestines of ruminants	66 - 90[g]
termites	6 - 42[b]/2-5[c]
exploitation of natural gas	30 - 40[h]
coal mining	35[h]
biomass burning	55 -100[h]
other nonbiogenic	1 - 2[h]
total	338 -714 [d] /334 -677[e]
Total sinks	400 -590[i]

[a] Aselmann and Crutzen (1989); [b] Fraser et al. (1986); [c] Seiler (1984); [d]-[e] total emission accounting for emission rates by termites given by Fraser et al. (1986) and Seiler (1984) resp.;[f] Bingemer and Crutzen (1987); [g] Crutzen et al. (1986); [h] Seiler in Bolle et al. (1986). [i] Cicerone and Oremland (1988) and Wahlen et al. (1989) respectively.

4.2.2 Methanogenesis in soils

Fermentation is the major biochemical function of organic matter degradation in a flooded soil. The main products of this fermentation process are ethanol, acetate, lactate, propionate, butyrate, molecular hydrogen, methane and carbon dioxide. More nitrogen gas and less methane are found in wetland soils planted to rice at later growth stages of rice than in an unplanted rice field. H_2 usually does not accumulate in a significant amount in a flooded soil (Yoshida, 1978).

The reduction of CO_2 with H_2, fatty acids or alcohols as hydrogen donor, and transmethylation of acetic acid or methyl alcohol by methane producing bacteria are considered as the major pathways to methane gas (Takai, 1970):

1. $CO_2 + 4\ H_2A \quad \text{--->} \quad CH_4 + 2H_2O + 4A$
2. $RCOOH \quad \text{--->} \quad RH + CO_2$

CH_4 is formed preferentially from acetic acid (Takai, 1970; Neue and Scharpenseel, 1984).

The addition of nitrate retards the methanogenesis. The competition of methane producing bacteria with amongst others denitrifiers and the high redox potential of nitrate or ferric iron are possible causes of this reduction of methane production (Yoshida, 1978).

The reduction of soil proceeds in two major steps: first reduction of NO_3^- Mn^{4+} and Fe^{3+} takes place (facultative anaerobic step) in that order. In the second step sulphate reduction and CH_4 formation by strictly anaerobic bacteria takes place in that order. The ratio of the final products of organic matter decomposition in anaerobic soil (CO_2 : CH_4) is regulated by the ratio of oxidizing capacity (= the amount of reducible O_2, NO_3^- Mn^{4+} and Fe^{3+}) to reducing capacity (Watanabe, 1984).

4.2.3 Controlling factors for methane production in soils

Redox potential (E_h)
The redox potential of a soil decreases after flooding. This is due to a decrease in activity of the oxidized phase and increased activity of the reduced phase. A positive correlation exists between the soil reduction potential E_h and methane emission, whereby the methanogenesis appears to follow the sequential usage of oxygen, nitrate, ferric iron and sulphate (Patrick et al., 1981; Jacobsen et al., 1981, Cicerone and Shetter, 1983). The relation which describes the redox potential for the equilibrium:

$$Ox + ne + m\,H^+ \longleftrightarrow Red$$

is:

$$E_h = E_o + 2.303\ RT/nF \log (Ox)/(Red) - 2.303\ RTm/nF\ pH$$

where:
R = gas constant;
T = temperature (K);
F = 96500 Coulomb eq^{-1};
E_o= equilibrium potential.

For Fe^{3+}/Fe^{2+} the reaction is:

$$Fe(OH)_3 + 3\,H^+ + e \longleftrightarrow Fe^{2+} + 3\,H_2O$$

for T = 25°C: $E_h = 1.06 - 0.059 \log (Fe^{2+}) - 0.177\ pH$

Thus, at increasing Fe^{2+} activity or higher pH the E_h will be low. However, not all soils have the same steady state E_h under flooded conditions. High levels of available Fe^{3+}, high content of organic matter, low NO_3^- MnO_2 and O_2 and high temperatures favor E_h decrease (Ponnamperuma, 1981). Swarup (1988) observed a sharp drop of the E_h due to addition of organic matter. The type of manure also had an effect on the fall of E_h.

Soil reaction (pH)
In acid soil flooding will have an increasing effect on the soil pH, while flooding will decrease the pH of alkaline soils. In acid soils the increase of pH is due to the reduction of Fe^{3+} to Fe^{2+} mainly. In alkaline soils accumulation of CO_2 causes a decrease of pH (Ponnamperuma, 1985). Williams and Crawford (1984) showed that the optimum pH for methanogenesis is 6.0 for peat soils with actual soil pH values of 3.8 to 4.3.

Substrate and nutrient availability
Fluxes may correlate with availability of oxidizible substrate (DeLaune et al., 1986), or peat depth and nutrient enrichment from e.g. groundwater (Harriss and Sebacher, 1981) or nitrogen fertilizer (Cicerone and Shetter, 1983). Holzapfel-Pschorn and Seiler (1986) reported that in early stages of rice growth methane emission is similar to emissions from unplanted fields. In both situations the flux shows a peak short after inundation due to mineralization of soil organic matter. A second maximum

occurs in the planted fields in the physiologically most active period of the rice plants (Holzapfel-Pschorn and Seiler, 1986; Schutz et al., 1989). This second peak was attributed to a supply of organic matter in the form of root exudates. These exudates are easily decomposable substrates and are preferentially released during the vegetative stage of growth. Results presented by Swarup (1988) indicate that the second peak may also be related to the E_h of the soil, which was at its minimum after 30 days of crop growth. Schutz et al. (1989) reported that CH_4 emission rates are increased considerably in the early stage after application of organic matter. Very high applications did not increase CH_4 emission, probably due to the formation of toxic products of fermentation. Schutz et al. (1989) found reduced methane fluxes after incorporation of $(NH_4)_2SO_4$ or urea in the soil. In the case of $(NH_4)_2SO_4$ this can be attributed to the presence of sulfate. Due to sulfate methanogens may be outcompeted by sulfate reducing bacteria. For the reduction of the methane emission due to urea there is no simple explanation.

Temperature
Methane flux was poorly correlated with temperature in the study by Sebacher et al. (1986). Fluxes from Northern areas were higher than could be expected from other experiments in temperate regions. An explanation proposed by Sebacher et al. is the presence of low temperature adapted methanogens, a phenomenon earlier reported by Svensson (1984). Holzapfel-Pschorn and Seiler (1986) report a marked influence of soil temperature on the CH_4 flux. They even report a doubling of emission rates at a temperature increase of 20 to 25°C, which was later confirmed by Schutz et al. (1989). Diurnal variation of the CH_4 emission is correlated with temperature. Amplitudes are high in early stages of growth and low during the second half of the growing season when the soil is shaded by rice plants (Schutz et al., 1989).

Sulphate concentration and presence of sulphate reducing bacteria
Sea water sulphate and sulphate reducing bacteria may interact with methanogenesis by competition with and/or inhibition of methanogenesis in sediments. Sulphate reducing bacteria may even oxidize CH_4 causing low concentrations in the soil or water column and subsequently little flux to the atmosphere. Bartlett et al. (1985, 1987) found CH_4 and SO_4^{2-} concentrations to be negatively correlated. A possible conclusion is that methane emission from salt wetlands, which usually contain considerable amounts of SO_4^{2-}, are lower than from fresh water wetlands.

Nitrate and sulphate or their reduction products repressed methane formation (Jacobsen et al., 1981). The effect of nitrate is twofold: first it delays methane formation until the reduction of nitrate is complete and the redox potential is lowered sufficiently for further anaerobic reactions to proceed. Secondly nitrate exerts a toxic effect on methanogenesis. Sulphate is toxic to methane formation (Jacobsen et al., 1981) and/or sulphate reducing bacteria use methane as electron acceptor.

Organic matter in paddy soils
The organic matter content of rice soils may range from over 30% in peaty soils to 0.8% and less in certain mineral soils. In Africa the organic matter content is used as an indicator of natural fertility for rice cultivation. The presence of too much organic matter may become a limiting factor. Organic matter in the form of manure is not often applied in rice fields, but is receiving increasing interest. Organic matter application to wetland soils may lead to organic acid production which has an injurious effect. Adding ammonium sulphate may reduce this latter effect (Yoshida, 1978).

Changes in soil organic matter content have been reported under wet rice cultivation. These changes are most marked in soils which in their natural status are more or less freely drained, but are only minor or absent in the majority of poorly drained lowland soils under rice (Moormann, 1981).

Anaerobic environments such as anaerobic digestion sludges, rumens, silages, lake and ocean sediments and paddy soils have many common characteristics. The patterns of anaerobic digestion are quite similar, as follows:
- nitrate is easily denitrified, while ammonium is a stable product of N-metabolism;
- accumulation of volatile fatty acids (VFA), mainly acetic, propionic and butyric acids;
- the formation of CH_4 and sulfide after the accumulation of VFA.

The differences of paddy soils from other aquatic environments are:
- the paddy system is not continuous due to incorporation of organic matter, green manure, crop residues, etc.;
- the paddy system is heterogeneous: the reduced layer of flooded rice is situated between the oxidized surface soil and the oxidized subsoil. Within the reduced layer there are patches of aerobic sites due to excretion of oxygen by rice roots.

4.2.4 Factors determining the methane flux

Water depth
An aspect which may control the methane flux is the water depth. Sebacher et al. (1986) found that water depths greater than 10 cm do not promote methane emission. Microbial CH_4 oxidation in aerobic water columns deeper than 10 cm may occur (de Bont et al., 1978; Delaune et al., 1983). Sebacher et al. (1986) also found that emission rates were linearly correlated with water depths up to about 10 cm.

Profile of the anaerobic environment
In steady state undisturbed methanogenic ecosystems show a characteristic stratification. The top layer is aerobic, the second layer is more reduced with Fe^{3+}, Mn^{4+} and NO_3^- still present. Below that a zone of sulphate reduction is found and finally a zone of bicarbonate reduction is found. This layering coincides with the value of the redox potential, which is discussed above. In the oxidized layer oxidation may occur of the methane produced in the zone of bicarbonate reduction (see below). The CH_4 concentration in interstitial water increases with depth in peat soils (Williams and Crawford, 1984; Dinel et al., 1988). Williams and Crawford (1984) suggested that the CH_4 measured at some depth was formed many years ago and now inhibits further methanogenesis.

Methane oxidation
Although methane flux rates appear to be a function of the total amount of methane in the soil, the vertical distribution of the gas also plays a role. Both the magnitude and the depth in the sediment of maximum methane concentration appear to increase as sediment temperatures increase, suggesting that the balance between the microbial processes of methanogenesis and methane consumption or oxidation controls near surface methane concentrations and thus in large part the flux of methane to the atmosphere (Bartlett et al., 1985). Holzapfel-Pschorn et al.. (1986) report that during a rice crop 67% of the methane produced is being oxidized and only 23% is actually emitted. In the absence of rice plants 35% of methane production is actually emitted, but methane production was much lower.

Process of methane release into the atmosphere
possible ways are:
- ebullition: methane loss as bubbles from the sediment should be a common and significant mechanism accounting for between 49 and 64% of total flux (Bartlett et al., 1988) to 70% (Crill et al., 1988). This contrasts findings of Seiler (1984) and Holzapfel-Pschorn et al. (1986) who reported that in a rice paddy 90 to 95% of the total methane release is through diffusive transport through the aerenchyma of rice plants.
- diffusion: diffusional loss of methnae across a water surface is a function of surface water concentration of methane, wind speed and methane supply to the surface water (Sebacher et al., 1983).
- transport through plants: typically in aquatic plants parts of the parenchyma break down leaving large lacunae for gas storage and transport. The phenomenon of methane transport through aerenchyma has been reported for rice (de Bont et al., 1978; Seiler, 1984; Cicerone and Shetter, 1983) and for other aquatic plants by Sebacher et al. (1985). More than 95% of the total methane release from paddy soils is through diffusive transport through the aerenchyma system of the rice plants (Seiler, 1984) and not through diffusion or escape of bubbles across the air-water interface. Transport of CH_4 from paddy soils into the atmosphere by rising bubbles is only important for

unplanted fields. Studies by de Bont et al. (1978) show, that the presence of rice plants enhances the escape of methane from soil. Older rice plants at the ripening stage released about 20 times more CH_4 than 2 week old seedlings. Holzapfel-Pschorn et al. (1986) reported that rice plants, and not weeds, have a stimulatory effect on methane emission. Rice paddies emitted about twice as much methane as unplanted fields. Kimura et al. (1984) found, that H_2 production was more active at younger roots, while CH_4 production was higher at older roots.

High concentrations of dissolved methane in root mats of floating meadows suggest high in situ methanogenesis with restricted flow where water can become oxygen deficient; other possible causes are trapping and subsequent dissolution of bubbles (Bartlett et al., 1988). Air samples from gas spaces in stems and leaves indicate elevated methane concentrations over concentrations in ambient air. Emission through plants would also be expected to show great diurnal variations tied to environmental changes and variations in respiration and photosynthesis rates.

Textural stratification in rice paddies
Puddling, the wet tillage of rice soils, causes a stratification of the soil material. Coarse materials settle first and are covered by finer silt and clay. Medium textured soils show a clear stratification upon puddling. In sandy soils the clayey cover is thin or absent. In very fine soils stratification may be difficult to observe. Trapping of gases may cause a build up of a vesicular structure. The presence of fine clay layers and algae at the soil surface restricts the escape of gases and contributes to the formation of vesicles.

4.2.5 Geographic distribution of paddy soils

The FAO (1985) estimated the harvested area of paddy rice at 144×10^6 ha of which 95% is located in the Far East (see Table 4.8). This area corresponds to approximately 9.5% of the total global cultivated area. The area harvested of paddy rice has increased from 86×10^6 to 144×10^6 ha between 1935 and 1985 (see Table 4.8) which is an annual average increase of 1.05%. Between 1950 and 1985 the average annual increase has been 1.23%. The last few years however, the expansion of the total acreage of paddy rice is decreasing. In appendix II the geographic distribution of paddy rice cultivation in Asia is presented. The area of paddy rice includes so called wet rice and dryland rice. Wet rice is grown in puddled soil and may be irrigated and continuously inundated or rainfed and almost permanently inundated. In Appendix III the increase of the irrigated area of paddy rice in selected countries is presented.

Rice is grown on a wide variety of soils, predominantly belonging to the Gleysols, Fluvisols, Luvisols, Acrisols, Nitosols (for a definition of these soil concepts is referred to FAO/Unesco, 1971-1981). Vertisols and Histosols are of minor importance. Special mention should be made of the Gleysols, which occur to a great extent in wetland areas and are wet during part of the year. This circumstance makes them attractive for rice cultivation. More detailed information on the distribution and classification of paddy soils can be found in Moormann and van Breemen (1978) and Moormann (1981).

Table 4.8 Global area harvested of wet and dryland rice (107 m2)

	1935	1950	1960	1970	1980	1985
Africa	1850	2900	2880	3960	4894	5467
N/C America	540	1040	1280	1428	2076	1914
S America	1190	2300	3880	5741	7258	6122
Asia	82000	87600	110940	122302	128393	129977
Europe	220	300	350	395	366	388
Oceania	10	30	40	50	123	140
USSR	148	nd	100	356	637	667
World	85958	94170	119470	134232	143747	144675

From: FAO (1952-1986) Production yearbooks 1951-1985. (nd = no data)

4.2.6 Fluxes of CH_4 from rice paddies

Over 80% of all atmospheric methane is of biogenic origin. 33 to 49 % of all methane is attributed to release from the world's rice paddies. Nitrogen fertilization may affect methane escape (Cicerone et al., 1983; Holzapfel-Pschorn and Seiler, 1986) possibly due to its effect on the growth of rice plants; further research on the effect of different types and doses of N-fertilizers is required.

Cicerone et al. found in an in situ experiment average daily emissions of 0.25 g CH_4 m^{-2} day^{-1}, the total accumulated emission over the 100-day growing season being 22 to 28 g CH_4 m^{-2}. During a 2 to 3 week period before the harvest the emissions reached 5.0 g CH_4 m^{-2} day^{-1}. Dramatic variation in methane flux through the growing season was found. Seiler et al. (1984) report emissions of 12 g CH_4 m^{-2} over the growing season in a Spanish rice paddy. Their low estimate was attributed to inflow of sulphate containing mediterranean water, which would have inhibited methanogenesis. Measurements in an Italian rice paddy by Holzapfel Pschorn and Seiler (1986) yielded a figure of 27-81 g CH_4 m^{-2} during a vegetation period. Holzapfel-Pschorn et al. (1986) report an emission of 36.3 g m^{-2} for an Italian rice paddy.

Methane emission rates as presented by a number of researchers are shown in Table 4.9.

Extrapolation of these emission rates to a global scale is difficult, since the effect of variations in agricultural practices, number of crops per year and other factors discussed above are highly uncertain. In Table 4.10 the geographic information from Table 4.9 and Appendix II are combined with flux estimates provided by Schutz et al. (1989). The latter authors gave methane release rates in dependence of temperature. The methane emission for intermediate and deep water rice is taken from Bartlett et al. (1988) for a tropical open water lake. About 90% of the world's harvested area of paddy rice is located in Asia. Of the total harvested area in Asia about 50% is irrigated (permanently wet) and another 39% is wetland rainfed paddy rice (almost continuously wet). Assuming that in the latter fields wet conditions prevail during 80% of the growing season and assuming that these figures can be extrapolated to other continents as well, the percentage of the global harvested area of paddy rice that is inundated is about 80%. In the calculation presented in Table 4.10 the period of inundation of rainfed paddies is assumed to be 80% of the growing period.

Table 4.9. CH_4 emission during the growing season (about 100 days) from wet rice soils as determined by a number of researchers.

emission rate ($gCH_4m^{-2}d^{-1}$)	mean annual flux ($g\ m^{-2}y^{-1}$)	fertilizer treatment ($kg\ ha^{-1}$)	reference
	210		1
0.15-0.18	42	140 kg ha^{-1} $(NH_4)_2SO_4$	2
0.22-0.28	25	220 kg ha^{-1} $(NH_4)_3PO_4$-$(NH_4)_2SO_4$ (16-20-0)+topdressing of 113 kg urea-N at planting time	3
0.1	12	160 kg N as urea + 40 kg N as NH_4NO_3 after tillering	4
0.2-0.58	54	various treatments/unfertilized	5
0.15-0.42	36.3	not documented	6
$8x10^{-3}$-$8x10^{-4}$			7
0.16-0.38	17-42	unfertilized, over 4 years	8
0.47-0.60	53-68	6-12 T straw+200 kgN urea/$(NH_4)_2SO_4$	8
0.1-0.3	35	$CaCN_2$	8
0.12-0.15	14-16	200 kg N as $(NH_4)_2SO_4$	8
0.18-0.21	19-22	100/50 kg N as $(NH_4)_2SO_4$	8
0.19-0.38	21-42	100-200 kg N as urea	8
0.23-0.68	24-77	3-12 T rice straw ha^{-1}	8

[1] Koyama (1963); [2] Cicerone and Shetter (1981); [3] Cicerone et al. (1983); [4] Seiler et al. (1984); [5] Holzapfel-Pschorn and Seiler (1986); [6] Holzapfel-Pschorn et al. (1986); [7] Minami and Yagi (1987); [8] Schutz et al. (1989).

Table 4.10 Release rates of CH4 from rice paddies per continent.

continent		irrigated	rainfed	CH4 release (g m-2d-1)**	growing period	Total (T g)
Asia	irrigated	64163		0.5-0.8	90-120	28.9-61.6
	rainfed, 0-30cm		32354	0.5-0.8	90-120	11.6-24.8
	rainfed, 30-100cm		11587	0.5-0.8	90-120	4.2-11.1
	rainfed, > 1m		5290	0.03	150?	0.2
Africa		3650	1820	0.5-0.8	90-120	2.3-4.9
N/C America			1914	0.3-0.6	120-150	0.9-1.7
S America		6122		0.5-0.8	90-120	2.8-5.9
Rest		1195		0.3-0.6	120-150	0.4-1.1
Total		**77044**	**51051**			**51-111**

* for rainfed rice a period of inundation of 80% of the growing period was assumed;
** from Schutz et al. (1989).

The estimated 1983 global emission from rice paddies based on the data in Table 4.10 is 50 to 110 T g CH_4 y^{-1}. In this estimate the areas of dryland rice for continents outside Asia are included. The annual average increase of paddy rice area of 1.05% results in an increase of the emission of 0.7 T g CH_4 y^{-1}. The 1989 release of CH_4 from paddy rice land is then 55 to 120 T g CH_4 y^{-1}. The resulting global emissions from rice paddies give a narrower range and as a whole are somewhat lower than estimates by e.g. Seiler in Bolle et al. (1986), whose estimate was 70-170 Tg y^{-1}, Aselmann and Crutzen (1989), who reported a range of 60-140 Tg based on temperature dependent emission rates, see Table 4.7) and Schutz et al. (1989). The latter authors estimate the global CH_4 flux from rice paddies to be 47-145 T g y^{-1}. The reason for the lower estimate in Table 4.10 is probably the separate calculation of rainfed and irrigated areas. In rainfed rice considerable extents have deep and intermediate water depths. Methane release rates are much lower from these fields than for the well managed shallow rice paddies where measurements have been made (see Table 4.9). Scharpenseel (1988, pers. comm.) estimated from organic matter additions and decomposition that the global CH_4 release from rice paddies is 110 Tg y^{-1}.

4.2.7 Geographic distribution of natural wetlands

Prior to estimating areal extents of wetlands, it is interesting to note that the scale of the data set used is of influence on the result. Van Diepen (1985) analysed the extents of soils with hydromorphic properties from the 1:5,000,000 Soil Map of the World (FAO,1978-1981) and from 1:2,000,000, 1:500,000, 1:200,000 and 1:50,000 maps respectively. The % hydromorphic soils were 0, 3, 9, 11 and 17-29 % respectively. This indicates the degree by which small scale maps may deviate from more detailed maps. On the other hand, wetlands or hydromorphic soils mapped as such on small scale maps may also contain aerobic or well drained soils. In that way unmapped wet soils in dry areas would be compensated.

Matthews and Fung (1987) distinguished 5 types of wetlands. The areas are presented in Table 4.11. Aselmann and Crutzen (1989) distinguished 6 types of natural wetlands including lakes. Although the total area corresponds well between the estimates, regional disagreement between the data suggests that the present knowledge of wetlands is still incomplete.

Table 4.11 Areas of wetlands by different authors.
(areas in $10^{10}m^2$)

a. Matthews and Fung (1987)	
Forested bog	208
nonforested bog	90
forested swamp	109
nonforested swamp	101
alluvial formations	19
total	526
b. Aselmann and Crutzen (1989)	
bogs	187
fens	148
swamps	113
marshes	27
floodplains	82
shallow lakes	12
total	569

It is important to realize, that part of the wetlands are only temporarily flooded and therefore are only active in methanogenesis during part(s) of the year. Matthews and Fung (1987) estimate the total area of wetlands at 2750×10^6 ha. Of the 2750×10^6 ha about 1383×10^6 ha is distinguished exclusively on Operational Navigation Charts (1:1,000,000 scale). The remainder, or 1367×10^6 ha, is the area where wetlands coincide on Matthews (1983), FAO (Zobler, 1986) and the ONC charts on the same location (corroboration cases). This figure corresponds with the 1283×10^6 ha of not cultivated hydromorphic soils and wetlands (Table 4.12). About 30% of this area is in the tropics, and this correlates well with the 35% calculated by Matthews and Fung (1987). However, a further analysis is required to check regional agreement between all data sources.

About 19% or 530×10^6 ha of the total global wetland area estimated by Matthews and Fung (1987) is actually inundated. The duration of the flooding period and the extension of the flooded area depend on the prevailing climatic and hydrologic conditions. Large parts of the marshlands in the Amazon and Congo Basins are only flooded during half of the year and even shorter flooding periods may occur in other swamp areas. Furthermore, considerable portions of the marshes may in reality consist of unvegetated open water, with lower CH_4 emission rates than vegetated areas (Cicerone et al., 1983). The actual area of inundation estimated by Matthews and Fung is 0.17 for wetlands derived from FAO Soil Map of the World to 0.32 for wetlands ditinguished on Matthews (1983).

Yet another, independent analysis made by Van Dam and Van Diepen, shows that the global area of soils with hydromorphic properties is 1261×10^6 ha. This figure was derived from dominant soils from the FAO Soil Map of the World (1971-1981). A higher figure would result if associated soils and inclusions would also be taken into account. And use of more detailed maps would probably also result in more extensive wetlands (Van Diepen, 1985).

In Table 4.12 wetland data from the FAO/Unesco Soil Map of the World are compared with data from other sources. The total extent of marshes and swamps is 210×10^6 ha (Matthews and Fung, 1987; and Table 2.2), a figure higher than that used by Bolle et al. (1986) who used 160×10^6 ha, but lower than the 260×10^6 and 250×10^6 ha given by Clark (1982) and Olson et al. (1983) respectively. Tundras occupy 695×10^6 ha (Matthews, 1983; and Table 2.2), the total extent of wetlands being 905×10^6 ha. However, Histosols (peat soils) cover a larger area than that covered by wetlands since those soils are not necessarily in swamps or may be drained. Other hydromorphic soils and potential sources of methane are the Gleysols and Fluvisols. For these soils the period during which anaerobic conditions prevail determines whether they are potential sources or not. Part of the Gleysols and Fluvisols are cultivated and either drained and not potential CH_4 sources, or are considered under wet paddy rice.

Table 4.12 Areas of soils with hydromorphic properties estimated from FAO/Unesco (1971-1981), Matthews (1983) and FAO (1983).

type	tropics	temperate	total area (10 10 m2)
marshes + swamps	120	90	210
other histosols (excl. those in marshes and swamps)	25	175	200
Gleysols	104	51	155
tundras		545	545
Other wet soils:			
Fluvisols (not cultivated)	102	71	173
total	**351 (27%)**	**932 (73%)**	**1283**
Fluvisols (cultivated)	78	47	125
Gleysols (cultivated)	52	93	145

4.2.8 Fluxes of CH_4 from wetlands

Harriss et al. (1982) found that sweet water peat soils in waterlogged conditions are a net source of methane to the atmosphere with seasonal variations in the emission rates of less than 0.001 to 0.02 g CH_4 m^{-2} day^{-1}. During drought conditions they found that swamp soils consume atmospheric methane at rates of less than 0.001 to 0.005 g CH_4 m^{-2} day^{-1}. This illustrates the complexity of processes which regulate the net flux of methane between wetland soils and the atmosphere. Their results raise questions concerning the generally accepted estimates of global methane emissions from wetlands.

Methane is usually found at very low concentrations in reduced soils if sulphate concentrations are high. Possible reasons for this phenomenon are (Mitsch and Rosselink, 1986):
1. competition for substrates between sulphate reducing bacteria and methanogens;
2. inhibitory effect of sulphate or sulfide on methanogens;
3. a possible dependence of methanogens on the products of sulfate reducing bacteria;
4. recent evidence (e.g. Valiela, 1984) suggests that methane may actually be oxidized to CO_2 by sulfate reducing bacteria.

Some authors suggested that methane emission is higher from fresh water environments than from saline water (Smith et al., 1982). A possible explanation for this is the lower sulphur concentration in freshwater environments so that sulphate reduction is less important in these environments. Other authors (Oremland et al., 1982) reported that competition between methanogens and sulphate reducing bacteria may not be important in salt marsh sediments due to the utilization of non-competitive substrates such as methanol and methylated amines or dihydrogen as electron donors (Wiebe et al., 1981). Williams and Crawford (1984) reported inhibition of methanogens by acetate. They suggested that methane and other products in peats at some depth may have been produced many years befors and that its presence is inhibitory to further methanogenesis.

In Table 4.13 rates of methane emission are compared. As in rice paddies the temporal and spatial variation is extremely high (Harriss et al., 1982; Harvey et al., 1989). Soil water content, temperature and other sesasonal climatological factors are all potentially critical factors in determining whether a wetland soil acts as a source or sink of atmospheric methane.

Recent global flux estimates for natural wetlands are 110 T g y^{-1} (Matthews and Fung, 1987) and 40-160 (Aselmann and Crutzen, 1989). This wetland source may change if estimates of geographic extents of different types of natural wetlands improve and the number of flux measurements increases, particularly in undersampled regions. Harriss (1988, pers. comm.) noted that future global warming in northern areas may increase flux rates from natural wetlands considerable due to increased phytomass production and accelerated fermentation.

Table 4.13 CH_4 emission rates from natural wetlands.

		10-3g m-2 d-1 range	mean	g m-2y-1	reference
FRESHWATER ENVIRONMENTS					
swamps					
swamp, waterlogged	Michigan, USA		110	40*	Baker-Blocker (1977)
cypress swamp, waterlogged	S. Carolina, USA	4.6 13	10	3.7	Harriss and Sebacher (1981)
cypress swamp, waterlogged	Georgia, USA	11 256	90	32.9	Harriss and Sebacher (1981)
cypress swamp, fertilized, waterlogged	Florida, USA		970	354	Harriss and Sebacher (1981)
cypress swamp, unfertilized	Florida, USA	8.2 265	67	24.5	Harriss and Sebacher (1981)
Dismal swamp, waterlogged	Virginia, USA	-20 1			Harriss et al. (1982)
Dismal swamp, drought conditions	Virginia, USA	-1 -5			Harriss et al. (1982)
tropical flooded forest	Manaus, Brazil	164 219	192	70*	Bartlett et al. (1988)
tropical flooded forest	Brazilian Amazon	2 760	110	40*	Devol et al. (1988)
marshes					
marsh, panicum spp.	Louisiana, USA		440	160	Delaune et al. (1983)
marsh, spartina cynosuroides	Virginia, USA		49*	18.2	Bartlett et al. (1987)
boreal marsh	Alaska, USA	101 111	106	39	Sebacher et al. (1986)
tropical floating grass mats	Manaus, Brazil	158 302	230	84*	Bartlett et al. (1988)
tropical floating grass mats	Brazilian Amazon	0 5200	590	215*	Devol et al. (1988)
bogs					
bog	Minnesota, USA	126 206	156	57*	Harriss et al. (1985)
bog	Minnesota, USA	19 174	47	17*	Harriss et al. (1985)
bog	Minnesota, USA	33 468	194	71*	Harriss et al. (1985)
fens					
fen	Minnesota, USA	3 5	3	1*	Harriss et al. (1985)
Fen + Bog	Minnesota, USA	60 1943	419	151*	Harriss et al. (1985)
shoreline fen	Minnesota, USA	158 171	165	60*	Harriss et al. (1985)
sedge meadow	Minnesota, USA		664	242	Harriss et al. (1985)
alpine fen	Alaska, USA	274 301	289	105	Sebacher et al. (1986)
subarctic fen	Quebec, Canada	19* 46*		0.1-0.6	Moore and Knowles (1987)
tundras					
coastal tundra, waterlogged	Alaska, USA	34 266	119	43*	Sebacher et al. (1986)
moist tundra, not waterlogged	Alaska, USA	0.3 12.5	4.9	2.0*	Sebacher et al. (1986)
meadow tundra, waterlogged	Alaska, USA	9.4 77.6	40	15*	Sebacher et al. (1986)
eriophorum (1987)	Alaska, USA			8.1	Whalen and Reeburgh (1988)
carex (1987)	Alaska, USA			4.9	Whalen and Reeburgh (1988)
intertussock (1987)	Alaska, USA			0.6	Whalen and Reeburgh (1988)
moss (1987)	Alaska, USA			0.5	Whalen and Reeburgh (1988)
eriophorum (1988)	Alaska, USA			11.4	Reeburgh (1989, pers. comm.)
carex (1988) (1988)	Alaska, USA			0.8	Reeburgh (1989, pers. comm.)
intertussock (1988)	Alaska, USA			3.9	Reeburgh (1989, pers. comm.)
moss (1988)	Alaska, USA			4.4	Reeburgh (1999, pers. comm.)
OPEN WATER					
tropical lake	Manaus, Brazil	-3 57	27	9.9*	Bartlett et al. (1988)
open water	Brazilian Amazon	0 830	120	43.8*	Devol et al. (1988)
tidal creek waters	Virginia, USA		2.2*	0.82	Bartlett et al. (1985)
SALT WATER ENVIRONMENTS					
salt marsh, tall spartina	Georgia, USA		1.2	0.4	King and Wiebe (1978)
salt marsh, interm.spartina	Georgia, USA		15.8	5.8	King and Wiebe (1978)
salt marsh, short spartina	Georgia, USA		145.2	53.1	King and Wiebe (1978)
salt marsh	Louisiana, USA		15	5	Smith et al. (1982)
brackish spartina patens	Louisiana, USA		200	73	Delaune et al. (1983)
salt spartina alterniflora	Louisiana, USA		12	4.3	Delaune et al. (1983)
salt meadow	Virginia, USA		1*	0.43	Bartlett et al. (1985)
short spartina alterniflora	Virginia, USA		4*	1.3	Bartlett et al. (1985)
tall spartina alterniflora	Virginia, USA		3*	1.2	Bartlett et al. (1985)
brackish Spartina cynosuroides	Virginia, USA		79*	29.0	Bartlett et al. (1987)
salt sp. alterniflore+cynosuroides	Virginia, USA		16*	5.6	Bartlett et ai. (1987)

* = calculated from data in the cited article.

4.2.9 CH_4 production by herbivorous animals

termites

In addition to deforestation and CO_2 evolution there are other indirect and compensatory effects of forest removal on the production of greenhouse gases, in particular of methane. Agricultural activities following deforestation, such as clearing, burning and cultivation, influence the activity and abundance of termites. Termites occur on about 68% of the earth's land surface (Zimmermann, 1982). Human activities such as clearing of tropical forests and conversion of forests to grazing land and arable land tend to increase the density of termites. The ecological areas that should have the largest methane emissions from termites are tropical wet savannas, areas that have been cleared or burned, and cultivated land in the (sub-) tropics.
Methane has been found in the guts of various xylophagous insects including scarab beetles, wood-eating cockroaches and various lower termites (Reticulitermes, Cryptotermes, Coptotermes). The digestion of these insects is primarily dependent on anaerobic decomposition by symbiotic bacteria in the higher termites (family Termiditidae) and by Protozoa in the lower termites (all other families). Their digestion efficiency is usually 60 % (Zimmermann et al., 1982).

A first estimate of the potential production of CH_4 by termites was made by Zimmermann et al. (1982). Their annual production of 150 T g CH_4 y^{-1} was based on the laboratory measured ratio of total gas evolved to food consumed by the termites. The species used were Reticulitermes tibialis, fam. Rhinotermitidae; Gnathamitermes perplexus, fam. Termitideae; Nasutitermitinae (unidentified), fam. Termitidae. Accompanying releases of other greenhouse gases are $5.4x10^{15}$ g C y^{-1} as CO_2 and $7x10^{11}$ g dimethylsulfide. Zimmermann et al. also calculated that with an assumed uncertainty of 50% and additional uncertainty in the significance of termites in the various ecosystems of the world, the methane emissions could range from 75 to 310 T g CH_4 y^{-1}. According to Zimmermann et al. (1982) the global area occupied by termites accounts for 68% of the earth's land area with 77% of the terrestrial net primary production of biomass. Zimmermann et al. estimate the world's termite population to be $2.4x10^{17}$. This population processes $33x10^{15}$ g dry weight of organic matter which is the equivalent of 28% of the earth's annual net primary biomass production and an average of 37% of the net primary production in areas where termites occur.

Rasmussen and Khalil (1983) found a methane production of 50 T g y^{-1} (ranging between 10 and 90 T g y^{-1}). They based their estimate on laboratory measured emission rates for the species Zootermopsis angusticollis only which is found in the American Rockies. They state that the uncertainty in the estimates of CH_4 emission by termites is in the production per termite and in the global number of termites and suggest that the disagreement with the data of Zimmermann et al. (1982) may be due to the different species observed. Their conclusion was that the CH_4 emission by termites is probably not more than 15% of the total global yearly emissions.

Seiler (1984) reported a much lower methane production by termites (2 to 5 T g CH_4 y^{-1}). This estimate was based on in vivo measurements with several species, including soil feeders, grass feeders, wood and dung feeders, grass harvesters and fungus grown termites. Seiler calculated the ratio of methane emitted to carbon ingested of $6x10^{-5}$ to $2.6x10^{-5}$ (depending on the species) and a used total consumed biomass of $7x10^{15}$ g dry matter (about 1/4 of the figure used by Zimmermann et al.,1982). Collins and Wood (1984) give a global figure for the dry matter consumption by termites of $3.4x10^{15}$ g y^{-1}.
Seiler also reported characteristic values for the CH_4 to CO_2 emission ratio which appear to be typical for each termite species. Seiler concludes, that since the bulk of termites lives in ecosystems not affected by humans (contrary to Zimmermann et al., 1982), it is unlikely that the total methane emission has changed significantly during the last decades.

Fraser et al. (1986) arrived at a global production of 14 T g CH_4 y^{-1} by termites (with a range of 6 to 42 T g). Fraser et al. measured in the laboratory (using methods described by Khalil and Rasmussen, 1983) the per termite methane production. The species used by Fraser et al. are Mastotermes darwiniensis, Nasutitermes exitiosus, Coptotermes acinaciformis, Coptotermes lacteus, Zootermopsis angusticollis and Coptotermes formosanus. All the above species are wood feeders, none are fungus builders.

The highest estimates of termite methane production may be exaggerated. Collins and Wood (1984) state that the subfamily of Macrotermitinae (fungus growing termites) is dominant in many ecosystems of the Ethiopian and Indo-Malayan regions. Since most of their digestion is performed aerobically by fungi, this subfamily is unlikely to produce much methane. Soil feeders, which occur in most tropical regions, use degraded soil organic matter. Therefore, if methane is produced by soil feeders, this is likely to occur in very small amounts.

Air turbulence occurring during the measurements is known to cause increased activity and higher CO_2 and probably CH_4 production (pers.comm Dr. O. Bruinsma, 1987). The lowest figures (Seiler, 1984; Collins and Wood, 1984; Fraser et al., 1986) are probably more realistic. Crutzen et al. (1986) estimated tentatively that the world's CH_4 emission by insects has an upper limit of 30 T g CH_4. One aspect not accounted for in all the experiments with termites is the possible microbial breakdown of CH_4 occurring in the soil of termite mounds.

ruminants
Crutzen et al. (1986) estimated that the CH_4 production by domestic animals is 74 Tg CH_4, with an uncertainty of 15%. Cattle contribute 74% (54 Tg), buffaloes (6 Tg) and sheep (7 Tg), the remainder stems from camels, mules, asses, pigs and horses. Humans produce less than 1 Tg CH_4. The world's wild ruminants may produce between 2 and 6 T g CH_4 y^{-1}. The total global emission from the domestic and wild animals is thus 66 to 90 Tg. Since the data in the case of livestock are comparably well documented, the CH_4 contribution from livestock is a fairly safe estimate.

4.2.10 CH_4 emission due to biomass burning

One of the major sources of abiogenic CH_4 is the methane formation during the burning of biomass such as agricultural wastes, savanna fires, burning due to shifting cultivation, etc. Crutzen et al. (1979) measured CH_4 to CO_2 ratios in several fire plumes and estimated the total CH_4 emission due to burning to be 25 to 110 T g y^{-1}. If data for the burning of agricultural wastes are included the resulting CH_4 to CO_2 ratio in fire plumes is 1:53 resulting in a global methane production of 53 to 97 T g CH_4 y^{-1} if the total amount of $48x10^{14}$ to $88x10^{14}$ g dry matter of biomass burned annually is applied (Seiler, 1984). Biomass burned in 1950 and 1980 is $37x10^{14}$ to $76x10^{14}$ and $42x10^{14}$ to $67x10^{14}$ g dry matter y^{-1}.

4.2.11 Oxidation of methane in dry soils

Only few measurents of the methane uptake in soils have been carried out. Soil methanotrophic bacteria can grow with methane as their sole energy source. Other soil bacteria which consume methane are e.g. Nitrosomonas species (Seiler and Conrad, 1987, quoting various authors). The uptake of methane occurs in well aerated soils. Harriss et al. (1982) observed that a methane emitting a swamp changed into a sink for methane after it dried up. Seiler (1984) quoting research data by Seiler et al. observed a destruction of methane at the soil surface in semi arid climates. The destruction rates varied between $3x10^{-4}$ and $24x10^{-4}$ g m^{-2} h^{-1} during the dry season with soil temperatures of 20 to 45°C. Keller et al. (1983) observed CH_4 uptake in temperate and tropical rainforest. For higher latitudes they reported loss rates of $1.2x10^{10}$ to $1.6x10^{10}$ molecules cm^{-2} s^{-1} with an average daily uptake of $2.5x10^{-4}$ g CH_4 m^{-2}. Methane decomposition was also observed at the surface of several types of soil in Germany. Seiler (1984) estimated, that the global methane decompostion must be at least 20 T g y^{-1}. Seiler and Conrad (1987) estimated a global methane consumption in soils of 32 ± 16 T g y^{-1}. Hao et al. (1988) measured methane fluxes in tropical savannas in the dry season. They found no methane consumption there, in contrast to the above literature.

4.3 CARBON MONOXIDE

4.3.1 Introduction

As was mentioned before, carbon monoxide does not interact in the atmospheric radiative balance, but it influences the concentrations of other atmospheric greenhouse gases such as CH_4, CH_3Cl, CH_3CCl_3 and $CHClF_2$ (F22). Moreover, the oxidation of CO is an important source of CO_2.

A growth of tropospheric CO concentrations would lead to a decrease of tropospheric OH. (Khalil and Rasmussen 1984a, 1984b, 1985) and an increase of ozone (see section 1.2.2). As the major tropospheric sink for many gases (esp. hydrocarbons and chlorinated hydrocarbons) is oxidation by OH., a rise in the CO mixing ratio would cause enhanced concentrations of these gases in the troposphere and a greater transfer to the stratosphere. The latter process would enhance ozone destruction in the stratosphere.

The major sources of CO are known, but their magnitudes are still uncertain as indicated in Table 4.14. The background concentration of CO is increasing at a rate of 2 to 6 % per year, but fluctuations of sources and sinks and the relatively short residence time of CO in the atmosphere make the estimates uncertain (Khalil and Rasmussen, 1984b).

4.3.2 Sources and sinks of CO

Table 4.14. Estimates of sizes of possible sources and sinks of carbon monoxide by various authors.

	T g CO yr^{-1} range	average	reference
sources			
vegetation	20- 200	110	Crutzen (1983)
	50- 200	130	Logan et al.(1981)
soils	3- 30	17	Conrad and Seiler(1985)
biomass burning	145-2015	660	Logan et al.(1981)
	240-1660	840	Crutzen et al. (1979)
	400-1600	800	Crutzen (1983)
oceans	20 -80	40	Logan et al. (1981)
fossil fuel burning	400-1000	450	Logan et al. (1981)
oxidation natural NHMC[1]	280-1200	560	Logan et al. (1981)
oxidadion anthropeg. NHMC[1]	0- 180	90	Logan et al. (1981)
oxidation of CH_4	400-1000	810	Logan et al. (1981)
		400	Khalil and Rasmussen, 1984a; 1984b)
sinks			
oxidation of CO to CO_2		3000	Crutzen (1983)
-do-	1600-4000	3170	Logan et al. (1981)
transportto stratosphere	190- 580	170	Crutzen (1983)
soil uptake	190- 580	450	Crutzen (1983)
-do-		250	Logan et al. (1981)

[1] NHMC = non methane hydrocarbons

Estimates of the sources and sinks of carbon monoxide made by various authors are listed in Table 4.14. The estimates for CO production due to biomass burning are, although quite similar, based on completely different statistics. Logan et al. (1981) applied clearing rates due to shifting cultivation of forest and woodland of $8x10^{10}$ to $36x10^{10}$ m^2 yr^{-1} and $5x10^{10}$ to $32x10^{10}$ m^2 yr^{-1}

respectively. Crutzen (1983) used a range of $21x10^{10}$ to $62x10^{10}$ m^2 yr^{-1} for burning due to shifting cultivation and $8.8x10^{10}$ to $15.1x10^{10}$ m^2 for deforestation due to colonization. Crutzen based his calculations on a ratio of CO/CO_2 production of 0.14. A recent study (Sachse et al., 1988) revealed increasing atmospheric CO concentrations over the Amazon, apparently from biomass burning.

