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SUMMARY

1. **Land cover changes**

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 $20x10^{10}$ 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.

2. **global biogenic 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 1980 is $5.2x10^{15}$ 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}. 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 $3.3x10^{15}$ 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 (1.43%) and the increase of methane emission by termites due to shifts in land use (1.5 to 4%) correlate well to the atmospheric increase. This indicates, that the increase in atmospheric CH_4 is most likely related to anthropogenic activities.
The present annual release rate is $329x10^{12}$ to $654x10^{12}$ g y^{-1} of methane according to various authors quoted in table 4.6. Individual sources are rice paddies (70-170), wetlands (25-70), 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.

<u>Carbon monoxide (CO)</u>
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 5700×10^{12} g CO yr^{-1} with an average of 2920×10^{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 1960×10^{12} and 4750×10^{12} g CO yr^{-1} averaging 3600×10^{12}. The model is not completely balanced, indicating the uncertainty in the estimates.

<u>Nitrogenous greenhouse gases (N_2O and NO_x)</u>
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 soils (11.1×10^{12} to 54.7×10^{12} g N y^{-1}) and emissions from other soil and land use and related sources (1×10^{12} to 2×10^{12} g N y^{-1}) yield a likely total global biogenic emission of 12.6 $\times 10^{12}$ to 57.2×10^{12} g N y^{-1}. In this estimate it is assumed that the emission induced by N-fertilization (0.7×10^{12} to 3.0×10^{12} g N y^{-1}) is included in the flux rate given for cultivated fields. All N_2O is eventually tranferred to the stratosphere where it reacts with ozone.
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.

3. <u>The influence of land cover changes on the energy balance</u>.
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.

4. <u>The influence of land cover changes on the hydrologic cycle</u>.
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, of imbalance to 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 to study 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

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 vegetation classification with time series analysis of data from the growing satellite data set (e.g. LANDSAT, NOAA-AVHRR) should be given more attention.

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 life zone. 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 net 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 understanding 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 this 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 done complementary to point measurements. The continued basic study of cycles of C and N is a prerequisite for the understanding of these gas fluxes.

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.

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.

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 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 of eminent importance, not only due to its role in cloud formation and rainfall, but also due to the effect of clouds on the incoming radiation. Vapour may also play a role in the energy transport between equatorial and temperate regions. Human interference in the land cover also bring about changes in the earth's albedo. The albedo plays a role in the surface energy balance and may therefore have large (mostly 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 concentrations of photochemical oxidants are. These gases have asterixes in the source columns 3 or 5 and in effect column B or C. The resulting gases are: CO_2, CH_4, NO_x and 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 and 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:

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

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

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.

	Sources of disturbances											Effect					
Chemical constituent	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 (NMHC[1])	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													
Organic S	o		*		*												o
Halocarbons											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).

In appendix II 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.

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

-quantification of the global spatial distribution of soils and their land cover;
-quantification of the emission and immision rates of greenhouse gases, evapotranspiration and albedo for the world soils and their land cover types;
-mechanisms and modelling of gas emissions and immisions, 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 and immission 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 the Soil Map of the World at scale 1:5,000,000 (FAO/UNESCO, 1971-1978). 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 Data on the major greenhouse gases concerning their rise and contribution to the greenhouse effect.

	annual rise (%)	heat absorbing capacity (CO_2 = 1)	contribution to global warming (%)
CO_2	0.5-1.0	1	55
CH_4	1.1	20	20
N_2O	0.25	200	5
O_3	1.5-3.0	2000	5
CFC's	5	up to 1000	15

Source: Tweede Kamer der Staten Generaal, vergaderjaar 1986-1987, 20 047, nrs 1-2.

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:

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

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

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

A second source of OH. are oxygenated organic compounds,

e.g. :

$$HCHO \xrightarrow{hv} H. + CHO.$$
$$H. + O_2 \xrightarrow{m} HO_2.$$
$$CHO. + O_2 \xrightarrow{} HO_2. + CO$$

followed by either:

$$HO_2.+HO_2. \xrightarrow{} H_2O_2 + O_2$$
$$H_2O_2 \xrightarrow{hv} OH. + OH.$$

or:

$$HO_2. + NO \xrightarrow{} NO_2 + OH.$$

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 hydro- carbons). 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

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

$$CH_4 + OH. \rightarrow CH_3. + H_2O$$
$$CH_3. + O_2 \rightarrow CH_3O_2.$$

Possible subsequent reaction paths are:

$$CH_3O_2. + HO_2. \rightarrow CH_3O_2H + O_2$$
$$CH_3O_2H + OH. \rightarrow CH_3O_2. + H_2O$$

Reaction path with NO>10pptv:

$$CH_3O_2. + NO \rightarrow CH_3O. + NO_2$$
$$CH_3O. + O_2 \rightarrow HCHO. + HO_2.$$
$$HO_2. + NO \rightarrow OH. + NO_2$$
$$2[NO_2 \xrightarrow{hv} NO + O.]$$
$$2[O. + O_2 \xrightarrow{m} O_3]$$

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net:

$$CH_4 + HO_2. \rightarrow HCHO + 2\ O_3 + H_2O$$

Reaction path with NO<10pptv:

$$CH_3O_2. + HO_2. \rightarrow CH_3O_2H + O_2$$
$$CH_3O_2H \xrightarrow{hv} CH_3O. + OH.$$
$$CH_3O. + O \rightarrow HCHO + HO_2.$$

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net:

$$CH_4 + O_2 \rightarrow HCHO + H_2O$$

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

$$HCHO \xrightarrow{hv} H. + HCO.$$
$$H. + O_2 \xrightarrow{m} HO_2$$
$$HCO. + O_2 \rightarrow CO + HO_2.$$
$$CO + OH. \rightarrow H. + CO_2$$

$$H. + O_2 \xrightarrow{m} HO_2.$$
$$3[HO_2. + NO \rightarrow NO_2 + OH.]$$
$$3[NO_2 \xrightarrow{hv} NO + O.]$$
$$3[O. + O_2 \xrightarrow{m} O_3]$$

$$HCHO + 6O_2 \rightarrow CO_2 + 3O_3 + 2OH.$$

$$2[H. + O_2 \rightarrow HO_2.]$$
$$3[HO_2. + O_3 \rightarrow OH. + 2O_2]$$

$$HCHO + 3O_3 \rightarrow CO_2 + 3O_2 + 2OH.$$

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:

$$RH + OH. \longrightarrow R. + H_2O$$
$$R. + O_2 \xrightarrow{m} ROO.$$

Two possible consecutive paths exist:

$$ROO. + R'OO. \longrightarrow ROOR' + O_2$$
$$ROOR' \xrightarrow{hv} RO. + R'O.$$

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

Carbon monoxide

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

with> 4-20 pptv NO:

$$CO + OH. \longrightarrow H. + CO_2$$
$$H. + O_2 \longrightarrow HO_2.$$
$$HO_2. + NO \longrightarrow OH. + NO_2$$
$$NO_2 \xrightarrow{hv} NO + O.$$
$$O. + O_2 \xrightarrow{m} O_3$$

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net:

$$CO + 2O_2 \longrightarrow CO_2 + O_3$$

with < 4-20 pptv NO

$$CO + OH. \longrightarrow H. + CO_2$$
$$H. + O_2 \longrightarrow HO_2.$$
$$HO_2. + O_3 \longrightarrow OH. + 2O_2$$

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net:

$$CO + O_3 \longrightarrow CO_2 + O_2$$

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

NO_x

As can be deducted from the above reactions in which methane and carbon monoxide are involved, nitrogen oxide 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:

$$NO_2 + OH. \xrightarrow{m} HNO_3$$

during nighttime a different path is followed:

$$NO_2 + O_3 \longrightarrow NO_3 + O_2$$
$$NO_3 + NO_2 \longrightarrow N_2O_5$$
$$N_2O_5 + H_2O_{aq} \longrightarrow 2\ HNO_3$$

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:

$$CH_3C[O]O_2NO_2 \longrightarrow CH_3C[O]O_2. + NO_2$$

thereby liberating NO_x radicals.