With the natural NHMC (nonmethane hydrocarbons) are meant isoprenes and terpenes (C_5H_8 and $C_{10}H_{16}$ resp.), which are produced in forest environments.

Bartholomew and Alexander (1981) found, that absorption of carbon monoxide occurs in most soils. Dry soils, that are producing CO turn into a net sink of CO after irrigation (Conrad and seiler, 1982). CO absorption was stopped after heat-sterilization of the soil material, while the CO production was enhanced by heat-sterilization. Apparently the CO production is a chemical process, while the CO oxidation in soils results from microbial activity. Therefore, the CO producing soils can be found in arid and semi-arid zones, i.e. zones with predominantly Yermosols and Xerosols.

Since both CO production and oxidation occur simultaneously, it is very difficult to estimate the fluxes separately. Conrad and Seiler (1985) developed a technique to estimate the CO production and consumption rates at different soil temperatures and moisture contents. Measurements in arid subtropical soils demonstrated a strong dependence of CO production on the soil surface temperature, while the CO consumption was independent of surface temperature. This indicates that the CO production occurs at the surface while consumption occurs predominantly in subsurface layers at lower temperatures. In temperate climates where relatively humid soil conditions prevail, the CO production is insignificant and CO consumption very active. No data are available for humid tropical soils. Seiler and Conrad (1987) expected that soils in these regions are net sinks of CO. The global production of CO by soils is 17 T g y^{-1} (ranging from 3 to 30 T g of which 1 to 19 T g y^{-1} is produced in dry tropical areas). CO consumption ranges from 300 to 530 T g y^{-1} of which 70 to 140 T g y^{-1} is oxidized in the humid tropics (Seiler and Conrad, 1987).

The global sources range between 1270 and 5700 T g CO yr^{-1} with an average of 2920 T g. The global sink strength ranges between 1960 and 475 T g CO yr^{-1} averaging 3600 T g. The model is not completely balanced, indicating the uncertainty in the estimates given in Table 4.14.

4.4 CONCLUSIONS SECTIONS 4.2 AND 4.3

There is strong evidence, that the total CH_4 source has increased during the last decades. The annual increase of the area of paddy rice cultivation (1.23%) and the growth of the world population correlate well to the atmospheric rise of the CH_4 concentration. This indicates, that this increase is most likely related to anthropogenic activities (Bolle et al., 1986). The role of termites appears to be over-emphasized in the past.

The production of methane in soils and wetlands is extremely sensitive to environmental conditions. Therefore, as for the nitrogenous trace gases, the variation in time and space of methane fluxes is extremely high. The currently used methods of measuring methane fluxes are point measurements. Extrapolation of results from such measurements to smaller scales is fraught with potential errors. The development of methods to measure methane flux over larger, ecologically homogeneous areas would greatly improve the quantification of the methane sources. Remote sensing techniques are the most promising methods.

Since the a number of ecosystems are undersampled, more flux data is needed. There are a number of fields where the present knowledge is inadequate:
- the geographic distribution of soils used for wet rice cultivation;
- the role of soil parameters in the production and emission of methane;
- the intensity of rice cropping (number of crops per year);
- soil and water management and their relation to methane fluxes from rice paddies;

- the relation between type and quantity of organic and anorganic fertilizer applied to rice and the methane flux;
- the geographic distribution of the different types of salt and sweet water wetlands;
- the relation between the type of wetland and methane fluxes;
- other aspects such as water depth, temperature, influence of plants and their stage of development both for rice and natural wetlands;
- fluxes from landfill sites, especially quantities of organic waste which is decomposed anaerobically in landfill sites are virtually unknown. Furthermore quantities of methane lost from biogas installations in both developed and developing countries.

Such information, linked to intensified measurement of fluxes, should yield more reliable estimates of regional and global methane production than currently possible. A secondary result of the above investigations is the capability to assess consequences of climate change for methane fluxes, especially those from natural wetlands.

Soils and vegetation are no major sources of carbon monoxide. Soils, however appear to absorb CO in considerable quantities. So far estimates are no more than rough guesses. More research is needed in this respect.

4.5 NITROUS OXIDE (N_2O)

4.5.1. Introduction

N_2O is capable of absorbing infrared radiation, but it is inert in the troposphere. In the stratosphere N_2O is destroyed photochemically to N_2 and NO. The atmosphere contains about 1500 T g N_2O-N (Terra gram; 1 Terragram = 10^{12} g). McElroy and Wofsy (1986) report an accumulation of 2.8 Tg, and removal by stratospheric protolysis of 10.5 Tg. As the lifetime of N_2O in the atmosphere is 100 to 200 years, changes in production will have a long term effect. The following budget shows a slightly different figures for the stratospheric loss, but it gives a good overview of sources and sinks.

Table 4.15 Global budget of tropospheric Nitrous oxide (figures in T g N y^{-1})

Sources			
	fossil fuel burning	2	±1
	biomass burning	1.5	±0.5
	oceans, estuaries	2	±1
	fertilized soils	1.5	±1
	natural soils	6	±3
	plants	<0.1	
	gain of cultivated land	0.4	±0.2
Total production		14	±7
sinks			
	stratospheric loss	9	±2

Source: Seiler and Conrad (1987).

Kaplan (1984) and McElroy and Wofsy (1986) suggest that tropical forest soils are probably a major source with an annual emission of 7 to 8 T g N2O-N y^{-1}. Release of oxides of nitrogen from soils (N_2O, NO and NO_2) is known to occur during biological denitrification (see 4.5.2), chemical denitrification (4.5.3) and nitrification (4.5.4). Recent evidence has shown that nitrification is a source of considerable quantities of NO and N_2O. Other biogenic sources of the three gases will be discussed only briefly. The controlling factors of N_2O fluxes will be discussed in 4.5.5., their spatial variability in 4.5.6 and N_2O flux measurements from 4.5.7 onwards.

4.5.2 Biological denitrification

Biological denitrification is the dissimilatory reduction of nitrate (NO_3^{-}) or nitrite (NO_2^{-}) to gaseous froms of nitrogen by essentially anaerobic bacteria producing molecular N or oxides of N when oxygen is limiting (adapted from SSSA, 1984). Denitrification occurs only at low oxygen pressures. It has been widely accepted as the cause of poor efficiency of nitrogen use in flooded soils, but it may play a significant role in generally well aerated soils. Between 10 and 30% of the applied nitrogen is commonly lost by gaseous loss mechanisms of which denitrification is believed to be the major. However, recent research on NH_3 volatilization from flooded soils and research on gaseous nitrogen suggest that denitrification and nitrification proceed concurrently with NH_3 loss and that the relative importance of these two loss mechanisms may change substantially over very short periods (Simpson and Freney, 1986).

The general path of the reduction of nitrate during denitrification is (see e.g. McKenney et al., 1982):

$$NO_3^- \dashrightarrow NO_2^- \dashrightarrow NO \dashrightarrow N_2O \dashrightarrow N_2 + H_2O$$

The energy for these reactions is supplied by the decomposition of carbohydrates.

Many microorganisms are capable of reducing nitrate (NO_3^-) to nitrite (NO_2^-), but not all are able to denitrify. Most but not all denitrifying bacteria can reduce N_2O to N_2. Ammonia may inhibit the further reduction of NO_2^- by denitrifiers (see 4.5.3).

There is still uncertainty about NO being an intermediate product of denitrification (see e.g. Poth and Focht, 1985). The highly reactive nature of NO makes its detection difficult. NO and N_2O are gases - they may escape from the soil before being reduced. The ratio of N_2 to N_2O in the gases evolved from soil depends on such factors as soil pH, moisture content, redox potential (E_h), temperature, nitrate concentration and content of available organic C. Formation of NO and NO_2 by chemical denitrification of nitrite may occur during both nitrification and biological denitrification.

4.5.3 Chemodenitrification

Chemical denitrification is the reduction of nitrite or nitrate to gaseous forms of N by chemical reductants producing molecular N or oxides of N (adapted from SSSA, 1984).
High NO_2^- concentrations have been attributed to inhibition of the nitrite oxidation presumed to result from ammonia toxicity to Nitrobacter. The possibility that gaseous loss of nitrogen (via NO, N_2O or N_2) may accompany temporary NO_2^- accumulation has been noted by several investigators. High concentrations of NO_2^- are sometimes found in anaerobic soils where NH_3 or NH_4^+ type fertilizers are applied at higher doses (Stevenson et al., 1970). Nitrite accumulation is enhanced by the application of phosphate (Minami and Fukushi, 1983). Nitrite ions react chemically with organic molecules forming nitroso-group (-N=O) which are unstable. Gaseous products (N_2, N_2O) can be formed from nitroso-groups. Another reaction path followed is the dismutation of HNO_2 whereby NO or NO and NO_2 may be formed:

$$2\ HNO_2 \longrightarrow NO + NO_2 + H_2O \text{ (Slemr et al., 1984)}$$

or:

$$3\ HNO_2 \longrightarrow 2NO + HNO_3 + H_2O \text{ (Stevenson et al., 1970)}$$

For both reactions acidic conditions are required.

Keeney et al. (1979) found a nitrite accumulation at high temperatures in anaerobic systems and attributed this to thermophyllic species of Bacillus and Clostridium, which are nitrate respirers but not denitrifiers. The gas production consisted predominantly of NO in the closed anaerobic system. Keeney et al. concluded, that chemodenitrification reactions do not contribute significantly to N_2O production under anaerobic conditions.

4.5.4 Nitrification

Nitrification is the biological oxidation of ammonium (NH_4^+) to nitrite or nitrate (NO_2^-, NO_3^- resp.), or a biologically induced increase in the oxidation state of nitrogen (adapted from SSSA, 1984). Except for poorly drained or submerged soils, NH_3 formed through ammonification is readily converted to NO_3^-:

$$NH_3 \underset{-H^+}{\overset{+H^+}{\longleftrightarrow}} NH_4^+ \overset{\text{Nitrosomonas}}{\longleftrightarrow} NO_2^- \overset{\text{Nitrobacter}}{\longleftrightarrow} NO_3^-$$

where subreactions are:

$-1/2O_2$ $+1/2O_2$
$NH_4^+ \longleftrightarrow NH_2OH \longleftrightarrow NOH \longleftrightarrow NO_2^- \qquad \longleftrightarrow NO_3^-$
$-H^+$ $-2H^+$ $-H^+$
/ \
chemical denitrification biological denitrification
\ /
N_2O+NO

Evidence for the formation of NH_2OH as an intermediate was provided by Yoshida and Alexander (1970). Minami and Fukushi (1986) suggest that NH_2OH may react with NO_2^- in well aerated soils forming N_2O. The reaction may occur chemically as well as biochemically. The reaction path however, is not the main mechanism for N_2O production.

The energy released during the formation of nitrite (65 kcal) and nitrate (19 kcal) is used by the Nitrosomonas and Nitrobacter organisms for carrying out their life functions. The relative populations and their past histories and changing conditions in the soil (e.g. temperature) may introduce temporary accumulation of nitrite in the soil (C.C. Delwiche, 1987, pers. comm). There is a differential effect of temperature on the nitrifiers, with Nitrobacter being more sensitive to low temperatures than is Nitrosomonas. In cold conditions this may lead to a NO_2^- accumulation in the soil, which may have a toxic effect on plants.

Nitrifying organisms may contribute significantly to NO and N_2O emissions from soils. In experiments with nitrifying and denitrifying bacteria (Nitrosomonas europeae and Alcaligenes faecalis respectively) Levine et al. (1984) showed, that nitrification rather than denitrification is the primary biological process leading to the formation of N_2O and NO. The ratio of NO : N_2O for aerobic conditions (with low oxygen pressures as found in soil) found by Levine et al. was 2 to 1; in air the ratio was 0.13 to 0.29 and at an oxygen mixing ratio of 0.5% the ratio NO : N_2O ranged between 2 and 4. These results are similar to those reported for nitrifiers by Lipschultz et al., 1981. This ratio was found to be 0.01 in denitrification.

Bremner and Blackmer (1981) report that soils evolve N_2O even when the moisture content is low and are well aerated (i.e. under conditions known to inhibit denitrification) and that N_2O emissions from aerated soils are correlated not with NO_3^-concentration but with nitrifiable N-contents (see e.g. Minami and Fukushi, 1983). In well aerated soils emissions of N_2O are greatly increased by the addition of nitrifiable forms of N such as ammonia, urea, alanine, etc. (Bremner and Blackmer, 1978), but are not significantly affected by the addition of nitrate, glucose, or both. This finding agrees with e.g. Breitenbeck et al. (1980) and Seiler and Conrad (1981), but conflicts with amongst others, Mulvaney et al. (1984).

Several studies have indicated, that N_2O is a by-product of NH_4^+ oxidation as well as of nitrate reduction by heterotrophic microorganisms. Yoshida and Alexander (1970) showed that N_2O could be produced by Nitrosomonas europaea. Focht (1974) doubted this could actually occur in soils where present Nitrobacter bacteria would immediately oxidize NO_2^- to NO_3^- thereby limiting the possibility of formation of N_2O.

Poth and Focht (1985) found that Nitrosomonas bacteria produce N_2O under oxygen limiting conditions and that N from nitrite but not nitrate is incorporated into nitrous oxide. In their study nitrification was not a direct source of N_2O. Nitrosomonas is a nitrifier which, under conditions of oxygen stress, uses nitrite as a terminal electron acceptor to produce nitrous oxide. They refer to this process as nitrifier denitrification.
Denmead et al. (1979b) reported a simultaneous increase of soil NO_3^- and N_2O production after moistening the soil leading them to conclude that nitrification and N_2O production occur simultaneously. Parton and Mosier (1988) similarly report a simultaneous increase of NO_3^- and N_2O production and suggest that nitrification or nitrifier denitrification (see Poth and Focht, 1985) is dominant over denitrification until soils become very wet or saturated.

4.5.5 Factors controlling N_2O fluxes

Partial ogygen pressure
The partial oxygen pressure in the soil atmosphere depends on the oxygen exchange with the atmosphere. Thus the soil's oxygen status is closely related with the soil water status on one hand and oxygen consumption by plant roots and microorganisms on the other hand. Denitrification is negligible at soil moisture contents below about 2/3 of the water holding capacity, but is appreciable in flooded soils. The process may occur in anaerobic microsites within the otherwise aerobic medium in well drained soils, such as pores filled with water or sites within structure aggregates (Dowdell and Smith, 1974). Anaerobic microsites can also exist in locations of high microbial activity where oxygen is being consumed and CO_2 produced, thus making conditions for denitrification more favourable (Parkin, 1987).

A tendency for increased N_2O release with improved aeration status suggests, that the reduction of N_2O may be slowed (but not stopped) in soil at low oxygen pressures (Fillery, 1983). Competition for electrons, preferential inhibition of N_2O reduction (by N_2O reductase) or a general slowdown of the denitrification process (allowing N_2O to move away unreduced) are all possible explanations for this phenomenon. Alternatively, it might be argued that where the oxygen status of a soil is low, this is because the microbial demand for electron acceptors (of which oxygen is energetically the most favourable) outstrips the diffusive supply. N_2O may then act as an alternative electron-acceptor with diffusive properties little different from oxygen. The reduction of N_2O (and hence a reduction in the N_2O fraction) will be favoured by just those factors which lead to a low oxygen status. Another possibility is the stimulation of nitrification of NH_4 whereby N_2O may be one of the products.

Soil water status

Letey et al. (1980) found that the release of N_2O from the soil environment to the atmosphere is stimulated under fluctuating oxygen pressures (alternating drying and wetting cycles). When the soil is wetted N_2O will be produced more rapidly than it is reduced; if the soil dries fast enough, N_2O reduction to N_2 is prevented and rapid diffusion is possible. Mean N_2O fluxes were generally higher during periods of irrigation in summer (alternating dry and wet) than in periods of frequent rains in winter (continuously wet). Parton and Mosier (1988) saw N_2O and NO_3^- production increase with increasing soil water simultaneously at low soil water contents; at very high soil water contents only N_2O production increased on adding water.

Mosier et al. (1981, 1986) showed brief peaks in the N_2O flux after precipitation events and Denmead et al. (1979b) found a marked response of N_2O emission to small additions of water. This would indicate that N_2O production takes place close to the soil surface, where the essentially aerobic conditions are unlikely to favor denitrification. In contrast, Mosier and Parton (1985) report only low N_2O emissions after rain events and Cates and Keeney (1987a) found no N_2O flux peaks following precipitation events in three prairie soils. They attributed this to nitrification being the dominant process involved.

Probably nitrifying and denitrifying microorganisms both contribute to N_2O production in aerobic soils (Minami and Fukushi, 1987) whereby nitrification is dominant in the topsoil (Seiler and Conrad, 1981; Denmead et al., 1979) and denitrification in the subsoil during periods of high soil water content (Goodroad and Keeney, 1985).

Flooding

Terry et al. (1981) found, that the major gaseous product prior to flooding was N_2O, while the major product after flooding was N_2. These results support the conclusion of Denmead et al. (1979a) and Sahrawat and Keeney (1986) that flooded soils contribute less N_2O to the atmosphere than drained soils (see also Goodroad and Keeney, 1984 in Table 4.16). On the contrary, Minami (1987) found appreciable N_2O emissions from flooded rice fields. Colbourn and Harper (1987) showed that drainage decreases total production of nitrogenous gases, but the balance was shifted towards N_2O in autumn and winter in a clay soil. The net result was a higher N_2O evolution from drained soils (see Table 4.16 and 4.17).

Terry et al. (1980) report that N_2O emission by an organic soil is reduced to virtually zero after flooding. Drained histosols however, may contribute considerably to atmospheric nitrous oxide (see also data from Goodroad and Keeney, 1984 in Table 4.16). The data in Table 4.16 suggest that open water and undrained wetlands are only small sources of N_2O.

Diurnal and seasonal variation; influence of temperature

The optimum temperature for denitrification is about 25°C. In nitrification the production of NO_3^- decreases with temperatures below 35oC (Stevenson, 1986). Important losses may also occur at lower temperatures, especially in grassland above 8°C and when cattle slurry is applied, i.e. with an abundant and mobile C-source even at lower temperatures (S.C. Jarvis, 1987, pers. comm). Keeney et al. (1979) found, that while the rate of denitrification was low at temperatures below 15°C, the amount of N_2O (44-50% of the total gas production) was equivalent to that evolved at 25°C. Denitrification during the late autumn and early spring in temperate climatic zones could account for a significant portion of the N_2O released over the year, particularly in water saturated soils or soils with high water contents (Schmidt et al., 1988).

Apart from a spatial variability there is a strong diurnal variation of N_2O emission rates (Ryden et al., 1978; Denmead et al., 1979b; Keeney et al., 1979; Blackmer et al., 1980; Conrad and Seiler, 1983; Minami, 1987) and seasonal variability (e.g. Keeney et al., 1979; Bremner et al., 1980). Generally the amplitude of variation is greater at higher temperatures and with higher fertilizer gifts. This indicates that also the timing of measurements may influence the result to a great extent. Blackmer et al. (1982) reported that there is no single time during a 24 hour period that is always satisfactory for assessing the amount of N_2O evolved during that period.

Both the rate of emission and the form of the products of denitrification / nitrification depend on temperature. Several authors (Bremner et al., 1980; Duxbury et al., 1982; Goodroad and Keeney, 1984; Goodroad and Keeney, 1985; Schmidt et al., 1988) reported appreciable emissions during the spring thaw. N_2O trapped under the frozen layer may be released (Bremner et al., 1980; Goodroad and Keeney, 1985) but N_2O production during denitrification at low temperatures may also be considerable.

Distribution of soil organic matter

A good supply of readily decomposable organic matter for energy supply is a prerequisite for the occurrence and speed of nitrification and denitrification processes.

Chemical status of the soil

The rate of denitrification is low in acid soils and rapid under slightly alkaline conditions. Soil pH also profoundly affects the products of denitrification. N_2O reduction is reduced at a low soil pH. The effect of NO_3^- on denitrification is strongly interrelated with the soil pH. Inhibition of the N_2O reduction to N_2 occurs at all NO_3^- levels at low pH; at a higher soil pH the inhibition of N_2O reduction is temporary, although N_2O remains as a significant product for a longer period at the higher nitrate levels (Fillery, 1983). Possibly a delay of N_2O reductase occurs until a threshold N_2O level is reached (Fillery, 1983).

Yoshida and Alexander (1970) showed that both phosphate and high soil pH enhance N_2O production in cell suspension of Nitrosomonas europaea. Bremner and Blackmer (1981) and Minami and Fukushi (1983) demonstrated that N_2O production increases in soils treated with ammonium-yielding fertilizers. The latter authors also showed that in soils treated with ammonium, phosphate has an additional effect on the N_2O emission. Minami and Fukushi concluded that calcium carbonate and phosphate create favourable conditions for both enzyme activity responsible for N_2O production and the nitrifying population. Sahrawat et al. (1985) concluded that an elevation of pH in acid forest soils enhances nitrification, but the ratio N_2O produced to NO_3^- produced is not influenced or decreases after liming, depending on the soil properties.

Keller et al. (1988) observed increased N_2O production immediately after fertilizating a terra firme forest on Oxisols (Latosolo amarelo; Manaus, Brazil) with NO_3 and NH_4. Losses of applied fertilizer were much lower for the NH_4 fertilizer than for the NO_3 fertilizer (see Table 4.17). The reaction to P fertilization was only small due to fixation of phosphate in these tropical oxide rich soils.

Land use

Plants may reduce denitrification (and the N_2O fraction) by depleting the anorganic N-pool (Haider et al., 1985). Where there is little nitrate available microorganisms may be compelled to reduce more N_2O. On the other hand root exudates may stimulate denitrification and roots may also create anaerobic conditions by depleting oxygen in the rhizosphere. An important interaction exists between leaching of nitrate and denitrification, whereby the land use plays an important role. Stimulation of one process will reduce the other.

4.5.6 Spatial variability of N_2O fluxes

Spatial variability of N_2O emission from soils has been recognized by several investigators (Ryden et al., 1978; Rolston et al., 1978; Breitenbeck et al., 1980; Bremner et al., 1980; Mosier et al., 1981; Duxbury et al., 1982; Folorunso and Rolston, 1984; Colbourn et al., 1984; Goodroad and Keeney, 1985; Parkin, 1987; Colbourn and Harper, 1987). Duxbury et al. (1982) found that the spatial variability is reduced when fluxes were summed over important flux periods. The accuracy of the field measurements of N_2O emission is more limited by sampling methods than by analytical problems. Folorunso and Rolston (1984) calculated that in a Typic Xerorthent (USDA, 1975) 350 measurements are required to estimate the N_2O emission within 10% of the true mean in a 3 to 30 m plot. To achieve an accuracy of 50% the number of measurements should be 14 for N_2O. The variability however, is also dependent on the size of the chamber ("box") used in measuring field emissions. Contrary to the above variability, Conrad et al. (1983) found the spatial variability of N_2O flux insignificant compared to the effects of fertilization. They found the N_2O emissions variable but never more than a factor 4. Cates and keeney (1987a) reported that spatial variability of the N_2O flux from prairie soils is generally much lower than the values reported for agricultural soils.

High specific rates of denitrification may be associated with anaerobic microsites resulting from high O_2 consumption rates. This suggests that the patchy distribution of particulate organic matter or of the humus layer is a significant factor influencing the high spatial variability of N_2O fluxes from soils (Parkin, 1987; Schmidt et al., 1988).

4.5.7 N_2O emission rates; general

The presented studies have generated a considerable amount of data. The greatest difficulty in the assessment of N_2O emissions conditions is the extrapolation of the measurements to field conditions, because of the complex and time dependent interactions of temperature, microbial populations, supply of organic carbon, oxygen diffusion, water content, nitrate concentration and the root systems. Important parameters, which are very difficult to measure in the field are the amount of soil in a field profile that is anoxic, and the effect of soil organic carbon on microbial processes.
Several studies have not provided for frequent sampling of individual study sites or for intensive sampling after rainfall; most studies are only concerned with maximum N_2O fluxes after fertilization or after irrigation rather than average emission rates. In many cases even the source or process involved is not determined. Nevertheless, in the following sections an attempt will be made to assess the flux rates for soils and various land use types. As suggested by Bremner and Blackmer (1981) and many other researchers, well drained and well aerated soils must be considered in addition to waterlogged soils when N_2O emission rates are evaluated.

4.5.8 Natural undisturbed soils

Table 4.16 lists published data for N_2O flux from undisturbed natural soils.
Uundisturbed tropical moist forest areas may be a significant source of N_2O (Kaplan, 1984; Keller et al., 1986). Nitrous oxide released from soils near Manaus (Brasil) is much higher than the global average, reflecting the rapidity of the N-cycle in tropical areas. Keller et al. (1988) report that between 2.2 and 2.4% of all N in litterfall is emitted as N_2O in terra firme forest and Tabanuco forest respectively. The emission data in Table 4.16 suggest a variation of 10 to 30 μ g N_2O-N $m^{-2}h^{-1}$, or 0.09 to 0.26 g N_2O-N $m^{-2}y^{-1}$. These figures are much higher than those reported for tropical savannas and temperate forests and suggest that tropical forests are the major source of atmospheric N^2O.

Dry season emissions for savanna regions reported by various authors correlate well and range between 0.01 to 0.03 g N in a dry period of about 7 months. Assuming that the data provided by Hao et al. (1988) for a watered savanna soil are representative for wet season conditions, the wet season flux would be 0.06 g N_2O-N m^{-2} y^{-1}, giving a total of 0.07 to 0.1 g N_2O-N m^{-2} y^{-1}. Unfortunately no figures for different soils and climatic conditions are available.

Forests produce more N_2O than prairies. Goodroad and Keeney (1984) report lower emissions for deciduous forest than for coniferous forests. The data from Schmidt et al. (1988) are in line with Goodroad and Keeney (1984) and suggest that there is a marked influence of soil type and thickness of the humus layer. However, the spread is too large to generalize the presented figures. The two very poor paleo-tropical soils in the analysis by Schmidt et al. (1988) show lower flux rates than the other soils. The figures are similar to those presented by Goodroad and Keeney (1984) for deciduous forest.

In general undrained marshes show the lowest N_2O fluxes followed by native prairie soils, although Mosier et al. (1981) measured flux rates from prairies which are considerably higher than those reported by other authors. Drained marshes however may produce considerable quantities of N_2O. Goodroad and Keeney (1984) report fluxes of 65 to 149 μ g N_2O-N m^{-2} h^{-1}, Duxbury et al. (1982) presented extremely high fluxes for drained cultivated organic soils, with fluxes of between 180 and 1900 μ g N_2O-N m^{-2} h^{-1} (see Table 4.17).

4.5.9 Cultivated soils and fertilizer induced N2O losses

Table 4.17 shows a number of emission rates from cultivated soils by a number of researchers. As might be expected the results are highly variable. In Table 4.18 the losses calculated as a % of the fertilizer are listed. It is interesting to note that some authors measured negative fluxes in cases where the soil (Ryden, 1981; Ryden, 1983) or water (Terry et al., 1981; Minami and Fukushi, 1984) apparently acted as a sink of N_2O. In Tables 4.17 and 4.18 an effect of the type of fertilizer is not apparent.

Cultivated fields show a wide flux range of about 2 to 1880 μ g N m^{-2} h^{-1} considering all the figures in Table 4.17. Extremely high figures were presented by Ryden et al. (1978), Ryden et al (1979) and Ryden and Lund (1980). A common aspect of the latter measurements is that they were measured in soils kept at field capacity during the measurements. Other extreme high emission rates in Table 4.17 are for drained organic soils. Other extreme values were measured on fields receiving fertilizer gifts higher than 500 kg N ha^{-1}. On grasslands gifts of 450 kg N ha^{-1} are rare, but may occur in Western Europe.
If extreme values are omitted, the range in fluxes would be much smaller: about 10 to 50 μ g N $m^{-2}h^{-1}$ for unfertilized soils and 0.2 to 1.0 μ g N $m^{-2}h^{-1}$ for fertilized soils. A further analysis of the data presenteed in Table 4.17 on N_2O emissions from cultivated fields and in Table 4.18 on fertilizer induced N_2O losses is presented in the Annex to chapter 4.

Table 4.16 N2O emission rates from uncultivated and natural lands.

soil / texture	ecosystem	mean flux range µ g N m-2h-1		annual emission range (kg N ha-1)		method*	reference
TEMPERATE ECOSYSTEMS							
	uncropped land		2.0		0.2		CAST (1976)
fine sandy loam over clay	grassland (dry)		2.1		0.2	o -	Denmead et al. (1979b)
fine sandy loam over clay	grassland (moist soil)	25.0	104.0	2.2	9.1	o -	Denmead et al. (1979b)
fine loamy mixed Haplargid	native prairie (summer)		10.0		0.9	c -	Mosier et al. (1981)
loess loam	grass	0.5	2.5	0.0	0.2	c -	Seiler and Conrad (1981)
silt loam	native prairie	0.8	2.9	0.1	0.3	c -	Cates and Keeney (1987a)
loess over glacial till	burned tall grass prairie	2.0	2.0	0.2	0.2	c -	Goodroad and Keeney (1984)
loess over glacial till	unburnt tall grass prairie	3.0	2.0	0.3	0.2	c -	Goodroad and Keeney (1984)
sand	meadow	2.0	13.0	0.2	1.1	c -	Seiler and Conrad (1981)
?	wet meadow	31.0	31.0	2.7	2.7	c -	Goodroad and Keeney (1984)
Organic soil	drained marshes	65.0	149.0	5.7	13.1	c -	Goodroad and Keeney (1984)
Organic soil	undrained marsh	1.0	1.0	0.1	0.1	c -	Goodroad and Keeney (1984)
organic soil	undrained marsh	4.0	6.0	0.4	0.5	c -	Smith et al. (1983)
open water		1.0	4.0	0.1	0.4	c -	Smith et al. (1983)
loess	mixed forest	1.0	3.0	0.1	0.3	c -	Seiler and Conrad (1981)
sand, 1-3 cm humus layer	temperate deciduous forest	4.5	10.5	0.4	0.9	c -	Schmidt et al. (1988)
Grey-brown podzol, 1-5 cm humus layer	temperate deciduous forest	3.5	9.5	0.3	0.8	c -	Schmidt et al. (1988)
Pseudogley soil, 1-3 cm humus layer	temperate deciduous forest	5.5	75.0	0.5	6.6	c -	Schmidt et al. (1988)
Grey brown podzol, 1-2 cm humus layer	temperate deciduous forest	5.5	7.5	0.5	0.7	c -	Schmidt et al. (1988)
Typic Dystrochrepts(old tropical Acrisols)#	temperate deciduous forest	1.5	3.5	0.1	0.3	c -	Schmidt et al. (1988)
Typic/Dystric Eutrochrepts(Vertic Cambisols)#	temperate deciduous forest	2.5	4.5	0.2	0.4	c -	Schmidt et al. (1988)
loess over glacial till	deciduous forest	5.0	15.0	0.4	1.3	c -	Goodroad and Keeney (1984)
loess over glacial till	coniferous forest	28.0	36.0	2.5	3.2	c -	Goodroad and Keeney (1984)
TROPICAL ECOSYSTEMS							
sandy loam (60-66% sand)	tropical savanna, dry season		4.0		0.4	c -	Johansson et al. (1988)
sandy loam/sand	trop. savanna, dry season (gra	2.0	6.0	0.2	0.5	c -	Hao et al. (1988)
sandy loam/sand	trop. savanna, watered soil		16.0		1.4	c -	Hao et al. (1988)
Oxisol (Latosolo amarelo)	tropical forest (undisturbed)		30.0		2.6	c -	Keller et al. (1986)
Oxisol (Latosolo amarelo)	trop. secundary forest (20-30		25.0		2.2	c -	Keller et al. (1988)
Oxisols/Ultisols, ridges/slopes	tropical Terra firme forests		11.0		1.0	c -	Livingston et al. (1988)
Spodosols/Psamments, valleybottoms	tropical Campinarana forest		10.0		0.9	c -	Livingston et al. (1988)

*c = closed chamber; o = open chamber; l = N2 and N2O measured (C2H2 inhibition); 2= 15N labelling; - = N2O measurement exclusively

*** the 1st number refers to summer-autumn 1979; the 2nd number to March-November 1980 emissions

50% probability with 95% confidence range

present author's interpretations

Table 4.17 N_2O emission rates from grasslands and croplands in relation to the fertilizer type and application rate.

soil / texture	remarks	crop	fertilizer application kg N ha-1	mean flux range µg N m-2h-1		annual emission range kg N ha y-1		method*	reference
GRASSLAND / GRAZING LAND									
no fertilizer									
fine loamy, mixed Typic Hapludalf ?		grazing land	0		3.8		0.3	c -	Cates and Keeney (1987b)
clay loam		rye grass	0		14		1.3	c -	Breitenbeck et al (1980)
sandy loam over clay loam	stagnogley, 4.1% C	grassland	0	9	31	0.8	2.7	g	Eggington and Smith (1986b)
clay loam/silt loam		grass sward	0	9	10	0.8	1.0	c -	Webster and Dowdell (1982)
sandy loam over clay loam	stagnogley, 4.1% C	rye grass	0	15	66	1.3	5.8	g	Eggington and Smith (1986a)
sandy loam over clay loam	stagnogley, 4.1% C	rye grass	0	2.2	8	0.2	0.7	g	Eggington and Smith (1986a)
organic (peat) soil		grass	0	183	1107	16.0	97.0	c -	Duxbury et al. (1982)
NO3									
sandy loam over clay loam	stagnogley, 4.1% C	grassland	250	30	128	2.6	11.2	g	Eggington and Smith (1986a)
sandy loam over clay loam	stagnogley, 4.1% C	grassland	250	6	23	0.5	2.0	g	Eggington and Smith (1986a)
loam; Typic Xerorthent	constantly wet,23oC	rye grass	300	21	49	1.8	4.3	c 2	Rolston et al (1978)
silt loam	2.3 % C; well drained	rye grass	400	50	70	4.0	6.0	c -	Webster and Dowdell (1982)
clay loam	4 % C; drainage restricted	rye grass	400	70	90	6.0	8.0	c -	Webster and Dowdell (1982)
sandy loam over clay loam	stagnogley, 4.1% C	grassland	700		153		13.4	g	Eggington and Smith (1986b)
NH4NO3									
heavy clay, Denchworth s. (stagnogley)	Nov'79-June'80	grassland	210		7		0.6	c 1	Colbourn et al. (1984a)
loam, ochraqualf		rye grass	250		40		3.5	o 1	Ryden (1983)
loam, ochraqualf		rye grass	500		91		8.0	o 1	Ryden (1983)
Urea									
loam, fine loamy mixed Haplargid		grassland	450		10		0.8	c -	Mosier et al. (1981)
organic									
sandy loam over clay loam	stagnogley, 4.1% C	grassland	298	6	23	0.5	2.0	g	Eggington and Smith (1986a)
sandy loam over clay loam	stagnogley, 4.1% C	grassland	700		38		3.3	g	Eggington and Smith (1986b)
sandy loam over clay loam	stagnogley, 4.1% C	grassland	1230	30	128	2.6	11.2	g	Eggington and Smith (1986a)
CROPS									
no fertilizer									
loam, Ochraqualf	soil acted as a sink	weeds	0		-7		-0.6	c -	Ryden (1983)
Ustic Torriorthents		barley	0		14		1.2	c -	Mosier et al. (1982)
silt loam		alfalfa	0	26	48	2.3	4.2	c -	Duxbury et al. (1982)
silt loam		weeds	0	10	19	0.9	1.7	c -	Duxbury et al. (1982)
organic (peat) soil		sugar cane	0	180	550	7.0	48.0	c -	Duxbury et al. (1982)
organic (peat) soil		fallow	0	670	1880	59.0	165.0	c -	Duxbury et al. (1982)
NO3									
clay loam	high organic C (5%) (96 days)	uncropped	125		16		1.4	c -	Breitenbeck et al. (1980)
clay loam	high organic C (5%) (96 days)	uncropped	250		16		1.4	c -	Breitenbeck et al. (1980)
loam; Typic Xerorthent	constantly wet,23oC	uncropped	300	7	24	0.6	2.1	c 2	Rolston et al (1978)
loam; Typic Xerorthent	constantly wet,23oC	uncropped	300+	62	110	5.4	9.6	c 2	Rolston et al (1978)
fine loamy pachic Haploxeroll	irrigated	artichokes	430	220	310	19.6	26.9	o 1	Ryden and Lund (1980)
fine loamy pachic Haploxeroll	irrigated	lettuce-celery	620	230	480	20.2	41.8	o 1	Ryden and Lund (1980)
fine loamy pachic Haploxeroll	irrigated	cauliflower	680	310	330	26.8	29.2	o 1	Ryden and Lund (1980)
NH3/NH4									
fine monmorillonitic Aridic Argiustolls		corn	200		29		2.5	c -	Mosier and Hutchinson (1981)
clay loam	high organic C (5%) (96 days)	bare soil	125		24		2.1	c -	Breitenbeck et al (1980)
clay loam	high organic C (5%) (96 days)	bare soil	250		27		2.4	c -	Breitenbeck et al (1980)
fine loamy pachic Haploxeroll	irrigated	celery	335	70	105	6.1	9.2	o 1	Ryden et al (1979)
urea									
clay loam	high organic C (5%) (96 days)	bare soil	125		22		1.9	c -	Breitenbeck et al (1980)
clay loam	high organic C (5%) (96 days)	bare soil	250		27		2.4	c -	Breitenbeck et al (1980)
NH4NO3									
Ustic Torriorthents	sandy loam, 1% C	barley	56		25		2.2	c -	Mosier et al. (1982)
heavy clay soil, Lawford series	ploughed, 2 yrs.	****	70/140***	6	11	0.5	1.0	c -	Burford et al. (1981)
heavy clay, Denchworth s. (stagnogley)	ploughed, 2 yrs.	****	70/140***	10	64	0.9	5.6	c -	Burford et al. (1981)
heavy clay soil, Lawford series	dir. drilled, 2 yrs.	****	70/140***	17	24	1.5	2.1	c -	Burford et al. (1981)
heavy clay, Denchworth s. (stagnogley)	dir. drilled, 2 yrs.	****	70/140***	61	98	5.4	8.6	c -	Burford et al. (1981)
Ustic Torriorthents	sandy loam, 1% C	barley	112		28		2.5	c -	Mosier et al. (1982)
silt loam		corn	132	25	33	2.2	2.9	c -	Duxbury et al. (1982)
organic (peat) soil		sweet corn	170	860	1740	76.0	152.0	c -	Duxbury et al. (1982)
organic (peat) soil		onions	170	970	820	85.0	72.0	c -	Duxbury et al. (1982)
heavy clay, Denchworth s. (stagnogley)	Nov'79-June'80	winterwheat	210	19	23	1.7	2.0	c 1	Colbourn et al. (1984a)
Ustic Torriorthents	sandy loam, 1% C	barley	224		38		3.3	c -	Mosier et al. (1982)
organic									
Ustic Torriorthents	sandy loam, 1% C	barley	71		29		2.5	c -	Mosier et al. (1982)
		corn	130	27	43	2.4	3.8	c -	Duxbury et al. (1982)
fine loamy, mixed mesic Hapludalf	168/13 N in manure/NH4NO3	maize	181		41		3.6	c -	Cates and Keeney (1987b)
fine loamy, mixed mesic Hapludalf	168/13/56 N in manure/NH4NO3/urea; maize		237		59		5.2	c -	Cates and Keeney (1987b)
Ustic Torriorthents	sandy loam, 1% C	barley	356		113		9.9	c -	Mosier et al. (1982)
		routinely fertilized cropland		2	16	0.2	1.4	c -	Seiler and Conrad (1981)

*c = closed chamber; o = open chamber; 1 = N2 and N2O measured (C2H2 inhibition); 2= 15N labelling; - = N2O measurement exclusively;
g= gradient (diffusion) method
** in this experiment measurements were made immediately following irrigation
*** the first number refers to the period November 1977- June 1978; the second number refers to November 1978- June 1979 resp.; fertilizer applications 70 kg N and 140 kg N as NH4NO3 in spring 1977 and spring 1978 respectively. Winterwheat sown in 1977, oilseed rape in 1978.
**** winterwheat sown in October 1977, oilseed rape in October 1978.
fertilizer gift spread over 2 years

Table 4.18 N2O loss induced by fertilizer use as a % of the fertilizer gift for various types of fertilizer.

soil / texture	remarks	crop	Fertilizer kg N ha-1	induced loss % of fertilizer application	method*	reference
UREA						
silt loam	flooded	rice	90-180	0.01-0.05	c -	Smith et al (1982)
clay loam	high organic C (5%)	bare soil	125	0.14	c -	Breitenbeck et al. (1980)
clay loam	high organic C (5%)	bare soil	250	0.12	c -	Breitenbeck et al. (1980)
fine loamy mixed Ustollic Haplargid		shortgrass prairie	450	0.6	c -	Mosier et al (1981)
NH3						
silt loam; fine silty mixed Pachic Ultic Haploxerolls		fallow	55	0.05	o -	Cochran et al (1981)
silt loam; fine silty mixed Pachic Ultic Haploxerolls		fallow	110	0.07	o -	Cochran et al (1981)
fine montmorillonitic Argiustolls	Low org.C (0.76%)	corn	200	1.25	c -	Mosier and Hutchinson (1981)
silt loam; fine silty mixed Pachic Ultic Haploxerolls		fallow	220	0.09	o -	Cochran et al (1981)
NH4						
andosol		wheat	80	0.24	c -	Minami (1987)
alluvial soil		wheat	80	0.18	c -	Minami (1987)
andosol		wet rice	90	0.33	c -	Minami (1987)
sandy clay loam	loess	grass	100	0.025	c -	Conrad et al. (1983)
sandy loam	loess, 2-2.6 % C	meadow	100	0.03	c -	Seiler and Conrad (1981)
pararendzina (sandy clay loam)	loess	meadow, grass	100	0.053	c -	Conrad et al. (1983)
sandy clay loam	loess	clover	100	0.065	c -	Conrad et al. (1983)
sandy loam/sandy clay loam	loess loam, 0.8 % C	grass	100	0.07	c -	Seiler and Conrad (1981)
sandy loam	loess	grass	100	0.07	c -	Conrad et al. (1983)
sand		mixed forest	100	0.09	c -	Seiler and Conrad (1981)
brown soil, sandy loam	loess	beet, plants removed	100	0.153	c -	Conrad et al. (1983)
brown soil, sandy loam	loess	beet, plants removed	100	0.216	c -	Conrad et al. (1983)
alluvial soil		wet rice	100	0.33	c -	Minami (1987)
sandy clay loam	loess	grass	100	0.378	c -	Conrad et al. (1983)
andosol		wet rice	100	0.55	c -	Minami (1987)
andosol		rape	150	0.09	c -	Minami (1987)
alluvial soil		rape	150	0.08	c -	Minami (1987)
clay loam	high organic C (5%)	bare soil	125	0.18	c -	Breitenbeck et al. (1980)
clay loam	high organic C (5%)	bare soil	250	0.12	c -	Breitenbeck et al. (1980)
clay; Oxisols (Latosolo amarelo)		tropical forest	200	0.1	c -	Keller et al (1988)
andosol		carrot	200	0.26	c -	Minami (1987)
alluvial soil		carrot	200	0.31	c -	Minami (1987)
fine montmorillonitic aridic Agriustolls; clay loam		barley	200	0.4	c 2	Mosier et al. (1986)
fine montmorillonitic Aridic Argiustollsirrigated		corn	200	1.3	c -	Hutchinson and Mosier (1979)
fine montmorillonitic aridic Agriustolls; clay loam		corn	200	1.5	c 2	Mosier et al. (1986)
NH4NO3						
Ustic Torriothents	sandy loam, 1% C	barley	56	0.7	c -	Mosier et al. (1982)
Ustic Torriothents	sandy loam, 1% C	barley	112	0.4	c -	Mosier et al. (1982)
Ustic Torriothents	sandy loam, 1% C	barley	224	0.4	c -	Mosier et al. (1982)
loam; Ochraqualf		grassland	250	1.3	o 1	Ryden (1981)
loam; Ochraqualf		rye grass	250	1.4#	o 1	Ryden (1983)
sandy loam			336	0.25		McKenney et al (1980)
loam; Ochraqualf		rye grass	500	1.6#	o 1	Ryden (1983)
NO3						
sandy clay loam	loess	clover	100	0.001	c -	Conrad et al. (1983)
sandy clay loam	loess	grass	100	0.007	c -	Conrad et al. (1983)
sandy loam/sandy clay loam	loess loam, 0.8 % C	grass	100	0.01	c -	Seiler and Conrad (1981)
sand		mixed forest	100	0.01	c -	Seiler and Conrad (1981)
sandy loam	loess	grass	100	0.017	c -	Conrad et al. (1983)
brown soil, sandy loam	loess	beet, plants removed	100	0.018	c -	Conrad et al. (1983)
sandy loam	loess, 2-2.6 % C	meadow	100	0.05	c -	Seiler and Conrad (1981)
sandy clay loam	loess	grass	100	0.071	c -	Conrad et al. (1983)
pararendzina (sandy clay loam)	loess	meadow, grass	100	0.073	c -	Conrad et al. (1983)
clay loam	high organic C (5%)	bare soil	125	0.04	c -	Breitenbeck et al (1980)
loamy sand		grassland	200	0.11	c -	Armstrong (1983)
clay loam		grassland	200	0.15	c -	Armstrong (1983)
clay; Oxisols (Latosolo amarelo)		tropical forest	200	0.5	c -	Keller et al (1988)
clay loam	high organic C (5%)	bare soil	250	0.01	c -	Breitenbeck et al (1980)
sandy loam over sandy clay loam	stagnogley,4.1% C	grassland	500##	0.32-1.34#	g	Eggington and Smith (1986a)
ORGANIC MANURES						
fine loamy mixed Typic Hapludalf	168/13 kg N from manure/NH4NO3; maize		181	1.80#	c -	Cates and Keeney (1987b)
fine loamy mixed Typic Hapludalf	168/13/56 kg N from manure/NH4NO3/urea; maize		237	2.05#	c -	Cates and Keeney (1987b)
Ustic Torriorthents	sandy loam, 1% C	barley	71	0.8	c -	Mosier et al. (1982)
Ustic Torriorthents	sandy loam, 1% C	barley	356	1	c -	Mosier et al. (1982)
sandy loam over sandy clay loam	stagnogley,4.1% C	grassland	1528##	0.01-0.07#	g	Eggington and Smith (1986a)

*c = closed chamber; o = open chamber; 1 = N2 and N2O measurement (C2H2 inhibition); 2= 15N labelling; - = N2O measurement exclusively;
g= gradient (diffusion) method.
\# calculated with the data provided by the author(s)
\## fertilizer gifts spread over 2 years (Eggington and Smith, 1986a) and 2 years (Eggington and Smith, 1986b).

4.5.10 N_2O emission from other biogenic sources

The loss of N_2O due to the increase of land area used for agriculture may not be significant since the natural N_2O emission from tropical rainforests (where most clearing and conversion to cropland is taking place) is comparable to emissions from arable land. However, recent experiments indicate that tropical forest land cleared for pasture produces much more N_2O than the original forest (Vitousek and Matson, 1989, per. comm.).

N_2O emission due to biomass burning is estimated at 1 T to 2 T g N y^{-1} (Crutzen ,1983). The latter figure is a tentative estimate, since N_2O release rates, quantities of biomass burnt annually and conditions under which biomass is burnt are uncertain.

4.5.11 Global sources and sinks of nitrous oxide

Most studies presented in section 4.4 are most useful to gain insight in the processes of biological and chemical denitrification and nitrification in soils. However, field conditions are so variable and influenced by so many factors, that the presented data on N_2O fluxes from soils are very difficult to interpret and extrapolate to a smaller scale.

The total global emission from all possible sources including oceans and freshwater, fossil fuel burning and lightning is 12 - 15 T g N y^{-1} (according to Bolle et al., 1986), to and 14 ± 7 T g N y^{-1} (according to Seiler and Crutzen, 1987). An example of one budget is given in Table 4.15. There is evidence that tropical forests are the major N_2O source. Kaplan (1984) and McElroy and Wofsy (1986) suggest that tropical forests, in particular tropical wet and moist forests, are emit 7 to 8 T g N_2O-N annually. The use of fossil fuel as a source of N_2O is uncertain since Kranlich and Muzio (1988) reported that measurements made so far are incorrect. Probably the fossil fuel combustion produces considerably less than the 1.5±1 T g N2O-N y-1. The analysis in the Annex to this chapter suggests that cultivated fields emit 2.4 to 3.7 T g N2O-N y-1. The above figures would make the budget presented in Table 4.15 look quite different.

4.6 NITRIC OXIDE AND NITROUS OXIDE (NO_x)

4.6.1 General

Nitric oxide (NO) and nitrous oxide (NO_2) have no absorption bands in the infrared part of the spectrum. They may however, be involved in a number of atmospheric reactions, which affect the concentrations of other gases contributing to the greenhouse effect are influenced. NO and NO_2 catalyze the ozone destruction in the atmosphere. Furthermore, NO and NO_2 influence the OH. concentration. The latter radical is involved in other atmospheric chemical processes, e.g. the oxidation of CH_4. The major measurement techniques are chamber methods and micrometeorological techniques.

4.6.2 NO_x sources and sinks

The main sources of NO_x are combustion of fossil fuels (40%), biomass burning (25%), the balance coming from lightning and microbial activity in soils. The global sources are listed in Table 4.19.

Table 4.19 Global tropospheric sources of NO_x (10^{12} g N y^{-1})

surface sources	mean	range
1. fossil guel combustion	21	14 - 28
2. biomass burning	5.1	3.6- 6.7****
3. nitrification/denitrification in soils	8	4 - 16*
atmospheric sources		
1. lightning	8	2 - 20.*
2. NH_3 oxidation		1 - 10.**
3. from stratosphere	0.5	
4. jet aircraft	0.25***	
total	50	25 - 99

All data are from Logan (1983);* Levine et al. (1984) estimated biogenic production at 10 T g N y^{-1}; ** Levine et al. (1984) gave a source of 1.8 T to 18 T g N y^{-1}; *** Crutzen (1983) estimated that 10% of all atmospheric NH_3 is oxidated, i.e. 12-15 T g N y^{-1} (see section 4.7); **** Galbally (1985).

4.6.3 NO production in soils

Three mechanisms may lead to production of NO in soils: microbial denitrification and nitrification and chemodenitrification. Photolysis of nitrite has been suggested as a mechanism for NO production in flooded rice soils (Galbally et al., 1987). It is uncertain which mechanism is responsible for NO or N_2O production under field conditions. There is however evidence that nitrification is the main process in many soils (e.g. Slemr and Seiler, 1984; Johansson and Granat, 1984; Anderson and Levine, 1987; Kaplan et al., 1988; Johansson et al., 1988). The NO/N_2O production ratio is 1-5 by nitrifying bacteria (Lipschultz et al., 1981; Anderson and Levine, 1986) and 0.01 in denitrification.

4.6.4 NO emission rates

The extent of biogenic NO_x production is highly uncertain. Lipschultz et al. (1981) estimated the biogenic NO source at 15 T g N y^{-1} on the basis of laboratory experiments with nitrifying bacteria found in soils. Levine et al. (1984) used both nitrifying and denitrifying bacteria. They calculated, that soil nitrifiers rather than denitrifiers are the major source of NO and NO_2. Assuming a production ratio of NO : NO_2 equal to 2 : 1 at low oxygen levels as usual in soils, and a global N_2O production of 5 T g N y^{-1}, the estimated NO emission is 10 T g N y^{-1}. The authors call their estimate conservative. Using the global N_2O production from Table 4.15, the NO_x production would be much higher. The production ratio of NO:N_2O is a highly uncertain factor in all such calculations.

So far the number of NO measurements is low compared with the data set available for N_2O. In Table 4.20 a list of emission rates is presented. Most measurements show that NO from soils may contribute a large portion to the global source of NO_x in the atmosphere. Table 4.20 shows that site to site measurements differ by more than an order of magnitude. In some areas a net uptake of NO_x was measured. Most of this variation can be expained by extreme spatial, temporal and diurnal variability and possibly also by difficulties of measurement techniques and timing of measurements. From the figures in Table 4.20 it becomes clear that fluxes in the tropics are 3 to 30 times higher than in temperate ecosystems. As for N_2O fluxes, rates are highest in wet periods (Johansson and Sanhueza, 1988). Since NO_x fluxes are important at a local to regional scale, a global budget cannot be made. Moreover, the data available is too incomplete to present regional budgets.

4.6.5 NO_x deposition rates

The process of NO_x deposition to plants and uptake of NO_x by plants is not well understood so far. Probably both processes occur at a local scale only. NO_x deposition is measured with micrometeorological techmniques. The deposition rates show a large scatter. Galbally and Gillett (1988) estimated that NO_x is deposited within 1000 to 2000 km from the place where it was emitted, in the wet and dry season, respectively. Regional budgets for a number of continents for NO_x was presented by the latter authors.

Table 4.20 NO emission rates from cultivated and natural land in relation to climate

soil / texture	crop/ecosystem	treatment	flux (μ g N m-2 h-1) range		mean	reference
CULTIVATED LAND						
1. Temperate climates						
	cropland, Sweden	unfertilized	1.1	61.2	2.2	Johansson and Granat (1984)
	cropland, barley	120 kg N ha-1	36.0	68.0		Johansson and Granat (1984)
	cropland, USA	--	-32.4	100.8		Delany et al. (1986)
	cropland, USA	unfertilized	0.01	241.2	6.1	Anderson and Levine (1986)
	cropland, USA	fertilized			23.8	Anderson and Levine (1986)
	cropland, wheat	70 kg N ha-1	14.0	163.0	54.0	Duffy et al. (1988)
	cropland, sunflowers	80 kg N ha-1		234.0	20.0	Duffy et al. (1988)
	cropland, bare soil	70 kg N ha-1	-37	497.0	94.0	Duffy et al. (1988)
flooded soil	flooded rice, Australia	80 kg N ha-1 as urea	0.7	3.4		Galbally et al. (1987)
	grassland, Sweden	200 kg N ha-1	40.0	61.0		Johansson and Granat (1984)
various soils	grassland,Australia	grazed	5.4	26.3	12.6	Galbally and Roy (1978)
	grassland, Germany	100 kg N ha-1	-50	500.0	140.0	Slemr and Seiler (1984)
	pasture, U.K.	fertilized		129.6	28.8	Colbourn et al. (1987)
	sward, U.K.	unfertilized	-43.2	93.6	1.8	Colbourn et al. (1987)
2. Subtropical climates						
	bare soil, Spain	fertilized, 100 kg N	0	12600	796	Slemr and Seiler (1984)
NATURAL / UNCULTIVATED LAND						
1. Temperate climates						
various soils	ungrazed grassland, Australia		2.2	9.4	5.8	Galbally and Roy (1978)
	coniferous forest, Sweden		0.4	2.9	1.4	Johansson (1984)
	grassland, Germany		-21.6	50.4	7.9	Slemr and Seiler (1984)
	grassland, USA		0.1	234	10.8	Williams et al. (1987)
2. Tropical climates						
Oxisol (Latosolo amarainforest, rainy season			33.1	57.6	39.6	Kaplan et al. (1988)
	savannna, dry season		10.8	54.0	28.8	Johansson et al. (1988)
	savanna, wet season		7.2	900		Johansson and Sanhueza (1988)
	savanna, 1 year		34.0	148	36-234	Johansson and Sanhueza (1988)
	tropical cloud forest		0.4	7.2	1.8	Johansson et al. (1988)

4.7 AMMONIA (NH_3)

Ammonia (NH_3) can absorb infrared radiation, but its role in the radiative balance is not significant due to its short residence time. It is known however, as a major air pollutant with a major role in the process of acid precipitation.

4.7.1 Sources of NH_3

Possible sources of NH_3 are animal excretion, natural soils, mineral fertilizer use, biomass burning, coal burning and emissions during industrial N-fertilizer production.

- Industrial emissions occur mainly during the production of ammonia and fertilizers. Emission rates for NH_3 fertilizer production are 0.7 kg N/ton NH_3 produced. For NPK fertilizer production the emission is 10 kg N/ton N produced. The global annual N-fertilizer production is 74 T g N; with an assumed average emission of 4 kg N/ton fertilizer N, the global emission amounts $29x10^{10}$ g N y^{-1} as NH_3.
- coal burning. On the basis of an emission factor of $2x10^3$ g N per ton of coal and an annual use of coal of 3000 T g, the anual emission is 4 -12 T g N (Soderlund and Svensson, 1976).
- biomass burning. nitrogen in biomass is present in a reduced form largely as proteins. Crutzen (1983) assuming an N-content of 1% estimated the maximum NH_3 emission at 60 T to 70 T g y^{-1}. This figure is highly uncertain and represents only the upper limit.
- N-fertilizer use. About 30% of all N-fertilizer is produced as urea and 20% as salts of ammonia. The rest consists of various compound and mixed fertilizers. NH_3 emission depends on such factors as urease activity, soil temperature, soil moisture content, soil pH, rate of absorption at the cation exchange complex, wind speed, type of fertilizer and way of application. Emission generally decreases in the order $(NH_4)_2SO_4$, NH_4NO_3, $(NH_4)_3PO_4$. Since measurement of the NH_3 emission from fertilized fields is difficult, Crutzen (1983) assumed an average emission factor of 5% giving a global annual emission of 3.7 T g N.
- Natural soils. Denmead et al. (1976) showed, that in a situation of natural vegetation the leaves absorb almost all NH_3 emitted from the soil surface. There are no data for NH_3 emission from natural soils.
- animals. In areas with intensive animal husbandry the emissions of NH_3 have a local effect predominantly. Important NH_3 volatilization occurs both directly from deposited urine in the fields by grazing animals (Hutchinson et al., 1982) and also from the returns of stored slurry from housed animals. Only a small part is transported through the troposphere and most of the NH_3 is re-deposited (dry and wet deposition). Soderlund and Svensson (1976) estimate the total human and animal NH_3 emission at 20 T to 35 T g N y^{-1}.
- plants. Evolution of ammonia and amines from senescing plants has been reported by several authors. O'Deen and Porter (1986) found a crop loss of 1.6 kg N ha^{-1} accounting for 3.1% of the total N in the plant material. Too little data is available to make a safe global emission, however. The total global annual production of NH_3 ranges from 117 T to 150 T.

4.7.2 Sinks of NH_3

Crutzen (1983) estimated, that about 10% of all atmospheric NH_3 (12 T to 15 T g N) reacts with OH. to from NO or NO_2. The annual dry deposition is 72 T to 151 T and wet deposition amounts to 30 to 60 T g N.

4.8 CONCLUSIONS SECTIONS 4.4, 4.5, 4.6 and 4.7

The production of nitrogenous oxides in soils is highly dependent on environmental conditions. In the first place there is a diurnal variation, secondly a seasonal variation. Moreover, spatial variability has been reported to be extremely high. From all the presented data much can be said about the degree of variability. However, an estimate of fluxes of nitrous and nitric oxides for ecosystems or land utilization types cannot be given with sufficient reliability.

Extrapolation of data from current measuring methods to smaller scales is fraught with potential errors. Current measurements are point measurements. To give statistically reliable estimates of fluxes from ecologically uniform areas, other measurement techniques should be developed, such as remote sensing techniques and eddy correlation techniques.

Furthemore, on an ecosystem level, much can be learnt from the spatial variability of soil conditions such as soil drainage. The relation between soil morphometric properties, soil taxonomic classes and fluxes should be given more attention to improve estimates of these fluxes. For this the basic study of the cycles of C and N is fundamental.

Ammonia is probably only of local importance. And where high emissions of ammonia occur, its influence will be more through its acidifying effect than through absorption of infra red radiation since its residence time in the atmosphere is only very short.

4.9 MEASUREMENT TECHNIQUES FOR TRACE GAS FLUXES

With regard to instrumentation the measurements of trace gas fluxes the following division can be made:

- chamber techniques;
- tower techniques;
- aircraft techniques;
- satellite techniques;
- inversion technique through modelling of atmospheric circulation.

In the chamber technique a sealed box is placed on or inserted into soil. The confined atmosphere above the soil is sampled and its composition determined. For N_2O this method can be combined with acethylene (C_2H_2) inhibition of N_2O reduction in soil to measure total N-loss (N_2+N_2O) or with isotope techniques. Chambers may be closed, vented or with a continuous flow of air. In other methods the box is placed over the soil surface for short periods only to prevent errors occurring in open or closed systems. During the past few years measurements of trace gas fluxes have been made almost exclusively with chamber techniques.

Chamber methods are used for point measurements of fluxes. Ecological chamber methods yield the most detailed information on the underlying processes of trace gas fluxes. Since the spatial variability of trace gas fluxes is extremely high, chamber methods apply to very small areas only. Depending on the degree of variability and the size of the chamber the coverage is up to 10 m. Chambers can be used repetitively to extend their time coverage.

For both tower and aircraft techniques the following micrometeorological methods are available at present:
- for vertical fluxes: eddy correlation, flux gradient methods, Bowen ratio (energy balance) and eddy accumulation. The backgrounds of these methods are discussed in detail in Andreae and Schimel (1989).
- for measuring horizontal fluxes the mass balance method is used. In this method the horizontal flux from an emission area is measured across a vertical plane. It can be considered as a plume measurement (Denmead, 1983).

Tower measurements apply to areas of up to 100 m, while aircrafts cover spatial areas of up to 10-100 km. Aircrafts give averages over large areas, but over short time periods (up to 1 h). Aircrafts are costly in use. Examples of aircraft measurements are MacPherson et al. (1987); Austin et al. (1987). Towers enable measurements over prolonged time periods, but their disadvantage is their immobility.

In summary, chamber methods are used primarily for process studies. They can be used repetitively to increase temporal coverage and in array to extend spatial coverage. Towers address a larger scale than do chambers (in the order of 100 m) and can also be applied in array. But fundamentally both methods are suitable for process studies only.
Aircraft measurements apply to larger areas, but are limited in temporal coverage.

Another method is atmospheric inversion. This method relates atmospheric concentrations to source strngths through modelling of atmospheric circulations. An extensive review of all the methods in use at present can be found in Andreae and Schimel (1989).

Other Methods for measuring gaseous losses in general from soils are isotope methods, soil gas concentration gradient methods and mass balance methods:

- Direct measurement of ^{15}N-gaseous emission: in this method the isotope ^{15}N is used to label nitrogen additions to the soil and measure the ^{15}N labelled emitted gas. Direct methods using ^{15}N can be used only where substrate for denitrification is added at a high level of ^{15}N enrichment and require accumulation of evolved gases into a confined atmosphere. The costs of this method are high, but its sensitivity of estimating N-loss is greater as compared to methods not using N-tracer (Hauck, 1986).
- Nitrogen balance: deduction of denitrification losses from the balance of a nitrogen budget, accounting for crop uptake, soil residue and leaching. This method again needs ^{15}N and has the same characterisitcs as the first.

- Soil gas concentration gradients: measurement of vertical concentration gradients in the soil profile and diffusion coefficients of N_2O. The data presented by Seiler and Conrad (1981) and Denmead et al. (1979) suggest however, that N_2O production takes place at or close to the soil surface and that vertical N_2O profiles cannot be used for reliable determination of N_2O emission rates.

Annex to chapter 4

ANALYSIS LITERATURE DATA ON N_2O FLUXES FROM FERTILIZED FIELDS

Table 4.17 in chapter 4 and Figures 1 - 4 show emission rates from cultivated soils reported in literature. As might be expected the results are highly variable. It is interesting to note that some authors measured negative fluxes in cases where the soil (Ryden, 1981; Ryden, 1983) or water (Terry et al., 1981; Minami and Fukushi, 1984) apparently acted as a sink of N_2O. In Table 4.17 and Table 4.18 an effect of the type of fertilizer is not apparent.

Cultivated fields show a wide flux range of about 2 to 1880 μ g N m^{-2} h^{-1} considering all the figures in Appendix 1. Extremely high figures were presented by Ryden et al. (1978), Ryden et al. (1979) and Ryden and Lund (1980). A common aspect of the latter measurements is that they were measured in soils kept at field capacity during the measurements. Other extreme high emission rates in Appendix 1 are for drained organic soils. Other extreme values were measured on fields receiving fertilizer gifts higher than 500 kg N ha^{-1}. On grasslands gifts of 450 kg N ha^{-1} are rare, but may occur in Western Europe.
In Figure 1 the N_2O fluxes from both cropped fields and grasslands are presented. Apparently there is no correlation at all between the level of fertilizer application and the N_2O flux. In Figure 2 the data for grasslands are presented separately, with different symbols for the type of fertilizer and soil drainage. Most of the published flux data are for poorly drained soils, conditions which would favour denitrification. Apparently there is no good correlation between type or level of fertilizer gift and the N_2O flux. However, if extreme fertilizer applications of over 500 kg N ha^{-1} are excluded and the high flux data for 0 N fertilizer levels, the correlation seems to be better. However, it was decided not to calculate a regression line.

In Figure 3a and 3b the data for crops are graphed. Most of these data are for well drained plots, conditions to be known to favour nitrification. In cropped fields usually less than 200 kg N ha^{-1} is applied. In figure 3b the figures for N_2O fluxes for fertilizer levels of over 250 kg N ha^{-1} are excluded as well as N_2O fluxes from poorly drained cropped fields. Thus the vertical variation is much less. The mathematical relation which can be calculated with the method described by Snedecor and Cochran, (1980:165) from this reduced data set is:

$$N_2O = 1.878536 + 0.00417*N \qquad (1)$$

95% confidence limits are: $t_{0.05}*1.030366*\ [0.04 + \{N-126\}^2/170826]^{0.5}$

where: N_2O= N_2O emission (kg N_2O-N ha^{-1} y^{-1});
N= N fertilizer level (kg N ha^{-1}).
$t_{0.05}$= student t.

Figures 2 and 3a seem to indicate that the type of fertilizer is of only little influence on the resulting N_2O flux. If there is a relation, it is not apparent from the presented data. Figure 4 shows the data for grass and crops combined, excluding the fertilizer applications of over 500 kg n ha^{-1} for grass, of over 250 kg N ha^{-1} for crops and poorly drained soils used for crops. The regression line for this data set is:

$$N_2O = 1.454114 + 0.007496*N \qquad (2)$$

95% confidence limits are: $t_{0.05}*1.966012*\ [0.02 + \{N-164\}^2/1016329]^{0.5}$

FERTILIZER INDUCED N_2O LOSSES

The data presented in Table 4.19 have been graphed in Figure 5.

Loss of fertilizer N as N_2O usually occurs within a few weeks after fertilization. Some authors suggest that the emission rate is higher for ammonia yielding mineral fertilizers than for nitrate (Conrad et al., 1983; Breitenbeck et al., 1980; Bremner and Blackmer, 1981). This latter effect would be independent of the type of counter ion. However, this effect is not apparent in the data presented in Table 4.18. Conrad et al. (1983) concluded, that the level of fertilizer application does not influence the percentage N-loss. Other authors report lower % losses at for high fertilizer applications than for low gifts (e.g. Breitenbeck et al. (1980). Mosier et al. (1986) report a marked influence of the growth and development pattern of the crop on the loss of fertilizer N as N_2O.

Bolle et al. (1986) assumed average N_2O losses induced by fertilizer use of 0.04% for nitrate, 0.15-0.19% for ammonium and urea, and 5 % for anhydrous ammonia. These values would be independent of climate and should thus be globally representative. Bolle et al. (1986) assumed average loss rates of 0.5-2% based on fertilizer consumption figures. The same amount being emitted by N_2O flux from groundwater or surface water from leached fertilizer. Thus the total N_2O loss is 1-4% of the global 67 Tg of N-fertilizer consumption.

GLOBAL ESTIMATE OF THE N_2O FLUX

Based on the total production rates of mineral fertilizers, the global loss of mineral fertilizers according to Bolle et al. (1986) in the form of N_2O is estimated at 0.5 to 2.0%. With a total nitrogen fertilizer consumption of about 67 T g in 1985 (FAO, 1985), the total N_2O emission due to application of nitrogen fertilizers amounted to 0.7 to 3.0 T g N y^{-1}. However, conditions under which N_2O is lost from fertilizers are as variable as the conditions in cultivated fields. The rates at which fertilizer N is applied is usually less than 200 kg N ha^{-1}. At lower application rates the % loss shows a lower range of 0.1 +/- 0.08% independent of the type of fertilizer. In that case the 1985 loss induced by global N-fertilizer use would be 0.02 to 0.15 T g y^{-1}. In addition to this N_2O loss there is a basal or natural N_2O emission from the cultivated lands. This basal emission is 1 to 2 kg N_2O-N ha^{-1} y^{-1}. With a cultivated global area of 1500x10^6 ha the total emission is 1.5 to 3 Tg. Adding this to the fertilizer induced loss yields 1.52 to 3.15 Tg N_2O-N y^{-1}. However, this basal estimate may as well be higher. Therefore, this the reliability of this estimate is only low.

The solubility of N_2O in water is relatively high. Considerable N_2O fluxes may occur from surface water draining from fertilized agricultural fields (Dowdell et al., 1979) and even cleared forests (Bowden and Bormann, 1979). Minami and Fukushi (1984) measured losses as N_2O of 200 to 1190 μ g N m^{-2} h^{-1} in drainage water when calculated for the total draining surface. Minami and Fukushi also demonstrated that at low N_2O concentration standing water in rice paddies may act as a sink for N_2O. Thus, an amount comparable to the N_2O gas emission may be lost through denitrification / nitrification of mineral fertilizers leaching from fields into groundwater or surface freshwater ecosystems (Bolle et al., 1986).

The above relations and statistics on N-fertilizer use (FAO, 1987) have been used to estimate the global N_2O flux from cultivated fields in Table 4.2.1 below.

Table 4.2.1 Global Nitrogenous fertilizer consumption, land use data and estimated N_2O emission.

	Total N use (10^9 kg)	arable land* 1000 ha	pasture 1000 ha	kg N/ha arable land*	kg N/ha agric. area**	equation 1**** crops arable land*
Africa	1874271	183214	779134	10.2	1.9	0.24 -0.46
N & C America	12842574	274122	359590	46.8	20.3	0.46 -0.68
S America	1072937	138878	456431	7.7	1.8	0.18 -0.35
Asia***	25362015	455220	644349	55.7	23.1	0.80 -1.13
W Europe	10331546	87079	64054	118.6	68.4	0.20 -0.21
E Europe***	5266801	53476	21504	98.5	70.2	0.11 -0.13
USSR	10292000	232290	373133	44.3	17.0	0.38 -0.58
Oceania	295405	48182	455157	6.1	0.6	0.06 -0.12
World	66906529	1472401	3153352	45.4	14.5	2.44 -3.65

sources: N-fertilizer consumption: FAO fertilizer yearbook vol. 34, 1984;
Land cover data from World Resources Institute (1988).
* arable land includes permanent crops.
** agricultural area includes arable land, permanent crops and permanent pastures.
*** excluding USSR.
**** equation 1 is applied to data for crops only and the arable land area.

Equation 2 (not used in Table 4.2.1) is used with data for the total agricultural area (arable land, permanent crops and pastures), giving an average annual emission of 5.6 Tg N_2O-N. It must be realized that much of the area in the permanent pastures is natural, uncultivated land. Fluxes for these lands are thus probably much lower than for cultivated lands. The last two colums show equation 1 used for estimating the emission from arable land + permanent crops. This figure is probably a better estimate of global N_2O emission from total cultivated lands, since crops receive higher doses of fertilizer except for Western Europe, where grasslands may also receive very high N- applications. This is however, not general. The estimate of 3.0 Tg N_2O-N for the annual emission from cultivated land is probably more realistic than the above higher figure. Moreover, it correlates well with the estimate derived from the fertilizer induced N_2O losses.

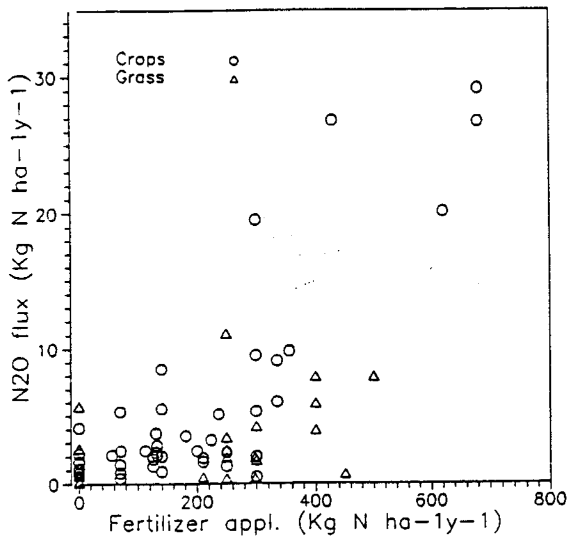

Figure 1. The effect of the level of N-fertilizer application in cultivated fields. In Figure 1 all the available data for crops and grass are presented.

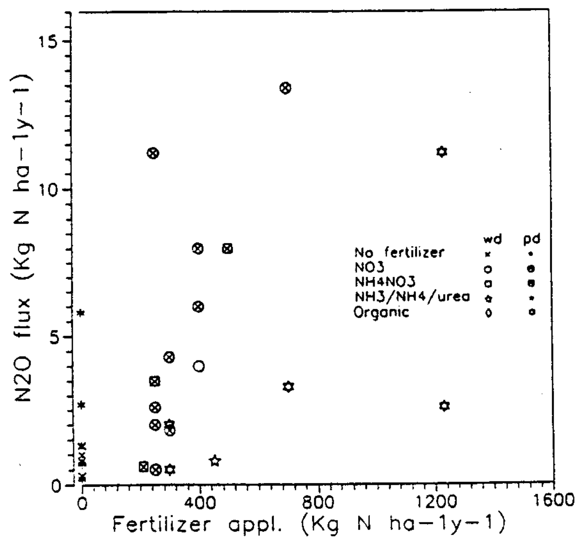

Figure 2. The effect of the level and type of N-fertilizer application on N_2O emission for grasslands only.

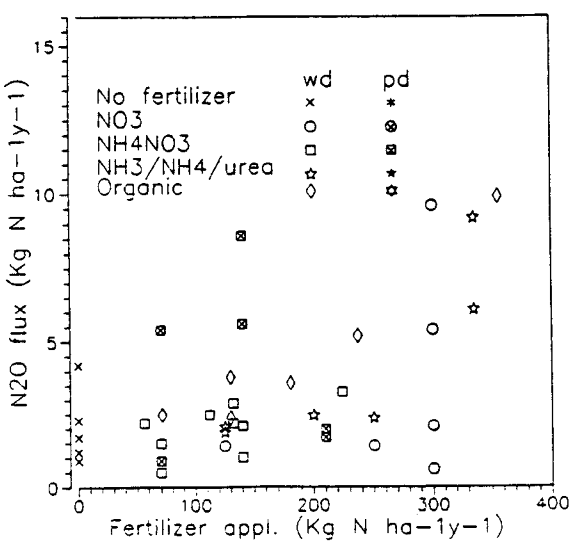

Figure 3a. The effect of the level and type of N-fertilizer application on N_2O emission for crops only.

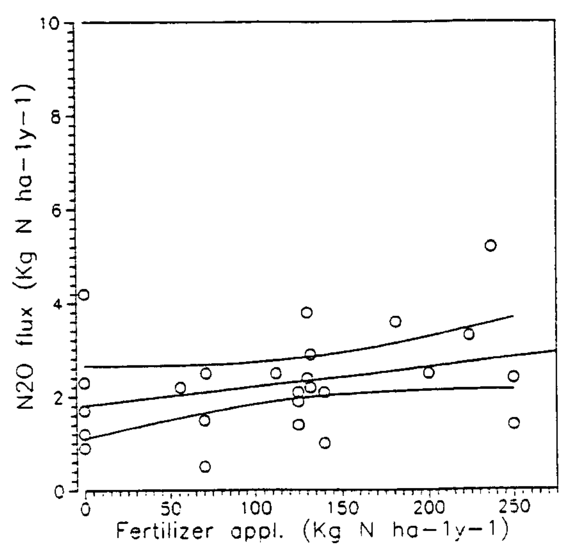

Figure 3b. The effect of the level of N-fertilizer application on N_2O emission for crops only. The data for fertilizer application of over 250 kg n ha^{-1} and data for poorly drained soils are excluded. The curves show regression line and confidence interval on the basis of this reduced data set.

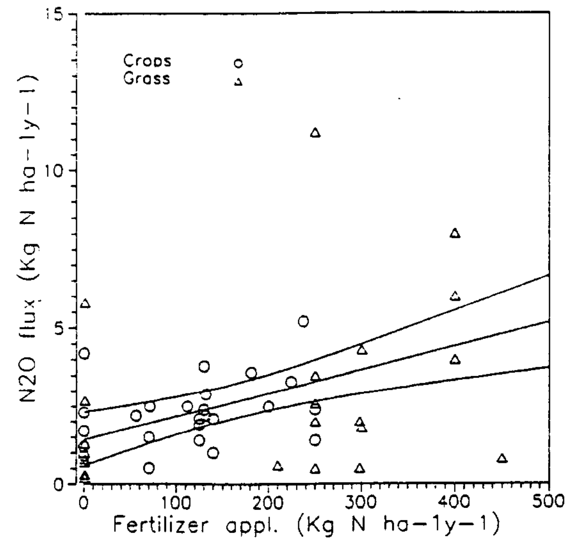

Figure 4. The effect of the level of N-fertilizer application on N_2O emission for crops and grass. Data for fertilizer applications of over 500 kg n ha^{-1} for grass, of over 250 kg n ha^{-1} for crops and data for crops on poorly drained soils are excluded. A regression line with confidence interval is drawn on the basis of this reduced data set.

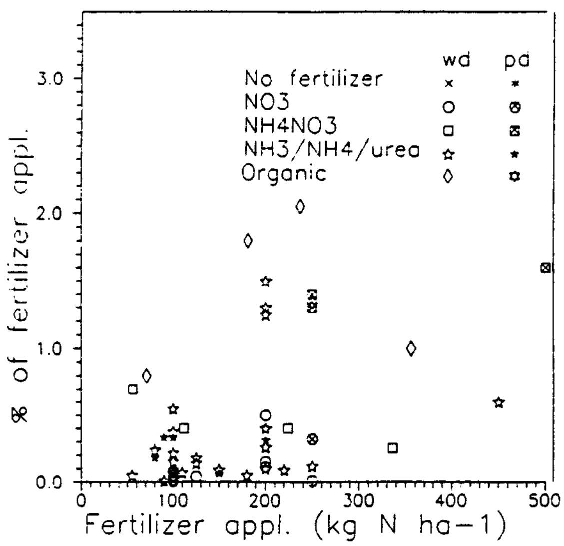

Figure 5. The effect of fertilizer level on the N_2O loss as % of the fertilizer application.

5. THE IMPACT OF CHANGING LAND COVER ON HYDROLOGIC PROCESSES

5.1 GENERAL

The role of vegetation, in particular forests, in the hydrologic cycle is multiple. Forests protect the soil surface against rainfall (and wind) erosion and they improve soil infiltration and thus reduce the rate of run-off. Furthermore forests transpire and evaporate water through their stomata into the lower atmosphere. Water vapour has strong absorption bands in the infrared region of the spectrum and is therefore an important greenhouse gas.
The vapour produced may form clouds after condensation. There is evidence that forests recycle up to 50% of all precipitation.

Evapotranspiration closes the loop in the hydrological cycle over land by returning water to the atmosphere and, correspondingly, returns an amount equal to about 20% of the solar energy absorbed in the atmosphere plus surface over land (Dickinson, 1986). Evapotranspiration also controls the disposition of rainfall. Part of the water balance studies have been directed towards investigations of evapotranspiration, particularly for determining a "correct value" of evapotranspiration for the area in question. In certain cases the water balance method has been supplemented with an attempt to quantify the energy budget of the watershed and to compare values of evapotranspiration obtained with the two methods.

When the actual evapotranspiration equals the potential rate, the values derived from water balance studies can be compared directly to values obtained from e.g. the Penman formula (Penman, 1948) for evapotranspiration. This contrasts strongly with conditions in dry areas, where the actual rate of evapotranspiration falls far below potential. In those cases an analysis of the surface energy balance is required to complement the water budget.

The water balance equation for a certain area can be written in general form as (according to e.g. Pereira, 1973):

$$E = R - Q - dS - dG - L \qquad (5.1)$$

where: E = evapotranspiration
R = precipitation
Q = streamflow
dS = change in water stored in the root zone
dG = change in water stored below the root zone
L = net loss of groundwater other than the streamflow

Equation 5.1 neglects possible effects of the type of land cover on the recurrence of rainfall and other feedback mechanisms that exist between e.g. soil moisture and vapour pressure deficit on one hand and evapotranspiration on the other hand. Assuming that on the long run there are no changes in soil water storage in the root zone and no loss of groundwater, three variables remain to be quantified: the evapotranspiration, streamflow or runoff and rainfall. This simplified form of equation 1 shows that there are two approaches of studying the effects of land cover changes on the hydrologic cycle. The first one considers the changes in surface runoff and in the second method the evapotranspiration is measured.
An attempt will be made in the next paragraphs to describe the possible effects of land cover changes, particularly deforestation, on the water balance through changes in the amount of vapour emitted as evapotranspiration into the lower atmosphere.

5.2 CHANGE OF RUNOFF DUE TO DEFORESTATION

One way of examining possible changes in the hydrologic cycle due to deforestation is the study of runoff rates and the incidence of floods and extreme discharges. Forests have generally been held responsible for reductions both in the water yield due to higher evapotranspiration losses and in the flood peaks due to reduced storm flows.

In a review of catchment experiments Bosch and Hewlett (1982) concluded, that no reduction in run off is found when the vegetation cover (e.g. by logging) is removed. Furthermore, they suggest the following:
- coniferous and eucalypt cover types give an increase of about 40 mm in water yield per 10% reduction in cover.
- deciduous hardwoods give a 25 mm increase in water yield per 10% cover reduction.

The above conclusions were drawn on the basis of analysis of 94 catchment experiment. Moreover, Bosch and Hewlett (1982) saw, that the effect of clearcutting of forests is greatest in high rainfall areas, but the effect is shorter due to more rapid regrowth.

Forests impose a positive effect on soil structure. This phenomenon and the sponge effect of the forest litter layer cause an increase of the infiltration. Forest harvesting appears to result in some cases in an increase in water table (e.g. Peck and Williamson, 1987; Burch et al., 1987) until the regrowing forest has resumed precutting levels of interception and evapotranspiration (literature review by Hamilton and King, 1983). Reduction in canopy area results in reduced evapotranspiration, increased throughfall due to decreased interception. Precipitation reaching the forest floor will either infiltrate the soil, contribute to surface runoff or evaporate.

Conversion of forests to annual cropping changes the annual water balance. Annual crops make a direct demand on water resources and seasonally deplete soil moisture reserves.

Many studies indicate, that the water outflow is increased after conversion of forest to annual cropping, particularly in areas of high precipitation (Russel, 1981; Edwards and Blackie, 1981). Replacement of rainforest by tea in Kenya resulted in reduced water use. The runoff however, was not influenced (Edwards and Blackie, 1981). The principal cause for low water discharge in rivers and streams from forested areas relative to areas with pasture is the high rate of evapotranspiration from wet forest canopies. Interception expressed as a fraction of the annual precipitation decreases with increasing rainfall (Calder and Newson, 1979). In their study Calder and Newson found that, assuming 50% canopy coverage, the increase in loss of water outflow amounts to 43% to 77% in relation to grazing land. They used a simple evapotranspiration model. The increase in loss is greatest in areas of high rainfall; but as runoff also increases in areas of high rainfall, the reduction in water outflow is conservative.

Morton (1984) compared water budget estimates of evapotranspiration for a number of large basins in agricultural areas with those for forested areas in two regions in Canada. He found that the runoff from forested areas was less than that from agricultural areas. The generally accepted idea that forest evapotranspiration exceeds water use by agricultural crops is not necessarily true however, for all situations.

Very little is known on the effect of shifting cultivation on runoff rates, but many authors agree on a negligible effect of shifting cultivation as long as the cultivation system is extensive (Hamilton and King, 1983).

The amount of water, that evaporates is further influenced by:
- The process of clearing, which has a decreasing effect on the soil's infiltration rate (Lal, 1981).
- Type of vegetation replacing the forest: infiltration rates of rainwater in pasture are considerably lower than in primary forests (Salati et al., 1983).

An interesting case for discussion is the Amazon basin, where deforestation is continuing at a high rate for some decades. The removal of the Amazon jungle can be expected to produce extreme local and regional climatic effects. The area of the Amazon is about 700 million ha, its water outflow is

$5.5*10^{12}$ m^3 yr^{-1}, which corresponds to only 80 cm of rainfall per year. The water outflow contributes about 20% of the total fresh waterflow on the earth (Friedman, 1977).

Gentry and Parody (1980) report an increasing height of the flood crest of the river Amazon at Iquitos during the period 1970-1978. During that period there had been no significant changes in precipitation, so they concluded that the change in Amazonian waterbalance is the result of increased runoff due to deforestation in the Peruan and Ecuadorian part of the Amazon basin. Their conclusions were debated by Nordin and Meade (1982), who conclude that for the Rio Negro at Manaus (Brasil) during 1942-1956 river flood stages were high, between 1957 and 1969 average flood stages were low, while between 1970 and 1979 the measured flood stages were high again. Their conclusion was, that statistically no increased runoff and consequent flood stages occur at Manaus. A review of the literature on increased runoff and increased erosion hazard due to forest clearing is given by Lal (1987).

5.3 CHANGE IN TOTAL EVAPOTRANSPIRATION DUE TO DEFORESTATION

5.3.1 General

In considering the world's major land use changes, i.e. deforestation and consequently conversion to agricultural land, the forest evapotranspiration should be compared to the vapour production from agricultural lands. Forest total evapotranspiration is the sum of dry canopy evaporation (transpiration, a physiologically controlled process) and wet canopy evaporation of intercepted rainfall. Recent investigations (e.g. Shuttleworth and Calder, 1979) have shown that wet canopy evaporation from forests is of the same order of magnitude as dry canopy evaporation. Both quantities will be discussed separately in the following sections.