NO_x catalyses the destruction of ozone:

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

Below 40 km ozone is formed via the reaction path:

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

Below 25 km ozone is formed as follows:

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

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.

<u>N_2O</u>

In the statosphere <u>N_2O</u> is oxidized to NO via:

$$\begin{array}{lll} O_3 & \xrightarrow{hv} & O. + O_2 \\ O. + N_2O & \longrightarrow & 2\ NO \end{array}$$

<u>NH_3</u>

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.

$$OH. + NH_3 \longrightarrow NH_2 + H_2O$$

The subsequent chemistry of NH_2 is uncertain:

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

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:

$$NH_{3(g)} + HNO_{3(g)} \longleftrightarrow NH_4NO_{3(s\ or\ aq)}$$

2. Distribution of the soils of the world and the major land cover types

As a basis for this study the extents of the major soil units as defined by the FAO/UNESCO soil map of the world 1:5000000 (1972-1978) and the corresponding area of the different ecosystems have been calculated from the soil maps and reports. For information on the ranges of characteristics of the concepts of soil is referred to the legend (Volume I) of the FAO/UNESCO soil map of the world (1972-1978) and Fitzpatrick (1983).

The procedure for compiling the global soil and vegetation distribution was summing the extent of the major soil units and their corresponding land cover type as given in the FAO/UNESCO soil map of the world Volumes (II-X) whereby the land use/vegetation types were classified according to the scheme given by Whittaker and Likens (1975). The extent of each map unit was calculated substracting the extents of the associated soils (30% for 1 associated soil, 40% if 2 associated soils were mapped as associations to the major soil) and inclusions (20% of the map unit area). As far as possible the land use type of these associated soils and inclusions was accounted for in summing the respective major units. The above procedure was followed for those soil reports, that have elaborated soils and land use tables (Volumes II, III, V, VI and X). For the remaining Volumes the generalized soil and vegetation maps as present in all reports were overlayed to estimate the areas; the resulting estimates were then supplemented with land use and vegetation data from the reports. The resulting figures were compared with data from World Resources Institute (1987) and Brown and Lugo (1982) on cultivated land, forest and woodland and other ecosystems and where necessary corrections were made. The result of this exercise can be found in table 2.1. The distributions for the map sheets seperately are presented in Appendix IV.1-IV.5

The resulting global extent of the major ecosystems from this study are compared with the area coverage presented by other authors in chapter 3, table 3.2.

The distribution given in table 2.1 will be used as a reference for further study in the following sections. It is important to realize, that the soil maps used are based on data which are 20-30 years old and may not reflect the present scientific knowledge of the world's soils. Furthermore, the major soil units given in the world distribution are generalized broad concepts of soil, that may include a wide variation in soil characteristics and they may occur in totally different climates. The general concepts however, can serve in this paper as a framework for the assessment of effects of land use changes. In the future a further subdivision into the minor soil units should be useful for further study in this field.

Table 1. The distribution of the soils and ecosystems (a) of the world (FAO/UNESCO soil map of the world VOL.I-X) (1000000 ha) (c)

Soil type (b)	Ecosystem 1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols	237.2	202.7	27.6	26.8	5.0	92.3	76.8	7.7	58.0	9.0		115.7	6.8		865.6
Cambisols		43.3	108.5	142.9	153.7	35.8	55.6	14.3	95.8	3.6		132.4		91.5	877.4
Chernozems			7.0	11.1	11.0	7.5			47.6			131.7		0.2	216.1
Podzoluvisols			22.1	19.7	29.0				66.3			71.9			209.0
Rendzinas		7.0	12.2	11.7		1.2	0.3		6.2			6.3		4.5	49.4
Ferralsols	395.9	147.8				54.0	80.2	13.0				74.7	2.2		767.8
Gleysols	52.9	5.4	30.6	43.8	195.4	8.2	36.8	8.5	35.6			145.9	74.9	127.3	765.3
Phaeozems		9.9	7.5	7.5	5.4	7.8	34.5	1.0	49.1			39.6			162.3
Lithosols	15.0	85.1	93.4	203.4	24.1	134.3	92.3	4.6	104.3	158.6	229.4	24.6	6.0	331.8	1506.9
Fluvisols	9.5	29.8	17.9	10.4	18.1	10.2	56.7	5.0	17.5			123.0	21.2		319.3
Kastanozems			17.1	33.0	5.0	17.7	9.5		185.9			119.6		42.7	430.5
Luvisols		51.8	93.5	66.8	9.2	103.1	82.0	20.8	93.7	37.4		193.2			751.5
Greyzems			2.3	2.3	9.7				11.2			10.0			35.5
Nitosols		56.1	18.9			11.0	26.0	18.0	1.9			52.2			184.1
Histosols	0.3		43.5	12.9	113.7		1.3	5.0	21.2			4.4	90.0	23.7	316.0
Podzols			166.3	62.3	69.5	5.4	1.2		30.0			44.1	8.0	20.0	406.8
Arenosols		40.9	3.8			56.2	165.3	14.6		193.4	51.4	9.5			535.1
Regosols		2.0	36.1	13.5	40.5	60.3	126.4	14.0	112.7	295.2	51.0	90.3		37.0	879.0
Solonetz						46.1	7.5	14.0	55.0	2.0	14.8	11.5			150.9
Andosols	0.2	8.5	17.1	9.6	10.0	2.0	2.1	2.3	9.5			18.2		8.5	88.0
Rankers		0.4	4.2	5,7		1.0			1.7						13.0
Vertisols						39.7	109.6	33.7	15.2	7.7		23.4			229.3
Planosols		17.4	1.0		2.0	16.5	26.3	35.0	6.1			14.0	1.0	7.5	126.8
Xerosols							23.7			290.3	44.4	39.3			397.7
Yermosols							30.5			166.9	782.6	73.5			1053.5
Solonchaks		2.4				7.0	24.9		22.2	36.0	83.9	8.6			185.0
Miscellaneous															1521.0
Total	711.0	710.5	730.6	683.4	701.3	717.3	1069.5	211.5	1046.7	1200.1	1257.5	1577.6	210.1	694.7	13042.8

[a] acc. to FAO/UNESCO soil map of the world (FAO, 1971-1978) VOL II-X.

[b] The presented acreages are estimated values of the extents of the soils, whether occurring as dominant soil, associated soil or inclusions, according to the FAO/UNESCO soil map of the world (VOL.II-X).

[c] ecosystems according to Whittaker and Likens (1975).
(1=tropical rainforest 6=woodland/shrubl. 11=extreme desert
2=tropical seasonal forest 7=savanna 12=cropland
3=temperate evergreen forest 8=trop.grassland 13=swamp/marsh
4=temperate deciduous forest 9=temp.grassland 14=tundra/alpine
5=boreal (taiga) 10=desert/semi-desert scrub)

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/Unesco Soil Map of the World at scale 1:5,000,000 (FAO/Unesco, 1971-1978) and complemented with other data (e.g. World Resources Institute, 1987). The published data in the literature deal 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. The combination of soil and land cover data bases would prove very useful in future.

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 space is by repeated mapping. The latter approach has not been used as widely as it could have been, but the approach is gaining importance through the development of remote sensing techniques.

Classification of vegetation is done 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 gives a good overview of the value of the Holdridge classification system.