5.3.2. Models for describing dry canopy evaporation

Dry canopy evaporation or transpiration is regulated by the opening and closing of stomata. The stomata respond not only to internal stress caused by constraints in the water supply from the soil, but also to external environmental factors such as solar radiation, temperature, vapour pressure deficit and carbon dioxide concentration. Below the Penman-Monteith equation and the Priestley-Taylor method for estimating dry canopy evaporation will be discussed. Both methods describe the process of or estimate dry canopy evaporation under conditions of non-limiting water supply, respectively.

Penman-Monteith equation
The latent heat flux (LE) or dry canopy evaporation is commonly described as a combination of the energy balance and vapour and heat transport equations (Penman-Monteith equation; Monteith, 1965):

$$LE = \frac{s(R_n - G) + \rho\, C_p\, (e(T_z) - e_z)/r_A}{s + \gamma\, (1 + r_C/r_A)} \qquad (\mathrm{W\ m^{-2}}) \tag{5.2}$$

where: E =evaporation(kg m^{-2})
L =latent heat of vaporization(J kg^{-1})
R_n=net radiation(W m^{-2})
G =soil heat flux (W m^{-2})
s =slope of saturation vapour pressure curve (mbar K^{-1})
ρ =air density (kg m^{-3})
C_p=specific heat of air at constant pressure (J $kg^{-1}K^{-1}$)
$e(T_z)$=vapour pressure at temperature T_z (mbar)
e_z=vapour pressure at reference height z (mbar)

r_A=aerodynamic resistance (s m^{-1})
r_C=canopy resistance (s m^{-1})
γ =psychometric constant (mbar K^{-1})
z =reference height (m)
T_z=temperature (°K) at reference height z.

The resistance r_A is dependent on the roughness length of the surface and on windspeed. The canopy resistance r_C depends on many variables, such as vapour pressure deficit, temperature, radiation, carbon dioxide concentration, leaf area index and soil moisture status. The relationship between r_C and these environmental quantities varies from species to species and depends also on soil type. Mostly r_C is determined by using an independently measured LE in the Penman-Monteith equation. Applying r_C's non-zero minimum value for situations in which the water supply in the root zone is optimal, equation 5.2 yields the potential dry canopy evaporation.

The Penman-Monteith equation is a descriptive model and should not be used to predict evapotranspiration (McNaughton and Jarvis, 1983), since a number of interrelations exist:

1. When transpiration is reduced because of a decrease in the availability of water, the sensible heat flux from the surface increases in relation to the latent heat flux, i.e. the Bowen ratio (see chapter 6) increases. On a basin scale this causes an increase in temperature and a decrease in the humidity of the overpassing air, thereby producing an increase in the transpiration estimated with the Penman-Monteith equation. Thus, evapotranspiration and vapour deficit are interrelated. (Bouchet, 1963; Morton, 1983, 1985).
2. When atmospheric vapour pressure deficits increase, the stomatal apertures of many plant species decrease in size. With this mechanism plants try to keep transpiration independent of the vapour pressure deficit. A high vapour pressure deficit will result in a high transpiration estimate according to the Penman-Monteith equation, but at the same time causes a low transpiration in the canopy resistance part of the equation. Consequently forest dry canopy evaporation is constrained within a comparatively narrow range (see e.g. Roberts, 1983; McNaughton and Jarvis, 1983).

A number of applications of the Penman-Monteith equation take the dependence of stomatal resistance of environmental conditions into account using phenomenological models (e.g. Stewart and De Bruin, 1985; Dolman et al., 1988; Stewart, 1988). Recently Dickinson (1984) and Wetzel and Jy-Tai Chang (1987) developed methods to estimate evapotranspiration during the drying phase (soil moisture limiting conditions). Most of these models are not applicable on a routine basis, however.

The Priestley-Taylor equation
Micrometeorological observations over well watered temperate arable crops show that evapotranspiration depends strongly on the available energy. In an attempt to provide a more convenient model to calculate the potential evapotranspiration, Priestley and Taylor (1972) proposed the following equation:

$$LE = \alpha \frac{s}{s + \gamma} (R_n - G) \qquad (5.3)$$

where: α = constant

In conditions of non-limiting water supply the actual evapotranspiration equals the potential rate calculated according to equation 5.3 with α in the order of 1.2 to 1.3. Net radiation is well correlated with global radiation (except in winter time). On the basis of this correlation Makkink (1957) proposed a formula similar to the Priestley-Taylor method.

At first sight equations 5.3 and the Makkink formula are purely empirical. However, recently a similar relation was found by taking into account that evapo(transpi)ration E and saturation deficit D are dependent variables. If at the surface water vapour and heat are brought into the lower atmosphere, the saturation deficit D is changed and in turn this affects E. On a regional scale the evapotranspiration of a well watered terrain with a short vegetation is primarily determined by net radiation and also by the temperature through the term $s/s+\gamma$. This means that equation 5.3 describes fairly well the evapotranspiration of e.g. grass on a regional scale if there is no short of water. A discussion on this issue is presented by De Bruin and Holtslag (1987).

5.3.3 Models for the Planetary Boundary Layer (PBL)

To describe the interrelation between evapotranspiration and vapour deficit, an additional model for the planetary boundary layer (PBL) was developed by Perrier (1980) and McNaughton and Jarvis (1983). The PBL is a well mixed layer above the surface layer, implying that the potential temperature and specific humidity are constant with height. Within the surface layer which is typically 1/10th of the PBL height, the Penman-Monteith equation applies (De Bruin, 1987). The PBL height depends on surface heating i.e. on the sensible heat flux. It shows a diurnal variation from 100 to 200 m in the early morning to heights up to 1 to 2 km in the late afternoon in summertime. For cloudy conditions and also in wintertime the variation of the height is much less. A description of the model can be found in McNaughton and Jarvis (1983) and De Bruin and Holtslag (1987). In short the relations for sensible and latent heat fluxes for a fixed height are as follows:

$$\delta T_m/\delta t = H/\rho C_p h \text{ and } \delta q_m/\delta t = E/\rho h \qquad (5.4)$$

where: Tm = potential temperature;
H = sensible heat flux density (see also eq. 6.1, chapter 6);
t = time;
h = height;
q_m = specific humidity

From equation 5.4 the following relation is derived for variable height of the PBL, which resembles equation 5.3 and the Makkink formula:

$$LE = \alpha \frac{s}{s+\gamma} \times (R_n - G) \qquad (5.5)$$

Inputs into equation 5.5 are estimates of the surface fluxes E and H. The parameter α is dependent on many variables such as the vapour deficit, H and LE. De Bruin (1983) showed that $\alpha = 1.3$ for $r_C = 0$; $\alpha = 1$ for $r_C = 60\text{-}90 sm^{-1}$; $\alpha < 1$ if $r_C > 100 sm^{-1}$. McNaughton and Spriggs (1986, quoted in Jarvis and McNaughton, 1986) presented similar results. In general the conclusion may be that a two-fold increase in r_C causes the evapotranspiration to decrease by only 10-20%. For ranges of values of r_C is referred to Table 5.1.

5.3.4 Wet canopy evaporation

In the case of wet canopy evaporation the term r_C/r_A in equation 5.2 falls to zero. Thus the evaporation of intercepted rainfall can be described in purely physical terms (Rutter et al., 1971):

$$LE = \frac{s(R_n - G) + \rho\, C_p(e(T_z) - e_z)/r_A}{s + \gamma} \quad (W\ m^{-2}) \qquad (5.6)$$

Wet canopy evaporation from forests is found to exceed the evaporation equivalent of the net radiation by a large amount during the winter months in England and South Wales. When averaged over several years the total evapotranspiration loss from the forested area in two forest stands in the U.K. is about 12% higher than the total radiant energy input (Shuttleworth and Calder, 1979). Morton (1984) concludes that wet canopies in these regions persist for only a fraction of the time and because significant energy is also used for dry canopy evaporation, either gross precipitation is too high, or the sum of net precipitation and streamflow is too low, or there is another source of energy or not all the intercepted water evaporates, but is absorbed by the leaves and transported towards the root zone.

The energy required for wet canopy evaporation is attributed to advection (Shuttleworth and Calder, 1979; Calder, 1982; McNaughton and Jarvis, 1983; Morton, 1985). Even during the nighttime very high rates of wet canopy evaporation were observed (Pearce et al., 1980). A positive temperature gradient with height, observed when canopies are wet or in the early morning on dry days, shows that downward flux of sensible heat occurs (Shuttleworth et al., 1985). Downslope winds can maintain a constant saturation deficit if they continuously lose sensible heat and gain latent heat (McNaughton and Jarvis, 1983). Calder (1986) reported very high rates of wet canopy evaporation occur in both areas of high and low relief. Morton (1985) states that wet canopy evaporation in excess of net radiation on a basin scale is only possible in areas of high relief.

Wet canopy evaporation rates appear to be many times the dry canopy evaporation rates. Interception and wet canopy evaporation from forests is usually 5 to 20 % of the total rainfall and is inversely related to intensity and duration of the rainfall (Dickinson, 1980). Recent investigations show that wet canopy evaporation may in certain conditions account for up to 70% of total rainfall (Dolman and Oosterbaan, 1986). The duration of rainfall is a much more important constraint on wet canopy evaporation than the net radiation (Morton, 1984). The quantity of water lost through wet canopy evaporation also depends on the canopy storage, which ranges from less than 2 to more than 8 mm (Calder et al., 1986).

Calder (1982) suggested that very high rates of wet canopy evaporation may not be supportable from large scale forests such as those in the Amazon Basin. Shuttleworth and Calder (1979) concluded that interactive feedback between wind in the well mixed boundary layers occurring during rainstorms over wet forests of medium size (≈10km) is slight. Wet canopy evaporation rates from forests of greater dimensions (≈100km) will be reduced, however. The medium sized forests are typical for European conditions.

Shuttleworth et al. (1984b) conclude that in wet months (precipitation 270 mm per month) evapotranspiration from Amazonian rain forest will approach and possibly exceed the accepted estimates of potential evapotranspiration. The evapotranspiration in dry months (precipitation 30 mm per month) will fall to about 70% of the potential rates.

5.3.5 Evaporation in (semi) arid regions

In dry areas 'actual' and 'potential' evaporation rates are poorly related. Just after rainfall events one may expect actual rates to increase and the potential rates to be low due to a decrease of the saturation deficit. In dry periods potential evaporation will be high while actual rates fall to low values. De Bruin (1988) mentions a number of features important for dry regions:

1. Vegetation is sparse in arid regions. As a consequence soil evaporation is important, particularly just after rainfall events. Presently there are no methods to estimate evaporation and heat flux from bare soils requiring routine data only. Examples of work in this field are Menenti (1984), who studied evaporation in desert areas and based his estimates on vapour transport in the soil layer between the surface and the water table; and ten Berge (1986) who presented a coupled PBL-surface model and considered application of remote sensing for e.g. the estimation of evaporation (see also chapter 7).
2. Soil heat flux is of the same order as LE and cannot be neglected as is often done for temperate regions.
3. Total evaporation is small compared to the potential rate. This is caused by the low leaf area index.
4. The use of the concept of potential evaporation determined with the 'old' Penman equation is questionable in (semi) arid regions.

5.3.6 Water recycling in forests

All hydrometeorological studies in tropical rainforests indicate, that recycling of water vapour is an important component in the hydrological cycle. Values in literature vary from 75% to 15% and in the case of the Amazon Basin between 50% and 35% (various authors quoted in Salati et al., 1983; Salati, 1987). Salati and Vose (1984) and Salati (1987) show that over the Amazon Basin about half of the rainfall is returned to the atmosphere by evapotranspiration.

Friedman (1977) in viewing part of the Amazon from the air, observed that great pillars of clouds sometimes appear to arise from the very top of the forest and where the trees thin out there is less cloud. Presumably cloud formation is in part due to the high surface area presented by the forest canopy and high leaf induced unstability. The moist atmosphere beneath it is possibly of influence as well (Salati, 1983).

5.3.7 Evapotranspiration rates

Dry canopy evaporation may be less than evaporation from bare soils or short vegetation due to the retardation caused by the stomata and the turbulence around the canopy (which causes an increase of the sensible heat flux). The transpiration rate from field crops is more sensitive to net radiation than to saturation deficits (various authors quoted in McNaughton and Jarvis, 1983).

Shuttleworth et al. (1984b) found that the fraction of incoming solar energy used in dry canopy evaporation from a tropical rainforest is only 70%. Calder et al. (1986) measured that 100% of incoming radiation is used for total evapotranspiration in a rainforest in West Java. 60% of the net solar radiation was required for dry canopy evaporation. They suggest to use this relation as a simple rule to estimate the total evapotransprration from net radiation.

Baumgartner (1965, 1970) and Baumgartner and Kirchner (1980) calculated the evapotranspiration for a number of land cover types. Their data are shown in table 6.2 in chapter 6. As was discussed in paragraph 6.2 (equation 4), the net radiation is related to the albedo value of the land's surface cover. Although data in table 6.2 are generalized, they indicate the role of the different cover types in the hydrology.

Many estimates of evapotranspiration have been made, especially for agricultural crops in the estimation of crop water requirements. Roberts (1983) studied dry canopy evaporation rates from European coniferous and deciduous forests. He concluded that under different conditions the rates are very similar and in the order of 280 to 430 mm y^{-1}. Dry canopy evaporation from grassland is not dissimilar from forest dry canopy evaporation. McNaughton and Jarvis (1983) showed that very high rates of evaporation from extensive areas of grassland are impossible. Forests have a much higher roughness length and can evaporate much more water under wet conditions. On the basis of a literature review on wet and dry canopy evaporation McNaughton and Jarvis (1983) compiled the following table 5.1, which is illustrative of processes involved in forest and crop evapotranspiration.

The rainfall intensity and duration of each storm determine the rate of wet canopy evaporation (Rutter et al., 1971). The relative amounts of water transpired and evaporated over a year depend on the proportion of the time that the canopy is wet. When the canopy is kept wet by frequent small storms, the evaporation of intercepted water is by far the larger evaporative component of the water balance and may be up to 2.5 times the loss through transpiration. Table 5.1 shows that forest water use in summer may in certain situations be less than water use by agricultural crops.

Table 5.1 Generalized summertime values of dry canopy evaporation LE, wet canopy evaporation E_W, $LE/(R_n\text{-}G\text{-}S)^{**}$, Bowen ratio (B), canopy resistance (r_c) and aerodynamic resistance (r_a) for wet and dry canopies of temperate forest and crops (including grassland and field crops).

	forest		crops	
measure	dry	wet	dry	wet
LE or E_W (mm h^{-1})	0.3(0.7)*	0.2(0.9)	0.6(1.4)	0.1(0.6)
$LE/(R_n\text{-}G)$	0.2-0.6	0.6-4	0.7-1.2	0.7-1.2
B	0.5-4	± 0.5	± 0.5	± 0.5
minimum r_c (s m^{-1})	40-100	<5	20-60	<5
r_a (s m^{-1})	5-10	5-10	20-200	20-200

* maxima in parenthesis; ** S = heat stored in and below the canopy.
Source: McNaughton and Jarvis (1983)

5.4 MODELS FOR PREDICTING DRY AND WET CANOPY EVAPORATION RATES

The simplest method of predicting the effects of changes in land cover on the evaporation and transpiration is that of transferring empirical results from one catchment to other catchments has proved unreliable (e.g. Perreira, 1973). McNaughton and Jarvis (1983) discuss an attempt to predict evaporation and transpiration from forest using meteorological measurements made over grassland and known properties of the forest canopy. Large errors are possible using this approach.

Recently a number of models have been developed which simulate the transpiration and evaporation from vegetation. A number of these simulation models will be discussed below.

The first model is the Simple Biosphere Model designed by Sellers (e.g. 1987) which is based on the Penman-Monteith equation. The land's surface is represented by a vegetation canopy and soil surface which may be bare or covered by ground cover and litter. This structure may be adapted to describe crudely the morphological characteristics of the major vegetation formations by adjustment of heights, densities and cover fractions of the two vegetative layers. Two prognostic variables, i.e. the crop and ground cover temperature, fluctuate over the model time steps and are associated with the fast process of radiation interception and partitioning into latent heat, sensible and storage heat terms. The remaining five prognostic variables are moisture stores: intercepted precipitation at the canopy and at the ground cover, and three moisture stores: the surface store, the rooting zone and the store below the rooting zone. These variables are updated between the time steps. Forcing variables in the simple biosphere model are incident radiative flux, precipitation, wind speed, air temperature and vapour pressure at reference height. In testing the model Sellers and Dorman (1987) found realistic predictions of net radiation, evapotranspiration and sensible heat for barley, wheat, maize and Norway spruce sites.

Dickinson (1986) and Wilson et al. (1987a, 1987b) developed a land surface parameterization sheme (BATS) of the NCAR Communtiy Climate Model (CCM). This scheme includes soil characteristics (porosity, soil suction, wilting point, saturated hydraulic conductivity, thermal conductivity and albedo) and a vegetation canopy with simulated roughness length, vegetation cover fraction, albedo, stomatal resistance, leaf area index and sensitivity to visible radiation. Their GCM modelling results were most sensitive to soil texture variation.

A number of models were developed to predict the dry canopy evaporation from simple meteorological data including environmental effects on the surface resistance as discussed in 5.2.2 (e.g. Dolman and Stewart, 1987, 1988; Dolman, 1988; Stewart, 1988). In Stewart's model (Stewart, 1988) the stomatal resistance depends on solar radiation, specific humidity deficit, temperature and soil moisture deficit. Although in comparison with simpler models the errors produced by this model were significantly smaller, further research is needed to test the model under various conditions and for longer time periods.

Models to predict wet canopy evaporation can be classified as practical (e.g. Gash, 1979; Calder and Newson, 1979; Mulder, 1985) and research models (e.g. Rutter et al., 1971). Practical models are most interesting in this context. Mulder (1985) recognized the distribution of storms as a major determining factor for wet canopy evaporation (see also 5.3). His model requires thrice daily observations of air temperature and relative humidity, daily means of wind run, daily totals of precipitation and the number of rainy hours and of bright sunshine, which are all available from standard meteorological observations. Vegetation parameters required are saturation storage capacity, free throughfall coefficient, zero plane displacement height and roughness length. Conditions influencing wet canopy evaporation are extremely variable and most models will have local validity only.

5.5 CLIMATIC IMPACT OF TROPICAL DEFORESTATION

The apportionment of absorbed solar radiation into sensible and latent heat is dominated by the availability of water at the earth's surface. Since the flux of latent heat associated with evaporative losses is dominant, greater vegetation cover should be associated with measured latent heat loss and therefore a reduction in surface temperature. The amount and type of vegetation covering a surface influences the moisture exchange with the atmosphere in a number of ways:

- Plants are able to extract water at various depths to sustain a flow of water towards the stomata, whereby soil hydraulic properties, soil moisture status and atmospheric demand are major determinants.
- Aerodynamic diffusion resistance and canopy resistance to water vapour transport are also important parameters in the energy exchange process of well vegetated surfaces. Aerodynamic diffusion resistance is determined by factors such as wind velocity, canopy height and -structure. Canopy resistance is a function of leaf area index. Both resistances are related to the amount and nature of the vegetation cover of a surface. In general an increase in vegetation is associated with a reduction of both resistances and with greater evapotranspiration.

The above discussed consequences of deforestation (decrease in infiltration, increase in runoff and decrease of the water use) would seem to indicate, that even with sufficient rainfall, much of the water would not remain to evaporate back to the atmosphere. If there is a reduction of water returned to the atmosphere, this will result in a reduction in cloud cover, increase in light intensity and heat, which in turn would have the effect of increasing potential evapotranspiration but decreasing the actual evapotranspiration. If forest clearing leads to a reduction in water returning to the atmosphere, the effect will vary according to the scale of clearing, whereby the effect may be observable immediately downwind (precisely focused) or more regional within a more diffuse system.

Devegetated areas have reduced turbulence and will be warmer, have a larger upward infrared radiation (reflection and emission), increased sensible and drastically reduced latent heat flux. The turbulence around the forest canopy accelerates the exchange of sensible heat and water vapour with the atmosphere. These mechanisms remove heat so efficiently, that despite being the site of positive radiation balance (see chapter 6), forest canopies are cooler than cultivated or open land in the summertime (Henderson-Sellers, 1980).

Variations in the amount of vapour condensing in the higher part of the troposphere may also influence climate on a global scale. During evapotranspiration solar energy is transformed into latent heat. This heat is subsequently released in the atmosphere where the water vapour condenses to form clouds. This energy is partly responsible for the circulation in the upper troposphere. On the other hand part of this vapour is transferred to higher latitudes where upon condensation energy is released. Hence energy is transported from equatorial to polar zones. Deforestation and consequently a reduction of the amount of evapotranspiration may thus affect the atmospheric general circulation (Salati, 1987).

In the remainder of this section a number of studies of the climatic impact of deforestation using climatic models will be discussed.

Potter et al. (1975) used the two dimensional (zonal) atmospheric model (ZAM2) to assess the impact of tropical deforestation. They assumed albedos of 0.07 and 0.25 of rainforest and cleared forest respectively. The chain of consequences predicted by the atmospheric model was: deforestation - increased surface albedo - reduced surface absorption of solar energy - surface cooling - reduced evapotranspiration and sensible heat flux from the surface - reduced convective activity and rainfall - reduced release of latent heat, weakened circulation, cooling in the middle and upper tropical troposphere - increased precipitation in the latitude bands 5 to 25°N and 5 to 25°S and a decrease in the Equator-pole temperature gradient - reduced meridional transport of heat and moisture out of equatorial regions - global cooling and a decrease in precipitation between 45 and 85°N and at 40 and 60°S. The results for the cases deforestation + albedo change and for albedo change only (no reduced evapotranspiration) are in table 5.3.

Table 5.3. Predicted climate changes acc. to different scenarios.

	control	complete deforest.	albedo change only
atm. water vapour content (g cm^{-2})	2.4731	2.4032	2.394
evaporation (g cm^{-2} d^{-1})	0.33323	0.32983	0.33016
rainfall (g cm^{-2} d^{-1})	0.33305	0.32996	0.33169
cloudiness	0.493	0.490	0.491

Source: Potter et al. (1975).

Henderson-Sellers and Gornitz (1984) reported smaller effects of deforestation using a simple climate model, but they assumed smaller albedo changes due to deforstation than Potter et al. (1975). Henderson-Sellers and Gornitz also analysed the effect of complete deforestation of the Amazon Basin using a 3-dimensional global climate model. They found no significant global temperature effect. Despite the fact, that the albedo increased, the temperature did not decrease. The cooling effect due to the albedo increase was apparently balanced by a reduced evaporation and reduced cloud cover. Local implications according to the model are: no change in temperature, decrease of the precipitation by around 0.6 mm d^{-1}, decrease of evaporation by 0.4-0.6 mm d^{-1}, planetary albedo increase by 1-1.5 % as a combined result of increased surface albedo and decreased cloud cover.

Dickinson and Henderson-Sellers (1988) included the land surface scheme (BATS) developed by Dickinson (1986) and Wilson et al. (1987) (see 5.4) in a version of the NCAR-CCM to study the local effect of complete tropical deforestation of the South-American Amazon. All the tropical forest was replaced by a cover having characteristics of impoverished scrub-grassland, intended to be typical of the parts of the Amazon converted to cattle ranching. The major changes imposed by the deforestation were:

- soil compaction was simulated by making soil texture finer. Soil colour was made lighter to simulate organic matter loss.
- albedo after forest clearing increased from the low forest albedo value of (4% and 20% for wavelengths of < 0.7 μm and > 0.7 μm, respectively) to a higher albedo (8% and 30% for wavelengths of < 0.7 and > 0.7 μm, respectively), composed of partial surface albedo of the grass-scrub cover and a fractional soil albedo.
- decreasing vegetation cover and decreasing roughness length of the land cover (roughness length decreases from 2.0 to 0.05 m in the BATS parameteriziation scheme);

- deceasing soil depth (expressed as the rooting ratio= ratio upper to total soil layers, decreasing from 12 to 10) through erosion and increase in run-off;
- increase of the relative density of roots in the soil's surface layer;
- decreasing sensitivity of the vegetaton to visible radiation;

A decrease in roughness length would reduce the turbulent exchange and hence potentially reduces energy transfer between the surface and the atmosphere, causing a temperature rise, and also reduces the rainfall interception. The modelling results presented by Dickinson and Henderson-Sellers are supported by actually measured changes in surface run-off (see 5.2.1) and are consistent with e.g. Ghuman and Lal (1987).

The above results should be viewed primarily as indications of possible change. They differ from the modelling results reported by Potter et al. (1975) and Henderson-Sellers and Gornitz (1984). It is however, difficult to compare modelling results. Furthermore, it should be remembered, that the GCM's only consider the direct effects of deforestation and do not account for indirect effects such as increasing carbon dioxide concentration, increasing concentration of other greenhouse gases and atmospheric particulates. Early versions of the GCM's ignored the existance of vegetation.

5.6 CONCLUSIONS

The presented models and equations for describing evapo(transpi)ration have only a limited validity. The Penman-Monteith equation and Priestley-Taylor equation can predict dry canopy evaporation for non-water limiting conditions and only in summertime. The models are not suited for arid and semi- arid conditions.

The prediction of wet canopy evaporation is still in development. However, conditions responsibl e for wet canopy evaporation are so variable, both in space and time, that models will never have a general applicability. The study of wet canopy evaporation will be very important for the understanding of the effect of (tropical) deforestation. Especially in the tropics, where most deforestation occurs (see chapter 3) it appears that there is a water recycling by forests. The process would be essential for maintaining the fragile balance of the tropical rainforest ecosystem.

Denudation of once forested land causes the ratio of sensible heat flux to latent heat flux to incre ase. The ultimate result of this change in the energy balance is that water originally lost through evapotranspiration now has to drain superficially. Apart from the erosion caused by this surface runoff a consequence is that less vapour will be emitted into the atmosphere. This reduced vapour flux into the atmosphere from ecosystems has a direct effect on the radiative balance both by reduced cloud cover (increase of incoming radiation) and reduced greenhouse gas concentration (increased infra red absorption). Added to the above described shift in the energy balance (see also chapter 6) this means that a decline in the degree of coverage provided by vegetation will provide a positive feedback to any tendency to aridity, of imbalance to local rainfall and also to the global heat balance.

In the last number of years scientists attribute an important microscale influence to forests and ar e reconsidering macroscale effects. Present consensus tends to the opinion, that the overall effects of tropical deforestation are more likely to be observed through the loss of biological heritage, regional climate stability, soil degradation and erosion, reduced CO_2 uptake and water loss, rather than through modification of global climate. However, during recent years some researchers have suggested that climatic changes in mid latitudes were a consequence of anomalies in the tropics.

6 SURFACE REFLECTANCE PROPERTIES

6.1 INTRODUCTION

Until recently the land cover was assumed to be solely the result of climatic forcing with minimal feedback onto the climate itself. The interactions between land surface and the atmosphere however, are significant and complex. The land cover influences the atmosphere via radiation, transfer of momentum and transfer of sensible and latent heat.

The albedo is one of the factors in the energy balance which determine the partitioning of the sun's incoming net radiation. The longwave terms in the radiation balance are the key factors in the greenhouse effect. Infrared (thermal) radiation is absorbed by atmospheric greenhouse gases.

Several studies have shown that denudation of once vegetated areas has an effect on the soil and the lower atmosphere as well as on the reflected radiation. On a regional or global scale the increased albedo caused by deforestation will have a cooling effect (Otterman, 1974; Schanda, 1986) which will result in decreased lifting of air necessary for cloud formation and precipitation. Other authors (Jackson and Idso, 1975) predicted that denudated surfaces should be warmer than vegetated ones. On a local scale the effects of deforestation are a warming of soil and air temperatures (Ghuman and Lal, 1987).

In this chapter the nature of incoming solar radiation and its partitioning over reflected global radiation and longwave radiation terms will be highlighted. In section 6.5 the albedo effect on climate will be discussed.

6.2 THE SURFACE ENERGY BALANCE

The processes at the earth's surface can be described by using the energy balance. This central equation sets boundary conditions to both the soil and the atmosphere subsystems. In a general form the surface energy balance can be written as:

$$R_n + H + LE + G_E = 0 \tag{6.1}$$

where: R_n = net radiation (net radiant flux density) (W m^{-2});
G_E = soil heat flux density (W m^{-2});
H = sensible heat flux density (W m^{-2});
LE = latent heat flux density (W m^{-2}).

A frequently used expression is the dimensionless "Bowen ratio" B, which is the ratio of sensible heat and latent heat: $B = H/LE$.
In the above equation net radiation (R_n) is adding energy to the surface (soil and canopy) and it is being repartitioned into soil heat flux (G_E), latent heat (LE) and sensible heat (H). The latter term represents the heat transfer by dry ventilation and cooling of the surface by air passing over it. A strong feedback exists between the fluxes in the energy balance and surface properties. For example, the radiation balance of bare soils is affected by soil properties such as soil moisture content, organic matter content and soil temperature, since these variables influence surface albedo, emissivity and emittance respectively (see also table 6.1).

Sensible and latent energy (LE) are not lost from the surface atmosphere as a whole. Changes in their relative importance (or changes in the Bowen ratio = ratio of sensible to latent heat flux densities) can be significant for the hydrological cycle and therefore affect regional climates (see also chapter 5). The soil heat flux (G_E in equation 6.1) is traditionally measured with sensors buried just beneath the surface. This flux is dependent on the moisture condition and the type and density of the vegetative cover. In dry soils the soil heat flux is considerable (up to 30% to 50% of net radiation) while in grassland and forested soils this term is much smaller than net radiation.

R_n can be divided in a short wave (R_{sw}) and a thermal contribution (R_{sky} and R_s):

$$R_n = (1-\alpha)\, R_{sw} + R_{sky} + R_s \tag{6.2}$$

where:

α = albedo (dimensionless);

R_{sky} = sky longwave incoming radiation (W m^{-2});

R_s = surface (longwave) emittance (W m^{-2});

R_{sw} = shortwave incoming radiation (global radiation; Wm^{-2}).

Emittance is a temperature dependent term (Bolzmann's law). Global radiation is the incoming shortwave radiation from the sun. The sun, a black body of 6000 K, emits radiation in the range of wavelengths of 0.2 to about 4 μm. The earth, a body with approximate temperature of about 300 K, emits radiation with wavelengths of 4 to 40 μm.

6.3 RADIATION

6.3.1 Shortwave radiation terms

- Global radiation (R_{sw})

R_{sw} is the major fraction of daytime incoming radiation. It is the shortwave radiant flux density (W m^{-2}) received at the surface resulting from the integration of radiance (W m^{-2} Sr^{-1}) over a solid angle 2πSr. R_{sw} can be estimated (not measured) from satellite data, whereby two different approaches exist:
- determination of statistical relationships between R_{sw} and cloudiness.
- calculation of atmospheric transmittance by a radiative transfer model, whereby some of the required parameters such as cloudiness, albedo and planetary reflectance are obtained from satellites.

- Net radiation (R_n)

The shortwave component of net radiation is the proportion of global radiation which is not reflected:

$$R_n = (1 - \alpha) \times R_{sw} \tag{6.3}$$

where: α = surface albedo.

Hummel and Reck (1979) defined the surface albedo as the ratio of wavelength averaged solar radiation reflected by the earth's surface to that incident on it. An increased use of satellite data to monitor the albedo of arid lands has arisen from the dual importance of albedo as a potential indicator of arid land degradation and as a physical parameter with possible impacts on climate.

For a given surface and wavelength, the sum of reflectivity, absorptivity and transmissivity equals unity. For opaque bodies the sum of absorptivity and reflectivity equals unity since transmission is 0 (Ten Berge, 1986). The following formulas describe the above relations:

$$r = 1 - a \tag{6.4a}$$

The absorbed energy equals the amount emitted (Kirchhoff's law):

$$e = a \tag{6.4b}$$

where: r = reflectivity (dimensionless);
a = absorptivity (dimensionless);
e = emissivity (dimensionless).

Reflectivity depends on the wavelength of the incoming radiation and in general increases with wavelengths up to 1.2 μm (Coulson and Reynolds, 1971; Ten Berge, 1986). As surface reflectivity is also dependent on azimuth and zenith (incidence) angles, it will be clear, that the overall fraction of shortwave radiation reflected by the surface in reality is not a constant, but is related to atmospheric conditions and the position of the sun and thus depends on time of the day and the date.

6.3.2 Longwave radiation terms

The longwave radiation terms are essential in the greenhouse effect since they are absorbed by atmospheric greenhouse gases and aerosols and cause global warming. Radiation of wavelengths in the range of 4 to 40 μm is emitted by the earth. The quantity of emitted energy depends on the surface temperature. This longwave radiation is partly absorbed in the atmosphere by the so called greenhouse gases earlier listed in chapter 1. The longwave emission can be divided in a number of terms:

- <u>Sky radiation R_{sky}</u>

Thermal sky radiation or longwave irradiance R_{sky} (W m^{-2}) also constitutes an important term in the surface energy balance. In analogy to global radiation, the longwave irradiance is defined as an integral over azimuth, zenith angle and wavelength. In practice it is often taken as (Stefan Boltzmann's law):

$$R_{sky} = \Sigma'_{sky} \times k \times T_a^4 \tag{6.5}$$

where: Σ'_{sky} = apparent emissivity of the air (K^{-3}s^{-1}m^{-2});
T_a = air temperature (K);
k = Stefan Boltzmann constant (5.67*10^{-8} W m^{-2} K^{-4}).

E'_{sky} can be estimated from air temperature and vapour pressure according to (Brutsaert, 1975):

$$E'_{sky} = 1.24\, (e_o/T_a)^{1/7} \tag{6.6}$$

where: e_o = vapour pressure in mbar;
T_a = air temperature (K).

- <u>Surface emittance R_s</u>

The longwave radiation leaving the surface consists of the terms emittance and reflection. Over the whole wavelength interval the <u>surface emittance</u> can be calculated as:

$$R_s = \Sigma \times k \times T_s^4 \tag{6.7}$$

where: Σ = emissivity of the emitting body;
T_s = surface temperature (K).

The surface emittance is particularly of interest in areas with bare soils. The emissivity (= the efficiency of emission of longwave radiation by a body) is a soil specific property, that ranges from 0.87 (sand) to 0.95 (material with higher organic matter content; Becker et al., 1985, 1987). The difference found between wet and dry soils usually amounts 0.02 to 0.04 (Ten Berge, 1986). Relatively few data are available on the emissivity of moisture conditions between dry and saturated (very wet) and the relation between emissivity, soil organic matter content and mineral composition. Differences in emissivity are small and hardly significant in the energy balance (equation 6.1) of bare soils, but they are of great importance in the interpretation of thermal infrared imagery. Differences in the emissivity Σ have been reported to make cool, wet sand appear warmer on surface imagery than warm, dry sand.

The reflection component of longwave radiation leaving the surface is:

$$(1-\Sigma)\ R_{sky} \tag{6.8}$$

where: Σ = emissivity

The emissivity equals the absorptivity (Kirchhoff's law). Except when the emissivity is equal to 1 the brightness temperature is always lower than the physical kinetic temperature of the emitting body. If the temperature is assumed to be equal to the brightness temperature, the real temperature is underestimated. The discrepancy increases with decreasing emissivities and with increasing surface temperatures (Thunnissen and Van Poelje, 1984; Schanda, 1986). Thus, the degree of underestimation is higher for bare soils than for vegetated surfaces.

Equations 6.7 and 6.8 can be combined in one equation:

$$R_s = \Sigma \times k \times T_s^4 + (1-\Sigma) \times R_{sky} \tag{6.9}$$

A surface which is partly covered by vegetation has a surface emittance (Thunissen and van Poelje, 1984):

$$R_s = f_c\Sigma_c T_c^4 + (1-f_c)\Sigma_s T_s^4 + f_c(1-\Sigma_c)R_{sky} + (1-f_c)(1-\Sigma_s)R_{sky} \tag{6.10}$$

where: f_c = vegetation coverage expressed as a fraction;
Σ_s = soil emissivity;
Σ_c = crop emissivity.

6.4 REFLECTANCE PROPERTIES

6.4.1 General

If one relies only on the albedo, one could not discern between even extremely different surfaces, such as grass fields, various rock types and the effect caused by thin cloud layers. Only by adding observables such as colour, size, shape, distribution and patterns, time (hours and seasons), solar incidence and polarization, does the discrimmination of the kinds of objects and some of their qualities become feasible.

6.4.2 Spectral reflectance characteristics of bare soils

About 99% of the solar radiation is received in the spectral range 0.3-1.9 μm and 83% in the range 0.3-1.1 μm. Hence an instrument sensing fluxes in the interval 0.3-2 μm provides representative values of reflected radiance. When a smaller band is selected, the relation between solar spectrum and soil spectral response must be taken into account, since in the visible and infra red spectrum the reflection is a consequence of scattering and has a definitive relation with incoming radiation.

As was stated above the dielectric properties of a medium determine the dependence of reflection on wavelength, angle of incidence, etc. The dielectric properties in turn, are governed by the characteristics of the medium, such as soil moisture status and soil water retention characteristics, organic matter content, mineralogic composition. Other factors influencing the reflectivity are the surface roughness and texture.

For literature values of the reflectance of different bare soil surfaces the measuring technique (zenith angle) and averaging procedure need not be the same. As a result of the daily cycle of the angle of incidence, the albedo (α) will decrease in the morning and increase in the late afternoon. Furthermore α is dependent on wavelength of incident radiation. α in the 0.4 to 1.1 μm spectral range (as in LANDSAT) is representative of a much wider spectral range of 0.2 to 3.5 μm. Knowledge of α-wavelength curves is helpful in classifying bare soil surfaces from LANDSAT-MSS data and in assessing soil moisture conditions. Broad band radiometers (0.5-1.1 μm) as carried in satellites allow a good determination of α. When dealing with multispectral scanners, one should account for the percentage of solar energy in each band to calculate α from the various single band reflectances.

The dielectric properties and reflectivity of soils are strongly related to soil moisture content. An increasing soil moisture suction or in other words, a decreasing free energy level of soil moisture, causes reflection to decrease in the optical and infra red wavelength range. Specific moisture absorption bands are 1.45 and 1.95 μm (Schanda, 1986). Soil organic matter also influences the reflectance (e.g. Baumgardner et al, 1970; Stoner et al, 1980; Latz et al, 1984; Stoner and Baumgardner, 1984; Al-Abbas et al, 1972; Menenti, 1984). An increase of organic matter content causes an increase of absorption (decrease of reflection) over the complete wavelength traject. Besides its proper reflectance characteristics, organic matter plays a role in soil structure as a binding agent. In that way it influences the surface roughness and -texture. Furthermore, organic matter may influence the soil moisture retention properties.

Other factors, that influence the spectral reflectance characteristics of bare soil surfaces are the surface roughness, salt content, salt crusts, sealing of soil's surface, anorganic coatings on mineral parts, erosion class (Latz et al, 1984).

6.4.3 Spectral reflectance charactersitics of vegetation

The spectrum of a plant leaf can be divided into the following ranges (Mulders, 1987):

- 0.4-0.7 μm: intense absorption of incident radiation by pigments in the plant leaves with major absorption bands of about 0.43-0.45 μm and 0.66-0.64 μm (Schanda, 1986).
- 0.7-1.3 μm: low absorption, high reflection.
- 1.3-2.6 μm: high absorption by water in the leaf; in this part of the spectrum the absorption of energy is related to moisture content and leaf thickness (maximum absorption at 1.45 and 1.95 μm).

Although some vegetative cover types have significantly different spectral response patterns, many tree species and many agricultural crops have spectral response curves, that are very similar (Hoffer, 1984). However, spectral differences between major cover types such as green vegetation, dry and dead vegetation are significant and distinct. The NOAA-AVHRR derived normalized difference vegetation index is a representation of the 0.4-0.7 and 0.7-1.3 μm areas. Therefore, this vegetation index cannot be used to discern between different species and vegetation types, but it can be useful for monitoring seasonal and annual patterns of change.

Table 6.1.Energy balance terms (annual averages) for earth surfaces with different land cover types.

land cover	R_n (Wm^{-2})	H (Wm^{-2})	L (Wm^{-2})	E (mm)	α (%)	Bowen ratio
coniferous forest	80	27	53	1000	10	0.5
deciduous forest	67	20	47	900	15	0.4
open land, moist	80	20	60	1000	20	0.3
savanna	67	27	40	800	25	0.6
grassland	60	20	40	750	20	0.5
cropland	67	27	40	800	25	0.7
bare sand	47	27	20	600	30	1.3
urban area	47	27	14	600	30	2
desert	47	40	20	600	30	6

source: Baumgartner (1965, 1970)

Visual colour is a strong indicator of pigmentation in vegetation. About 5% of the total light is still transmitted through 8 stacked leaves, and even in the reflected radiance a small difference may be recognized between stacks of 6 and 8 leaves. The status of pigmentation, inner structure and water content of a leaf can be observed almost independently. Recent developments show the ability to distinguish between vegetation species and to identify water stress (Schanda, 1986).

6.4.4 Surface albedo values

Baumgartner (1965, 1970) demonstrated the interrelations between net radiation, albedo, sensible heat and latent heat. Although his figures as presented in table 6.1 are generalizations, they indicate that the climatic parameters vary considerably between land cover types. The dependence of net radiation on the albedo, which was discussed in section 6.2 (equation 4) becomes clear in table 6.1.

In the literature one finds mention of two types of deninitions of albedo, i.e. the surface albedo and the planetary albedo. The surface albedo is the desired quantity in climatological studies of the heat balance of the earth. Hummel and Reck (1979) developed a very useful albedo model, which was already mentioned in chapter 3. Their data as presented in table 6.2 illustrate the seasonal variation of albedo. Another seasonal global albedo dataset is the one produced by Matthews (1985) using a 1x1° grid. The latter data set is based on the vegetation types and does not include snow and ice cover. Gutman (1988) developed a method to determine the monthly mean albedo of land surfaces using AVHRR data (see chapter 7).

In contrast to the data in Table 6.2, Pinker et al (1980) found a mean albedo of 13% for tropical evergreen forest in Thailand, ranging from about 11% around midday to 18 to 19% in the early morning or late afternoon. The variation is largely suppressed on overcast days and the midday albedo is higher on overecast days than on clear days. Shuttleworth et al (1984a) reported a mean albedo value of about 12.15±0.2% for the Amazon rain forest and less variation than found by Pinker et al (1980). For a tropical forest in Nigeria Oguntoyimbo (1970) reported an albedo value of 13%.

Table 6.2 Land cover types, extent and seasonal (Northern Hemisphere) albedo values.

albedo (%)		period			
land cover type	extent ($10^{12}m^2$)	January-March	April-June	July-September	October-December
Arable land with intensive farming					
now in winter	5.24	50	16	15	27
no winter snow	15.98	14	15	15	14
Grazing and marginal farming lands					
snow in winter	5.49	50	18	16	20
heavy snow in winter	0.38	50	39	16	20
no winter snow	25.20	20	16	18	20
Rice lands or areas where paddies dominate	2.63	12	12	12	12
Other irrigated land					
winter/fall snow	0.17	50	39	17	32
winter snow	0.20	50	17	17	26
no winter snow	1.25	20	20	20	20
Coniferous forests					
heavy snow	10.19	47	47	14	47
snow in winter	2.63	36	16	14	11
no winter snow	0.42	16	16	16	16
Deciduous forest					
snow in winter	0.35	33	14	14	19
no winter snow	4.65	19	14	14	19
Mixed coniferous and deciduous forests					
snow in winter	2.03	35	15	15	19
no winter snow	2.12	15	15	15	15
Tropical woodlands and grasslands	6.64	16	16	16	16
Rain forests	15.13	7	7	7	7
Deserts					
shrubland with winter snow	0.08	36	22	22	22
shrubland with no winter snow	14.47	22	22	22	22
sand	4.77	42	42	42	42
Marshes					
snow in winter	0.85	55	35	14	55
no winter snow	1.95	10	10	10	10
Tundra	11.72	82	82	17	82

Source: Hummel and Reck (1979)

6.5 THE EFFECT OF CHANGING ALBEDO ON CLIMATE

Although it is difficult to consider climatic effects of albedo change separately, some of the literature on this subject will be reviewed below. The climatic effects of deforestation through changing patterns of sensible and latent heat is further discussed in section 5.5.