Existing vegetation is mapped usually on the basis of physiognomy or floristics. Physiognomy provides a better estimate of phytomass since it into account 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. Since it would be difficult to show the climax for cultivated parts, which cover a large part of the earth's surface, 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. 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, South America 1:8,000,000); White (1981; Africa 1:5,000,000); Küchler (1965, North America 1:7,500,000); the Mediterranean Region (Unesco, 1:5,000,000); etc. Apart from the above continental maps mention must be made of the maps produced by the Institut de la Carte Internationale de la Vegetation (1988), which 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)

Table 3.2 List of global vegetation maps and digital data sets

reference	map scale	classification used	No.of types
Schithüsen (1968)	1: 25,000,000	physiognomic-environmetnal	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
Olson et al, 1983)	1:30,000,000 ($0.5x0.5^o$ grid)	physiognomic-environmental	43
Matthews (1983)	1^ox1^o grid	physiognomic (Unesco, 1973)	32
Emmanuel (1985)	$0.5^ox0.5^o$ grid	bioclimatic (Holdridge, 1967)	37

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 biomes. 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 biome types or, in the terminology of Walter (1985), zonobiomes.

In this study the same classification has been adopted as that proposed by Whittaker and Likens (1975) which in actual fact is the Unesco scheme in its modified form. The same scheme has been used by Ajtay et al (1979) and many other authors. Furthermore it has been applied in the map unit descriptions of the FAO/Unesco soil map of the world (see chapter 2) and this makes linking to soil geography an easy task. Additionally, since the Mattews vegetation data set was also based on the Unesco classification method, comparison between the various studies is possible without any further data interpretation and conversion.

Although a classification has been chosen, 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.

3.2.2 <u>Global land use classification</u>

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^o to 0.9^o 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^{o}x0.5^o 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. This makes comparison difficult, but a number of land cover types show a great similarity between studies. Houghton (1983) and World Resources Institute (1987) (table 18.3, 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.

Although there is good agreement about the carbon content of the soils of the world, there is wide 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); F. This study.

Land cover type	area (10^{12} m^2)					
	A	B	C	D	E	F
1. tropical rainforest	17.0	10.3	{	12.3	11.6	7.1
2. trop. seasonal forest	7.5	4.5	{12.1	6.7	7.7	7.1
3. temperate forest	12.0	7.0	12.0	{20.3	10.0	14.1
4. boreal forest	12.0	9.5	11.5	{	11.4	7.0
5. woodland, shrubland and interrupted woods	8.5	4.5	4.7	25.2	14.8	7.2
6. savanna(+trop.grassl.)	15.0	22.5	24.6[3]	{	12.5	12.8
7. temperate grassland	9.0	12.5	6.6	{27.4[4]	17.7	10.5
8. tundra, alpine	8.0	9.5	13.6	7.3	7.0	6.9
9. desert, semidesert	18.0	21.0	8.5	{	{21.3	12.0
10.extreme desert	24.0	24.5	16.4	{15.6	{	12.6
11.cultivated land	14.0	16.0	15.9	17.6	15.0	15.8
12.swamps, marshes and coastal land	2.0	2.0	2.5	—	—	2.1
13.bogs and peatland	—	1.5	—	—	—	—
14.other	2.0	4.0	—	2.4	5.2[5]	15.2[6]
total	149.0[2]	149.3[2]	129.7	134.8	133.6	130.4

[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; [6] incl. miscellaneous areas, urban areas, rocks, etc.

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 would be 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

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.

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.).

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

author	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 (1987)*				
closed forest	271.0	668.4	217.6	1157.0
open forest	27.0	204.7	458.5	690.2
total	298.0	873.1	676.1	1847.2

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

3.2.4 Temperate forest resources and distribution

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

	Area*						
	temp. Africa	North America	temp. S.Amer.	temp. Asia	USSR	Europe	temp. Oceania
closed forest	3.4	459.9	14.5	185.3	791.6	137.0	52.1
open forest	21.4	275.1	34.7	65.0	137.0	21.9	67.8
total	24.8	734.0	49.2	250.3	928.6	158.9	119.9

*refer to Appendix III for lists of countries in each region.
Source: World Resources Institute (1987)

Data for the temperate forest resources were taken from World Resources Institute (1987). 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 as is shown in table 3.6. 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 highlighted briefly later 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	shifitng 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, etc. 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.

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

	area*						
	trop.America		Trop.Africa		Trop.Asia		Total
	76-80	81-85	76-80	81-85	76-80	81-85	81-85
closed broad-leaved 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

*refer to appendix III for a list of countries in each region.
source: Lanly (1982)

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 3.9. 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 (1987) 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 reforestation rate for all temperate countries for which data are available is $13.5x10^{10}m^2y^{-1}$. 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 (10^7 m^2) in temperate zones.[1] (n.a. = no data available).

	temp. Africa	temp. N.Amer.	temp. S.Amer.	temp. Asia	USSR	Europe	temp. Oceania
closed forests	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
refo-restation[2]	102[3]	2529	93	5228[4]	4540	1020[5]	111

[1] countries for each zone are listed in Appendix III;[2] Reforestation includes reforestation 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 preliminary results of a SCOPE conference on acidification of tropical ecosystems (held in April 1986) 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:

1. 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;
2. ozone, direct leaf injury due to high ozone concentrations in the troposphere;
3. 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;

4. 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.

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).

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).

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 20×10^{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 dividing vegetation. 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 as 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 carbonaceous and nitrogenous greenhouse gases between soils and the atmosphere.

4.1 Carbonaceous 'greenhouse' gases

4.1.1 Carbon dioxide

4.1.1.1 Introduction

The understanding of the global carbon cycle has improved in recent years. Alterations in the carbon cycle have created a considerable debate among environmentalists and climatologists. 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: 835×10^{15} g C (Whittaker and Likens, 1975), 544×10^{15} g C (Ajtay et al, 1979), 584×10^{15} g C (present study with net primary biomass production taken from Ajtay et al, 1979), 594×10^{15} g C (Goudriaan and ketner, 1984).

The other carbon pools are: atmosphere (700×10^{15} g C), oceans (38000×10^{15} g C), fossil reserves (6000×10^{15} g C) (Goudriaan and Ketner, 1984) and caliche (petrocalcic horizons in arid and semi-arid regions; $780\text{-}930 \times 10^{15}$ g C; Schlesinger, 1982, 1984a).

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

approach	reference	carbon pool	cumulative loss	annual release
vegetation	Bolin (1977)	700	10-40[2]	0.3
	Schlesinger(1977)	1456		0.85
	Bolin et al(1979)	1672	—	1-2
	Ajtay et al(1979)	1635	—	—
	Houghton et al(1983)	—	45-76[1]	0.87
	Houghton et al(1985, 1987)	—	—	0.2-0.5
Soil	Bohn(1976)	3000	—	—
	Ajtay et al (1979)	2070	—	—
	Post et al (1982)	1395	—	—
	Schlesinger(1984b)	1515	36[2]	0.8
	Buringh(1984)	1477	537[1]	1.5-5.4
Modelling	Meentemeyer et al (1981)	1457	—	—
	Goudriaan and Ketner (1984)	1400	62.8	---
	Goudriaan (1988)	2000[3]	30	

[1]Since prehistory;[2]Since mid 1800's;[3]incl. biomass.

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. 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). Global estimates of soil carbon are much more accurate than estimates of the area assigned to the different ecosystems. Other causes of uncertainty are the estimates of the quantity of carbon present in and the annual release rate from the histosols (peat soils) of the world. In section 4.1.1.8 the soil carbon pool and the annual CO_2 release according to this study will be presented.