Under conditions of unlimited moisture thermal emissions will be inversely related to the amount of vegetation. This relationship however, may be complicated by the effect of albedo, since increased albedo results in reduced surface heating. The results by Goward et al (1985) support the above theory. They concluded, that a satellite designed especially to observe the relation between vegetation index and thermal emission is needed.

The earth absorbs energy in the zones of small albedo in the visible depending on the incidence angle of the sun, thanks to the atmospheric transparency in this part of the spectrum. It loses energy rather uniformly over the globe due to the partial transparency in the thermal infrared. The extraordinary importance of water in the heat budget of the earth is obvious. For steep solar incidence, i.e. in the tropical region, the albedo of water is very low, hence absorption of the sun's radiation high. The polar ice caps with high albedo reflect the energy influx incident in the visible spectrum. However, their lower albedo in the infrared causes effective radiation of thermal radiation. The strong dependence of emitted radiation on temperature (Boltzmann's law) balances the earth's temperature to a well defined mean and limits the ocean surface temperature to about 30°C (Schanda, 1986).

Simply altering ground cover affects all the terms of the energy balance: the albedo and the Bowen ratio (= the ratio of sensible to latent heat transport). Furthermore, the surface winds and runoff rate are influenced by denudation of soils. These changes in turn cause soil moisture, temperature and erosion rates to change. The Bowen ratio, which has values of 0.3 to 0.6 for different vegetation covers, will change in extreme cases of devegetation to 2.0 (see table 6.1). This implies that areas that previously were sources of latent heat would become foci for the generation of large amounts of sensible heat (Salati and Vose, 1984).

Ghuman and Lal (1987) show that maximum air temperature above the soil surface was always 2 to 5° lower under forest than in the open during daytime in a Nigerian rainforest. However, the forest at 1m was relatively warm at night. With the advance of the dry season the difference between the maximum air temperatures in the open and forest decreased. Some experiments (e.g. Ribeiro et al, 1982) show that in areas where vegetation cover is reduced, the relative humidity is reduced and the temperature is higher. Lal (1987) gives an extensive literature review on tropical forest microclimate and changes due to deforestation for various parts of the world.

A decline in the surface density of vegetation will increase the surface albedo resulting in a net radiative loss, which gives rise to general subsidence and drying over the area, thereby inhibiting or reducing the convection necessary for rain (Lo, 1987). This process will cause further vegetation declines through reduced precipitation (Sagan, et al, 1979). When high albedo soils are denuded, the resultant increase in albedo causes lower surface temperatures which in turn reduce the heat input in the lower atmosphere and reduce convective activity necessary for rainfall.

Walker and Rowntree (1977) reported, that soil moisture and albedo provide a positive feedback to any tendency in aridity. It may be of importance to climatic stability both with respect to local rainfall and -because of the albedo influence of soil moisture- to the global heat balance.

The albedo of a bright desert sandy soil can be estimated to be 0.37, whereas the albedo of the same soil with appreciable vegetation cover would be 0.25 (Ottermann, 1974). Such large differences in albedo can have large environmental effects. Higher albedos cause lower surface temperatures (Ottermann, 1974). Jackson and Idso (1975) however, reported an adverse relation.

Sagan et al (1979) found the global temperature change over the past several thousand years to be -1K, due primarily to desertification. The temperature decrease over the past 25 years was 0.2K in their estimate. Henderson-Sellers and Gornitz (1984) in modifying the estimates given by Sagan et al (1979) considered the effect of complete deforestation of the Amazon basin. The $5x10^{12}$ m^2 of tropical moist forest was replaced by a grass vegetation cover. They arrived at maximum planetary albedo increase of between 0.00033 and 0.00064, corresponding to a global temperature decrease of 0.06 to 0.09 K, which is well below the larger period variability. The major difference with the estimates by Sagan et al (1979) are reduced albedo changes. Henderson-Sellers and Gornitz (1984) concluded, that albedo changes induced by current levels of tropical deforestation appear to have a negligibly small effect on the global temperature. And if there is an effect, then this will (partly) counterbalance the warming effects of the so called greenhouse gases discussed earlier in this paper.

6.6 CONCLUSIONS

In this chapter an attempt has been made to summarize the role of the earth's reflection properties (albedo) in the total radiation balance. In the greenhouse effect the radiation terms which are of particular interest are the longwave radiation terms as discussed in 6.3.2. Net radiation consists of shortwave and longwave terms. The portion of global shortwave incoming radiation which is not reflected (i.e. 1 minus the albedo value) is the shortwave fraction. Longwave terms are sky radiation, which depends on atmospheric vapour pressure and temperature, and surface emittance, which depends on the surface temperature. Surface temperature is determined by its albedo value and incoming radiation.

A decline in the surface density of vegetation or a complete denudation will increase the surface albedo. This will have an effect on net radiation directly by determining the shortwave component, and indirectly by determining the surface's temperature and longwave emittance. Especially the latter type of radiation is of significance in the greenhouse effect.
In other words, net radiation and the ratio of sensible heat flux to latent heat flux will increase following denudation. This implies that areas which in their original state are sources of latent heat, become foci for the generation of large amounts of sensible heat. This phenomenon has also been observed after partial deforestation. Especially in areas with tropical rainforest the albedo effect of denudation is expected to have a great impact of local, regional and possibly global climate.

In this respect the albedo value of the original land cover type previous to clearing in relation to the albedo of the remaining cover type or bare soil is of interest. So far few attempts have been made to map albedo values. The studies referred to in the text concern albedo values of natural vegetation. It would be interesting to map the changes in the radiative balance seasonally as occurring globally and regionally on a scale which can be adopted by climate models. Inclusion of these data in global climate sensitivity models has received little attention so far. The attempts that have been made have not produced satisfactory results.

7. ASSESSMENT OF REMOTE SENSING TECHNIQUES FOR VEGETATION MONITORING AND FOR ESTIMATING EVAPOTRANSPIRATION

7.1 INTRODUCTION

Global and regional estimates of attributes such as the area of forests and the year by year changes in the area is basic information for the global carbon cycle, but it is also essential with respect to changes in albedo, evaporation, run off rates and the area of agricultural land. Changes such as the transition from forest to non forest cause large changes in albedo, that are easily detected using predominantly empirical remote sensing models.

For the detection of land cover changes different approaches exist. Strahler et al (1986) divide remote sensing models into discrete and analogue (continuous) scene models and into deterministic and empirical models. Discrete models have elements with abrupt boundaries whereas in continuous models energy flows are taken to be continuous and there are no sharp or clear boundaries in the scene. Deterministic remote sensing formulates relations concerning real processes of energy and matter interaction (emissivity, scattering, absorption and properties related to thermodynamics). Empirical remote sensing models aim at associating sensor measurements with scene elements, typically in a statistical fashion. In reality, methods that are basically deterministic are often formulated with empirical components (Strahler et al, 1986).

In 7.2 techniques of remote sensing will be discussed followed by a review of the applications for the measurement of evapotranspiration and for the monitoring of land cover. In 7.3 a number of applications are discussed.

7.2 REMOTE SENSING TECHNIQUES; GENERAL ASPECTS

Under clear sky conditions short wave reflectance measured from satellites will be only slightly different from values obtained on the ground (Menenti, 1984), but under cloudy conditions satellite measurements of reflectance in the visible part of the spectrum will be impossible.

Long wave (thermal) radiation emitted by the surface can in principle be obtained by measuring the emittance in the infrared 8 to 14 μm spectral range. Use of radiometers installed in an airplane or satellite is a suitable experimental technique, which allows the coverage of large areas.

For radiometrical observation of the ground it is important to use a wavelength range where absorption and scattering by the atmosphere are very low, otherwise the desired information is hidden by a strong attenuation by amongst others atmospheric absorption and upwelling radiation created in the atmosphere itself (Schanda, 1986).

For so called <u>active</u> methods of remote sensing of the earth's surface (radar and lidar) it is obvious that the scatter behaviour of rough surfaces in general and features of the portion of radiation which is reflected back to the sensor in particular, are of fundamental interest. But also for observation of natural radiation of an object the scatter and absorption properties of rough surfaces the albedo and emissivity are fundamental. The utilization of surface scattering concepts are much more dev eloped in the microwave than in the optical range of the spectrum. This is because radar systems are all wheather day and night sensors, which measure backscattering of artificial radiation and allow sensitive discrimination between different surface properties.

In relation to monitoring of vegetation at a global scale various remote sensing techniques are applicable depending on the purpose and desired detail. A summary of the various techniques will be given in this paper following the order: aerial photography, multispectral scanning techniques (MSS), Advanced Very High Resolution Radiometer (AVHRR) and microwave sensing (SLAR = side looking air borne radar; SAR = synthetic aperture radar). The different types and characteristics of the satellites, that carry environmentally useful sensors, are listed in Lo (1987), Curran (1986), Billingsley (1984), Hoffer (1984) and Mulders (1987).

7.2.1 Aerial Photography

In aerial photography four types of film are commonly used: black and white panchromatic, black and white infrared, colour and colour-infrared. Use of aerial photography offers the advantage of stereoscopic view which enhances recognition of objects and the interpretation for soil and vegatation mapping.

As a group, panchromatic films have the best resolution of any of the film types, which makes them useful for such measurements as heights of trees or diameters of crowns. But since panchromatic film is only sensible to the visible part of the spectrum (0.4-0.7 μm) it is not highly suitable for distinguishing between species. Black and white infrared films show less contrast, but can be used for multiband photography using different filters. In some applications a Wratten 89B filter (which filters all the visible wavelengths) is used to provide contrast between deciduous and coniferous forests. Wratten 25B filters sensitize films to both visible and reflected infrared wavelengths.

Table 7.1. Utility of different scales of aerial photography for vegetation mapping

type/scale of imagery	general level of plant discrimination
earth-satelite imagery	separation of extensive masses of evergreen vs. deciduous forest
1:25000-1:100000	broad vegetation types, recognition by inferential process.
1:10000-1: 25000	direct identification of major cover types and species occurring in pure stands.
1: 2500-1: 10000	identification of individual trees in pure stands.
1: 500-1: 2500	identification of individual range plants and grassland types.

Source: Hoffer (1984)

With colour films individual species of trees can be identified far better than with panchromatic film. Colour-infrared is sensitive to the visible and near infrared part of the spectrum (not thermal infrared) and can be used to detect spectral differences, which may be very small in the visible wavelengths but distinct in the near infrared. Another advantage of colour-infrared film (and also of black and white infrared film) is, that atmospheric penetration is better than for true colour film.

The type of film used and the scale depend on the degree of detail involved and the accuracy required. A summary of the scale of photography and the obtained degree of detail is presented in table 7.1.

7.2.2 Multispectral Scanner Systems (MSS, TM)

In satellite-borne multispectral scanner systems the energy reflected or emitted from a small area on the earth's surface at a given moment is reflected from a rotating or oscillating mirror through an optical system, which disperses the energy spectrally to an array of detectors. The motion of the mirror allows the energy along a scan line, which is perpendicular to the direction of flight, to be measured while the forward movement of the aircraft or spacecraft brings successive strips of terrain into view. The detectors simultaneously measure the energy in the different wavelength bands and the output signal from the detectors is amplified and recorded on magnetic tape. The latter feature makes this technique suitable for computer aided analysis techniques.

The spatial resolution of scanner systems is dependent on the characteristics of the scanner and its altitude. Usually the spatial resolution is not as good as can be obtained from photographs, but its spectral resolution is far better. Another aspect is the wide range of wavelengths that are accessible including part of the thermal region of the spectrum.

Table 7.2. wavelength ranges of the 7 bands of the Thematic Mapper (TM)

Band	range(um)	Description
1	0.45- 0.50	designed for water body penetration, useful for coastal water mapping and differentiating soils from vegetation and deciduous from coniferous flora.
2	0.51- 0.60	designed to measure visible green reflectance of vegetation for vigor assessment.
3	0.63- 0.69	chlorophyll absorption band important for vegetation discrimmination.
4	0.76- 0.90	useful for determining biomass content and delineation of water bodies.
5	1.55- 1.75	indicative of vegetation moisture content and soil moisture and useful for differentiation of snow from clouds.
6	10.40-12.50	thermal infrared band of use in vegetation stress analysis, soil moisture discrimmination and thermal mapping.
7	2.08- 1.35	band selected for its potential for discrimminating rock types and for hydrothermal mapping

Source: Lo (1987)

The multispectral scanner installed in the LANDSAT 1-3 collects data in 4 bands: 0.5-0.6 μm (green); 0.6-0.7 μm (red); 0.7-0.9 μm and 0.8-1.1 μm (refl. infrared). The data are handled in frames each covering a ground area of 185 x 185 km. A frame contains 2340 scan lines, each having 3236 resolution elements (pixels). Each pixel represents an area of 0.46 ha. The thematic mapper in LANDSAT-4, launched in 1982 as part of a complete end-to-end highly automated earth monitoring system, obtains greater quantities of data since it has a higher spatial resolution (30 m, exc. band 6 with 120 m) and the thematic mapper multi spectral scanning system (TM) has 7 bands. The wavelengths in each band are presented in table 7.2.

Both manual interpretation (e.g. Crapper and Hynson, 1983) and computer (Hoffer, 1984) aided analysis techniques have been used with LANDSAT imagery. LANDSAT data allow for delineation of major cover types and also disturbed forest lands can be reliably defined. However, brushland which develops after clear-cutting, is difficult to distinguish.

Use of LANDSAT data from different years offers potential for monitoring historical and actual deforestation in many critical areas of the world. Where available, coupling of LANDSAT data with NOAA-AVHRR (see 7.2.3) or radar imagery (see 7.2.4) yields more information than can be obtained from either of the three separately. In this respect the repeat cycle of a satellite is important, especially in areas of frequent cloud cover. LANDSAT has a repeat cycle of 18 days. A few strategically placed days of poor viewing conditions can render a data set virtually useless. SPOT has a somewhat longer cylce of 26 days. The ERS-1 (Earth Resources Satellite, launch planned in 1990) will have a repeat cycle of 3 days while the repeat cycle of NOAA is 1/2 day.

7.2.3 Low spatial resolution scanners (AVHRR, CZCS)

The primary sensor for coarse resolution land remote sensing is the advanced very high resolution radiometer on board the National Oceanographic and Atmospheric Administration's (NOAA) seri es of polar orbiting sun synchronous meteorological satellites (Townshend and Justice, 1986). The spectral range of NOAA-9 (launched in 1984) is 0.58-0.69 μm (band 1), 0.725-1.1 μm (band 2), 3.5-3.93 μm (band 3), 10.3-11.3 μm (band 4) and 11.5-12.5 μm (band 5). It provides low spatial resolution high radiometric resolution multispectal data for the entire surface of the earth on a daily basis (and at a low cost). It is planned for continued operation through the 1990's. The applications of AVHRR for environmental sciences are discussed in 7.6.3.

AVHRR offers local area coverage (LAC), global area coverage (GAC) and a global vegetation index data set (GVI). The LAC data have a resolution of 1.1 km (at nadir). As the name implies, these data are available for geographically limited areas on any one day. The GAC are generated from the 1.1 km resolution data on a daily basis for the whole globe. The GAC is obtained from LAC values by taking 4 out of 5 pixels in 1 out of 3 lines, which are averaged to represent GAC. The resulting resolution is approximately 4 km. This procedure is far from ideal, but in future more satisfying methods will be introduced.

The global vegetation index (GVI) data set is generated on receipt of the GAC data, whereby one GAC pixel is selected to represent the value of 4 pixels. The resulting resolution is about 8 km at the equator. The normalized difference vegetation index (NDVI) is calculated from the red and near infra red bands according to: NDVI = (band 2- band 1)/(band 2 + band 1). This ratio yields a measure of photosynthetic capacity such, that the higher the value of the ratio, the more photosynthetically active the cover type (Sellers, 1985). For specific periods of time (e.g. 7 days) a technique of maximum value compositing (MVC) is applied to correct for cloud contamination of the image. For specific geographic regions the highest NDVI-pixel is retained. The maximum value compositing technique is such, that near nadir pixels will usually be chosen, thereby reducing atmospheric attenuation and surface directional effects (Holben, 1986). On a continental broad sc ale there is close similarity between NDVI images obtained for different years in the same month. Sellers (1985) found a near linear relation between NDVI and IPAR (intercepted photosynthetically active radiation), but concluded that NDVI is an insensitive measure to estimate the leaf area index or biomass in the following cases:

- when the leaf area index exceeds 2 or 3;
- when there are patches of bare soil in the sensor field of view;
- when there is an unknown quantity of dead material in the canopy; or:
- when the leaf angle distribution is unknown and the solar elevation is high.

The NOAA-AVHRR (and also the Coastal Zone Color Scanner, CZCS, on board the NIMBUS-7) do not suffer from LANDSAT's limitations with respect to the repeat cycle. NOAA and NIMBUS have cycles of 1/2 day and 6 days respectively. The sensors of AVHRR, LANDSAT and CZCS, with spectral responses in the visible and infra red regions of the energy spectrum, are found to respond comparably to incident radiation from agricultural targets (Cicone and Metzler, 1984). The principal variation in the signals of the 3 sensors is found to reside in two dimensions, that are highly correlated between sensors. These dimensions, called brightness and greenness, are related to targ et albedo and vegetative green leaf biomass.
The potential of AVHRR and CZCS for assessment of overall vegetation condition on a large area or small scale basis may exceed that of LANDSAT due to favourable temporal and data volume

attributes. The spatial resolution of AVHRR and CZCS (1100 m and 825 m resp.) however, do not favourably compare to MSS at 79 m resolution.

7.2.4 Radar systems (SLAR, SAR)

Radar operates in the microwave area of the electromagnetic spectrum. In radar radio signals are transmitted from a radar antenna, whereby the length of time required for the signal to travel to the target and be reflected back to the antenna, allows the distance of the target to be determined. The signals are pulses lasting only 10^{-7} seconds producing a range resolution of 15 m (range resolution is perpendicular to the flight-line). The along track resolution is proportional to the width of the beam of the microwave signal, which is inversely proportional to the length of the antenna.

The radar equation relating transmitted and received power shows a dependence of the received power on the fourth power of the distance. This is a severe limitation for satellite-borne remote sensing applications, which cannot always be overcome by increasing antenna aperture (Schanda, 1986). The antenna length can be artificially improved through the synthetic aperture systems (SAR), which have an along track resolution independent of the range, allowing for high resolution imagery of objects miles away. This phenomenon is due to the fact, that with increasing target distance the object remains in the beam of the antenna for a longer period of time. Thus, the effective length of the synthetic antenna is proportional to the range to the target, and the resolution is inversely proportional to the range. The result is a constant resolution.

One advantage of radar is the all-weather and day-or night capability. Radar systems are usually side looking (SLAR; Side Looking Air-borne Radar), viewing the terrain from an oblique angle. For vegetation mapping differences between physiognomic classes can be enhanced; differences in moisture content also influence the radar signal reflection (reflection of certain bands of microwaves is influenced by the dielectric properties of the medium, which in turn are strongly related to moisture content). Radar patterns are marked by the presence of 'shadows' due to the reflection characteristics of the objects (facetted rocks, railways or even wiring). Areas behind tall terrain features facing the radar antenna often are in a radar 'shadow'. These objects do not return the radar signal and will appear black on the imagery. Mountainsides and slopes facing the radar antenna provide a much higher return than areas of similar cover types on flat terrain, thereby making the interpretation of radar imagery a difficult task. A basic rule for using radar imagery for change detection is radar look direction consistency for features having a predominant orientation (urban areas, cultivated areas), situated on non-level land or surfaces composed of non-isotropic scatterers. The interpreter needs to have a basic understanding of the nature of the terrain and of the nature of the interaction of radar energy and the surfaces imaged.

Changes such as recent clearing of forest areas are very distinct on radar imagery. Some wavelength bands penetrate vegetation which may be an important feature for the mapping of topographic features in forest areas. Synthetic aperture radar enables a high spatial resolution, e.g. 10m at a distance of 100km. Current radar systems usually operate in the K- and X-band of approximately 0.83-2.8 and 2.8-5.2 cm wavelength.

About 1200 LANDSAT scenes will be required to cover the tropical forest area of the earth. Many of these areas have never been imaged, because of the pervasive cloud cover. This gap can be filled with an active all-weather system and with coarse spatial resolution scanners (AVHRR) with very short repeat cycles.

The registration of SAR imagery to maps and to LANDSAT images is not simple due to differences in the appearance of tie-points and the relief distortion due to large off-nadir angles. Efforts are under way to solve these difficulties.

A number of unexplored possibilities of microwave sensing are (adapted from Billingsley, 1984):
- measurement of stand density, which may show better on radar imagery than on LANDSAT-MSS imagery.
- area delineation
- species differentiation
- tropical forest inventory
- tree height
- detection of tree stress and susceptibility to fire
- measurement and identification of interrelationships between soil type, soil moisture, surface roughness and vegetation cover

7.3 APPLICATIONS OF REMOTE SENSING FOR ESTIMATION OF EVAPOTRANSPIRATION, BIOMASS PRODUCTION AND SOIL MOISTURE

7.3.1 Estimation of the soil moisture status

No operational satellite measures any aspect of the soil water balance directly at the present time. Geostationary satellites like METEOSAT can provide information about the precipitation and the soil moisture status in dryland areas. Information about the soil moisture condition can be derived from the thermal inertia and surface reflection (Milford, 1987). Thermal inertia is the ease by which the soil surface temperature is changed by heating from above. It is derived from measurements of the diurnal temperature and heat flux variations at the surface. Surface reflection changes on drying or wetting and is of no use for operational monitoring, because the shallow surface layer, which is observed in remote sensing, dries out very quickly after rain. The following equation describes the method:

$$B = A \times DHC \tag{7.1}$$

where:
A = amplitude of the surface temperature variation (K)
B = amplitude of the diurnal variation of the heat flux into the soil
DHC= diurnal heat capacity

The amplitude A can be estimated from thermal infrared radiometer data, the heat flux is found as the residual in the surface energy balance (equation 3) or it may be estimated as a constant fraction of R_n or of the total solar irradiation. With equation (7.1) the value of DHC can be calculated. From a run of DHC values of e.g. a year, the extreme values will give information on the dry conditions (minimum diurnal heat flux) and about field capacity (maximum diurnal heat flux). Only when coupled with ground data on soil moisture profiles, interpretations can be made. This method has only potential in dryland areas with a minimum of both clouds and vegetation and in these regions the method can yield a classification of the water content in the upper topsoil. Combined with information from other satellites (e.g. on changes in biomass) or with surface observations the thermal inertia method might be improved (Milford, 1987).
Wetzel and Woodward (1987) used infrared surface temperature observations taken from the GOES-satellite (Geostationary Operational Environmental Satellites) to predict the soil moisture status. The morning surface temperature change turned out to be especially sensitive to the soil moisture status. The temperature change was linearly related to the square of the soil moisture deficit and to the remotely sensed (NOAA-AVHRR) vegetation index NDVI.

7.3.2 Estimation of evaporation

The evapotranspiration of water from vegetated surfaces is one of the less understood aspects of the hydrological cycle. One reason for this is, that the measurement of evapotranspiration at a regional scale is very difficult. In recent years agricultural remote sensing has been mainly concerned with developing fundamental relationships for assessing plant condition and development on the basis of the emitted and reflected radiation from the plant canopy. Emitted thermal radiation from plant canopies has been related to evapotranspiration and plant water status (Monteith and Szeicz, 1962; Idso et al, 1977). Emitted thermal infrared and reflected visible and near infrared have been used to estimate total phytomass production (Asrar et al, 1985) and to asses evapotranspiration (Jackson et al, 1977; Nieuwenhuis et al, 1985; Reginato et al, 1985).

A relation between evaporation and crop canopy temperature can be derived from the energy balance equation (6.1). Combining equations 6.1, 6.2 and 6.7 (chapter 6) the relation between evapotranspiration and crop temperature T_c can be found:

$$LE = \rho \times C_{p} \times \frac{T_a - T_c}{r_{ah}} + (1-a_0)\, R_{sw} + R_{sky} - \Sigma x k x T_c^4 - G_E \qquad (7.2)$$

T_c can be remotely sensed by thermal infrared remote sensing. When T_a, r_{ah}, a_0, R_{sw}, R_{sky}, E and G_E are known (or estimated) LE can be calculated. The turbulent diffusion resistance to heat transport r_{ah} depends on wind velocity, roughness length z_0 of the crop surface and atmospheric stability. In general T_a, R_{sw}, R_{sky} and the wind velocity may be taken to be constant over a regional area, implicating that standard meteorological measurements can be used.

A major limitation of equation (7.2) is that calculating sensible heat flux from temperature differences between the air and plant leaves is only valid for uniformly evaporating surfaces. The formula will not yield satisfactory results in the case of partial cover, e.g. shrubs whose roots may reach a water table and transpire at potential, with large areas of bare soil between the shrubs. For non-uniform surface conditions it may not be possible to specify the factors required to compute sensible heat flux with sufficient accuracy to yield acceptable values of evaporation (Jackson et al, 1987).

The remote measurement of radiation temperature is strongly influenced by the atmosphere. But the absolute values of crop radiation temperatures are of little importance. Nieuwenhuis et al (1985) state, that differences in radiation temperature are a practical tool for determining regional evapotranspiration.

Jackson et al (1977) proposed to relate the 24 hour evaporative flux LE^{24} to 24 hour net radiation R^{24} and the instantaneous temperature difference near midday $(T_c-T_a)^i$, i.e.

$$LE^{24} = R_n^{24} - B\,(T_c-T_a)^i \quad (W\,m^{-2}) \qquad (7.3)$$

where: B = calibration constant ($W\,m^{-2}\,K^{-1}$)

Equation (7.3) was developed for Phoenix, Arizona (USA). Nieuwenhuis et al (1985) propose a simpler model for the Netherlands with intermittent cloudiness may occur frequently:

$$LE^{24} = LE_p^{24} - B'\,(T_c-T_c^*)^i \quad (W\,m^{-2}) \qquad (7.4)$$

where:
LE_p^{24} = potential 24-hour evapotranspiration
T_c^* = temperature of a crop evaporating at the potential rate
B' = calibration constant ($W\,m^{-2}\,K^{-1}$)

LE_p^{24} can be calculated according to various methods (e.g. Monteith, 1973, quoted in Nieuwehuis et al, 1985). The calibration constant is calculated on the basis of the assumption that the ratio of daily evaporation and instantaneous evaporation equals the ratio of daily and instantaneous net radiation and that windspeed is relatively constant over the daily period (Jackson et al, 1987). This allows for extrapolation of instantaneous to daily evaporation and yield satisfactory results.

Reginato et al (1985) evaluated the evapotranspiration from cropped land by combining remotely sensed reflected solar radiation (using a multiband radiometer) and surface temperatures with ground meteorological data (incoming solar radiation, air temperature, windspeed and vapour pressure) and to calculate net radiation and sensible heat flux G_E as a fraction of net radiation. (G_E = [0.1-0.042h]R_n, h being crop canopy height after Clothier et al., 1986). They compared remotely sensed evaporation with measured water extraction rates. Their results suggest, that ET maps of relatively large areas could be made using this method with data from airborne sensors. The extent of the area covered appeared to be limited by the distance over which air temperature and windspeed can be extrapolated. These results are confirmed by observations made by Desjardins et al. (1986) and Schuepp et al. (1987) who measured sensible heat flux densities using aircraft based eddy correlation techniques. In their experiments repeated passes in 4 distances were used to examine the variability of flux densities over areas of 20-500 km^2.

7.3.3 Estimation of phytomass production

Desjardins et al. (1987) showed that photosynthesis measured as CO_2 flux, measured with airborne sensors are related to vegetation indices (0.6-0.7 μm band/0.7-0.8 μm band; or red/near infrared) from LANDSAT D-MSS.

Asrar et al (1985) estimate phytomass production by:

$$M = \sum_{i=1}^{n} e_c \times e_i \times e_s \times S \times C \qquad (7.5)$$

where:
- M = dry phytomass
- e_c = photochemical efficiency factor
- e_i = fraction absorbed PAR(photosynthetically active portion of solar radiation)
- e_s = fraction of energy in the PAR region of the electromagnetic spectrum
- S = total incident solar radiation(J m^{-2} d^{-1})
- C = crop stress index (ratio actual : potential evapotranspiration)

The factor C is a function of temperature (air and canopy temperature), windspeed and net radiation. In this way a combination of measured reflected visible and near infrared and emitted thermal radiation (see equation 6.2 for the calculation of R_n) is used to estimate total above ground phytomass production. The results of this method showed, that at sparse canopies the yield estimates were high due to improper partitioning of solar energy in the radiation and energy balances of the canopy, but with dense canopies the method proved very accurate (Asrar et al, 1985).

7.3.4 Global vegetation monitoring and estimation of primary biomass production

The NOAA-AVHRR (see 7.2.3) is the primary sensor for monitoring of land cover. Although the spatial resolution is low (1.1 km at nadir), thanks to the repeat cycle of 1 day there will mostly be a cloud-free coverage of any geographic area for a given period of time.

Justice et al (1985) showed, that the full resolution LAC images can be used successfully to delineate deforestation in Rondonia, Brazil. Individual fields however, could not be delineated. Nelson and Holben (1986) used AVHRR LAC data for delineating colonization clearings in Rondonia with Landsat MSS data as a ground reference. When and where available MSS data can provide the thresholds necessary for discrimmination of the cleared areas from forest. Nelson and Holben also found, that GOES-VISSR imagery is of little value in this respect due to excessive data noise. Malingreau and Tucker (1988) selected the best NOAA-LAC images for each year and used channel 3 of NOAA (3.5-3.9 μm). Tentative classifications in forest-nonforest categories were compared with maps, radar images, Landsat and Space Shuttle images and field checks and the extent of deforestation could be determined within Acre, Rondonia and Mato Grosso (Amazon Basin, Brasil). Their results were also discussed in chapter 3.

AVHRR data for vegetation monitoring can be interpreted only on the basis of a thorough knowledge of the distribution of the cover types, since the same normalized difference vegetation index (NDVI) value may represent very different conditions for different vegetation communities (Townshend et al, 1986). If the NDVI- data are integrated over time, the results can be correlated with the total amount of biotic activity during the integrating period (Tucker et al, 1983; Tucker et al, 1985; Justice et al, 1986). A high value of the annual integrated NDVI corresponds to high net primary production (Sellers, 1985; Holben, 1986). In comparing the integrated NDVI image with the ecoclimatic map of East Africa (Pratt and Gwynne, 1977), Justice et al (1986) found a good correspondence. Especially in semi-arid vegetation types with a considerable interannual variation controlled primarily by rainfall amount and distribution, the AVHRR data will give a good indication of vegetation conditions. AVHRR data can also be used to estimate the length of the growing season by assuming a threshold value for discrimminating between presence and absence of photosynthesis and counting the number of days with NDVI values higher than the threshold (Justice et al, 1986). Hatfield et al (1984) found a good correlation between NDVI and greenness (calculated as a function of the MSS bands 4,5,6 and 7) on one side and IPAR (Intercepted Photosynthetically Active Radiation). NDVI correlated significantly better with IPAR than greenness for all planting dates.

At vegetation covers between 20 and 75% Huete et al (1985) found that greenness became strongly dependent upon soil background effect. This background is composed of both a soil spectral and a soil brightness effect. Normalization of soil background to a constant ratio or a perfect one-dimensional soil line only removed bare soil spectral influences and not the greater soil brightness influence.

Tucker et al (1986) related the AVHRR derived NDVI to the seasonal atmospheric variation of the CO_2 concentration as measured at Point Barrow, Alaska and Mauna Loa, Hawai. The monthly variations in atmospheric CO_2 concentrations and terrestrial NDVI correlate well. The authors concluded, that refining of the model with oceanic uptake and release, respiration and decomposition processes will greatly enhance the understanding of the global carbon cycle.

Gutman (1988) proposed that AVHRR data can be used to estimate surface albedo values. For that purpose AVHRR data were reorganized by 9 day repeat cycle, screened for clouds, averaged and atmospheric corrections were carried out.

7.4 CONCLUSIONS

Two factors, that influence the global climate, deforestation and desertification, can be monitored by remote sensing. Accuracy in identifying changes in vegetation cover may be enhanced by proper selection of spectral bands. Narrower bands will be available in future. LANDSAT-4 provides improved sensing spatially and spectrally. For global monitoring the NOAA-AVHRR sensor data are highly suited and new AVHRR applications are being researched.

The low spatial resolution data of the AVHRR can be supplemented with high spatial resolution MSS data when and where available. SAR (synthetic aperture radar) promises all weather sensing, although operational spacecraft are still in the future and proper correlation with e.g. LANDSAT is still in development. Deterministic remote sensing models to estimate biomass and evapotranspiration are being refined and have already proved their value for the environmental sciences. Calibration is still a problem in multitemporal comparison.

For estimation of regional evaporation from agricultural land and grassland remote sensing techniques are available. However, for surfaces not uniformly evaporating, i.e. surfaces with partial coverage, and areas where evaporation is dependent on other factors than net radiation only (such as forest areas), remote sensing techniques are as limited value as their basic theoretical equations.

As a tool for estimating stand density of vegetation, species differentiation and forest inventory,microwave sensing is very promising but needs further exploration.

Remote sensing techniques for measuring (trace) gas concentrations are available. However, for most gases this technique is not suited (yet) since the concentration gradients usually show a sharp drop at short distance above the emitting surface. Measurements usually take place at greater heights above the surface and not in the track where the gradient is.

REFERENCES

Ajtay, G.L, P. Ketner and P. Duvigneaud (1979): Terrestrial primary production and Phytomass. In: Bolin, B., E.I. Degens, S. Kempe and P. Ketner (Eds.): The global carbon cycle, p 129-181. SCOPE Vol. 13. Wiley and Sons, New York.

Al Abbas, A.H., P.H. Swain and M.F. Baumgardner (1972): Relating organic matter and clay content to the multispectral radiance of soils. Soil Science 114:477-485.

Alexander, M. (1977): Introduction to soil microbiology (2nd edition). 467 p. Wiley and Sons, New York.

Allison, F.E. (1973): Soil organic matter and its role in crop production. Developments in Soil Science 3. 637 pp. Elsevier, Amsterdam.

Andersen, B. and L. Moseholm (1987): Effects on forests and other vegetation. In: H.M. Seip, Acid precipitation literature review, Nordisk Ministerrad, Kopenhagen, Miljorapport 1987:5.

Anderson, I.C. and J.S. Levine (1986): Relative rates of nitric oxide and nitrous oxide production by nitrifiers, denitrifiers and nitrate respirers. Applied Environmental Microbiology 51:938-945

Anderson, I.C. and Levine J.S. (1987): Simultaneous field measurements of nitric oxide and nitrous oxide. Journal of Geophysical Research 92:965-976.

Andreae, M.O. and D.S. Schimel (1989): Exchange of trace gases between terrestrial ecosystems and the atmosphere. Report of Dahlem workshop, February 19-24, 1989. (in preparation).

Andriesse, J.P. (1988): Nature and Management of tropical ped soils. 165 pp. FAO Soils Bulletin 59, FAO, Rome.

Armentano, T.V. (1980): Drainage of organic soils as a factor in the world carbon cycle. BioScience 30:825-830.

Armentano, T.V. and E.S. Menges (1981): Patterns of change in the carbon balance of organic soil-wetlands of the temperate zone. Journal of Ecology 74:755-774.

Armstrong, A.S.B. (1983): Nitrous oxide emissions from two sites in southern England during winter 1981/1982. Journal of Science Food and Agriculture 34:803-807.

Aselmann, I. and P.J. Crutzen (1989): Freshwater wetlands: Global distribution of natural wetlands and rice paddies: their net primary productivity, seasonality and possible methane emissions. Accepted for publication in Journal of Atmospheric Chemistry.

Asrar, G. and E.T. Kanemasu (1985): Estimation of total above ground phytomass production using remotely sensed data. Remote Sensing of Environment 17:211-220.

Austin, L.B., P.H. Schuepp and R.L. Desjardins (1987): The feasibility of using airborne CO_2 flux measurements for the imaging of the rate of biomass production. Agricultural and Forest Meteorology 39:13-23.

Banerjee, A.K. (1972): Evapotranspiration from a young Eucalyptus hybrid plantation of West Bengal. In: Proceedings and Technical papers Symposium man made forests in India (Society of Indian Foresters), quoted in: Unesco (1978).

Bartholomew, G.W. and Alexander, M. (1981): Soils as a sink for atmospheric carbon monoxide. Science 212:1389-1391.

Bartlett, K.B., D.S. Bartlett, R.C. Harriss and D.I. Sebacher (1987): Methane emissions along a salt marsh salinity gradient. Biogeochemistry 4:183-202.

Bartlett, K.B., P.M. Crill, D.I. Sebacher, R.C. Harriss, J.O. Wilson and J.M. Melack (1988): Methane flux from the central Amazonian floodplain. Journal of Geophysical Research 93:1571-1582

Bartlett, K.B., R.C. Harriss and D.I. Sebacher (1985): Methane flux from coastal salt marshes. Journal of Geophysical Research 90:5710-5720.

Baumgardner, M.F., S.J. Kristof, C.J. Johanson and A.L. Zachary (1970): Effects of organic matter on the multispectral properties of soils. Indiana Academy of Sciences Proceedings 79:413-422.

Baumgardner, M.F., L.F. Silva, L.L. Biehl and E.R. Stoner (1985): Reflectance properties of soils. Advances in Agronomy 38:1-44.

Baumgartner, A. (1965): Energetical bases for differential vaporization from forest and agricultural land. In: Proceedings International Symposium For. Hydr., p 381-389. London, Pergamon Press.

Baumgartner, A. (1970): Water and energy balances of different vegetation covers. In: Proceedings IASH-Unesco Symposium World Water balance, p 56-65. Reading, 1970.

Baumgartner, A. and N. Kirchner (1980): Impacts due to deforestation. In: W. Bach, S. Pankrath and J. Williams (Eds.), Interactions of energy and climate, p 305-317. Reidel, Dordrecht.

Baumgartner, A. and E. Reichel (1975): The world water balance. Mean annual global, continental and maritime precipitation, evaporation and runoff. 179 pp. R. Oldenburg verlag, München, Wien.

Becker, F., F. Nerry, P. Ramanantsizehena and M.P. Stoll (1985). Emissivité dans l'infrarouge thermique: conditions d'une mesure par retrodiffusion et comparaison avec une méthode de mesure passive. In: Proceedings 3rd International Colloquium on Spectral Signatures of Objects in Remote Sensing, Les Arcs, France, 16-20 December 1985.

Becker, F., F. Nerry, P. Ramanantsizehena and M.P. Stoll (1987): Importance of a remote measurement of spectral infrared emissivities; presentation and validation of such a determination. Advanced Space Research 7:121-127.

Billingsley, F.C. (1984): Remote sensing for monitoring vegetation: an emphasis on satellites. In: G.M. Woodwell (Ed.), The role of terrestrial vegetation in the global carbon cycle. Measurement by remote sensing, p 161-180. SCOPE Vol. 23. Wiley and Sons, New York.

Bingemer, H.G. and P.J. Crutzen (1987): The production of methane from solid wastes. Journal of Geophysical Research 92:2181-2187.

Blackmer, A.M., S.G. Robbins and J.M. Bremner (1982): Diurnal variability in rate of emission of nitrous oxide from soils. Soil Science Society of America Journal 46:937-942.

Blake, D.R., Mayer, E.W., Tyler, S.C., Makide, Y., Montague, D.C. and F.S. Rowland (1982): Global increase in atmospheric methane concentrations between 1978 and 1980. Geophysical Research Letters 9:477-480.

Blake, D.R. and F.S. Rowland (1986): World wide increase in troposphere methane, 1978-1983. Journal of Atmoshperic Chemistry 4:43-62.

Bohn, H.L. (1976): Estimate of organic carbon in world soils II. Soil Science Society of America Journal 40:468-470.

Bolin, B. (1977): Changes of land biota and their importance for the carbon cycle. Science 196:613-615.

Bolin, B. (1986): How much CO2 will remain in the atmosphere ? The carbon cycle and projections for the future. In: Bolin, B., B.R. Döös, J. Jager, and R.A. Warric (Eds.), The greenhouse effect, climatic change and ecosystems, p 93-156. SCOPE Vol. 29. Wiley and Sons, New York.

Bolin, B., E.T. Degens, P. Duvigneaud and S. Kempe (1979): The global carbon cycle. In: Bolin, B., E.T. Degens, S. Kempe and P. Ketner (Eds.), The global carbon cycle, p 1-56. SCOPE Vol. 13. Wiley and Sons, New York.

Bolle, H.J., W. Seiler and B. Bolin (1986): Other greenhouse gases and aerosols. Assessing their role in atmospheric radiative transfer. In: Bolin, B., B.R. Döös, J. Jager and R.A. Warrick (Eds.), The greenhouse effect, climatic change and ecosystems, p 157-203. SCOPE Vol.29, Wiley and Sons, New York

Bont, J.A.M. de, K.K. Lee and D.F. Bouldin (1978): Bacterial oxidation of methane in a rice paddy. Ecological Bulletin 26:91-96.

Bormann, F.H., G.E. Likens, I.G. Siccana, R.S. Pierce and J.S. Eaton (1974): The export of nutrients and recovery of stable conditions following deforestation at Hubbard Brook. Ecological Monographs 44:255-277.

Bosch, J.M. and J.D. Hewlett (1982): A review of catchment experiments to determine the effect of vegetation changes on water yield and evapotranspiration. Journal of Hydrology 55:3-23.

Bouchet, R.J. (1963): Evapotranspiration reelle et potentielle, signification climatique. Proceedings IAHS 62:134-142.

Bouten, W., F.M. de Vre, J.M. Verstraten and J.J.H.M. Duysings (19..): Carbon dioxide in the soil atmosphere: simulation model parameter estimation from field conditions. Fysisch Geografisch Laboratorium, Dapperstraat 115, Amsterdam.

Bouwman, A.F. (1989a): The role of soils and land use in the Greenhouse Effect. Netherlands Journal of Agricultural Science 37:13-19.

Bouwman, A.F. (1989b): Modelling soil organic matter decomposition and rainfall erosion in two tropical soils after forest clearing for permanent agriculture. Accepted by Land Degradation & Rehabilitation.

Bouwman, A.F. (1989c): Land evaluation for rainfed Agriculture using global terrain information on a 1° longitude x 1° latitude grid. ISRIC Working Paper and Preprint 89/1. ISRIC, Wageningen.

Bowden, W.B. and F.W. Bormann (1986): Transport and loss of nitrous oxide in soil water after forest clear cutting. Science 233:867-869.

Brams, E.A. (1971): Continuous cultivation of West African soils: organic matter dimunition and effects of applied lime and phosphorous. Plant and Soil 35:401-414.

Breitenbeck, G.A., A.M. Blackmer and J.M. Bremner (1980): Effects of different nitrogen fertilizers on emission of nitrous oxide from soil. Geophysical Research Letters 7:85-88.

Bremner, J.M. and A.M. Blackmer (1978): Nitrous oxide: emission from soils during nitrification of fertilizer nitrogen. Science 199:295-296.

Bremner, J.M. and A.M. Blackmer (1981): Terrestrial nitrification as a source of atmospheric nitrous oxide. In: C.C. Delwiche (Ed.), Denitrification, Nitrification and atmospheric nitrous oxide, p 151-170. Wiley and Sons, New York.

Bremner, J.M., S.G. Robbins and A.M. Blackmer (1980): Seasonal variability in emission of nitrous oxide from soil. Geophysical Research Letters 7:641-644.

Brockman-Jerosch (1978): Die Einteilung der Pflantzengesellschaften nach ökologisch-physiognomischen Geschichtspunkten. Verlag Wilhelm Engelmann, Leipzig, 68 pp.

Broecker, W.S., T. Takahashi, H.J. Simpson and T.H. Peng (1979): Fate of fossil fuel carbon dioxide and the global carbon budget. Science 206:409-418.

Brown, S. and A.E. Lugo (1982): The storage and production of organic matter in tropical forests and their role in the global carbon cycle. Biotropica 14:161-187.

Brown, S. and A.E. Lugo (1984): Biomass of tropical forests: a new estimate based on forest volumes. Science 223:1290-1293.

Brutsaert, W.H. (1975): On a derivable formula for long-wave radiation from clear skies. Water Resources Research 11:742-774.

Bryan, M.L. and J. Clark (1984): Potentials for change detection using Seasat synthetic aperture radar data. Remote Sensing of Environment 16:107-124.

Budyko, M.I. (1958): Heat balance of the earth's surface. Transl. by M.A. Stevenson, Dept. of Commerce, Washington,D.C.

Burch, G.J., R.K. Bath, I.D. Moore and E.M. O'Loughlin (1987): Comparative hydrological behaviour of forested and cleared catchments in Southeastern Australia. Journal of Hydrology 90:19-42.

Buringh, P. (1979): Introduction to the study of soils in tropical and subtropical regions. 124 pp. Centre for Agricultural Publishing and Documentation. Wageningen, The Netherlands.

Buringh, P. (1984): Organic carbon in soils of the world. In: Woodwell, G.M. (Ed.)(1984), The role of terrestrial vegetation in the global carbon cycle. Measurement by remote sensing, p 91-109. SCOPE Vol. 23. Wiley and Sons, New York.

Burford, J.R., R.J. Dowdell and R. Crees (1981): Emission of nitrous oxide to the atmosphere from direct drilled and ploughed clay soils. Journal of the Science of Food and Agriculture 32:219-223.

Burton, M.A.S. (1987): Terrestrial ecosystems and biome types. A background for studying contaminants in global ecosystems. A Research Memorandum. Monitoring and Assessment Research Centre, King's College London, University of London, 42 pp.

Buyanovsky, G.A., G.H. Wagner and C.J. Gantzer (1986): Soil respiration in a winter wheat ecosystem. Soil Science Society of America Journal 50:338-344.

Calder, I.R. (1982): Forest evaporation, p 173-194. Proceedings Can. Hydr. Symposium 1981. National Research Council Canada, Ottawa, Ontario.

Calder, I.R. (1985): What are the limits on forest evaporation - comment. Journal of Hydrology 82:179-184.

Calder, I.R. (1986): What are the limits on forest evaporation - a further comment. Journal of Hydrology 89:33-36.

Calder, I.R. and M.D. Newson (1979): Land use and upland water resources in Brittain- a stategic look. Water Resources Bulletin 15:1628-1639.

Calder, I.R., I.R. Wright and D. Murdiyarso (1986): A study of evaporation from tropical rainforest - West Java. Journal of Hydrology 89:13-31.

Cates, R.L. and D.R. Keeney (1987a): Nitrous oxide emission from native and reestablished prairies in southern Wisconsin. The American Midland Naturalist 117:35-42.

Cates, R.L. and D.R. Keeney (1987b): Nitrous oxide production throughout the year from fertilized and manured maize fields. Journal of Environmental Quality 16:443-447.

Cicerone, R.J. and R.S. Oremland (1988): Biogeochemical aspects of atmospheric methane. Global Biogeochemical Cycles 2:299-327.

Cicerone, R.J., J.D. Shetter and C.C. Delwiche (1983): Seasonal variation of methane flux from a California rice paddy. Journal of Geophysical Research 88:11022-11024.

Cicone, R.C. and M.D. Metzler (1984): Comparison of Landsat MSS, Nimbus-7 CZCS and NOAA-7 AVHRR features for land use analysis. Remote Sensing of Environment 14:257-265.
Clark, W.C. (1986): Sustainable development of the biosphere: themes for a research program. In: W.C. Clark and R.E. Munn (Eds.), Sustainable development of the biosphere, p 5-48. IIASA, Laxenburg, Austria. Cambridge University Press.
Clements, F.E. (1916): Plant succession. An analysis of the development of vegetation. 512 pp. Carnegie Institute, Washington, D.C.
Clements, F.E. (1928): Plant succession and indicators. 453 pp. H.W. Wilson Co., New York, London.
Clothier, B.E., K.L. Clawson, P.J. Pinter Jr., M.S. Moran, R.J. Reginato and R.J. Jackson (1986): Estimation of soil heat flux from net radiation during the growth of alfalfa. Agricultural and Forest Meteorology 37:319-329.
C.O.L. (Commissie Onderzoek Luchtverontreiniging van de Vereniging Lucht Troposferische Chemie) (1987): Onderzoekprogramma t.b.v. de raad voor Milieu en Natuuronderzoek.
Cochran, V.L., L.F. Elliot and R.I. Papendick (1980): Nitrous oxide emissions fro ma fallow field fertilized with anhydrous ammonia. Soil Science Society of America Journal 45:307-310.
Colbourn, P. and I.W. Harper (1987): Denitrification in draied and undrained arable clay soil. Journal of Soil Science 38:531-539.
Colbourn, P., M.M. Iqbal and I.W. Harper (1984a): Estimation of the total gaseous nitrogen losses from clay soils under laboratory and field conditions. Journal of Soil Science 35:11-22.
Colbourn P., I.W. Harper and M.M. Iqbal (1984b): Denitrification losses from 15N labelled calcium fertilizer in a clay soil in the field. Journal of Soil Science 35:539-547.
Colbourn, P., J.C. Ryden and G.J. Dollard (1987): Emission of NOx from urine-treated pasture. Environmental Pollution 46:253-261.
Collins, N.M. and Wood, T.G. (1984): Termites and atmospheric gas production. Science 224:84-86.
Conrad, R. and W. Seiler (1982): Arid soils as a source of atmospheric carbon monoxide. Geophysical Research Letters 9:1353-1356.
Conrad, R., W. Seiler and G. Bunse (1983): Factors influencing the loss of fertilizer nitrogen in the atmosphere as N_2O. Journal of Geophysical Research 88:6709-6718.
Coster, C. (1937): De verdamping van verschillende vegetatie vormen op Java. Tectona 30:1-102. Quoted in Unesco (1978).
Coulson, K.L.and D.W. Reynolds (1971): The spectral reflectances of natural surfaces. Journal of Applied Meteorology 10:1285-1295.
Crapper, P.F. and K.C. Hynson (1983): Change detection using Landsat photographic imagery. Remote Sensing of Environment 13:291-300.
Crill, P.M., K.B. Bartlett, J.O. Wilson, D.I. Sebacher, R.C. Harriss, J. M. Melack, S. MacIntyre, L. Lesack and L. Smith-Morrill (1988): Tropospheric methane from an amazonian floodplain lake. Journal of Geophysical Research 93:1564-1570.
Crutzen, P.J. (1981): Atmospheric Chemical Processes of the oxides of nitrogen, including nitrous oxide. In: C.C. Delwiche (Ed.), Denitrification, nitrification and nitrous oxide, p 17-44. Wiley and Sons, New York.
Crutzen, P.J. (1983): Atmospheric interactions in homogeneous gas reactions of C, N and S containing compounds. In: Bolin, B. and R.B. Cook (Eds.): The major biogeochemical cycles and their interactions, p 67-112. SCOPE Vol. 21, Wiley and Sons, New York.
Crutzen, P.J., I. Aselmann and W. Seiler (1986): Methane production by domestic animals, wild ruminants, other herbivorous fauna and humans. Tellus 38B:271-284.
Crutzen, P.J. and L.T. Gidel (1983): A two dimensional model of the atmosphere 2: The tropospheric budgets of anthropogenic chlorocarbons, CO, CH4, CH3Cl and the effect of various NOx sources on tropospheric ozone. Journal of Geophysical Research 88:6641-6661.
Crutzen, P.J. and Graedel, T.E. (1986): The role of atmospheric chemistry in environment-development interactions. In: W.C. Clark and R.E. Munn (Eds.), Sustainable development of the biosphere, p 213-250. IIASA, Laxenburg, Austria. Cambridge University Press.
Crutzen, P.J., L.E. Heidt, J.P. Krasnec, W.H. Pollock and W. Seiler (1979): Biomass burning as a source of atmospheric CO, H_2, N_2O, NO, CH_3Cl and COS. Nature 282:253-256.
Curran, P.J. (1986): Principles of remote sensing. Longman, London, New York.
Dam Kofoed, A. (1982): Humus in long term experiments in Denmark. In: Boels, D., D.B. Davies and A.E. Johnston (Eds.), Soil degradation, p 241-258. Balkema, Rotterdam.
De Bruin, H.A.R. (1983): A model for the Priestley-Taylor parameter α. Journal of Climate and Applied Meteorology 22:572-578.
De Bruin, H.A.R. (1987): From Penman to Makkink. In: J.C. Hooghart (Ed.), Evaporation and Weather, Proceedings and Information No. 39, p 5-31. TNO Committee on Hydrological Research, The Hague.
De Bruin, H.A.R. (1988): Evaporation in arid and semi-arid regions. In: I. Simmer (Ed.), Estimation of Groundwater Recharge, p 73-88. D. Reidel Publ. Comp., Dordrecht.
De Bruin, H.A.R. and A.A.M. Holtslag (1987): Evaporation and weather: interactions with the planetary boundary layer. In: J.C. Hooghart (Ed.), Evaporation and Weather, Proceedings and Information No. 39, p 63-83. TNO Committee on Hydrological Research, The Hague.
Delany, A.C., D.R. Fitzjerrald, D.H. Lenschow, R. Pearson, C.J. Wendel and B. Woodruff (1986): Direct measurements of nitrogen oxides and ozone fluxes over grasslands. Journal of Atmospheric Chemistry 4:429-444.
Delaune, R.D., C.J. Smith and W.H. Patrick Jr. (1983): Methane release from Gulf coast wetlands. Tellus 35B:8-15.
Delaune, R., C.J. Smith and W.H. Patrick Jr. (1986): Methane production in Mississipi River deltaic plain peat. Organic Geochemistry 9:193-197.
Denmead, O.T. (1983): Micrometeorological methods for measuring gaseous losses of nitrogen in the field. In: J.R. Freney and J.R. Simpson (Eds.), Gaseous loss of nitrogen from plant-soils systems, Developments in plant and soil science Vol. 9, p 133-157. Martinus Nijhoff/Dr. W. Junk Publishers.
Denmead, O.T. (1983): Micrometeorological methods for measuring gaseous losses of nitrogen in the field. In: J.R. Freney and J.R. Simpson (Eds.), Gaseous loss of nitrogen from plant soil systems, p 133-158. Martinus Nijhoff, The Hague.
Denmead, O.T., J.R. Freney and J.R. Simpson (1976): a closed ammonia cycle within a plant canopy. Soil Biology & Biochemistry 8:161-164.
Denmead, O.T., J.R. Freney and J.R. Simpson (1979a): Nitrous oxide emission during denitrification in a flooded field. Soil

Science Society of America Journal 43:716-718.
Denmead, O.T., J.R. Freney and J.R. Simpson (1979b): Nitrous oxide emission from a grass sward. Soil Science Society of America Journal 43:726-728.
Dent, F.J. (1978): Guidelines for utilizing data bank profile field sheets. Prepared for the land capability Appraisal project at the Soil Research Institute, Bogor, Indonesia. Government of Indonesia, FAO.
Desjardins, R.L., E.J. Brach, J.I. MacPherson, P.H. Schuepp and L. Austin (1986): Regional measurements of evapotranspiration using aircraft mounted sensors. In: Proceedings ISLSCP Conference, Rome, Italy, 2-6 December 1985. p 381-386.
Desjardins, R., A. Mack, I. MacPherson and P. Schuepp (1987): Characterising crop conditions using airborne CO_2 fluxes measurements and Landsat-D MSS Data. In: Preprint of Proceedings of 18th Conference on Agricultural and Forest Meteorology and 8th Conference on Biometeorology and Aerobiology, Sept. 14-18, 1987, p 97-100. American Meteorological Society, Boston, USA.
Detwiler, R.P. and C.A.S. Hall (1988): Tropical Forests and the Global Carbon Cycle. Science 239:42-47.
Detwiler, R.P., C.A.S. Hall and P. Bogdonow (1985): Land use change and carbon exchange in the tropics: II. Estimates for the entire region. Environmental Management 9:335-344.
Devol, A.H., J.E. Richey, W.A. Clark and S.L. Kiny (1988): Methane emissions to the troposphere from the Amazon floodplain. Journal of Geophysical Research 93:1583-1592.
Dickinson, R.E. (1980): Effects of deforestation on climate. In: V.H. Sutlive, N. Altshuler and M.D. Zamora, Blowing in the wind: deforestation and long range implications. Studies in Third World societies. Publ. no. 14. Dept.of Anthropology, College of William and Mary, Williamsburg, Virginia.
Dickinson, R.E. (1986): Evapotranspiration in global climate models. Paper prsented at the 26th COSPAR Meeting, July 2-10 1986, Toulouse, France.
Dickinson, R.E. and A. Henderson-Sellers (1988): Modelling tropical deforestation: a study of GCM land surface parameterizations. Quarterly Journal of the Royal Meteorological Society 114:439-462.
Diez, Th. and G. Bachtaler (1978): Auswirkungen unterschiedlicher Fruchtfolge, Dungung und Bodenbearbeitung auf den Humusgehalt der Boden. Landwirt. Forschung 55:368-377.
Dinel,H., S.P. Mathur, A. Brown and M. Lévesque (1988): A field study on the effect of depth on methane production in peatland waters: equipment and preliminary use. Journal of Ecology 76:1083-1091.
Dolman, A.J. (1988): Transpiration from an oak forest as predicted from porometer and weather data. accepted for publication. Journal of Hydrology 1988.
Dolman, A.J. and W.E. Oosterbaan (1986): Grondwatervoeding, interceptie en transpiratie can de Castricumse boslysimeters. H_2O (19)No.9:174-175.
Dolman, A.J., J.B. Stewart and J.D. Cooper (1988): Predicting forest transpiration from climatological data. Agricultural and Forest Meteorology 42:339-353
Dowdell, R.J. and K.A. Smith (1974): Field studies of the soil atmosphere II. Occurrence of nitrous oxide. Journal Soil Science 25:231-238.
Dowdell, R.J., J.R. Burford, R. Crees (1979): Losses of nitrous oxide dissolved in drainage water from agricultural land. Nature 278:342-343.
Duffy, L., I. Galbally and M. Elsworth (1988): Biogenic NO_x emissions in Latrobe valley. Clean Air 22/4:196-199.
Duvigneaud, P. (1972): Morale et écologie. La Pensée et les hommes 7:286-303.
Duxbury, J.M., D.R. Bouldin, R.E. Terry and R.L. Tate III (1983): Emissions of nitrous oxide from soils. Nature 298:462-464.
Edwards, K.A. and J.R. Blackie (1981): Results of East-African catchment experiments, 1958-1974. In: R. Lal and E.W. Russell (Eds.), Tropical Agricultural Hydrology, p 165-188. Wiley and Sons, New York.
Edwards, N.T. (1975): Effect of temperature and moisture on carbon dioxide evolution in a mixed deciduous forest floor. Soil Science Society of America Journal 39:361-365.
Edwards, N.T. and B.M. Ross-Todd (1983): Soil carbon dynamics in a mixed deciduous forest following clear-cutting with and without residue removal. Soil Science Society of America Journal 47:1014-1021.
Eggington, G.M. and K.A. Smith (1986a): Nitrous oxide emission from a grassland soil fertilized with slurry and calcium nitrate. Journal of Soil Science 1986,37:59-67.
Eggington, G.M. and K.A. Smith (1986b): Losses of nitrogen by denitrification from a grassland soil fertilized with cattle slurry and calcium nitrate. Journal of Soil Science 1986,37:69-80.
Ehhalt, D.H. and U. Schmidt (1978): Sources and sinks of atmospheric methane. PAGEOPH 116:452-464.
Ehhalt, D.H., R.J. Zander and R.N. Lamontagne (1983): On the temporal increase of tropospheric CH4. Journal of Geophysical Research 88:8442-8446.
Ellenberg, H. (1973): Die Oekosysteme der Erde: Versuch einer Klassifikation der Oekosysteme nach funktionalen Gescichtspunkten. In: Ellenberg, H. (Ed.), Ökosystemforschung, p 235-265. Springer Verlag, New York, Heidelberg, Berlin.
Emmanuel, W.R., H.H. Shugart and M.P. Stevenson (1985): Climatic change and the broad-scale distribution of ecosystem complexes. Climatic Change 7:29-43.
Endo, M., K. Minami and S. Fukushi (1986): Effects of interception of near ultraviolet radiation on nitrifier activity and nitrification process in a fertilized andosol under field conditions. Soil Science and Plant Nutrition 32:365-372.
Esser, G. (1987): Sensitivity of global carbon pools and fluxes to human and potential climatic impacts. Tellus 39B:245-260.
Farquhar, G.D., R. Wetselaar and P.M. Firth (1979): Ammonia volatilization from senescing leaves of maize. Science 203:1257-1258.
FAO (1949-1983): Production yearbook Vol.1-37. FAO Statistics series No.55, FAO Rome, Italy.
FAO (1978-1981): Reports of the Agro-Ecological Zones Project. World Soil Resources Project Nº48, Wol.1 - Methodology and Results for Africa, Vol.2 - Southwest Asia, Vol3 - South and Central America, Vol.4 - Southeast Asia. FAO, Rome.
FAO (1983): A physical resource base (map). Technical cooperation among developing countries. Main climatic and soil divisions in the developing world, scale 1:25,000,000. FAO, Rome.
FAO (1985): Fertilizer yearbook, vol. 35. FAO Statistics series No.71, FAO, Rome, Italy.
FAO/UNEP (1981a): Tropical Forest Resources Assessment Project. Forest Resources of tropical Asia. FAO, Rome.
FAO/UNEP (1981b): Tropical Forest Resources Assessment Project. Forest Resources of tropical America. FAO, Rome.
FAO/UNEP (1981c): Tropical Forest Resources Assessment Project. Forest Resources of tropical Africa. FAO, Rome.

FAO/UNESCO (1971-1978): Soil Map of the World 1:5000000. Vol. I-X. FAO, Rome.
FAO/UNESCO (1988): Soil Map of the World. Revised Legend. World Resources Report 60. FAO, Rome.
Fearnside, P.M. (1987): Causes of deforestation in the Brasilian Amazon. In: R.E. Dickinson (Ed.), The Geophysiology of Amazonia. Vegetation and climate interactions, p 37-57. Wiley and Sons, New York.
Fillery, I.R.P. (1983): Biological Denitrification. In: Freney, J.R. and J.R. Simpson (Eds.), Gaseous loss of nitrogen from plant-soil systems. Developments in Soil Science Vol.9. Martinus Nijhoff/Dr.W.Junk Publishers, The Hague.
Fitzpatrick, E.A. (1983): Soils. Their formation, classification and distribution. 355 pp. Longman. London and New York.
Focht, D.D. (1974): The effect of temperature, pH and aeration on the production of nitrous oxide and gaseous nitrogen- a zero order kinetic model. Soil Science 118:173-179.
Folorunso, O.A. and D.E. Rolston (1984): Spatial variability of field measured denitrification gas fluxes. Soil Science Society of America Journal 48:1214-1219.
Fraser, P.J., P Hyson, R.A. Rasmussen and L.P. Steele (1986a): Methane, carbon monoxide and methylchloroform in the southern hemisphere. Journal of Atmospheric Chemistry 4:3-42.
Fraser, P.J., R.A. Rasmussen, J.W. Creffield, J.R. French and M.A.K. Khalil (1986b): Termites and global methane: another assessment. Journal of Atmospheric Chemistry 4:295-310.
Friedman, F. (1977): The amazon basin, another sahel. Science 197:7.
Friedman, H. (1986): The science of global change- an overview. In: T.F. Malone and J.G. Roederer (Eds.), Global Change, p 20-52. ICSU publ. Cambridge University Press.
Froment, A. (1972): Soil respiration in a mixed oak forest. OIKOS 23:273-277.
Galbally, I.E. (1985): The emission of nitrogen to the remote atmosphere: background paper. In: J.M. Galloway et al. (Eds.), The Biogeochemical cycling of sulfur and nitrogen in the remote atmosphere, p 27-53. D. Reidel Publishing Company, Dordrecht.
Galbally, I.E. and R.W. Gillett (1988): Processes regulating nitrogen compounds in the tropical atmosphere. In: H. Rodhe and R. Herrera, Acidification in tropical countries, p 73-116. SCOPE Vol. 36, Wiley and Sons, New York.
Galbally, I.E., J.R. Freney, W.A. Muirhead, J.R. Simpson, A.C.F. Trevitt and P.M. Chatte (1987): Emission of nitrogen oxides (NO_x) from a flooded soil fertilized with urea: relation to other nitrogen loss processes. Journal of Atmospheric Chemistry 5:343-365.
Galbally, I.E. and C.R. Roy (1978): Loss of fixed nitrogen by nitric oxide exhalation. Nature 275:734-735.
Garret, H.E. and G.S. Cox (1973): Carbon dioxide evolution from the floor of an oak-hickory forest. Soil Science Society of America Journal 37:641-644.
Gash, J.H.C. (1979): An analytical model of rainfall interception by forests. Quarterly Journal of the Royal Meteorological Society 105:43-55.
Gaussen, R. and F. Hädrich (1965): Atlas zur Bodenkunde. Bibliographisches Institut, Mannheim, West Germany.
Gentry, A.H. and J. Lopez Parodi (1980): Deforestation and increased flooding of the upper Amazon. Science 210:1354-1356.
Ghuman, B.S. and R. Lal (1987): Effects of partial clearing on microclimate in a humid tropical forest. Agricultural and Forest Meteorology 40:17-29.
Gildea and B. Moore (1985): FAO Global Soils Database. 1/2x1/2 degree Lou-Lat. University of New Hampshire, Science & Engineering Building, COmplex Systems Research Center, Durham, New Hampshire, USA.
Goodroad, L.L. and D.R. Keeney (1984): Nitrous oxide emission from forest, marsh and prairie ecosystems. Journal of Environmental Quality 13:448-452.
Goodroad, L.L. and D.R. Keeney (1985): Site of nitrous oxide production in field soils. Biology & Fertility of Soils 1:3-7.
Goudriaan, J. (1987): The biosphere as a driving force in the global carbon cycle. Netherlands Journal of Agricultural Science 35:177-187.
Goudriaan, J. (1988): Modelling biospheric control of carbon fluxes between atmosphere, ocean and land in view of climatic change. Manuscript draft, J. Goudriaan, Agricultural University, Wageningen, The Netherlands.
Goudriaan, J. and D. Ketner (1984): A simulation study for the global carbon cycle, including man's impact on the biosphere. Climatic change 6:167-192.
Goward, S.N., G.D. Cruickshanks and A.S. Hope (1985): Observed relation between thermal emission and reflected spectral radiance of a complex vegetated landscape. Remote Sensing of Environment 18:137-146.
Grainger, A. (1980): The state of the world's tropical forests. Ecologist 10:6-54.
Greenland, D.J. and P.H. Nye (1959): Increase in the carbon and nitrogen contents of tropical soils under natural fallows. Journal of Soil Science 10:284-299
Gregory, G.L., R.C. Harriss, R.W. Talbot, R.A. Rasmussen, M. Garstang, M.O. Andreae, R.R. Hinton, E.V. Browell, S.M. Beck, D.I. Sebacher, M.A.K. Khalil, R.J. Ferek and S.V. Harriss (1986): Air chemistry over the tropical forest of Guyana. Journal of Geophysical Research 91:8603-8612.
Groffman, P.M. (1985): Nitrification and Denitrification in conventional and no-tillage soils. Soil Science Society of America Journal 49:329-334.
Gutman, C. (1988): A sample method for estimating monthly mean albedo of land surfaces from AVHRR data. Journal of Applied Meteorology 27:973-988.
Haider, K., A. Mosier and O. Heinemeyer (1985): Phytotron experiments to evaluate the effect of growing plants on denitrification. Soil Science Society of America Journal 49:636-641.
Hager, H. (1975): Kohlendioxyd Konzentrationen, -Flusse und -Bilanzen in einem Fichtenholzwald. Wiss. Mitt. Meteor. Inst. München, Nr. 26. Quoted in: Baumgartner and Kirchner (1980).
Hall, C.A.S., R.P. Detwiler, P. Bogdonow and S. Underhill (1985): Land use change and carbon exchange in the tropics: I. Detailed estimates for Costa Rica, Panama, Peru and Bolivia. Environmental Management 9:313-334.
Hamilton, L.S. and King, P.N. (1983): Tropical forested watersheds. Hydrologic and Soils Response to major uses or conversions. 168 pp. Westview Press, Boulder, Colorado.
Hao, W.M., D. Scharffe and P.J. Crutzen (1988): Production of N_2O, CH_4 and CO_2 from soils in the tropical savanna during the dry season. Journal of Atmospheric Chemistry 7:93-105.
Harriss, R.C., E. Gorham, D.I. Sebacher, K.B. Bartlett and P.A. Flebbe (1985): Methane flux from northern peatlands. Nature 315:652-653.
Harriss, R.C. and D.I. Sebacher (1981): Methane flux in forested freshwater swamps of the southeastern United States. Geophysical Research Letters 8:1002-1004.
Harriss, R.C., D.I. Sebacher and F.P. Day Jr. (1982): Methane flux in the great dismal swamp. Nature 297:673-674.

Harvey, H.R., R.D. Fallon and J.S. Patton (1989). Methanogenesis and microbial lipid synthesis in anoxic marsh sediments. Biogeochemistry 7:111-129.
Hatfield, J.L., G. Asrar and E.T. Kanemasu (1984): Intercepted photosynthetically active radiation estimated by spectral reflectance. Remote Sensing of Environment 14:65-75.
Hauck, R.D. (1986): Field measurement of denitrification - an overview. In: R.D. Hauck and R.V. Weaver (Eds.), Field measurement of dinitrogen fixation and denitrification, p 59-72. Soil Science Society of America Special Publ. No. 18, Madison, Wisconsin, USA.
Henderson-Sellers (1980): The effects of land clearance and agricultural practices on climate. In: V.H. Sutlive, N. Altshuler and M.D. Zamora, Blowing in the wind: deforestation and long range implications. Studies in Third World Societies. Publication No. 14. Dept. of Anthropology, College of William and Mary, Williamsburg, Virginia.
Henderson-Sellers, A. (1987): Effects of land use on climate in the humid tropics. In: R.E. Dickinson (Ed.), The Geophysiology of Amazonia, p 463-493. Wiley Interscience Publ., Wiley and Sons, New York.
Henderson-Sellers, A., R.E. Dickinson and M.F. Wilson (1988): Tropical deforestation: important processes for climate models. Climatic Change 13:43-67.
Henderson-Sellers, A. and V. Gornitz (1984): Possible climatic impacts of land cover transformations, with particular emphasis on tropical deforestation. Climatic Change 6:231-256.
Henin, S. and M. Dupuis (1945): Essai de bilan de la matière organique du sol. Ann. Agron. 15:17-29.
Hoffer, R.M. (1984): Remote sensing to measure distribution and structure of vegetation. In: G.M. Woodwell (Ed.), The role of terrestrial vegetation in the global carbon cycle. Measurement by remote sensing, p 131-159. SCOPE Vol.23. Wiley and Sons, New York.
Holben, B.N. (1986): Characteristics of minimum value composite images from temporal AVHRR data. International Journal of Remote Sensing 7:1417-1434.
Holdridge, L.R. (1967): Life Zone Ecology. Tropical Science Center, San Jose, Costa Rica, 206 pp.
Holzapfel-Pschorn, A., R. Conrad and W. Seiler (1986): Effects of vegetation on the emission of methane from submerged rice paddy soil. Plant and Soil 92:223-233.
Holzapfel-Pschorn, A. and W. Seiler (1986): Methane emission during a cultivation period from an Italian rice paddy. Journal of Geophysical Research :91:11803-11814.
Houghton, R.A., J.E. Hobbie, J.M. Melillo, B. Moore, B.J. Peterson, G.R. Shaver and G.M. Woodwell (1983): Changes in the carbon content of terrestrial biota and soils between 1860 and 1980: a net release of CO2 to the atmosphere. Ecological Monographs 53:235-262.
Houghton, R.A., R.D. Boone, J.M. Melillo, C.A. Palm, G.M. Woodwell, N. Myers, B. Moore III and D.L. Skole (1985): Net flux of carbon dioxide from tropical forests in 1980. Nature 316:617-620.
Houghton, R.A., R.D. Boone, J.R. Fruci, J.E. Hobbie, J.M. Melillo, C.A. Palm, B.J. peterson, G.R. Shaver, G.M. Woodwell, B. Moore, D.L. Skole and N. Myers (1987): The flux of carbon from terrestrial ecosystems to the atmosphere in 1980 due to changes in land use: geographical distribution of the global flux. Tellus 39B:122-139.
Huke, R.E. (1982): Rice area by type of culture: South, Southeast and East Asia. International Rice Research Intstitute, Los Baños, Laguna, Philippines, 32 pp.
Hueck, K. (1966): Die Wälder Südamerikas. Fischer Verlag, Stuttgart. 422pp.
Hueck, K. and P. Seibert (1972): Vegetationskarte von Südamerika. Gustav Fischer Verlag, Stuttgart.
Huete, A.R., R.D. Jackson and D.F. Post (1985): Spectral response of a plant canopy with different soil backgrounds. Remote Sensing of Environment 17:37-53.
Hummel, J. and R. Reck (1979): A global surface albedo model. Journal of Applied Meteorology 18:239-253.
Hutchinson, G.L. and A.R. Mosier (1979): Nitrous oxide emissions from an irrigated corn field. Science 205:1125-1127.
Hutchinson, G.L. and A.R. Mosier (1981): Improved soil cover method for field measurement of nitrous oxide flux. Soil Science Society of America Journal 45:311-316.
Hutchinson, G.L., A.R. Mosier and C.E. Andre (1982): Ammonia and Amine emissions from a large cattle feedlot. Journal of Environmental Quality 11:288-293.
Huttel, C. (1962): Estimation du bilan hydrique dans une foret sempervirente de basse Cote d'Ivoire. In: Radioisotopes in soil plant nutritiion studies, p 125-138. Vienne, AIEA, 1965. Quoted in Unesco (1978).
Idso, S.B., R.D. Jackson and R.J. Reginato (1977): Remote sensing of crop yields. Science 196:19-25.
Institut de la Carte Internationale de la Vegetation (1988): Catalogue 1988, Cartes, Livres. Université Paul Sabatier, 39 Allées Jules Guesde, 31062 Toulouse Cedex-France.
Jackson, P. (1983): The tragedy of our tropical rainforests. Ambio 12:252-254.
Jackson,R.D. and S.B. Idso (1975): Surface albedo, soils and desertification. Science 189:1012-1015.
Jackson, R.D., P.J. Pinter and R.J. Reginato (1985): Net radiation calculated from remote multispectral and ground station meteorological data. Agricultural and Forest Meteorology 35:153-164.
Jackson, R.D., M.S. Moran, L.W. Gay and L.H. Raymond (1987): Evaluating evaporation from field crops using airborne radiometry and ground-based meteorological data. Irrigation Science 8:81-90.
Jacobsen, P., W.H. Patrick Jr. and B.G. Williams (1981): Sulfide and methane formation in soils and sediments. Soil Science 132:279-287.
Janssen, B.H. (1984): a simple method for calculating decomposition and accumulation of 'young' soil organic matter. Plant and Soil 76:297-304.
Jarvis, P.G. and K.G. mcNaughton (1986): Stomatal control of transpiration, scaling up from leaf to region. In: Advances in Ecological Research 15:1-49. Academic Press London.
Jenkinson, D.S. (1977): Studies on the decomposition of plant material in soil. V. The effects of plant cover and soil type on the loss of carbon from ^{14}C labelled ryegrass decomposing under field conditions. Journal Soil Science 28:424-434.
Jenkinson, D.S. and A. Ayanaba (1977): Decomposition of carbon-14 labelled plant material under tropical conditions. Soil Science Society of America Journal 41:912-915.
Jenkinson, D.S. and J.H. Rayner (1977): The turnover of soil organic matter in some of the Rothamsted classical experiments. Soil Science 123:298-305.
Jenny, H. (1941): Factors of Soil formation. McGrawHill, New York.
Jenny, H., S.P. Gessel and F.T. Bingham (1949): Comparative study of decomposition rates of organic matter in temperate and tropical regions. Soil science 68:419-432.

Jenny, H., Salem, A.E. and J.R. Wallis (1968): Interplay of soil organic matter and soil fertility with state factors and soil properties. In: Study week on organic matter and soil fertility, p 5-36. Pontificea Academicae Scientinarum Scripta Varia 31. North Holland Publ. Co. Amsterdam, the Netherlands.

Johansson, C.L. (1984): Field measurements of emission of nitric oxide from fertilized and unfertilized forest soils in Sweden. Journal of Atmospheric Chemistry 1:429-442.

Johansson, C. and I.E. Galbally (1984): Production of nitric oxide in loam under aerobic and anaerobic conditions. Applied and Environmental Microbiology 47:1284-1289.

Johansson, C. and L. Granat (1984): Emission of NO from arable land. Tellus 36B:27-37.

Johansson, C., H. Rodhe and E. Sanhueza (1988): Emission of NO in a tropical savanna and a cloud forest during the dry season. Journal of Geophysical Research 93:7180-7192.

Johansson, C. and E. Sanhueza (1988): Emission of NO from savanna soils during rainy season. Journal of Geophysical Research 93:14193-14198.

Jones, M.J. (1972): The organic matter content of the savanna soils of West Africa. Journal of Soil Science 24:42-53.

Jong, E. de and H.J.V. Schappert (1972): Calculation of soil respiration and activity from CO_2 profiles in the soil. Soil Science 113:328-333.

Justice, C.O., B.N. Holben and M.D. Gwynne (1986): Monitoring East African vegetation using AVHRR data. International Journal of Remote Sensing. 7:1453-1475.

Justice, C.O., J.R.G. Townshend, B.N. Holben and C.J. Tucker (1985): Analysis of the phenology of global vegetation using meteorological satellite data. International Journal of Remote Sensing 6:1271-1318.

Kaplan, W. (1984): Sources and sinks of nitrous oxide. In: M.J. Klug and C.A. Reddy (Eds.), Current perspectives in microbial ecology, p 479-483. American Society of Meteorology.

Kaplan, W.A., S.C. Wofsy, M. Keller and J.M. de Costa (1988): Emission of NO and deposition of O_3 in a tropical forest system. Journal of Geophysical Research 93:1389-1395.

Keeney, D.R., I.R. Fillery, and G.P. Marx (1979): Effect of temperature on gaseous N products of denitrification in soil. Soil Science Society of America Journal 43:1124-1128.

Keepin, W., I. Mintzer and L. Kristoferson (1986): Emission of CO2 into the atmosphere. The rate of release of CO2 as a function of future energy developments. In: Bolin, B., B.R. Döös, J. Jager and R.A. Warrick (Eds.): The greenhouse effect, climatic change and ecosystems. p 35-91. SCOPE Vol.29. Wiley and Sons, New York.

Keller, M., W.A. Kaplan and S.C. Wofsy (1986): Emissions of H_2O, CH_4 and CO_2 from tropical forest soils. Journal of Geophysical Research 91:11791-11802.

Keller, M., W.A. Kaplan, S.C. Wofsy and J.M. Da Costa (1988): Emission of H_2O from tropical soils: response to fertilization with NH_4^+, NO_3^- and PO_4^{3-} Journal of Geophysical Research 93:1600-1604.

Keller, M., S.C. Wofsy, T.J. Goreau, W.A. Kaplan, M.B. McElroy (1983): Production of nitrous oxide and consumption of methane by forest soils. Geophysical Research Letters 10:1156-1159.

Khalil, M.A.K. and R.A. Rasmussen (1983a): Increase and seasonal cycles of nitrous oxide in the earth's atmosphere. Tellus 35B:161-169.

Khalil, M.A.K. and R.A. Rasmussen (1983b): Sources, sinks and seasonal cycles of atmospheric methane. Journal of Geophysical Research 88:5131-5144.

Khalil, M.A.K. and R.A. Rasmussen (1984a): The global increase of carbon monoxide. In: V.D. Aneja (Ed.), Transactions APCA Specialty Conference Environmental Impact of Natural Emmissions, March 1984. p 403-414.

Khalil, M.A.K. and R.A. Rasmussen (1984b): Carbon monoxide in the earth's atmosphere: increasing trend. Science 224:54-56.

Khalil, M.A.K. and R.A. Rasmussen (1985): Causes of increasing atmospheric methane: depletion of hydroxyl radicals and the rise of emissions. Atmospheric Environment 19:397-407.

Kimura, M., H. Wada and Y. Takai (1984): Studies on the rhizosphere of paddy rice. IX. Microbial activities in the rhizosphere of paddy rice. Japanese Journal of Soil Science and Plant Nutrition 55:338-343 (In Japanese).

King, G.M. and W.J. Wiebe (1978): Methane release from soils of a Georgia solt marsh. Geochimica et Cosmochimica Acta 42:343-348.

Klemedtsson, L., B.H. Svensson and T. Rosswall (1987): Dinitrogen and nitrous oxide produced by denitrification and nitrification in soil with and without barley plants. Plant and Soil 99:303-319.

Kononova, M.M. (1975): Humus of virgin and cultivated soils. In: Gieseling, J.E.(Ed.): Soil components. Vol.I. Organic components. p 475-526. Springer Verlag, Berlin.

Köppen, W. (1936): Das Geographische System der Klimate. In: Köppen, W. and Geiger, G. (Eds.), Handbuch der Klimatologie, Vol. 1, Part C. Gebr. Borntraeger, Berlin.

Kortleven, J. (1963): Kwantitatieve aspecten van humusopbouw en humusafbraak. Verslagen Landbouwkundige Onderzoekingen 69-1. 109 pp. PUDOC. Wageningen, the Netherlands.

Kovacs, G. (1987): Estimation of average areal evapotranspiration - proposal to modify Morton's model based on the complementary character of actual and potential evapotranspiration. Journal of Hydrology 95:227-240.

Kovda, V.A., Ye.V. Lobova, and B.G. Rozanov (1967): Classification of the world's soils. Soviet Soil Science 851-863.

Kovda, V.A., V.E. Lobova, G.V. Dobrovolsky, J.M. Ivanv, B.G. Rozanov and N.A. Solomatina (1975): Soil Map of the World. 1:10,000,000 (In Russian and English). Institute for Geodesy and Cartography, Moscow.

Koyama, T. (1963): Gaseous metabolism in lake sediments and paddy soils and the production of atmospheric methane and hydrogen. Journal of Geophysical Research 68:3971-3973.

Küchler, A.W. (1965): Potential natural vegetation map at 1:7,500,000. Revised Edition. USDA Geological Survey, Sheet No. 90.

Lal, R. (1981): Deforestation of tropical rainforest and hydrological problems. In: R. Lal and E.D. Russell (Eds.), Tropical Agricultural Hydrology. Watershed Management and land use, p 131-151. Wiley Interscience Publication. Wiley and Sons, New York.

Lal, R. (1986): Tropical ecology and physical edaphology. Wiley Interscience Publication. Wiley and Sons, New York.

Lanly, J.P. (1982): Tropical forest resources. FAO Forestry Paper No. 30. 106 pp. FAO, Rome.

Latz, K., R.A. Weismiller, G.E. Van Scoyoc and M.F. Baumgardner (1984): Characteristic variations in spectral reflectance of selected eroded alfisols. Soil Science Society of America Journal 48:1130-1134.

Leemans, R. (1989): World Map ofHoldridge Life Zones (digital data set) 0.5° resolution map, presenting the 38 Holdridge life zones. IIASA, Laxenburg, Austria.

Letey, J., M. Valoras, D.D. Focht and J.C. Ryden (1981): Nitrous oxide production and reduction during denitrification as affected by redox potenital. Soil Science Society of America Journal 45:727-730.

Levine, J.S., T.R. Augustsson, I.C. Anderson, J.M. Hoell jr. and D.A. Brewer (1984): Tropospheric sources of NO_x: lightning and biology. Atmospheric Environment 18:1997-1804.

Lipschultz, F., O.C. Zafiriou, S.C. Wofsy, M.B. McElroy, F.W. Valois and S.W. Watson (1981): Production of NO and N_2O by soil nitrifying bacteria. Nature 294:641-643.

Livingston, G.P., P.M. Vitousek and P.A. Matson (1988): Nitrous oxide flux and nitrogen transformations across a landscape gradient in Amazonia. Journal of Geophysical Research 93:1593-1599.

Lo, C.P. (1987): Applied remote sensing. Longman, Essex.

Logan, J.A. (1983): Nitrogen oxides in the troposphere. Global and regional budgets. Journal of Geophysical Research 88:10785-10807.

Logan, J.A., M.J. Prather, S.C. Wofsy and M.B. McElroy (1981): Tropospheric chemistry: a global perspective. Journal of Geophysical Research 86:7210-7254.

Lugo, A.E., M.J. Sanchez and S. Brown (1986): Land use and organic carbon content of some subtropical soils. Plant and Soil 96:185-196.

Lundegard, H. (1927): Carbon dioxide evolution of soil and crop growth. Soil Science 23:417-454. Quoted in: Buyanovsky et al (1986).

MacPherson, J.I., R.L. Desjardins and P.H. Schuepp (1987): Gaseous exchange measurements using aircraft-mounted sensors. In: Preprint of Proceedings 6th Symposium on Meteorological Observations and Instrumentation, January 12-16, 1987. American Meteorological Society, Boston, Massachussets.

Makkink, G.F. (1957): Testing the Penman formula by means of lysimeters. Journal Inst. Water Engineering London. 11:277-288.

Malaisse, F. (1973): Contribution a l'etude de l'ecosysteme foret clair (miombo). Note 8. Le Projet Miombo. Ann. University Abidjan., E., Vol.6, No.2, 1973, pp 227-250. Quoted in Unesco (1978).

Malingreau, J.P. and C.J. Tucker (1987): The contribution of AVHRR data for measuring and underestanding global processes: large scale deforestation in the Amazon Basin. Proceedings of IGARSS '87 Symposium, Ann Arbor, pp 443-448.

Malingreau, J.P. and C.J. Tucker (1988): Large scale deforestation in the Southeastern Amazon Basin of Brasil. Ambio 17:49-55.

Manabe, S. and R.T. Wetherald (1987): Large scale changes of soil wetness induced by an increase in atmospheric carbon dioxide. Journal of Atmospheric Science 44:1211-1235.