The estimated total CO_2 emission from fossil fuel combustion was $5.2x10^{15}$ g C y^{-1} in 1980 (Keepin et al, 1986) while the net 1980 emission from the terrestrial biota is estimated at $1x10^{15}$ to $2.6x10^{15}$ g C y^{-1} (Houghton et al, 1985). Projections for the year 2050 for the emission from fossil fuel combustion range from $2x10^{15}$ to $20x10^{15}$ 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. Sinks for atmospheric carbon dioxide are the atmosphere (55%), oceans (30%) and the terrestrial biota (15%) (Adema, 1988). The resulting increase of the atmospheric concentration is 0.5% or $3.3x10^{15}$ g C y^{-1}.

The emissions from terrestrial biota can roughly be divided into emissions due to deforestation and emissions due to the release of CO_2 from soils after deforestation. An additional effect is the reduction of the respiration as a consequence of the reduction of forest areas. Forests resynthesize 10 to 20 times more carbon per unit area than land in crops or pastures.

The emission of CO_2 from the world soils will be evaluated in 4.1.1.2 - 4.1.1.8. Net annual release rate of CO_2 from soils will be discussed in 4.1.1.6. The world soil and land cover distribution as presented in chapter 2 will be applied to give estimates of the soil carbon pool (see 4.1.1.8). In 4.1.1.9 a carbon bookkeeping model will be discussed to illustrate the total CO_2 emission from vegetation and soils.

4.1.1.2 The occurrence of organic matter in soils

Carbon fixation by terrestrial photosynthetic organisms is $110x10^{15}$ g C y^{-1}, of which about 50% finds its way into soils (Oades, 1987). Organic matter in soils is thus 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 an incompletely decomposed component (organic matter) and a stable component (humus). Humus in turn can be subdivided into strictly humus substances (humic and fulvic acids) and other products synthesized by microbes. Since a large portion of the soil carbon is found above a depth of 50 cm, Schlesinger (1984b) 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% (average of 34% C and 17.5% N with C/N decreasing from 21.9 to 12.9 between 0 and 83 cm) 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) after Kortleven (1960) 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.

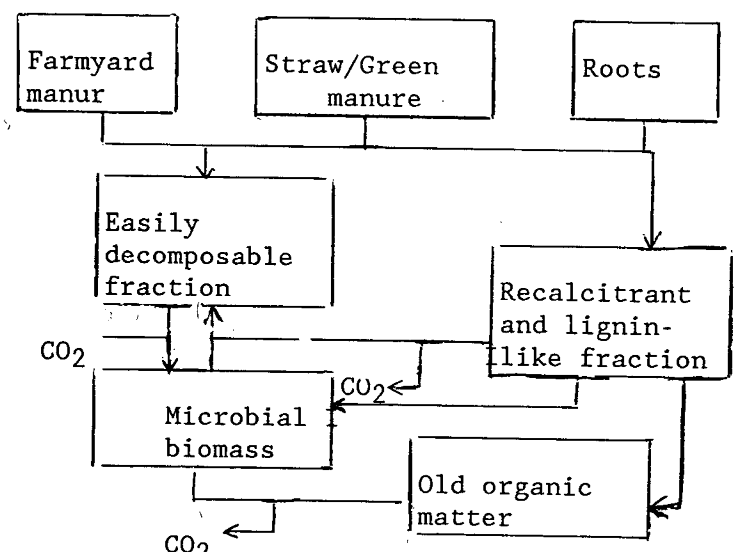

Figure 4.1. Schematic representation of the different pools of soil organic matter (Van der Linden et al, 1987).

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.

A variety of different types of models have been made to simulate the decomposition of soil organic matter (Jenkinson and Rayner, 1977; Paul and van Veen, 1978; Van Veen and Paul, 1981; Van Veen et al, 1984; Van der Linden et al, 1987; Parton et al, 1987). Van der Linden et al (see figure 4.1) separate the organic matter into a number of components: easily decomposable organic matter (sugars, amino acids), recalcitrant and lignin-like fraction, old organic matter and microbial biomass. In a less simplified model (Van Veen et al ,1984) the old organic matter has been subdivided into active (protected) organic matter and old organic matter; furthermore (dead) decomposable microbial material and recalcitrant microbial and plant material were distinguished in this model. Parton et al (1987) used a soil organic matter model with 3 fractions: 1. active soil organic matter (microbes, microbial products, organic matter with a high turnover); 2. slow soil organic matter with an intermediate turnover time; 3. passive soil organic matter with a long turnover time (200 to 1500 years). Plant residues are divided into structural pools (1 to 5 years turnover time) and metabolic pools (0.1 to 1 year turnover time).

4.1.1.3 Factors affecting the stability of organic matter in soils

-mineralogy

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, 1987). 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) also indicated, that in savanna soils parent material has an effect on soil carbon contents, but this effect may be attributed to the clay content which 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 m in diameter and would not be accessible to microorganisms (Oades, 1987).

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; physical aspects

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. Higher residual organic carbon in clay soils is explained by:

- the proportion of material from dead cells remaining in close proximity to surviving organisms;
- the protection from predation by larger organisms (amoeba, eelworms) because of restricted movement of these organisms;
- restricted movement of enzymes.

Van Veen et al (1987) and Schimel et al (1985a, 1985b) found, that finer, clayey soils tend to be more preservative than the coarser, sandy soils. In their experiments larger amounts of added ^{14}C were incorporated in microbial biomass and soil organic matter fractions in clayey soils than in sandy soils. Jenkinson (1977) indicated, that clay soils retain more organic carbon than sandy soils. The effect however, is 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 (Soil Survey Staff, 1975; FAO/Unesco, 1972) Jones indicates a linear relationship between soil carbon and clay content between 35 and 80% clay.

-<u>soil texture; chemical aspects</u>

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.

-<u>soil structure</u>

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.

-<u>soil fertility</u>

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, 1987). 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, 1987) 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.

-<u>vegetation-soils effect</u>

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 (1987) 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 (Janssen, 1986). 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. Under forest vegetation the organic carbon content is higher than under cultivation, since:

- the biomass produced on arable land is harvested;
- the organic substances of crops are less resistant against decomposition;
- the average temperature on the soil surface in a forest is lower than in cultivated fields;
- losses due to leaching are smaller in forests.

Substrates are less well decomposable and possibly more re-immobilization of N occurs in rangeland soils than in cultivated soils (Schimel, Coleman and Horton, 1985). Decomposition rates were higher for cultivated soils, indicating that microbial biomass had narrower C/N ratios. Apparently the microbial species composition was changed with the importance of fungi declining (see also Voroney et al, 1981).

-<u>climate</u>

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 soil moisture regime (Soil Survey Staff, 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 (moist) soil moisture regimes (Soil Survey Staff, 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; Scharpenseel, in prep.).

In dry tropical areas, the biological activity is reduced during parts of the year, thereby extending the period needed to reach equilibrium. It should be remembered, that in dry areas decomposition rates are lower than in moist regions. In the temperate regions the biological activity is greatly reduced during the winter . In the 78% of the tropics, that has an ustic or aridic soil moisture regime (Soil Survey Staff, 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, although this cannot be supported by experimental data. Schlesinger (1982) reports, that arid soils lose organic matter upon cultivation but may gain in carbonate carbon in calcrete or caliche.

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.

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).

-the soil's preservation capacity

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 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.

4.1.1.4 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.

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. The pattern of decomposition can be described according to a first order rate process in which organic matter decays exponentially (Henin and Dupuis, 1945; Stanford and Smith; 1972; Jenkinson, 1977; Jenkinson and Rayner, 1977) :

$$X_t = X_0 x e^{-kt}$$

where: X_t and X_0 are the amounts of organic material at time t and initially;
k = decay constant.

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 (Soil Survey Staff, 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 (Soil Survey Staff, 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 (1972) 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, whereas 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.

Shifting cultivation seldom results in substantial soil organic matter depletion; usually the carbon contents are maintained at 75% of natural equilibrium levels, 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.

4.1.1.5 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).

4.1.1.6 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 forest with only sawlogs removed and only 169 kg C ha^{-1} for clear cut forest with removal of all woody material, both in a mixed deciduous forest. This indicates, that a forest which is left to regrow after clearing and is not 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.

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.

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)

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.

4.1.1.7 Composition of generalized soil carbon profiles

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} (= 5×10^2g 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. For the calculation of the quantity of organic soil carbon the following formula is used:

$$\text{soil carbon} = \sum_{n}^{1} p \times d \times 10$$

where: p = bulk density of the soil material (kg m^{-3});
d = thickness of soil layer n (m)
soil carbon = total quantity of organic carbon in the soil profile between 0 and 100 cm.

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 is not included in soil carbon.
- the following conventional conversion factor for the calculation of carbon from organic matter: org. matter = 1.72 x C.

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 appear to be in accordance with Buringh (1983). 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.

Table 4.4. Generalized organic carbon profiles for the major soil units of the FAO/UNESCO (1972-1978) 1:5,000,000 soil map of the world

Soil type Generalized soil carbon profile tot. soil C (10^6g C ha^{-1})

	generalized soil carbon profile in natural (typical) ecosystem						forest		grass land	crop-land
	ecosystem	C (%) at depth (cm)					prim.	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	20	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	forest	depth33cm;bd=0.25					375	375	375	375
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

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.

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.

4.1.1.8 Soil carbon pool in world soils and annual release of CO_2

Using the generalized soil carbon profiles and the world distribution of soils and ecosystems (table 2.1 and 4.4 respectively) and data on the annual land cover changes as presented in Houghton (1983) (see Appendix I), the total global soil carbon pool and the annual CO_2 release from soils after interferences in ecosystems can be estimated. The carbon pool calculated this way is about 1700×10^{15} g C and the annual release is 0.75×10^{15} g C (see also table 4.1).

Using the land cover changes given in Houghton et al (1983) (see appendix II) and the world soil and land cover distribution in table 2.1, and data on biomass for the different ecosystems presented by Ajtay et al (1979) the annual release rate from vegetation is 1.05×10^{15} g C. The total global release from soils + vegetation is then 1.8×10^{15} g C.

4.1.1.9 Emission of CO_2 from vegetation and soil 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; Esser, 1987; and many other authors).

A number of dynamic carbon models have been developed, e.g. those by Goudriaan and ketner, 1984, Esser, 1987 and Goudriaan, 1988. A number of these models is discussed in Bolin (1986), pp 130-133. The most systematic analysis of the terrestrial source of atmospheric CO_2 has been conducted by Houghton et al (1983, 1987), who used a bookkeeping model. This model will be discussed in more detail since it uses the data bases already presented in chapter 2. The results of above dynamic models and Houghton (1983,1987) will be compared in table 4.5b.
The model is based on the FAO production yearbooks (1949-1978) and on data presented by Myers (1981) (see Houghton et al, 1983). The estimated total annual release rates from the terrestrial biota are 1.82×10^{15} and 4.70×10^{15} g C y^{-1} respectively. A third data set, the population based model, was developed by Houghton et al (1983) assuming 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 infood production was assumed to be achieved through increasing agricultural output on existing cropland. This third data set yielded a release rate of 2.6×10^{15} g 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% (see table 4.5 No.2). In another test 0 carbon loss was assumed following the clearing of forests (see table 4.5 No.3). 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) (biomass of tropical forests 46 to 183 tons ha^{1}). 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. In a later study Brown and Lugo (1984) presented even lower estimates of biomass per unit of area based on volumes. Their weighted carbon density of tropical forests is 53 tons C ha^{-1}, which is well out of range of the formerly proposed average densities of 124 tons C ha^{-1} by Brown and Lugo (1982) or 188 tons C ha^{1} proposed by Whittaker and Likens (1975) both for tropical forests.

Table 4.5a The release of carbon from terrestrial biota (figures in 10^{15} g C y^{-1}) based on the population based model.

	total	agricultural clearing	harvest/ regrowth of forest	Oxidation of wood removed	abandonment of agriculture	afforestation	clearing for pas-
1.^{x}	2.61	1.30	0.63	0.56	0.066	0.039	0.23
2.^{x}	2.35	1.01	0.63	0.56	0.041	0.039	0.23
3.^{x}	2.30	1.30	0.34	0.56	0.073	0.050	0.23
4.^{x}	1.98	1.03	0.43	0.49	0.066	0.039	0.14

^{x}1= population based; 2= reduced soil respiration following gricultural clearing; 3= no loss of soil carbon following harvesting/regrowth of forests; 4=reduced amount of carbon per unit area in vegetation (ref. Brown and Lugo, 1982).

Source: Houghton (1983)

Houghton et al (1985, 1987) repeated their analysis with reduced soil carbon loss rates (see 4.1.1.7) 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 2.6×10^{15} g C y^{-1}. In this figure the contribution of the tropics is 0.9 to 2.5×10^{15} g C y^{-1} of which 0.4 to 0.8×10^{15} g 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 2.6×10^{15} g C y^{-1}. The soils contribution to this emission is about 20% or 0.2 to 0.5×10^{15} g C y^{-1} (Houghton, 1987, pers.comm.).
Finally, estimates of fluxes made by a number of authors are compared in table 4.5b. Goudriaan and Ketner (1984) and Goudriaan (1987,1988) consider the CO_2 fertilizing effect responsible for the negtive flux to the biota. Esser (1987) estimates that in the 80's the emissions from terrestrial biota are balanced by CO_2 fertilizing effects.

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.

Table 4.5b Comparison of the annual flux of carbon from terrestrial biota by various authors.

reference	type of model	annual release (10^{15} g C y^{-1})
Houghton e.a.(1983)	bookeeping model of terr. biosphewre	1.8-4.7
Goudriaan and Ketner (1984)	dynamic model incl. oceans	- 0.9
Houghton e.a.(1987)	bookeeping model of terr. biosphere	1.0-2.6
Goudriaan (1987)		0 - -0.5

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 $13x10^{15}$ g C y^{-1} (Zimmermann et al, 1982). With all uncertainties the above emission of CO_2 by termites could range from $6.5x10^{15}$ to $26x10^{15}$ g C y^{-1}. Later studies by other authors revealed however, that the biomass comsumption rate assumed by Zimmermann et al 91982) 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 emissions from termites are tropical wet savanna, areas that have been cleared or burned and cultivated land in developing countries (see also section 4.1.2.7).

4.1.1.10 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.

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.

4.1.2 Methane

4.1.2.1 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).

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 last 10 years is about 18 ppbv per year or 1.1% per year (Bolle et al, 1986). The observed increase over 1970-1980 requires a yearly excess of 70×10^{12} g CH_4 of sources over sinks (Blake et al, 1982) to 50×10^{12} g CH_4 (Bolle et al, 1986). 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 (reaction with OH. is believed to be the major mechanism by which CH_4 is removed from the atmosphere, see 1.2.2).

Table 4.6. CH_4 emission rates from various sources.

Source	CH_4 emission ($\times 10^{12}$ g y^{-1})
waterlogged soils (paddy rice)	70-170[a]
organic soils (peat)	25- 70[a]
landfill sites	30- 70[f]
oceans/lakes/other biogenic	15- 35[a]
intestines of ruminants	66-90[a]
termites	6- 42[b]/2-5[c]
exploitation of natural gas	30- 40[a]
coal mining	35[a]
biomas	55-100[a]
other nonbiogenic	1- 2[a]
total	333-654[d]/329-617[e]

[a] Bolle et al (1986); [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).