Mann, L.K. (1986): Changes in soil carbon storage after cultivation. Soil Science 142:279-288.

Manning, M.R., D.L. Lowe, W.H. Melhuish, R.J. Sparks, G. Wallace and C.A.M. Breuninkmeijer (1989). The use of radiocarbon measurements in atmoshperic studies. Radiocarbon 31 (in press).

Marks, P.L. and F.H. Bormann (1972): Revegetation and forest cutting: mechanism for return to steady state nutrient cycling. Science 176:914-915.

Martel, Y.A. and E.A. Paul (1974): The use of radiocarbon dating of organic matter in the study of soil genesis. Soil Science Society of America Journal 38:501-506.

Matthews, E. (1983): Global vegetation and land use: new high resolution data bases for climate studies. Journal of Climate and Applied Meteorology 22:474-487.

Matthews, E. (1984): Vegetation, land-use and seasonal albedo data sets. Documentation of archived data tape. NASA Technical memorandum 86107.

Matthews, E. (1985): Atlas of archived vegetation, land-use and seasonal albedo data sets. NASA Technical memorandum 86199.

Matthews, E. and I. Fung (1987): Methane emission from natural wetlands: Global distribution, area, and environmental characterisatics of sources. Global Biogeochemical Cycles 1:61-86.

Matthews, E. and W.B. Rossow (1987): Regional and seasonal variations of surface reflectance from satellite observations at 0.6 μm. Journal of Climate and Applied Meteorology 26:170-202.

McElroy, M.B. and S.C. Wofsy (1986): Tropical forests, interaction with the atmosphere. In: G.T. Prance (Ed.), Tropical forest and World Atmosphere. AAAS symp. Westview Press, Boulder, Colorado.

McKenney, D.J., K.F. Shuttleworth and W.I. Findlay (1980): Nitrous oxide evolution rates from fertilised soils: effects of applied nitrogen. Canadian Journal of Soil Science 60:429-438.

McKenney, D.J., K.F. Shuttleworth, J.R. Vriesacker and W.I. Findlay (1982): Production and loss of nitric oxide from denitrification in anaerobic Brookston clay. Applied Environmental Microbiology 43:534-541.

Mc.Naughton, U.G. and P.G. Jarvis (1983): Predicting effects of vegetation changes on transpiration and evaporation. In: T.T. Kozlowski (Ed.), Water Deficits and Plant Growth, Vol. VII, p 1-47. Academy Press, New York.

Meentemeyer, V., E.O. Box, M. Folkoff and J. Gardner (1981): Climatic estimation of soil properties; soil pH, litter accumulation and soil organic content. Ecological Society of America Bulletin 62:104.

Menenti, M. (1984): Physical aspects and determination of evaporation in deserts, applying remote sensing techniques. 202 pp. PhD thesis and report No.10 of ILRI, Wageningen, The Netherlands.

Milford, J.R. (1987): Problems of deducing the soil water balance in dryland regions from METEOSAT data. Soil Use and Management 3:51-57

Minami, K.(1987): Emission of nitrous oxide (N_2O) from Agro-ecosystem. JARQ Vol.21:22-27.

Minami, K. and S. Fukushi (1983): Effects of phosphate and calcium carbonate application on emission of N_2O from soils under aerobic conditions. Soil Science and Plant Nutrition 29:517-524.

Minami, K. and S. Fukushi (1984): Methods for measuring N_2O flux from water surface and N_2O dissolved in water from agricultural land. Soil Science and Plant Nutrition 30:495-502.

Minami, K. and S. Fukushi (1986): Emission of nitrous oxide from a well aerated andosol treated with nitrite and hydroxylamine. Soil Science and Plant Nutrition 32:233-237.

Minami, K. and K. Yagi (1988): methode for measuring methane flux from rice paddies. Japanese Journal of Soil Science and Plant Nutrition 59:458-463. (In Japanese).

Minderman, G. (1968): Addition, decomposition and accumulation of organic matter in forests. Journal of Ecology 56:355-362.

Minderman, G. and J. Vulto (1973): Carbon dioxide production by tree roots and microbes. Pedobiologia Bd.18:337-343.

Mitsch, W.J. and J.G. Gosselink (1986): Wetlands. Van Nostrand Reinhold Company, NY.

Molion, L.C.B. (1987): Micrometeorology of an Amazonian rain Forest. In: R.E. Dickinson (Ed.), The Geophysiology of Amazonia. Vegetation and Climate Interactions, p 391-407. Wiley and Sons, New York.
Monteith, J.L. (1965): Evaporation and environment. Proceedings Symposium Society of Biology 19:205-234.
Monteith, J.L. and G. Szeics (1962): Radiative temperatures in the heat balance of natural surfaces. Quarterly Journal of the Royal Meteorological Society 88:496-507.
Monteith, J.L., G. Szeicz and K. Yabuki (1964): Crop photosynthesis and the flux of carbon dioxide below the canopy. Journal of Applied Ecology 6:321-337. Quoted in Buyanovsky et al (1986).
Mooney, H.A., P.M. Vitousek and P.A. Matson (1987): Exchange of materials between terrestrial ecosystems and the atmosphere. Science 238:926-932.
Moore, T.R. and R. Knowles (1987): Methane and Carbondioxide evolution from suarctic flux. Canadian Journal of Soil Science 67:77-81.
Moormann, F.R. (1981): The classification of "paddy soils" as related to soil taxonomy. In: Institute of Soil Science, Academia Sinica (Ed.): Proceedings of symposium on paddy soil, p 139-150. Science Press, Beijing, Springer Verlag. Berlin, Heidelberg, New York.
Moormann, F.R. and N. van Breemen (1978): Rice, Soil, Water, Land. International Rice Research Institute. Los Baños, Laguna, Philipines.
Morton, F.I. (1983): Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology. Journal of Hydrology 66:1-76.
Morton, F.I. (1984): What are the limits of forest evapotranspiration. Journal of Hydrology 74:373-398.
Morton, F.I. (1985): What are the limits on forest evaporation - reply. Journal of Hydrology 82:184-192.
Mosier, A.R. and G.L. Hutchinson (1981): Nitrous oxide emissions from cropped fields. Journal of Environmental Quality 10:169-173.
Mosier, A.R., G.L. Hutchinson, B.R. Sabey and J. Baxter (1982): Nitrous oxide emissions from barley plots treated with ammonium nitrate or sewage sludge. Journal of Environmental Quality 11:78-81.
Mosier, A.R. and W.J. Parton (1985): Denitrification in a shortgrass prairie: a modelling approach. In: D.E. Caldwell, J.A. Brierley and C.L. Brierley (Eds.), Planetary Ecology, p 441-451. Van Nostrand Reinhold Co., New York.
Mosier, A.R., M. Stillwell, W.J. Parton and R.G. Woodmansee (1981): Nitrous oxide emissions from a native shortgrass prairie. Soil Science Society of America Journal 45:617-619.
Mosier, A.R., W.D. Guenzi and E.E. Schweizer (1986): Soil losses of dinitrogen and nitrous oxide from irrigated crops in northeastern Colorado. Soil Science Society of America Journal 50:344-348.
Mueller-Dombois, D. (1984): Classification and mapping of plant communities: a review with emphasis on tropical vegetation. In: G.M. Woodwell (Ed.), The role of terrestrial vegetation in the global carbon cycle. Measurement by remote sensing, p 21-90. SCOPE Vol. 23, Wiley and Sons, New York.
Mulder, J.P.M. (1985): Simulating interception loss using standard meteorological data. In: B.A. Hutchison and B.B. Hicks (Eds.), The Forest Atmosphere Interaction, pp 177-196. D. Reidel Publ. Comp., Dordrecht.
Mulders, M.A. (1987): Remote sensing in soil science. Elseviers publ. comp. Amsterdam, New York.
Mulvaney, R.L. and L.T. Kurtz (1984): Evolution of dinitrogen and nitrous oxide from nitrogen-15 fertilized soil cores subjected to wetting and drying cycles. Soil Science Society of America Journal 48:596-602.
Myers, N. (1980): Conversion of tropical moist forests. National Academy of Science, Washington, D.C., 205pp.
Myrold, D.D. and J.M. Tiedje (1985): Diffusional constraints on denitrification in soil. Soil Science Society of America Journal 49:651-657.
Nelson, R. and B. Holben (1986): Identifying deforestation in Brazil using multiresolution satellite data. International Journal of Remote Sensing 7:429-448.
Neue, H.H., and H.W. Scharpenseel (1984): Gaseous products of decomposition of organic matter in submerged soils. In: International Rice Research Institute, Organic Matter and Rice, p 311-328. IRRI, Los Baños, Laguna, Philipines.
Nieuwenhuis, G.J.A., E.H. Smidt and H.A.M. Thunnissen (1985): Estimation of regional evapotranspiration of arable crops from thermal infrared images. International Journal of Remote Sensing 6:1319-1334.
Nordin, C.F. and R.H. Meade (1982): Deforestation and increased flooding of the upper Amazon. Science 215:426-427.
Nye, P.H., and D.J. Greenland (1960): The soil under shifting cultivation. Technical Communication Nº51, 156 pp. Commonwealth Bureau of Soils, Harpenden.
Oades, J.M. (1988): The retention of organic matter in soils. Biogeochemistry 5:35-70.
O'Deen, W.A.and K.L. Porter (1986): Continuous flow system for collecting volatile ammonia and amines from senescing winter wheat. Agronomy Journal 78:746-749.
Odum, E.P. (1959): Fundamentals of Ecolgy, 2nd edition, W.B. Saunders Company, 546 pp.
Olson, J. (1963): Energy storage and the balance of producers and decomposers in ecological systems. Ecology 44:322-331
Olson, J.S., J.A. Watts and L.J. Allison (1983): Carbon in live vegetation of major world ecosystems. ORNL 5862. Environmental Sciences Division, Publ. No.1997. Oak Ridge National Laboratory, Oak Ridge, Tennessee. National Technical Information Service. U.S. Dept. Commerce.
Oremland, R.M., L.M. Marsh and S. Polein (1982): Methane production and simultaneous sulfate reduction in anoxic marsh sediments. Nature 296:143-145.
Otterman, J. (1974): Baring high albedo soils by overgrazing: a hypothesized desertification mechanism. Science 186:531-533.
Palacpac, A.C. (1982): World Rice Statistics. International Rice research Institute, Dept. Agr. Economics, 152 pp.
Parkin, T.B. (1987): Soil microsites as a source of denitrification variability. Soil Science Society of America Journal 51:1194-1199.
Paul, E.A. and J.A. Van Veen (1978): The use of tracers to determine the dynamic nature of organic matter. Paper presented at the 11th International Congress of Soil Science, June 19-27 1978, Edmonton, Canada.
Paulson, C.A. (1970): Mathematical representation of wind speed and temperature profiles in the unstable atmospheric surface layer. Journal of Applied Meteorology 9:857-861.
Palm, C.A., R.A. Houghton and J.M. Melillo (1986): Atmospheric carbon dioxide from deforestation in Southeast Asia. Biotropica 18:177-188.
Parton, W.J, A.R. Mosier and D.S. Schimel (1988): Rates and pathways of nitrous oxide production in a shortgrass steppe. Biogeochemistry 6:45-48
Parton, W.J., D.S. Schimel, C.V. Cole and D.S. Ojima (1987): Analysis of factors controlling soil organic matter levels in Great Plains grasslands. Soil Science Society of America Journal 51:1173-1179.

Patrick, W.H. (1981): The role of inorganic redox systems in controlling reduction in paddy soils. In: Proceedings of Symposium on Paddy Soil, p 107-117. Science Press, Beijing, Springer Verlag, Berlin.

Pearce, A.J., L.K. Rowe and J.B. Stewart (1980): Nighttime wet canopy evaporation rates and the water balance of an evergreen forest. Water Resources Research 16:955-959.

Peck, A.J. and D.R. Williamson (1987): Effects of forest clearing on groundwater. Journal of Hydrology 94:47-65.

Penman, H.L. (1948): Natural evaporation from open water, bare soil and grass. Proceedings Royal Society of London Ser.A.193, p 120-145.

Penman, H.L. (1970): The water cycle. Scientific American 223:99-108.

Pereira, H.C. (1973): Land use and water resources. Cambrdige University Press. 246 pp.

Persson, R. (1974): World forest resources. Royal College of Forestry, Stockholm No.17, 261pp.

Ponnamperuma, F.N. (1981): Some aspects of the physical chemistry of paddy soils. In: Proceedings of Symposium on Paddy Soil, p 59-94. Science Press, Beijing, Springer Verlag, Berlin.

Ponnamperuma, F.N. (1985): Chemical kinetics of wetland rice soils relative to soil fertility. In: Wetland Soils: Characterization, classification and utilization, p 71-89. Proceedings of Workshop IRRI 1984. IRRI, Los Baños.

Post, W.M., W.R. Emanuel, P.J. Zinke and A.G. Staugenberger (1983): Soil carbon pools and world life zones. Nature 298:156-159.

Poth, M. and D.D. Focht (1985): ^{15}N kinetic analysis of N_2O production by Nitrosomonas europaea: an examination of nitrifier denitrification. Applied Environmental Microbiology 49:1134-1141.

Potter, G.L., H.W. Ellsaessen, M.C. MacCracken and F.M. Luther (1975): Possible climatic effects of tropical deforestation. Nature 258:697-698.

Pratt, D.J. and M.D. Gwynne (1977): Rangeland Management and ecology in East Africa. Hodder and Stoughton, London.

Rasmussen, R.A. and M.A.K. Khalil (1981): Atmospheric methane (CH_4): trends and seasonal cycles. Journal of Geophysical Research 86:9826-9832.

Rasmussen, R.A. and M.A.K. Khalil (1983): Global production of methane by termites. Nature 301:700-702.

Rasmussen, R.A. and M.A.K. Khalil (1986): Atmospheric trace gases: trends and distributions over the last decade. Science 232:1623-1624.

Reginato, R.J., R.D. Jackson and P.J. Pinter (1985): Evapotranspiration calculated from remote multispectral and ground station meteorological data. Remote Sensing of Environment 18: 75-89.

Reiners, W.A. (1968): Carbon dioxide evolution from the floor of three Minnesota forests. Ecology 49:471-483.

Reinke, J.J., D.C. Adriano and K.W. Mc Leod (1981): Effects of litter alteration on carbon dioxide evolution from a South Carolina pine forest floor. Soil Science Society of America Journal 45:620-632.

Ribeiro, M.M.G., E. Salati, N.A. Villa Nova, G.G. Demetrio (1982): Radicao solar disponivel em Manaus (Am) e su relacao com a duracao do brilho solar. Acta Amazon 12:339-346. Quoted in Salati (1987).

Rjabehakov, A..M. (?): World Map on Actual Land Use at 1:15,000,000 (with 80 types). Legend translated in English at ISRIC. Faculty of Geography, Moscow State University.

Rockwood, A.A. and S.K. Cox (1978): Satellite inferred surface albedo over northwestern Africa. Journal of Atmospheric Science 35:513-522.

Rodhe, H., and R. Herrera (Eds.) (1988): Acidification in tropical countries. SCOPE Vol. 36, 405 p., Wiley and Sons, New York.

Rolston, D.E. (1981): Nitrous oxide and nitrogen gas production in fertiliser loss. In: C.C. Delwiche (Ed.), Denitrification, nitrification and atmospheric nitrous oxide. Wiley and Sons, New York.

Rolston, D.E., D.L. Hoffmann and D.A. Goldhamer (1976): Denitrification measured directly from nitrogen and nitrous gas fluxes. Soil Science Society of America Journal 40:259-266.

Rolston, D.E., D.L. Hoffman and D.W. Toy (1978): Field measurement of denitrification:I. Flux of N_2 and N_2O. Soil Science Society of America Journal 42:863-869.

Rotty, R.M. (1987): A look at 1983 CO_2 emissions from fossil fuels (with preliminary data for 1984). Tellus 39B:203-208.

Rubinoff, I. (1983): Strategy for preserving tropical rainforest. Ambio 12:255-258.

Russell, E.W. (1981): Role of watershed management for arable land use in the tropics. In: R. Lal and E.W. Russell (Eds.), Tropical Agricultural Hydrology, p 11-16. Wiley and Sons, New York.

Rutter, A.J., K.A. Kershaw, P.C. Robbins and A.J. Morton (1971): A predictive model of rainfall interception in forests. 1. Derivation of the model from observations in a plantation of Corsican Pine. Agricultural and Forest Meteorology 9:367-384.

Ryden, J.C. (1981): N_2O exchange between a grassland soil and the atmosphere. Nature 292:235-237.

Ryden, J.C. (1983): Denitrification loss from a grassland soil in the field receiving different rates of nitrogen as ammonium nitrate. Journal of Soil Science 1983:355-365.

Ryden, J.C., L.J. Lund and D.D. Focht (1978): Direct in-field measurement of nitrous oxide flux from soils. Soil Science Society of America Journal 42:731-737.

Ryden, J.C., L.J. Lund and D.D. Focht (1979): Direct measurement of denitrification loss from soils II. Developmentand application of field methods. Soil Science Society of America Journal 43:110-118.

Ryden, J.C. and L.J. Lund (1980): Nature and extent of directly measured denitrification losses from some irrigated crop production units. Soil Science Society of America Journal 44:505-511.

Sachse, G.W., R.C. Harriss, J. Fishman, G.F. Hill and D.R. Cahoon (1988): Carbon monoxide over the Amazon Basin during the 1985 dry season. Journal of Geophysical Research 93:1422-1430.

Sadovnikov, Yn.N.(1979): Changes in spectral reflectivity over the profiles of the major soil genetic types. Moscow University Soil Science Bulletin, 1979 35,1: 34-39.

Sagan, C., B.T. Owen and J.D. Pollack (1979): Anthropogenic albedo changes and the earth's climate. Science 206:1363-1368.

Sahrawat, K.L. and D.R. Keeney (1986): Nitrous oxide emission from soils. Advances in Soil science Vol.4 pp 103-148. Springer Verlag, New York.

Sahrawat, K.L., D.R. Keeney and S.S. Adams (1985): Role of aerobic transformations in six acid climax forest soils and the effect of phosphorous and $CaCO_3$. Forest Science 31:680-684.

Salati, E. (1987): The forest and the hydrological cycle. In: R.E. Dickinson (Ed.), The Geophysiology of Amazonia, Vegetation and Climate Interactions, p 273-296. Wiley and Sons, New York.

Salati, E., T.E. Lovejoy and P.B. Vose (1983): Precipitation and water recycling in tropical rainforests. Environmentalist 3:67-74.

Salati, E. and P.B. Vose (1984): Amazon Basin: a system in equilibrium. Science 225:129-138.

Sanchez, P.A. (1976): Properties and management of soils in the tropics. Wiley Interscience Publication. Wiley and Sons, New York.
Schanda, E. (1986): Fysical fundamentals of remote sensing. Springer Verlag, Berlin, Heidelberg, New York, Toronto.
Scharpenseel, H.W. and H. Schiffmann (1977): Radiocarbon dating of soils, a review. Zeitschrift für Pflanzenernährung und Bodenkunde 140: 159-174.
Scharpenseel, H.W. (in prep.): Organic matter decompostion and recycling as contribution to nutrient supply in tropical soils. Ordinariat für Bodenkunde der Universität Hamburg. Allendeplatz 2, D-2000 Hamburg 13.
Schimel, D.S., D.C. Coleman and K.A. Horton (1985a): Soil organic matter dynamics in paired rangeland and cropland toposequences in North Dakota. Geoderma 36:201-214.
Schimel, D.S., M.A. Stillwell and R.G. Woodmansee (1985b): Biogeochemistry of C, N and P in a soil catena of shortgrass steppe. Ecology 66:276-282.
Schlesinger, W.H. (1977): Carbon balance in terrestrial detritus. Annual Revue of Ecology and Systematics 8:51-81.
Schlesinger, W.H. (1982): Carbon storage in the caliche of arid soils. a case study from Arizona. Soil Science 133:247-255.
Schlesinger, W.H. (1984): Soil organic matter: a source of atmospheric CO_2. In: Woodwell, G.W. (Ed.): The role of terrestrial vegetation in the global carbon cycle, p 111-127. SCOPE Vol. 23. Wiley and Sons, New York.
Schlesinger, W.H. (1985): The formation of caliche in soils of the Mojave Desert, California. Geochimica et Cosmochimica 49:57-66.
Schlesinger, W.H. (1986): Changes in soil carbon storage and associated properties with disturbance and recovery. In: Trabalka, J.R. and D.E. Reichle (Eds.), The changing carbon cycle. A global analysis. p 194-220. Springer Verlag.
Schmidt, J., W. Seiler and R. Conrad (1988): Emission of nitrous oxide from temperate forest soils into the atmosphere. Journal of Atmospheric Chemistry 6:95-115.
Schmithüsen, J. (1968): 1:25 million vegetation maps of Europe, North Asia, South Asia, South West Asia, Australia, North Africa, Southern Africa, North America, Central America, South America (northern part), South America (southern part). In: GroBes Duden-Lexikon, Vol. 8, 321-346. Bibliographisches Institut A.G., Mannheim.
Schroeder, F.G. (1983): Die thermischen Vegetationszonen der Erde. Ein Beitrag zur Präzisierung der geobotanischen Terminologie. Mit einer Vegetationskarte. Sonderdrück aus Tuexenia, Mitteilungen der Floristisch-soziologischen Arbeitsgemeinschaft, Neue Serie, Band Nr.3, Tuexenia, Göttingen.
Schuepp, P.H., R.L. Desjardins, J.I. MacPherson, J. Boisvert and L.B. Austin (1987): Airborne determination of regional water use efficiency and evapotranspiration: present capabilities and initial field tests. Agricultural and Forest Meteorology 41:1-19.
Schütz, H., A. Holzapfel-Pschorn, R. Conrad, H. Rennenberg and W. Seiler (1989). A three years continuous record on the influence of daytime, season and ferilizer treatment on methane emission rates from an Italian rice paddy field. Submitted to Journal of Geophysical Research April 1989.
Sebacher, D.I., R.C. Harriss and K.B. Bartlett (1983): Methane flux across the air-water interface: air velocity effects. Tellus 35B:103-109.
Sebacher, D.I., R.C. Harriss and K.B. Bartlett (1985): Methane emissions to the atmosphere through aquatic plants. Journal of Environmental Quality 14:40-46.
Sebacher, D.I., R.C. Harriss, K.B. Bartlett, S.M. Sebacher, S.S. Grice (1986): Atmospheric methane sources: Alaskan tundra bogs, an alpine fen, and a subarctic boreal marsh. Tellus 38B:1-10.
Seiler, W. (1984): Contribution of biological processes to the global budget of CH_4 in the atmosphere. In: M.J. Klug and C.A. Reddy (Eds.), Current perpectives in microbial ecology, p 468-477. American Society of Meteorology.
Seiler, W. and R. Conrad (1981): Field measurements of natural and fertiliser induced N_2O release rates from soils. Journal Air Poll. Contr. Ass. 31:767-772.
Seiler, W. and R. Conrad (1985): Exchange of atmospheric trace gases with anoxic and oxic tropical ecosystems. In: R. Dickinson (Ed.), Geophysiology of Amazonia, p 133-160. Wiley and Sons, New York.
Seiler, W. and R. Conrad (1987): Contribution of tropical ecosystems to the global budgets of trace gases, especially CH_4, H_2, CO and N_2O. In: R.E. Dickinson (Ed.), Geophysiology of Amazonia. Vegetation and Climate Interactions, p 133-160. Wiley and Sons, New York.
Seiler, W. and P.J. Crutzen (1980): Estimates of gross and net fluxes of carbon between the biosphere and the atmosphere from biomass burning. Climatic Change 2:207-247.
Seiler, W., A. Holzapfel-Pschorn, R. Conrad and D. Scharffe (1984): Methane emissions from rice paddies. Journal of Atmospheric Chemistry 1:241-268.
Sellers, P.J. (1985): Canopy reflectance, photosynthesis and transpiration. International Journal of Remote Sensing 6:1335-1372.
Sellers, P.J. (1987): Modeling effects of vegetation on climate. In: R.E. Dickinson (Ed.), The Geophysiology of Amazonia. Vegetation and climate interactions. p 297-339. Wiley and Sons, New York.
Sellers, P.J. and J.L. Dorman (1987): Testing the simple biosphere model (SiB) using point micrometeorological and biophysical data. Journal of Climate and Applied Meteorology 26:622-651.
Sextone, A.J., T.B. Parkin and J.M. Tiedje (1984): Temporal response of soil denitrification rates to rainfall and irrigation. Soil Science Society of America Journal 49:99-103.
Shields, J.A. and D.R. Coote (1988): SOTER procedures manual for small scale map and database compilation (for discussion). ISRIC Working Paper and Preprint 88/2. ISRIC, Wageningen.
Shuttleworth, W.J. and I.R. Calder (1979): Has the Priestley-Taylor equation any relevance to forest evaporation ? Journal of Applied Meteorology 18:639-646.
Shuttleworth, W.J., J.H.C. Gash, J.C.R. Lloyd, C.J. Moore, J. Roberts, A.de O. Marques Filho, G. Fisch, V. de Paula Silva Filho, L.C.B. Molion, L.D. de Abreu Sa, J.C.A. Nobre, O.M.R. Cabral, S.R. Patel, J.C. de Moraes (1984a): Observation of radiation exchanges above and below Amazonian forest. Quarterly Journal of the Royal Meteorological Society 110:1163-1169.
________________ (1984b): Eddy correlation measurements of energy partition for Amazonian forest. Quarterly Journal of the Royal Meteorological Society 110:1143-1162.
________________ (1985): Daily variations of temperature and humidity within and above Amazonian forest. Weather 40:102-108.

Simpson, J.R. and J.R. Freney (1986): Interacting processes of gaseous nitrogen loss from urea applied to flooded rice fields. Transactions of the XIIIth Congress of the International Soceity of Soil Science (ISSS), Hamburg, Vol. III, pp 968-969.

Simpson, J.R. and K.W. Steele (1983): Gaseous nitrogen exchanges in grazed pastures. In: J.R. Freney and J.R. Simpson (eds.), Gaseous losses of nitrogen from plant soil systems. Developments in Soil Science Vol.9. Martinus Nijhoff/-Dr.W.Junk Publishers, The Hague.

Slemr, F., R. Conrad, and W. Seiler (1984): Nitrous oxide emissions from fertilized and unfertilized soils in a subtropical region (Andalusia, Spain). Journal of Atmospheric Chemistry 1:159-169.

Slemr, F. and W. Seiler (1984): Field measurement of NO and NO_2 emissions from fertilised and unfertilized soils. Journal of Atmospheric Chemistry 2:1-24.

Smith, C.J., M. Brandon and W.H. Patrick, Jr. (1982): Nitrous oxide emission following urea-N fertilization of wetland rice. Soil Science and Plant Nutrition 28:161-171.

Smith, C.J., R.D. Delaune and W.H. Patrick Jr. (1982): Carbon and nitrogen cycling in a spartina alterniflora salt marsh. In: J.R. Freney and I.E. Galbally (Eds.), Cycling of Carbon, nitrogen, sulphur and phosphorous in terrestrial and aquatic ecosystems. p. 97-103, Springer Verlag, New York.

Smith, C.J., R.D. Delaune and W.H. Patrick, Jr. (1983): Nitrous oxide emission from Gulf Coast Wetlands. Geochimica et Cosmochimica Acta 47:1805-1814.

Snedecor, G.W. and W.G. Cochran (1980): Statistical methods (7th edition). 507 p, the Iowa State University Press, Ames, Iowa.

Soderlund, R. and B.H. Svensson (1976): The global nitrogen cycle. In: B.H. Svensson and R. Soderlund (Eds.), Nitrogen, Phosphorous and Sulphur- Global cycles, p 23-73. SCOPE Vol.7, Ecological Bulletin, Stockholm.

Spycher, G., P. Sollins and S. Rose (1983): Carbon and nitrogen in the light fraction of a forest soil: vertical distribution and seasonal patterns. Soil Science 135:79-87.

Steele, L.P, P.J. Fraser, R.A. Rasmussen, M.A.K. Khalil, T.J. Conway, A.J. Crawford, R.U. Gamuron, K.A. Masaric and K.W. Thoning (1987): The global distribution of methane in the troposphere. Journal of Atmospheric Chemistry 5:127-171.

Stoner, E.R. and M.F. Baumgardner (1981): Characteristic variations in reflectance of surface soils. Soil Science Society of America Journal 45:1161-1165.

Stoner, E.R., M.F. Baumgardner, L.L. Biehl and B.F. Robinson (1980): Atlas of soil reflectance properties. Dept. Agron. Purdue University, West Lafayette, In. USA. Research Bulletin Agr. Exp. Station, Purdue University, 1980, No.962 75pp.

SSSA (1984): Glossary of Soil Science Terms. Publ. of the Soil Science Society of America, Madison WI 53711 USA.

Staatsbosbeheer (1986): Verslag van het vitaliteitsonderzoek 1986. Utrecht, afdeling bosontwikkeling, rapport 1986-21.

Staatsbosbeheer (1987): Effecten van luchtverontreiniging op terrestrische ecosystemen, met name bossen. Afd. Planologie en Milieu, Staatsbosbeheer.

Stanford, G. and S.J. Smith (1972): Nitrogen mineralization potential of soils. Soil Science Society of America Journal 36:465-472.

Stevenson, F.J. (1986): Cycles of soil- carbon, nitrogen, phosphorous, sulphur, micronutrients. Wiley interscience publication. Wiley and Sons, New York.

Stevenson, F.J., R.M. Harrison, R. Wetselaar and R.A. Leeper (1970): Nitrosation of soil organic matter :III. Nature of gases produced by reaction of nitrite with lignins, humic substances and phenolic constituents under neutral and slightly acidic conditions. Soil Science Society of America Journal 34:430-435.

Stewart, J.B. (1988): Modelling dependence of surface conductance of Thetford forest on environmental conditions. Accepted for publication in Agricultural and Forest Meteorology 1988.

Stewart, J.B. and H.A.R. de Bruin (1985): Preliminary study of dependence of surface conductance of Thetford forest on environmental condition. In: B.A. Hutchison and B.B. Hicks (Eds.), The Forest Climate Interaction, pp 91-104. D. Reidel publ. Company, Dordrecht.

Strahler, A.H., C.E. Woodcock and J.A. Smith (1986): On the nature of models in remote sensing. Remote Sensing of Environment 20:121-139.

Swarup, A. (1988): Influence of organic matter and flooding on the chemical and electrochemical properties of sodic soil and rice growth. Plant and Soil 106:135-141.

Szeicz, G., Endrodi, G. and S. Tajchman (1969): Aerodynamic and surface factors in evaporation. Water Resources Research 5:380-394.

Takai, Y. (1970): The mechanism of methane fermentation in flooded paddy soil. Soil Science and Plant Nutrition 10:238-244.

Ten Berge, H.F.M. (1986): Heat and water transfer at the bare soil surface. Aspects affecting thermal imagery. Proefschrift ter behaling v.d. graad v. doctor in de Landbouwwetenschappen. (PUDOC) Vakgroep Bodemkunde en Plantevoeding. Landbouwuniversiteit Wageningen.

Terry, R.E., R.L. Tate III and J.M. Duxbury (1981): The effect of flooding on nitrous oxide emissions from and organic soil. Soil Science 132:228-232.

Thompson, K. (1980): Forests and climate change in America: some early views. Climatic Change 3: 47-64.

Thorntwaite, C.W. (1948): An approach toward a rational classification of climate. Geographical Review 38:55-94.

Thunnissen, H.A.M. and H.A.C. van Poelje (1984): Remote Sensing studieproject Oost Gelderland. Deelrapport 3. Bepaling van regionale gewasverdamping met remote sensing. Projectteam Remote Sensing Studieproject. I.C.W. Wageningen.

Tiessen, H. and J.W.B. Stewart (1983): Particle size fractions and their use in studies of soil organic matter. II. Cultivation effects on organic matter composition in size fractions. Soil Science Society of America Journal 47:509-514.

Townshend, J.R.G. and C.O. Justice (1986): Analysis of the dynamics of African vegetation using the normalized difference vegetation index. International Journal of Remote Sensing 7: 1435-1445.

Tsutsuki, K. and F.N. Ponnamperuma (1987): Behaviour of anaerobic decomposition products in submerged soils. Effects of organic material amendment, soil properties and temperature. Soil Science and Plant Nutrition 33:13-33.

Tucker, C.J. (1980): A critical review of remote sensing and other methods for non-destructive estimation of standing crop biomass. Grass Forage Science 35:177-182.

Tucker, C.J., B.N. Holben, J.H. Elgin and J.E. McMurtrey (1981): Remote sensing of total dry matter accumulation in winter wheat. Remote Sensing of Environment 11:171-189.

Tucker, C.J., C.L. Vanpraet, E. Boerwinkel and A. Gaston (1983): Satellite remote sensing of total dry matter production in the Senegalese Sahel. Remote Sensing of Environment 13:461-474.
Tucker, C.J., C.L. Vanpraet, M.J. Sharman and G. van Itersum (1985): Satellite remote sensing of total herbaceous biomass production in the Senegalese Sahel: 1980-1984. Remote Sensing of Environment 17:233-249.
Tucker, C.J., I.Y. Fung, C.D. Keeling and R.H. Gammon (1986): Relationship between atmospheric CO_2 variations and a satellite-derived vegetation index. Nature 319: 195-199.
Udvardy, M.D.F. (1975): A classification of the biogeographical provinces of the world, IUCN Occasional Paper 18, 48 pp. IUCN, Gland, Switzerland.
Unesco (1973): International classification and mapping of vegetation. Unesco, Paris. 93 pp.
Unesco (1978): Tropical forest ecosystems: a state of the knowledge report. Unesco, Paris.
USDA (1971): Soils: Soil moisture regimes (map 1:50,000,000). Soil Geography Unit, USCS, Hyattsville, USA.
USDA (1972): Soils of the World. Distribution of Orders and Principal Suborders (map 1:50,000,000). Soil Geography Unit, SCS, Hyattsville, USA.
USDA (1975): Soil Taxonomy. A Basic System of Soil Classification for making and interpreting soil surveys. Agr. Handbook 436. Soil Conservation Service, U.S. Dept. of Agriculture.
Van Dam, A.J. and C.A. van Diepen (1982): The soils of the flat wetlands of the world, their distribution and their agricultural potential. 47 pp. International Soil Museum, Wageningen, The Netherlands.
Van der Linden, A.M.A., J.A. Van Veen and M.J. Frissel (1987): Modelling soil organic matter levels after long term applications of crop residues, farmyard and green manures. Plant and Soil 101:21-28.
Van Diepen, C.A. (1985): Wetland soils of the world, their characterization and distribution in the FAO-Unesco approach. In: Wetland soils: characterization, classification and utilization, p 361-374, Proceedings of Workshop IRRI, 1984. IRRI, Los Baños.
Van Veen, J.A. (1987): The use of simulation models of the turnover of soil organic matter: an intermediate report. Transactions XIII Congress ISSS Vol. VI pp 626-635.
Van Veen, J.A. and E.A. Paul (1981): Organic carbon dynamics in grassland soils. 1. Background information and computer simulation. Canadian Journal of Soil Science 61:185-201.
Van Veen, J.A., J.N. Ladd and M.J. Frissel (1984): Modelling C and N turnover through the microbial biomass in soil. Plant and Soil 76:257-274.
Van Veen, J.A., R. Merckx and S.C. van de Geijn (1987): Plant and soil related controls of the flow of carbon from roots through the soil microbial biomass. Submitted to Plant and Soil.
Vitousek, R.M. (1983): The effects of deforestation on air, soil and water. In: B. Bolin and R. Cook (Eds.), The major biogeochemical cycles and their interactions, p 223-245. SCOPE Vol.21. Wiley and Sons, New York.
Voroney, R.P., J.A. van Veen and E.A. Paul (1981): Organic C dynamics in grassland soils.2. Model validation and simulation of the long-term effects of cultivation and rainfall erosion. Canadian Journal of Soil Science 61:211-224.
Walker, J. and P.R. Rowntree (1977): The effect of soil moisture on circulation and rainfall in a tropical model. Quarterly Journal of the Royal Meteorological Society 435:29-46.
Walter, H. (1973): Vegetation of the earth. Translation by Joy Wieser, 274 pp. Springer Verlag, New York, Heidelberg, Berlin.
Wanatabe, I (1984): Anaerobic decomposition of organic matter in flooded rice soils. In: International Rice Research Institute, Organic Matter and Rice. Los Baños, Laguna, Philipines.
Webster, C.P. and R.J. Dowdell (1982): Nitrous oxide emission from permanent grass swards. Journal of the Science Food and Agriculture 33:227-230.
Welch, R., T.R. Jordan and A.W. Thomas (1984): a photogrammetric technique for measuring soil erosion. Journal of Soil and Water Conservation 194:191-194.
Wetzel, P.J. and Jy-Tai Chang (1987): Concerning the relationship between evapotranspiration and soil moisture. Journal of Climate and Applied Meteorology 26:18-37.
Wetzel, P.J. and R.H. Woodward (1987): Soil moisture estimation using GOES-VISSR infrared data: a case study with a simple statistical method. Journal of Climate and Applied Meteorology 26:107-117.
Whalen, S.C. and W.S. Reeburgh (1988): A methane flux time series for tundra environments. Global biogechemical Cycles 2,4:399-409.
Whalen, M., N. Tanaka, R. Henry, B. Deck, J. Zeglen, J.S. Vogel, J. Southon, A. Shemesh, R. Fairbanks and W. Broecker (1989): Carbon-14 in methane sources and in atmospheric methane: the contribution from fossil carbon. Submitted to Science.
Whalen, M. and T. Yoshinari (1985): Oxigen isotope ratios in N_2O from different environments. Nature 313:780-782.
Whittaker, R.H. (1970): Communities and Ecosystems. Current Concepts in Biology Series. The MacMillan Company, London, 162 pp; 2nd edition (1975), 287 pp.
Whittaker, R.H. and G.E. Likens (1975): The biosphere and man. In: Lieth, H. and R.H. Whittaker (Eds.), Primary production of the biosphere. Ecological Studies 14. Springer Verlag, Heidelberg, Berlin, New York.
Wiebe, W.J., R.R. Christian, J.A. Hansen, G. King, B. Sherr and G. Steyring (1981): Anaerobic respiration and fermentation. In: L.R. Pomeroy and R.G. Wiegert (Eds.) The ecology of a salt marsh, p.137-159, Springer Verlag, New York.
Williams, E.J., D.D. Parrish and F.C. Fehsenfeld (1987): Determination of nitrogen oxide emissions from soils: results from a grassland site in Colorado, United States. Journal of Geographical Research 92:2173-2179.
Williams, R.J. and R.L. Crawford (1984): Methane production in Minnesota peatlands. Applied and Environmental Microbiology 47:1266-1271.
Wilson, M.F., A. Henderson-Sellers, R.E. Dickinson and P.J. Kennedy (1987): Sensitivity of the biosphere-atmosphere transfer scheme (BATS) to the inclusion of variable soil characteristics. Journal of Climate and Applied Meteorology 26:341-362.
Wilson, M.F., A. Henderson-Sellers, R.E. Dickinson and P.J. Kennedy (1987): Investigation of the sensitivity of the land surface parameterization of the NCAR Community Climate Model in regions of tundra vegetation. Journal of Climatology 7:319-344.
Witkamp, M. (1966): Rates of carbon dioxide evolution from the forest floor. Ecology 47:492-494.
Woodwell, G.M. (1978): The carbon dioxide question. Scientific American 238:34-43.
Woodwell, G.M. (1984): The carbon dioxide problem. In: Woodwell, G.M. (Ed.), The role of terrestrial vegetation in the global carbon cycle. Measurement by remote sensing. SCOPE Vol. 24. Wiley and Sons, New York.

Woodwell, G.M., R.H. Whittaker, W.A. Reiners, G.E. Likens, C.C. Delwiche and D.B. Botkin (1978): The biota and the world carbon budget. Science 199:141-146.

World Resources Institute (1986-1988): World Resources. An assessment of the resource base that supports the global economy (with data tables for 146 countries), World Resources Institute/International Institute for Environment and Development, Basic Books, New York.

Yoshida, T. (1978): Microbial metabolism in rice soils. In: International Rice Research Institute, Soils and Rice. Los Baños, Laguna, Philipines.

Yoshida, T. and M. Alexander (1970): Nitrous oxide formation by Nitrosomonas europaea and heterotrophic microorganisms. Soil Science Society of America Journal 34:880-882.

Young, A. (1976): Tropical Soils and Soil Survey. Cambridge University Press. Cambridge, London, New York, Melbourne.

Young, A. and A.C.S. Wright (1979): Rest period requirements of tropical and subtropical soils under annual crops. Consultants' Working Paper N°6. FAO, Rome.

Zimmermann, P.R., J.P. Greenberg, S.O. Wandiga and P.J. Crutzen (1982): Termites: a potentially large source of atmospheric methane, carbon dioxide and molecular hydrogen. Science 218:563-565.

Zobler, L. (1986): A world soil file for global climate modeling. NASA Technical Memorandum 87802, 32 pp.

APPENDIX I

The chemical composition of the atmosphere 1968 and 1985 and some relevant data of the constituents listed.

n.a.=no data available.

component	Concentrations[1] 1968	1985	annual increase (%)	atmospheric life span (y)
Nitrogen (N_2)	780900	780900	0	n/a
Oxygen (O_2)	209400	209400	0	n/a
Argon (A)	9300	9300	0	n/a
Neon (Ne)	18	18	0	n/a
Helium (He)	5.2	5.2	0	n/a
Krypton (Kr)	1	1	0	n/a
Xenon (Xe)	0.08	0.08	0	n/a
Carbon dioxide(CO_2)	315	345	0.5	± 7 y
Methane (CH_4)	(1.0-1.2)	1.65	1.1[4]	7-8 y[3]
Nitrous oxide (N_2O)	(0.5)	304ppb	0.25	100-200y
Hydrogen gas (H_2)	0.5		n.a.	n.a.
Nitrogendioxide(NO_2)	(0.02)	75ppt	n.a.	n.a.
ozone (O_3)	0.01-0.0	440ppb	2.0	n.a.
carbon monoxide (CO)	____	110ppb	2-6[2]	2 months
nitrogen oxides (NO_x)	____	100ppb	1.5 days	
inert hydrocarbons	____	5ppb	2.0	n.a.
PAN	____	100ppt	n.a.	n.a.
$CFCl_3$ (CFC-11)	____	215ppt	5.0	80 y
CF_2Cl_2(CFC-12)	____	370ppt	5.0	170 y
CCl_4	____	125ppt	2.0	n.a.
CH_3OCl_3	____	130ppt	7.0	8 y

The data between brackets are based on unreliable or non-representative measurements.

[1] ppm unless indicated otherwise
[2] Levine et al (1985) and Khalil and Rasmussen (1983) resp.
[3] Khalil and Rasmussen (1983)
[4] Bolle et al (1986)

Where no references are indicated, the data are from Commissie Onderzoek Luchtverontreiniging (1987) and Crutzen and Graedel (1986).

APPENDIX II

Areas of irrigated rice, deepwater rice, dryland rice in 10^7 m^2. shallow refers to water depths of up to 30 cm; intermediate to water depths of 30 cm to 1 m, both in bounded fields. Double cropped areas are counted twice. Source: Huke (1982).