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 4.1.2.2. The factors influencing the actual emission (i.e. production minus oxidation or consumption) are highlighted in 4.2.1.3. A list of possible sources and estimated volumes of emissions are presented in table 4.6.
CH_4 may also be consumed by microorganisms in aerobic soils in tropical regions (see 4.2.1.11).
Very few data are available about possible biological sinks of methane. In the present study the methane emission from paddy rice soils, wetlands and emission by herbivorous insects (termites) will be discussed in more detail. Methane production due to biomass burning and emission by ruminants are discussed only briefly.

In table 4.7 the increase of production of CH_4 by various sources is estimated for selected years.

Table 4.7. Estimated trend of CH_4 emission rates in 10^{12} g y^{-1}.

source	year					
	1940	1950	1960	1970	1975	1980
ruminants	53.5	58.0	68.5	78.5	84.0	86.0
rice paddies	64.5	74.5	89.0	105.0	115.0	117.0
swamps/marshes	79.0	73.0	63.0	54.0	51.0	47.0
other biogenic	15.5	18.0	20.5	23.0	24.0	25.0
biomass burning	49.0	57.0	65.0	71.0	75.0	79.0
natural gas	2.0	5.0	11.0	29.5	30.0	34.5
coal mining	19.0	19.0	26.0	29.0	31.0	35.0
total	283.0	305.0	343.0	385.0	410.0	423.0

Source: Bolle et al (1986)

4.1.2.2 Methanogenesis

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. $CO_2 + 4\ H_2A \dashrightarrow CH_4 + 2H_2O + 4A$
2. $RCOOH \dashrightarrow RH + CO_2$

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).

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 (Jacobsen et al, 1981, Cicerone and Shetter, 1983). Nitrate and sulphate or their reduction products repressed mathane 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.

4.1.2.3 Factors affecting 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.

substrate and nutrients
Fluxes may correlate with availability of 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).

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).

profile of methane concentration
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).

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) who reported that in a rice paddy more than 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).
- loss 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). 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 repiration and photosynthesis rates.

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 (sea water contains considerable amounts of SO_4^{2-}) are lower than from fresh water wetlands.

4.1.2.4 Distribution of paddy soils and other waterlogged soils

The FAO (1985) estimated the harvested area of paddy rice at $144x10^6$ ha of which 95% is located in the Far East (see table 4.8). This area corresponds to approximately 9.5% of the total cultivated area. The area harvested of paddy rice has increased from $86x10^6$ to $144x10^6$ 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 III a list of data concerning the geographic distribution of paddy rice cultivation are 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. The data for the harvested area of paddy rice world are in Appendix III.1 and III.2. The data for Asia separately are in Appendix III.3 and III.4.

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%.

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, 1971-1978). 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 thus are flooded 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 paddy rice (10^7 m^2)

	1935	1950	1960	1970	1980	1985
Africa	1850	2900	2880	3960	4894	5467
North/Central America	540	1040	1280	1428	2076	1914
South 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	n.a.	100	356	637	667
World	85800	94200	119500	134231	143746	144674

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

4.1.2.5 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.

4.1.2.6 Emission rates 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, 1985) 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.
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. Kimura et al (1984) found, that H_2 production was more active at younger roots, while CH_4 production was greater at the older roots.
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). The principal means of escape of methane is through the rice plants 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 those parts of the paddy fields, where no plants are grown.

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

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

author	emission rate (g CH_4 m^{-2})	remarks
Koyama (1963)	app. 21	laboratory
Ehhalt and Schmidt (1978)*	app. 140	laboratory
Cicerone et al (1983)	22- 28	California (in situ)
Holzapfel-Pschorn /Seiler(1985)*	45-110	Italy (in situ)

*quoted in Bolle et al(1986)

Extrapolation of the above 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. The data provided by Holzapfel-Pschorn and Seiler (1985) will be used to calculate the global emission rate. The annual increase of the area under rice and the related CH_4 emission is 1.23%. The estimated 1983 global emission from rice paddies based on the data in table 4.8 is $52x10^{12}$ to $126x10^{12}$ g CH_4 y^{-1}, the annual average increase being $0.7x10^{12}$ g CH_4 y^{-1}. The 1987 release of CH_4 from paddy rice land is then $55x10^{12}$ to $134x10^{12}$ g CH_4 y^{-1}. These estimates are slightly lower than those presented by Bolle et al (1986).

4.1.2.7 Distribution of wetlands

The total extent of wetlands is $210x10^6$ ha (table 2.1), a figure higher than that used by Bolle et al (1986) who used $160x10^6$ ha, but lower than the $260x10^6$ and $250x10^6$ ha given by Clark (1982) and Olson et al (1983) respectively. The (temporarily) waterlogged soils, such as gleysols ($91x10^6$ ha), fluvisols ($21.2x10^6$ ha) are considered under paddy soils and other waterlogged soils (see 4.1.2.4). About 50% of the area of wetlands is located in tropical and subtropical regions.

The Histosols (peat soils) cover a larger area than that covered by wetlands since those soils are not necessarily in swamps or may be drained.
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. 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).

4.1.2.8 CH_4 emission and absorption rates of 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 (Valiela, 1984) suggests that methane may actually be oxidized to CO_2 by sulfate reducing bacteria.

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. In table 4.10 rates of methane production are compared. As in rice paddies the temporal and spatial variation is extremely high (Harriss et al, 1982; Wiebe et al, 1981).

Table 4.10 CH_4 emission rates from wetlands

		CH_4 flux ($g\ m^{-2}\ y^{-1}$)
freshwater environments		
Michigan swamp, waterlogged[1]		40.0
Cypress swamp, waterlogged[2]		3.65
Georgia cypress swamp, waterlogged[2]		32.9
Florida cypress swamp; fertilized, waterlogged[2]		354.1
Florida cypress swamp, unfertilized[2]		24.5
Virginia Dismal swamp, waterlogged[4]		0.4 - -7.3
Virginia Dismal swamp, drought conditions		-0.4- -1.8
Louisiana Marsh, panicum spp[6]		440.0
Northern Minnesota peatlands[7]	bog	57(46- 75)
	bog	17(7 - 64)
	fen	1(1 - 2)
	bog	71(12-171)
	fen+bog	151(22-709)
	sedge meadow	242
	rice bed	180(46-322)
	shoreline fen	60(58- 62)
Alaska coastal tundra, waterlogged[9]		43 ± 10
Alaska moist tundra (not waterlogged)[9]		2 ± 1
Alaska meadow tundra, waterlogged[9]		15 ± 7
Alaska alpine fen, waterlogged[9]		105± 5
Alaska boreal marsh, waterlogged[9]		39 ± 2
Virginia marsh, spartina cynosuroides[10]		18 ± 6
Saltwater and brackish environments		
Georgia salt marsh; tall Spartina, waterlogged[3]		0.3
Georgia salt marsh; interm. Spartina, waterlogged[3]		11.0
Georgia salt marsh; short Spartina[3]		40.0
Louisiana salt marsh[5]		5.0
Brackish spartina patens[6]		73.0
Spartina alterniflora[6]		4.0
Virginia salt meadow[8]		0.43
Virginia; short Spartina alterniflora[8]		1.3
Viriginia; tall Spartina alterniflora[8]		1.2
Virginia Brackish Spartina cynosuroides[10]		29 ± 3
Virginia salt Spartina alterniflora+cynosuroides[10]		6 ± 1

[1] Baker-Blocker (1977);[2] Harriss and Sebacher (1981);[3] Wiebe et al (1981);[4] Harriss et al (1982);[5] Smith et al (1982);[6] Delaune et al (1983);[7] Harriss et al (1985);[8] Bartlett et al (1985); [9] Sebacher et al (1986); [10] Bartlett et al (1987).