	IRRIGATED wet season	IRRIGATED dry season	RAINFED shallow	RAINFED intermediate	DEEP WATER	DRY-LAND	TOTAL
Southeast Asia							
Burma	780	115	2291	1165	173	793	5317
Indonesia	3274	1920	1084	534	258	1134	8204
Kampuchea	214		713	170	435	499	2031
Laos	67	9	277			342	695
W. Malaysia	252	212	92			10	566
Philippines	892	622	1207	379		415	3515
Sabah	8	4	9			21	42
Sarawak	6	4	46	11		60	127
Thailand	866	320	5128	1002	400	965	8681
Vietnam	1326	894	1549	977	420	407	5573
Total SE Asia	7685	4100	12396	4238	1686	4646	34751
Bangladesh	170	987	4293	2587	1117	858	10012
Bhutan			121	40		28	189
India	11134	2344	12677	4470	2434	5937	38996
Nepal	261		678	230	53	40	1262
Pakistan	1710						1710
Sri Lanka	294	182	210	22		52	760
Total S. Asia	13569	3513	17979	7349	3604	6915	52929
S. Korea	1120		99			12	1231
N. Korea	500					150	650
China	33676*		1880			606	36162
Total Asia	56550	7613	32354	11587	5290	12329	125723

* Huke (1982) indicated 23968 and 9690 for 1st and 2nd crop, respectively.

APPENDIX III
Irrigated area (10^7 m^2) for rice for selected countries and for the period 1950-1980. Source: Palacpac 1982)

	1950	1960	1970
Burma	490	469	736
Indonesia	4781	5975	6679
Philippines	n.a.	960	1470
Thailand	n.a.	1363	1758
India	9844	12461	14917
Pakistan	968	1181	1503
Sri Lanka	n.a.	325	465
South Korea	n.a.	867	1071
total	n.a.	23601	28599

Appendix IV.1 Matrix of ecosystems for North and Central America.
The numbers on the horizontal axis represent bioclimatic zones according to Holdridge (1967), as listed in Appendix V. The numbers on the vertical axis represent the Major Soil grouping as listed in chapter 2, section 2.3. For each soil - bioclimate combination the area is given in 1,000,000,000 m^2. These areas are underlined.
Each column has two subcolumns with headings: "LCT" and "%". LCT stands for Land Cover Type as defined by Matthews (1983) and as listed in Table 2.2a. LCT No. 33 = "Other Types".
The subcolumns with heading "%" gives the % of the area occupied by each soil-bioclimate combination.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %
1														73 11 14 32 78 33 8	20 11 22 32 75 33 3				
2								234 8 12 10 9 32 78					75 32 100	477 32 92 33 8					
3				13 22 100		15 8 50 14 50	51 8 46 14 54	637 8 55 10 26 22 12 33 7	68 8 100		4 8 100	28 8 100	38 8 76 14 19 32 5	46 8 71 14 17 32 13					
4																			
5	1516 22 77 30 13 31 8	28 8 49 22 21 29 30	478 8 23 14 20 22 49 33 5	553 22 97 33 3		12 8 89 29 11	143 8 90 33 10	842 8 60 14 13 22 13 33 14	439 8 24 22 61 29 9 33 6	14 8 70 29 30				74 8 59 32 41					
6	338 22 100			269 14 13 22 87	88 14 18 22 82	2 14 100	60 8 25 14 35 29 37 33 3	351 8 34 14 36 22 17 33 13	325 8 7 14 62 22 19 33 12	207 8 22 29 15 31 48 33 15	6 14 67 32 33	15 14 37 22 25 32 38	2 22 100	49 8 23 14 31 22 20 28 27		17 8 70 31 30			11 21 100
7							18 14 100	160 8 12 25 34 32 47	4 8 100			98 8 11 18 17 25 52 33 13	1083 8 9 25 12 32 46 33 26	140 25 12 27 13 28 13 32 46 33 14	11 8 20 14 80			44 14 12 18 17 19 22 21 19 33 30	166 14 21 21 21 25 12 28 13 32 9
8		1 22 100	72 14 68 22 29 33 3	152 14 12 22 88		52 14 85 22 15	102 8 12 14 51 22 37	503 8 16 14 62 22 22	28 8 51 10 15 22 23 33 11			7 24 100	148 8 30 14 15 25 15 28 28	913 8 30 14 15 25 25 28 28 33 2	74 8 87 32 11 33 2	6 8 100			
9																			
10				40 14 19 22 81	21 14 42 22 58		7 14 49 22 51	467 8 35 14 36 22 29	1372 8 65 14 21 10 8 33 6	31 8 100		8 8 50 14 50	29 8 67 14 7 22 6 28 20	537 8 67 28 20 33 13	226 8 68 10 18 14 7 32 7	68 8 100			
11																			
12							9 14 50 25 50	17 8 57 14 43	10 8 100		12 8 60 25 40	22 8 67 18 23 22 11	160 8 36 14 16 25 14 28 19 32 15	46 8 36 14 16 25 14 28 19 32 12			2 25 100	17 8 14 19 44 25 14 28 7 32 21	95 6 19 17 16 28 29 32 29 33 5
13																			
14								17 8 12 14 88	31 8 67			2 8 100	16 8 51 14 49	58 8 51 14 49	22 8 64 14 36				29 12 10 14 10 21 24 32 56
15													3 32 100					9 2 33 14 67	23 2 25 14 37 32 38
16														47 32 100					
17							2 21 100	9 25 100				9 25 100	333 25 14 28 45 32 24 33 17						
18							2 19 100	29 8 17 14 24 18 17 21 34 33 8	5 8 50 19 50		48 17 25 18 5 20 10 21 60	208 18 23 19 49 21 10 33 18	280	5 19 49 25 51			12 19 41 21 59	67 18 25 19 33 21 16 25 19	188 18 11 19 19 21 15 25 36 33 19
	1854	27	550	1027	107	81	394	3266	2282	252	70	393	2145	2465	353	91	14	137	512

20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	totals
LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	
10 11 56 32 37 33 7	872 10 24 11 21 32 54						21 10 100	430 10 49 32 45 33 6	18 2 100							45 2 100			1289
44 32 100	37 32 98 33 2																		867
																			898
																			0
	82 10 13 11 12 32 75						17 9 85 32 15	232 10 74 32 19 33 7							31 1 19 9 28 23 44 33 9				4439
20 15 86 21 14					17 12 17 21 83	81 12 32 21 59 33 9	48 2 23 12 30 32 31 33 16	6 2 100							3 2 100				1893
219 32 80 33 20					111 14 10 21 75 33 15	213 12 17 14 26 21 28 28 12 32 6	145 12 11 23 22 32 60 33 7												2412
77 6 15 8 12 12 26 14 19 28 12	156 8 12 11 33 12 15 32 32 33 8	3 9 50 32 50			6 14 52 28 12 32 36	118 14 67 15 19 33 14	182 9 20 12 30 14 28 32 4 33 18	297 1 11 2 16 7 13 32 19 33 41	60 2 65 12 10 32 15 33 10	6 2 50 32 50			12 12 75 14 25	26 2 22 10 33 12 22 19 22	229 2 24 19 28 32 20 33 28	183 1 25 2 47 32 18 33 10	21 2 71 12 14 32 7 33 8		3434
								59 1 95 2 5	9 2 100						3 1 100	21 1 100			92
																			2808
																			0
142 6 17 28 11 32 47 33 25	5 8 12 21 50 32 38			3 19 100	3 17 100	10 17 50 21 50	79 10 20 24 14 32 62 33 4	33 10 47 32 53											665
																			0
40 32 86 33 14	3 28 25 32 75				3 28 100	25 12 38 28 33 32 20 33 9	11 14 50 32 44 33 6	40 14 43 32 48 33 9											297
9 12 100					73 2 40 12 24 14 36	34 2 25 12 25 14 34 24 16	121 11 18 12 12 24 29 32 28 33 13	61 10 65 32 28 33 7					6 14 100	6 12 50 24 50	3 12 100				348
																			47
																			353
13 19 20 24 20 32 60				142 19 31 21 61	258 12 15 17 12 21 62 33 11	336 19 23 21 54 28 13 33 10	24 23 45 32 55				2 19 100	14 21 100							1611
574	935	3	0	145	471	797	848	1158	87	6	2	14	18	32	269	249	21	0	

Total area: 21449

Appendix IV.2 Matrix of ecosystems for South America.
The numbers on the horizontal axis represent bioclimatic zones according to Holdridge (1967), as listed in Appendix V. The numbers on the vertical axis represent the Major Soil grouping as listed in chapter 2, section 2.3. For each soil - bioclimate combination the area is given in 1,000,000,000 m^2. These areas are underlined.
Each column has two subcolumns with headings: "LCT" and "%". LCT stands for Land Cover Type as defined by Matthews (1983) and as listed in Table 2.2a. LCT No. 33 = "Other Types".
The subcolumns with heading "%" gives the % of the area occupied by each soil-bioclimate combination.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %
1															6 1 52 8 48				
2														5 26 25 32 75					
3																			
4																			
5																		10 12 30 21 70	2 21 100
6							2 28 100	50 4 50 10 29 28 21	5 4 100	17 4 100	12 4 33 10 33 21 34	8 4 25 10 51 28 24	39 4 24 21 30 28 30 30 10 33 5	64 4 32 9 10 28 24 30 13 33 21	87 1 11 4 62 30 14 33 13	94 4 91 30 9	125 13 6 21 35 28 59	43 28 100	36 1 9 12 17 21 15 28 59
7											36 21 73 28 27	27 21 43 28 57	25 21 33 28 67	15 21 68 32 24					
8											18 21 100	18 21 100	23 21 100	9 4 49 21 51	4 4 100				
9																			
10																			
11																			
12														42 4 46 11 48 32 6	7 4 83 32 17	2 4 50 32 50	5 28 100	18 12 71 28 29	59 12 82 28 18
13								3 21 100			3			2 25 25 32 75			113 21 80 28 20	6 21 100	6 21 100
14	3 28 100		6 28 100	24 28 100			6 28 100	12 28 100	18 28 100	3 28 100	12 28 100	45 3 11 10 15 21 30 28 39 33 5	51 3 14 4 33 9 22 28 32	101 4 35 10 18 28 47	71 4 92 28 8	10 4 69 28 31	24 21 63 28 37	16 3 30 21 33 28 36	38 3 6 4 7 21 26 28 61
15																			
16																			
17																			22 21 100
18											186 21 95 28 5	100 21 77 28 23	33 21 6 28 94				46 21 14 28 86	181 21 91 28 9	199 3 5 21 85 28 10
total	3		6	24			8	65	23	20	267	198	171	238	175	106	313	274	362

20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	total
LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	
9 1 67 26 33	101 2 52 9 12 25 19 26 9 33 8	12 1 7 9 74 32 20			3 12 100		100 1 21 2 3 9 48 12 24 24 3	1924 1 63 9 13 32 1 33 23	388 1 92 33 8			9 1 67 12 33	30 1 50 9 30 12 20	73 1 8 2 4 9 17 12 71	740 1 56 24 29 33 15	2240 1 87 33 13	70 1 100		5705
169 25 2 26 50 32 48	80 25 75 26 6 32 19						94 3 5 12 11 21 22 26 14 32 48	105 3 31 12 10 23 10 25 30 32 17											453
3 1 100	52 2 42 9 16 26 21 32 16 33 5						369 1 3 12 52 23 42 24 2	2204 1 23 9 28 23 39 32 1 33 9	86 1 82 24 18					12 12 100	483 1 29 12 20 23 38 24 11	851 1 93 33 7			4060
21 1 30 3 25 12 45	17 1 36 9 28 32 36	3 9 25 32 75				67 1 8 12 79 21 8	151 3 10 12 44 23 13 32 6 33 27	257 1 51 9 18 12 11 25 11 32 6	25 1 62 24 38					21 1 57 12 29 32 14	97 1 39 9 29 24 31 32 2	166 1 67 9 26 24 7			837
118 1 16 12 5 21 28 28 43 32 6	61 1 59 21 5 28 20 32 15			35 12 9 21 51 28 40	43 21 79 28 21	34 12 27 21 37 28 27	178 1 33 28 37 32 2 33 28	210 1 58 28 39 32 1 33 2	86 1 82 33 18	30 1 40 28 60	15 21 100	3 21 100		8 1 50 24 50	87 1 52 9 10 21 14 28 24	28 1 100	9 1 100		1525
70 21 54 32 37 33 9						3 12 100	291 12 83 21 9 32 8	11 12 100						40 12 100	29 12 100				547
							21 24 100	18 1 50 9 33 24 17	21 2 79 32 21						68 1 22 9 43 24 35	30 1 60 2 25 32 15			230
								12 1 100							37 1 100	62 1 100			111
							33 12 55 23 27 24 18	324 1 22 9 21 23 49 33 8						6 12 100	225 9 6 12 16 23 53 24 25	110 1 58 12 20 24 22			698
33 4 7 12 83 32 9 33 1	2 4 100						89 12 100	40 12 100						73	77 12 100				447
78 26 73 32 27	13 26 55 32 45			20 21 100	6 21 100	6 21 50 28 50	39 12 70 21 8 28 22	30 3 36 12 64							9 12 100				334
61 12 9 28 87	12 28 100	8 28 100	3 28 100	54 21 39 28 61	12 21 50 28 50	9 21 34 28 66	41 12 14 28 86	24 28 100	3 28 100	8 28 100					26 28 100				697
3 12 100	44 25 43 32 57				8 12 100		11 12 28 26 72	120 12 11 25 55 26 15 32 18							20 12 43 23 57				204
							49 9 24 12 76	329 3 17 9 12 12 28 23 37 33 6							91 9 20 12 58 23 22				469
105 21 100						81 12 65 21 33 32 3	234 1 10 12 64 21 10 32 6 33 10	12 9 76 12 24						3 12 100	82 1 38 9 24 24 38				519
86 21 57 28 43				5 21 100	29 21 81 28 19	87 21 87 28 13	85 28 78 33 22								6 28 100				983
738	382	21	3	114	99	267	1765	5620	609	38	15	12	30	234	2057	3487	79		17819

Appendix IV.3 Matrix of ecosystems for Europe.
The numbers on the horizontal axis represent bioclimatic zones according to Holdridge (1967), as listed in Appendix V. The numbers on the vertical axis represent the Major Soil grouping as listed in chapter 2, section 2.3. For each soil - bioclimate combination the area is given in 1,000,000,000 m^2. These areas are underlined.
Each column has two subcolumns with headings: "LCT" and "%". LCT stands for Land Cover Type as defined by Matthews (1983) and as listed in Table 2.2a. LCT No. 33 = "Other Types".
The subcolumns with heading "%" gives the % of the area occupied by each soil-bioclimate combination.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %
1																			
2									6 10 17 11 8 32 75				94 15 2 20 6 23 2 25 16 32 70	77 10 17 11 2 25 11 32 68 33 2	2 10 25 32 75				
3																			
4																			
5				20 22 100				62 8 84 20 16	48 8 100					65 10 17 25 11 32 68 33 4	23 10 4 11 21 32 75				
6			3 22 100	13 20 39 22 61	10 22 100		1 22 100	18 8 29 20 6 22 64	77 8 15 20 15 22 70	8 8 24 20 12 22 64			5 6 25 32 75	41 8 14 10 30 11 19 32 33 33 4	25 8 23 10 51 32 18 33 8	2 10 100			6 32 100
7																			
8								26 8 9 10 33 32 57	98 8 19 10 37 11 13 32 27	23 8 27 10 31 20 12 32 31			132 10 24 11 2 13 9 25 1 32 59	1402 8 8 10 31 11 12 32 47 33 2	143 8 12 10 25 11 20 32 43	2 10 25 32 75			38 13 42 32 58
9																			
10	1 22 100		1 22 100	30 8 45 20 46 22 10	7 8 36 22 37 28 26			277 8 86 10 5 32 3 33 6	302 8 74 10 7 22 9 32 7 33 3	40 8 39 22 47 28 5 32 8			12 8 71 10 29	268 8 32 10 23 11 7 32 38	58 8 38 11 13 22 10 28 13 32 26	2 11 25 32 75			
11																			
12	2 22 100				4 10 100				8 10 64 32 36	6 10 100			90 10 25 13 21 25 5 32 43 33 6	277 10 45 13 11 32 35 33 9	36 10 86 13 6 32 8	6 10 100			53 6 4 9 8 13 18 32 70
13																			
14								7 20 33 22 67	46 20 8 22 87 31 5	25 22 55 31 45				13 10 25 32 75	2 10 25 32 75				
15																			
16									2 10 25 32 75				8 10 25 32 75	16 10 25 32 75					
17																			
18																			
Total	3		4	64	21		1	389	587	102			339	2159	289	12			97

* The total area for Europe calculated with the Holdridge, Matthews and soil data base is 5128x1000,000,000 m2.
However, the actual area is 4727. The difference is caused by scale differences and inaccuracies due to the coarse resolution of the Matthews and soil data bases.
All areas in this table are corrected values. The % distribution of the land cover types is not corrected.

20		21		22		23		24		25		26		27		28		29		30		31		32		33		34		35		36		37		38		Total
LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	
25 32	2 25 75																																					181
		10 32	2 25 75																																			219
11 13 32	48 4 35 61	13 32	2 25 75									32	16 100	11 13 32	11 38 50 12																							284
10 32 33	159 21 59 20	10 32 33	56 45 39 16									13 32	6 68 32	6 13 32	42 38 26 36																							2127
8	4 100	8	2 100											8	2 100																							1006
6 10 13 32 33	187 10 28 16 32 14	10 13 32 33	46 27 40 31 2							32	5 100	10 13 32	25 18 18 64	10 13 32	29 8 40 52																							773
10 32	10 26 75	10 32	8 25 75													10 32	2 25 75																					113
																																						24
	407		118								5		47		84		2																					

Total area*4727

Appendix IV.4 Matrix of ecosystems for Africa.
The numbers on the horizontal axis represent bioclimatic zones according to Holdridge (1967), as listed in Appendix V. The numbers on the vertical axis represent the Major Soil grouping as listed in chapter 2, section 2.3. For each soil - bioclimate combination the area is given in 1,000,000,000 m^2. These areas are underlined.
Each column has two subcolumns with headings: "LCT" and "%". LCT stands for Land Cover Type as defined by Matthews (1983) and as listed in Table 2.2a. LCT No. 33 = "Other Types".
The subcolumns with heading "%" gives the % of the area occupied by each soil-bioclimate combination.

	LCT 1	%	LCT 2	%	LCT 3	%	LCT 4	%	LCT 5	%	LCT 6	%	LCT 7	%	LCT 8	%	LCT 9	%	LCT 10	%	LCT 11	%	LCT 12	%	LCT 13	%	LCT 14	%	LCT 15	%	LCT 16	%	LCT 17	%	LCT 18	%	LCT 19	%
1																																						
2																																						
3																																						
4																																						
5																																						
6																																	21 30	19 43 57	21 30	18 15 85	28 30	28 18 82
7																																						
8																																					24 32	13 69 31
9																																						
10																																						
11																																						
12																							21 32	3 25 75			13 32	3 25 75					17 32	3 25 75	10 21 32	26 17 38 52	13 17 32	23 17 37 46
13																																						
14																																						
15																											32	3 100										
16																											23 28 32	5 40 50 10										
17																							17 21	5 50 50			24 32	9 80 20					21	5 100	17 21 28	47 11 78 11	17 21 28 32 33	143 14 50 14 5 17
18																									28	2 100							21 30	72 8 92	25 28 30	55 19 33 48	17 28 30 32	89 32 32 23 2
Total																								8		2		20						99		146		298

20 LCT %	21 LCT %	22 LCT %	23 LCT %	24 LCT %	25 LCT %	26 LCT %	27 LCT %	28 LCT %	29 LCT %	30 LCT %	31 LCT %	32 LCT %	33 LCT %	34 LCT %	35 LCT %	36 LCT %	37 LCT %	38 LCT %	Total
							3 15 80 32 20	134 1 27 2 13 15 22 32 20 33 18							420 2 29 15 11 23 13 32 27 33 20				557
	37 15 60 23 13 28 13 32 13	9 15 33 28 53 32 13				24 15 13 21 25 23 32 24 23 32 7	980 15 40 23 14 24 16 32 18 33 12	3012 1 38 15 18 23 16 32 14 33 24	163 1 23 2 31 23 17 32 21 33 12			3 24 80 32 20	25 21 13 23 38 24 40 32 10	64 21 14 23 19 24 18 25 27 32 16	798 23 12 24 50 32 19 33 19	414 1 53 2 13 23 16 32 12 33 6	6 2 40 23 40 32 20		5535
				8 17 100	8 23 40 25 50 32 10	9 23 27 28 66 32 7	249 15 44 23 14 28 16 32 16	292 1 49 15 24 32 12 33 15				15 25 60 28 32 32 8	88 23 31 25 28 28 22 32 19	111 23 56 32 28 33 16	138 2 10 23 27 25 14 32 35 33 14	74 1 84 32 16			988
43 28 39 32 50 33 11				874 21 11 30 80 33 9	95 21 18 25 33 28 13 30 25 33 11	79 21 13 25 57 32 12 33 18	35 15 27 23 27 32 23 33 23	75 2 48 15 10 23 24 32 18			1131 30 87 33 13	69 25 50 30 40 32 6 33 4	75 25 58 30 20 32 18	124 15 10 25 64 32 13 33 13	78 15 71 32 20 33 9	33 23 89 32 11			2776
8 13 62 32 38	11 25 78 32 17 33 5			3 24 25 32 75	37 23 49 24 19 30 8 32 17 33 7	87 2 11 13 25 24 15 32 32 33 17	144 13 10 21 13 24 15 25 30 32 19	192 25 49 27 21 28 12 33 18	9 28 100		9 2 93 32 7	3 13 80 32 20	12 13 52 30 25 32 22	24 12 51 15 10 25 22 32 11 33 6	82 9 15 25 29 27 36 28 10 33 10	32 25 63 27 23 32 5 33 9			646
39 21 25 23 14 24 25 32 33 33 3	101 1 34 21 31 24 12 32 16				3 23 100	61 15 15 19 21 23 37 32 24 33 3	1132 15 42 23 11 32 22 25	845 1 26 15 17 23 17 32 14 33 26	21 1 26 17 34 21 23 32 17				112 21 13 23 34 25 25 32 17	945 15 11 23 13 24 10 25 42 32 19	1078 15 16 23 16 24 31 32 23 33 14	24 1 56 23 12 32 31			4159
52 25 16 27 33 32 51	8 25 80 32 20			17 21 50 25 17 30 17 32 8 33 8	177 21 16 25 50 30 25 33 9	807 19 10 23 11 25 61 33 18	728 15 26 23 11 25 39 32 16 33 8	644 1 23 15 37 23 23 32 17 33			131 21 53 30 42 33 5	537 25 46 30 43 32 3 33 8	815 25 54 30 17 32 13 33 16	259 23 28 25 45 32 17 33 10	76 1 24 9 19 25 31 32 12				4049
92 21 14 23 13 28 11 32 48 33 14	8 17 32 23 17 32 51			8 17 17 21 34 32 50	37 17 21 21 36 32 43	162 21 16 23 10 25 29 32 30 33 15	352 23 11 25 28 32 39 33 22	47 2 32 15 12 23 18 25 10 32 27						31 25 79 32 21					795
				27 30 100	8 17 67 30 33	21 23 14 25 80 32 6	9 23 100					12 25 80 32 20	18 15 74 23 13 32 13	86 15 54 23 26 32 15 33 5					181
							6 27 80 32 20	6 27 80 32 20											12
47 32 95 33 5					3 21 80 32 20	12 13 20 21 17 25 43 32 17 33 3	157 21 14 23 15 25 17 28 39 32 15	21 15 40 23 57 32 3	3 15 100		15 21 20 30 64 32 16	47 21 64 30 20 32 16	164 21 32 23 20 25 23 32 20 33 5	317 25 30 28 32 32 24 33 14	122 25 10 28 57 32 20 33 13				911
70 21 15 23 13 27 15 28 24 32 33																			75
94 13 19 21 28 24 15 28 11 32 12	2 28 80 32 20			200 21 43 30 54 33 3	195 21 62 25 13 30 18 33 7	314 21 17 25 43 32 17 33 23	232 15 19 21 15 25 24 32 16 33 26	38 15 20 23 37 25 19 32 24			154 15 12 21 31 25 13 30 40 33 4	104 21 21 25 22 30 43 32 9 33 5	24 23 12 21 20 24 12 25 40 32 15	27 15 11 21 36 24 27 32 18	6 15 40 23 50 32 10				1597
20 17 13 28 82 32 5				3008 17 6 21 4 25 4 30 85 33 1	399 15 12 17 10 21 10 30 57 33 11	191 15 30 21 18 25 16 30 17 33 19	117 15 23 25 38 27 27 32 9 33 3				2751 30 88 33 12	308 12 11 21 12 25 17 30 49 33 11	159 12 14 15 52 21 15 24 10 33 9	108 15 62 21 11 25 11 28 11 32 4					7277
465	185	9		4143	960	1767	4144	5104	196		4191	1098	1290	2098	2776	577	6		29558

Appendix IV.5 Matrix of ecosystems for Asia and USSR.
The numbers on the horizontal axis represent bioclimatic zones according to Holdridge (1967), as listed in Appendix V. The numbers on the vertical axis represent the Major Soil grouping as listed in chapter 2, section 2.3. For each soil - bioclimate combination the area is given in 1,000,000,000 m^2. These areas are underlined.
Each column has two subcolumns with headings: "LCT" and "%". LCT stands for Land Cover Type as defined by Matthews (1983) and as listed in Table 2.2a. LCT No. 33 = "Other Types".
The subcolumns with heading "%" gives the % of the area occupied by each soil-bioclimate combination.

	1 LCT %	2 LCT %	3 LCT %	4 LCT %	5 LCT %	6 LCT %	7 LCT %	8 LCT %	9 LCT %	10 LCT %
1	0	0	0	0	0	0	0	0	0	0
2	0	0	0	0	0	0	26 28 23 32 70 33 7	748 8 15 11 14 23 29 32 27 33 15	28 8 94 33 6	0
3	0	0	4 10 66 22 34	17 10 62 22 38	0	0	7 8 100	1872 8 75 22 12 33 13	208 8 81 10 12 33 7	0
4	0	0	0	0	0	0	0	0	0	0
5	639 22 61 30 25 33 14	143 16 78 22 10 33 12	1081 16 50 20 10 22 34 33 14	384 16 27 20 14 22 33 33 26	0	20 11 96 33 4	508 11 72 16 23	1419 8 16 10 13 11 45 16 21 33 5	87 8 13 11 70 18 11 33 6	0
6	132 16 13 22 75 33 12	190 16 40 22 53 33 7	582 16 32 22 58 33 10	391 16 32 18 18 22 40 33 10	0	179 18 33 22 32 18 18 22 40	930 16 33 22 32 28 10 30 25	3000 8 20 10 11 11 32 16 18 33 19	504 8 26 10 12 11 47 33 15	0
7	0	0	0	0	0	0	290 25 58 28 29 33 13	132 23 48 25 14 28 25 32 13	0	0
8	0	0	0	0	0	0	0	88 8 11 10 33 11 31 23 10 32 15	22 8 75 23 15 33 10	0
9	0	0	0	0	0	0	0	0	0	0
10	0	0	14 8 17 20 49 22 34	211 8 42 16 21 22 14 33 23	0	0	0	378 8 89 33 11	285 8 97 33 3	0
11	0	0	0	0	0	0	0	0	0	0
12	0	0	0	0	0	0	0	0	0	0
13	0	0	0	0	0	0	0	87 8 30 10 15 11 39 22 15 33 1	0	0
14	0	0	0	6 18 75 20 25	22 18 50 20 50	0	0	43 11 12 18 42 20 27 22 20	151 8 24 18 24 20 36 22 15	17 8 11 18 79 20 11
15	0	0	0	0	0	0	0	0	0	0
16	0	0	0	0	0	0	0	0	0	0
17	0	0	0	0	0	2 25 100	132 21 19 25 77 33 4	0	0	0
18	16 23 14 30 86	0	7 30 100	9 21 73 23 27	0	17 25 100	74 21 15 23 13 25 15 30 58	18 21 48 24 26 30 26	0	0
Total	787	333	1648	1018	22	218	1967	7785	1285	17

	11 LCT %	12 LCT %	13 LCT %	14 LCT %	15 LCT %	16 LCT %	17 LCT %	18 LCT %	19 LCT %
1	0	0	0	0	11 2 45 5 25 32 30	0	0	0	0
2	0	2 28 25 32 75	760 23 10 28 26 32 52 33 12	763 11 22 28 19 32 41 33 18	0	0	0	0	0
3	0	0	0	726 8 31 10 45 32 21 33 3	0	0	0	0	0
4	0	0	0	0	0	0	0	0	0
5	0	8 23 100	52 25 71 32 17 33 11	325 10 35 11 26 32 33 33 6	3 2 50 32 50	3 2 50 32 50	0	0	31 10 23 32 70 33 7
6	365 28 71 30 26 33 3	1065 28 26 30 55 33 19	1478 25 11 28 33 32 17 33 39	1206 8 24 10 10 28 23 32 14 33 29	205 8 29 10 22 11 18 33 31	8 28 35 32 49 33 16	180 28 35 30 41 33 24	788 23 13 25 11 28 12 30 39 33 25	303 17 12 23 17 25 23 28 17 32 19
7	8 25 12 28 50 32 37	235 21 10 25 55 28 13 32 19	1585 25 42 28 26 32 27 33 5	103 23 59 32 28 33 13	0	0	0	0	4 10 25 32 75
8	0	25 21 42 25 49 33 9	235 25 31 32 41 33 28	714 8 13 10 32 11 13 32 28 33 14	107 7 13 9 18 10 24 32 27 33 18	24 2 23 5 22 7 12 10 13 32 30	19 13 13 28 13 30 74	21 13 12 21 43 25 12 28 34	35 10 14 19 36 32 43 33 7
9	0	0	0	0	3 32 100	0	0	0	0
10	0	0	0	22 8 66 11 34	0	0	0	0	0
11	0	0	0	0	0	0	10 21 100	64 21 57 25 39 33 4	2 21 100
12	0	2 17 100	85 6 10 18 14 23 14 32 46 33 16	16 11 28 14 14 18 28 23 29	0	0	0	8 13 33 32 67	20 8 24 17 22 32 36 33 18
13	157 21 22 28 20 30 49 33 9	180 21 42 25 16 28 16 30 17 33 9	87 21 15 25 78 33 7	0	0	0	29 21 76 30 24	91 21 31 25 34 30 18 33 17	84 17 12 23 18 25 40 33 30
14	0	0	0	7 8 65 32 35	31 8 15 10 29 32 56	0	0	0	0
15	0	0	2 17 25 32 75	0	0	0	0	0	5 17 25 32 75
16	0	0	7 8 17 13 8 32 75	2 8 25 32 75	0	0	0	0	0
17	76 21 33 24 28 28 29 33 10	452 21 32 25 32 28 11 30 14 33 11	493 21 21 25 48 32 10 33 21	25 17 38 21 17 30 26	0	0	0	57 30 43 32 56 33 1	74 8 10 21 22 30 31 32 29 33 8
18	694 21 64 30 23 33 13	881 21 43 25 13 30 38 33 8	447 21 53 30 26 33 21	7 8 71 21 29	18 2 57 8 29 32 11	0	273 21 74 30 18	390 21 48 30 28 33 24	71 17 18 23 19 28 14 30 27 32 13
Total	1300	2850	5231	3916	378	35	511	1419	629

20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	Total
LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	
55 2 10 10 11 14 47 32 32	557 2 22 5 25 7 23 32 21 33 9	8 1 29 7 64 32 7	0	0	0	0	70 2 25 5 44 32 27 33 4	1598 2 46 32 23 33 31	587 1 76 2 12 33 12	9 1 36 2 64	0	0	0	3 1 100	319 1 18 2 43 15 10 32 22 33 7	823 1 80 33 20	149 1 86 2 12 33 2	0	4167
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	2327
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	2834
0	0	0	0	0	0	0	0	85 9 45 15 11 32 39 33 5	15 15 56 32 44	0	0	0	0	0	18 9 33 15 13 32 53	30 1 20 9 10 15 37 32 28 33 5	3 1 100	0	151
339 10 33 32 67	31 10 43 32 50 33 7	3 2 50 32 50	0	0	0	0	28 2 11 5 39 9 25 32 25	431 1 32 2 11 32 38 33 19	87 1 64 15 14 32 23	17 9 33 32 67	0	0	0	0	158 15 24 32 54 33 22	508 1 72 32 24 33 4	0	0	6283
375 10 18 32 49 33 33	430 2 11 7 30 32 29 33 30	20 5 25 10 28 11 12 28 27 32 9	0	657 25 33 30 46 33 21	657 21 10 23 19 25 13 31 41 33 17	83 23 28 28 21 32 23 33 28	80 6 12 13 11 19 17 32 33 33 27	179 2 28 5 22 32 42 33 8	3 2 50 32 50	0	311 2 11 21 56 30 21 33 12	163 21 25 25 12 30 52 32 7 33 4	13 21 45 30 20 32 30 33 5	0	3 19 100	0	0	0	14420
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	2357
109 2 17 5 22 10 13 19 14 32 19	240 6 23 9 11 10 13 32 35 33 18	80 2 29 5 16 10 13 32 25 33 17	0	0	0	11 2 20 21 19 32 61	166 6 12 13 11 19 17 32 33 33 27	266 1 22 9 43 32 27 33 8	285 1 77 32 12 33 11	0	12 21 100	0	0	83 2 11 15 14 32 65 33 10	163 1 11 9 43 32 41 33 5	141 1 78 32 17 33 5	57 1 100	0	2903
0	9 2 42 32 58	3 32 100	0	0	0	18 9 16 12 13 32 63	88 2 26 9 24 32 44 33 6	174 2 29 9 33 32 30 33 8	60 1 40 2 40 32 20	12 2 25 9 50 32 25	0	0	3 2 100	112 2 24 9 25 32 44 33 7	335 2 10 9 49 32 34 33 7	83 1 57 2 21 32 14 33 8	0	0	900
0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	910
0	0	0	0	53 25 20 30 80	77 30 79 32 13 33 8	62 30 23 32 65	8 19 67 32 25 33 8	0	0	0	0	49 30 75 32 21 33 4	76 30 54 32 40 33 6	22 30 28 32 57 33 15	0	0	0	0	423
99 8 13 23 12 32 41 33 34	7 12 67 14 33	0	0	3 32 100	15 25 50 32 50	20 13 30 17 40 25 13 32 17	80 9 32 32 53 33 15	63 9 59 32 37 33 4	0	0	0	0	0	137 12 17 32 77 33 6	180 9 39 32 52 33 9	6 2 41 32 59	0	0	741
103 17 12 23 11 25 47 30 24 33 6	2 17 100	0	0	204 23 12 25 33 31 46 33 9	124 13 17 23 13 25 36 32 14 33 20	108 13 10 23 33 25 42 33 15	0	0	0	0	8 21 50 30 50	47 13 18 23 29 25 13 30 30 33 10	11 23 25 25 51 27 25	0	0	0	0	0	1320
0	0	0	0	0	0	0	3 32 100	27 1 34 32 64 33 2	21 1 14 32 86	0	0	0	11 21 25 32 75	8 9 25 32 75	15 32 95 33 5	34 32 100	0	0	394
0	86 5 24 10 48 32 28	0	0	0	0	31 32 68 33 32	197 9 15 15 12 32 65 33 8	98 9 75 32 25	0	0	0	0	14 9 10 27 59 32 30	202 9 10 32 72 33 18	281 9 64 32 34 33 2	0	0	0	878
10 13 19 32 75 33 6	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	19
63 13 22 32 48 33 30	2 17 100	0	0	0	22 25 45 30 10 32 45	34 21 14 25 14 32 68 33 4	10 21 12 32 88	0	0	0	0	0	0	0	0	0	0	0	1442
5 17 63 32 37	11 9 13 19 12 32 75	0	0	1220 21 12 31 77 33 11	437 17 12 25 11 30 52 32 9 33 16	118 17 12 25 11 30 52 32 10 33 15	60 19 16 32 69 33 15	18 12 13 32 75 33 12	0	0	581 30 83 32 11	417 30 83 32 11 33 6	44 19 41 27 13 32 28	29 19 25 32 75	3 19 25 32 75	0	0	0	5843
1158	1355	114	0	2137	1332	485	770	2935	1038	38	890	878	172	594	1455	1623	209	0	48310

Appendix IV.6 Matrix of ecosystems for Australia, New Zealand and Papua New Guinea.
The numbers on the horizontal axis represent bioclimatic zones according to Holdridge (1967), as listed in Appendix V. The numbers on the vertical axis represent the Major Soil grouping as listed in chapter 2, section 2.3. For each soil - bioclimate combination the area is given in 1,000,000,000 m^2. These areas are underlined.
Each column has two subcolumns with headings: "LCT" and "%". LCT stands for Land Cover Type as defined by Matthews (1983) and as listed in Table 2.2a. LCT No. 33 = "Other Types".
The subcolumns with heading "%" gives the % of the area occupied by each soil-bioclimate combination.

	1		2		3		4		5		6		7		8		9		10		11		12		13		14		15		16		17		18		19	
	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%	LCT	%
1																																						
2																																						
3																																						
4																																						
5																											 23	5 100										
6																																						
7																																						
8																											 4 28 32	143 26 52 17										
9																																						
10																																						
11																																			21 25 30	32 24 44	30 21 23 32	100 20 30 50
12																											 5 6 32	32 17 58 17									 21 23 32	55 20 30 50
13																									 13	2 100									 21 25 28	20 14 57 29	 21 32 33	75 32 57 11
14																																						
15																																					 21 32	41 20 80
16																																					 6 13 17 32	53 12 26 10 52
17																																			 21 30	52 78 22	 21 32	157 72 18
18																																					 21 32	11 50 50
Total																										2		181								145		409

The total area for Australia, New Zealand and Papua New Guinea calculated with the Holdridge, Matthews and soil data base is 7756 x1000,000,000 m2.
However, the actual area is 8338. The difference is caused by scale differences and inaccuracies due to the coarse resolution of the Matthews and soil data bases.
All areas in this table are corrected values. The % distribution of the land cover types is not corrected.

20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	Total
LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	LCT %	
	14 28 100						67 13 100	29 1 11 2 34 13 44 23 11						16 13 100	115 1 14 13 30 23 50 33 6				241
34 23 41 24 50 33 9	12 8 75 23 25						35 5 33 13 36 24 17 32 6	38 2 16 5 61 13 12 32 12											125
						10 24 100	81 13 69 24 31							148 13 48 24 52	135 2 10 13 64 15 17 33 9				353
23 3 11 28 89	62 4 24 28 73 32 3						44 1 21 13 79	31 1 40 13 60							3 13 100				306
					25 14 28 13 30 62 33 11	25 22 27 16 30 50 33 12	13 38 24 33 25 29	30 100				30 100	30 85 33 15	24 28 25 38 30 18 33 16	13 78 23 17 33 5				
225 5 14 6 31 13 10 32 27	19 6 85 23 15			3 21 100	148 21 65 28 35		42	9 5 33 6 33 23 34				3 21 100	46 21 20 28 80						583
105 6 12 13 23 21 13 23 25 28 27					347 25 44 30 36 33 20	9 21 31 28 35 30 34							8 28 100		16 28 100				582
5 28 100	22 4 25 28 75																		27
9 32 100					77 21 11 24 15 25 58 28 15	513 13 11 21 17 24 62 33 10	287 13 37 21 24 32 21 33 18	15 1 60 13 40					33 24 91 25 9	157 24 86 33 14	16 13 100				1128
116 6 55 17 34 32 11					9 17 100	68 13 13 17 30 32 54	31 6 27 17 64 32 9	3 6 100											277
16 18 33 21 50 32 13				40 21 93 25 7	572 21 75 25 11 33 14	218 13 22 21 57 32 13 33 8							22 25 43 28 57						1076
					922 21 57 28 18 30 13	375 13 16 21 41 32 16 33 27	153 13 67 21 17 24 10 32 6						77 28 88 33 12	3 13 100					1542
533	129			43	2770	1918	776	128				6	289	650	360				8338

APPENDIX V. Holdridge life zone classification; table presenting the bioclimatic zones indicated with Latitudinal and altitudinal belt, annual precipitation, the Potential evapotranspiration/precipitation ratio, mean annual temperature and the name of the bioclimatic zone according to Holdridge (1967).

	Latitudinal belt	Altitudinal belt	PET/P	P (mm)	T* (oC)	Life Zone (Holdridge (1967)
1	polar	nival	<1.5	0-750	<1.5	Polar desert
2	subpolar	alpine	1-2	<125	1.5-3	subpolar dry tundra
3	subpolar	alpine	0.5-1	125-250	1.5-3	subpolar moist tundra
4	subpolar	alpine	0.25-0.5	250-500	1.5-3	subpolar wet tundra
5	subpolar	alpine	0.125-0.25	500-1000	1.5-3	subpolar rain tundra
6	boreal	subalpine	2-4	<125	3-6	boreal desert
7	boreal	subalpine	1-2	125-250	3-6	boreal dry scrub
8	boreal	subalpine	0.5-1	250-500	3-6	boreal moist forest
9	boreal	subalpine	0.25-0.5	500-1000	3-6	boreal wet forest
10	boreal	subalpine	0.125-0.25	1000-2000	3-6	boreal rain forest
11	cool temperate	montane	4-8	<125	6-12	cool temperate desert
12	cool temperate	montane	2-4	125-250	6-12	cool temperate desert scrub
13	cool temperate	montane	1-2	250-500	6-12	cool temperate steppe
14	cool temperate	montane	0.5-1	500-1000	6-12	cool temperate moist forest
15	cool temperate	montane	0.25-0.5	1000-2000	6-12	cool temperate wet forest
16	cool temperate	montane	0.125-0.25	2000-4000	6-12	cool temperate rain forest
17	warm temperate	lower montane	8-16	<125	12-18	warm temperate forest
18	warm temperate	lower montane	4-8	125-250	12-18	warm temeprate desert scrub
19	warm temperate	lower montane	2-4	250-500	12-18	warm temperate thorn steppe
20	warm temperate	lower montane	1-2	500-1000	12-18	warm temperate dry forest
21	warm temperate	lower montane	0.5-1	1000-2000	12-18	warm temperate moist forest
22	warm temperate	lower montane	0.25-0.5	2000-4000	12-18	warm temperate wet forest
23	warm temperate	lower montane	0.125-0.25	4000-8000	12-18	warm temperate rain forest
24	subtropical	premontane	8-16	<125	18-24	subtropical desert
25	subtropical	premontane	4-8	125-250	18-24	subtropical desert scrub
26	subtropical	premontane	2-4	250-500	18-24	subtropical thorn woodland
27	subtropical	premontane	1-2	500-1000	18-24	subtropical dry forest
28	subtropical	premontane	0.5-1	1000-2000	18-24	subtropical moist forest
29	subtropical	premontane	0.25-0.5	2000-4000	18-24	subtropical wet forest
30	subtropical	premontane	0.125-0.25	4000-8000	18-24	subtropical rain forest
31	tropical		16-32	<125	>24	tropical desert
32	tropical		8-16	125-250	>24	tropical desert scrub
33	tropical		4-8	250-500	>24	tropical thorn woodland
34	tropical		2-4	500-1000	>24	tropical very dry forest
35	tropical		1-2	1000-2000	>24	tropical dry forest
36	tropical		0.5-1	2000-4000	>24	tropical moist forest
37	tropical		0.25-0.5	4000-8000	>24	tropical wet forest
38	tropical		0.125-0.25	>8000	>24	tropical rain forest

* mean annual temperature