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.

Assuming a period of flooding of 6 months per year and emission rates of 45 to 110 g of CH_4 m^{-2} $year^{-1}$ (Bolle et al, 1983), the estimated global emission range is then 50 to 120×10^{12} g CH_4 y^{-1} from the 210×10^{6} ha of wetlands. In the above estimate the possible occurring negative flux towards wetlands is not considered.

4.1.2.9 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 $150x10^{12}$ 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 $75x10^{12}$ to $310x10^{12}$ 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 $50x10^{12}$ g y^{-1} (ranging between $10x10^{12}$ and $90x10^{12}$ 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 ($2x10^{12}$ to $5x10^{12}$ 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 $14x10^{12}$ g CH_4 y^{-1} by termites (with a range of $6x10^{12}$ to $42x10^{12}$). 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. 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.

Literature data also contradict Zimmermann's (1982) increase in termite densities due to clearing of forests (Collins and Wood, 1984). Nevertheless, using the emission rates for the various ecosystem types given by Zimmermann et al (1982), the annual <u>increase</u> of the CH_4 emission by termites is about $0.08x10^{12}$ g CH_4 y^{-1}.
This calculation is based on the the distribution of soils and ecosystems from table 2.1 and statistics on land cover changes presented by Houghton et al (1983) (see Appendix I). This amount is small compared to the total estimated release and will not influence the CH_4 budget substantially.

<u>ruminants</u>

Estimates of the CH_4 production by ruminants based on the world population of domestic and non domestic ruminants made by Seiler (in Bolle et al, 1986) range between $72x10^{12}$ and $99x10^{12}$ g CH_4 y^{-1}. These figures are known to be more accurate than other global methane production estimates.

4.1.2.10 <u>CH_4 emission due to biomass burning</u>

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 $25x10^{12}$ to $110x10^{12}$ 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 $53x10^{12}$ to $97x10^{12}$ 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}.

4.1.2.11 Oxidation of methane in 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 $20x10^{12}$ g y^{-1}. Seiler and Conrad (1987) reported a global methane consumption in soils of $32 \pm 16x10^{12}$ g y^{-1}.

4.1.2.12 Global sources and sinks of methane

The data on the global emissions from various sources are uncertain. The only source which is known with some grade of accuracy is the world ruminants population. However, the global budget is well known and the major sources are known. The allocation of the total budget among the sources is still not well known.

The present annual release rate is $333x10^{12}$ to $654x10^{12}$ or $329x10^{12}$ to $617x10^{12}$ g y^{-1} of methane according to data presented in table 2.1 using the termite emission rates given by Rasmussen and Khalil (1986) and Seiler (1984) respectively. The average methane emission used by Bolle et al (1986) is $425x10^{12}$ g y^{-1}. The total atmospheric mass of CH_4 is about $4400x10^{12}$ g of CH_4, implying a residence time of about 7.5-10 years.

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):

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

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.

Adopting an average OH. concentration of $5x10^5$ molcules cm^{-3} (Crutzen and Gidel, 1983) gives a sink of $260x10^{12}$ g y^{-1}. About $60x10^{12}$ g CH_4 is transported to the stratosphere each year where it is oxidized and another $32x10^{12}$ g is decomposed in soils in semi arid regions by (Keller et al, 1983; Seiler, 1984; Seiler and Conrad, 1987). The total sink strength is then $350x10^{12}$ g CH_4 y^{-1}. Clearly, the CH_4 budget is not balanced in this estimate.

4.1.3 Carbon monoxide

4.1.3.1 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 4.11. 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).

4.1.3.2 Sources and sinks of CO

Table 4.11. Estimates of sizes of possible sources and sinks of carbon monoxide by various authors.

	10^{12} 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 of nat. NHMC[1]	280-1200	560	Logan et al (1981)
oxid.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)
	1600-4000	3170	Logan et al (1981)
transp.to stratosphere	190- 580	170	Crutzen (1983)
soil uptake	190- 580	450	Crutzen (1983)
		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 below in table 4.11. 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.

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 $17x10^{12}$ g y^{-1} (ranging from 3 to $30x10^{12}$ of which $1x10^{12}$ to $19x10^{12}$ g y^{-1} is produced in dry tropical areas). CO consumption ranges from $300x10^{12}$ to $530x10^{12}$ g y^{-1} of which $70x10^{12}$ to $140x10^{12}$ g y^{-1} is oxidized in the humid tropics (Seiler and Conrad, 1987).

The global sources range between 1270 and $5700x10^{12}$ g CO yr^{-1} with an average of $2920x10^{12}$. The global sink strength ranges 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 given in table 4.11.

4.1.3.3 Conclusions sections 4.1.2 and 4.1.3

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 (1.43%) and the increase of methane emission by termites due to shifts in land use (1.5 to 4%) correlate well to the atmospheric increase. The increase in CH_4 concentration correlates remarkably well with the increase of the human world population. This indicates, that the increase in atmospheric CH_4 concentration is most likely related to anthropogenic activities (Bolle et al, 1986). The role of termites seems to be overemphasized 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.

Apart from measurement techniques there are a number of fields where the present knowledge is inadequate:
- the geographic distribution of soils used for wet rice cultivation;
- 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 wetlands;
- the relation between the type of peat 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.

4.2 Nitrogenous greenhouse gases

There are 6 major channels of nitrogen loss from soils:

1. biological denitrification;
2. chemodenitrification;
3. nitrification
4. NH_3 volatilization;
5. leaching;
6. erosion.

The first two processes are known to produce quantities of N_2O. Nitrification is a process during which both NO and N_2O are emitted mainly from aerobic soils. The process of ammonia volatilization will be highlighted in section 4.2.3 while the processes of erosion and leaching are unimportant in the context of this paper. Section 4.2.1 deals primarily with the production of N_2O. 4.2.2 will evaluate the sources of NO_x ($NO + NO_2$).

4.2.1 Nitrous oxide (N_2O)

4.2.1.1 Introduction

The atmosphere contains about $1500x10^{12}$ g N_2O-N; the annual increase is about 0.25 % or $3.5\pm0.5x10^{12}$ g N y^{-1}. The principal sink for N_2O, 10.3 $\pm$ $3x10^{12}$ g N y^{-1}, is stratospheric protolysis (see chapter 1). As the lifetime of N_2O in the atmosphere is 100 to 200 years, changes in production will have a long term effect. A list of all the global sources is presented in table 4.12.

Table 4.12 Global budget of tropospheric Nitrous oxide (figures in 10^{12} g N y^{-1})

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

Source: Seiler and Conrad (1987).

Release of oxides of nitrogen (N_2O, NO and NO_2) is known to occur during biological denitrification (see 4.2.1.2), chemical denitrification (4.2.1.3) and nitrification (4.2.1.4). 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.

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.

4.2.1.2 Biological denitrification

Biological denitrification is the dissimilatory reduction of nitrate (NO_3^-) or nitrite (NO_2^-) to gaseous froms of nitrogen by essentially aerobic bacteria producing molcular 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. The importance of denitrification may therefore have been overestimated (Fillery, 1983). 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):

$$NO_3^- \text{ ----> } NO_2^- \text{ ----> } NO \text{ ---> } N_2O \text{ --> } N_2 + H_2O$$

There is still uncertainty about NO being an intermediate product of denitrification (see e.g. Poth and Focht). 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. Ammonia may inhibit the further reduction of NO_2^- by denitrifiers (see 4.2.1.3). 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, pers.comm).
The highly reactive nature of NO makes its detection difficult. Its existence as an intermediate product is still in doubt. Most but not all denitrifying bacteria can reduce N_2O to N_2. N_2O is a gas - it 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.

In the following paragraphs the term N_2O fraction is used to indicate that fraction of the nitrogenous gaseous products of denitrification and nitrification which evolves as N_2O.

Conditions that influence the products of denitrification.

- Oxygen pressure: the oxygen content of 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).

Few studies have evaluated the effect of the oxygen level on the fraction of N_2O. 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 (see also Letey et al ,1981) 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. The nitrification process is discussed in more detail in 4.2.1.4.

- 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. However, root exudates may stimulate denitrification and roots may also create anaerobic conditions by depleting oxygen in the rhizosphere.

- Temperature: the optimum temperature for denitrification is about 35^oC. Important losses may also occur at lower temperatures, especially in grassland above 8^oC and when cattle slurry is applied, i.e. with an abundant and mobile C-source even at lower temperatures (S.C. Jarvis, pers.comm). Keeney et al (1979) found, that while the rate of denitrification was low at temperatures below 15^oC, the amount of N_2O (44-50% of the total gas production) was equivalent to that evolved at 25^oC. 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.

- Nitrate concentration: 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, whereas 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. Possibly a delay of N_2O reductase occurs until a threshold N_2O level is reached (Fillery, 1983). 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.

- pH (soil reaction): 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. Another explanation is inhibition of NO_3^--reduction by affecting NO_2^--reductase.

- Soil organic matter content: a good supply of readily decomposable organic matter for energy supply is a prerequisite for the occurrence and speed of the denitrification process (see also Rolston, 1978).

4.2.1.3 Chemodenitrification

Chemical denitrification is the reduction of nitrite or nitrate to gaseous froms of N by chemical reductants producing molcular N or oxides of N (adapted from SSSA, 1984).
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:

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

or:

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

For both reactions acidic conditions are required.

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 (see also 4.2.1.2).

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.

4.2.1.4 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:

$$NH_4^+ \underset{-H^+}{\overset{-1/2O_2}{\Longleftrightarrow}} NH_2OH \Longleftrightarrow NOH \underset{-2H^+}{\overset{+1/2O_2}{\Longleftrightarrow}} NO_2^- \underset{-H^+}{\Longleftrightarrow} NO_3^-$$

```
                 / \
                /   \
     chemical  /     \  biological
  denitrification    denitrification
               \     /
               N2O+NO
```

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. 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.

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. Mean production of nitric and nitrous oxides were at least an order of magnitude lower for the soil denitrifier. The ratio of NO : N_2O at 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 by Lipschultz et al, 1981.

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.

Nitrifying organisms may contribute significantly to NO and N_2O emissions from soils. Bremner and Blackmer (1981) found in their studies, that:

1. soils evolve N_2O even when the moisture content is low and are well aerated (i.e. under conditions known to inhibit denitrification);
2. emissions of N_2O from aerated soils are correlated not with NO_3^--concentration but with nitrifiable N-contents (see also Minami and Fukushi, 1983).
3. emissions of N_2O from well aerated soils are greatly increased by the addition of nitrifiable forms of N (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). Addition of plant residues and other nitrogenous organic materials increases the emission of N_2O.
4. emissions of N_2O from soils amended with ammonium can exceed those from soils amended with nitrate even when the soils are saturated with water to promote denitrification.

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.
NO and NO_2 may also be formed during chemodenitrification (see 4.2.1.3).

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. Poth and Focht indicated that nitrification is not a direct source of N_2O. Nitrosomonas is a denitrifier 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) even found 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.

4.2.1.5 Factors affecting the variability of N_2O emission

Drainage and soil water status

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.

Reduction of N_2O to N_2 in soil is greater in anaerobic environments. Small amounts of nitrate stimulate the reduction of N_2O in anaerobic conditions. 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) that flooded soils contribute less N_2O to the atmosphere than drained soils (see also Goodroad and Keeney, 1984 in table 4.13). 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.

Terry et al (1980) studied the effect on N_2O emission of flooding of an organic soil and found that it is reduced virtually to zero. Drained histosols however, may contribute considerably to atmospheric nitrous oxide (see also Goodroad and Keeney, 1984 in table 4.13). Denmead et al (1979b) suggested, that flooded mineral soils in general may contribute less N_2O to the atmosphere than is commonly assumed. Sahrawat and Keeney (1986) concluded in their review that flooded conditions do not promote N_2O emission. Contrary to the above quoted literature, Minami (1987) found appreciable N_2O emissions from flooded rice fields of 0.12×10^{-4} to 0.19×10^{-4} g N m^{-2} h^{-1} averaged over the 120 day growing season. Smith et al (1983) reported fluxes of 0.04×10^{-4} to 0.06×10^{-4} g N m^{-2} h^{-1} from marshlands and 0.01×10^{-4} to 0.04×10^{-4}g N m^{-2} h^{-1} from open water.

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.

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).

<u>Diurnal and seasonal variation; influence of temperature</u>

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 time of the day of the measurement 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) 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.

<u>Distribution of soil organic matter</u>

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 is a significant factor influencing the high spatial variability of natural denitrification rates in soil (Parkin, 1987). The observed variability (see also tables 4.13 and 4.14) may be a result of the patchy distribution of organic matter.

<u>Chemical status of the soil</u>

Yoshida and Alexander (1970) showed that both phosphate and high soil pH enhance N_2O production in cell suspension of <u>Nitrosomonas europaea</u>. 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. This phosphate effect was greatest at 60% water holding capacity. 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.

<u>Spatial variability</u>

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. One general conclusion that may be drawn is that 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. For N_2 + N_2O emission rates the number is 156 to 4100 measurements. To achieve an accuracy of 50% the number of measurements should be 14 for N_2O alone and 7 to 165 for N_2 + 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. They conducted the experiments without replicates. 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.

4.2.1.6 Measurement techniques

Total gaseous losses from field soils can be measured by a number of different methods:

1. 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).
2. 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.
3. Chamber methods, with or without acetylene inhibition of N_2O reduction: in this method a sealed chamber is placed on or iserted into soil in which the confined atmosphere above the soil is sampled and its composition determined. This method can be combined with the acethylene (C_2H_2) inhibition of N_2O reduction in soil to measure total N-loss (=N_2 + N_2O). Most experiments deal with the total N-losses from soil. But in most cases measurements are also done without addition of acetylene. Chambers may be closed, vented or with a continuous flow of air. In other cases the chambers are kept over the soil surface for short periods to prevent errors occurring in closed or open chambers.
4. 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.
5. Micrometeorological approach: this method which is based on the turbulent diffusion theory. Concentration profiles are measured aboveground. Evolving gas flux intensity can be calculated from wind and gas profiles and thermal gradients. Other parameters include friction velocity, plant displacement height and correction for thermal instability. On the basis of the reported rates of N_2O emission, current instrumentation required for this micrometeorological technique is not sensitive enough to use under most

field conditions (Hauck, 1986). Nevertheless, micrometeorological techniques theoretically are the preferred approach for measuring gaseous flux since they:

- would minimize problems of spatial variability by permitting an average flux to be measured over a large area;
- measure gas fluxes without disturbing soil processes;
- permit rapid sequential measurements and facilitate study of environmental effects on gaseous fluxes. Denmead (1983) wrote a good review on the micro-meteorlogical approach. Examples of such studies are Hutchinson and Mosier (1979), Mosier and Hutchinson (1981) and Minami (1987).

During the past few years field estimates have been almost exclusively utilizing the chamber technique with or without acetylene blockage of N_2O reduction.

Table 4.13. Emission rates of N_2O from natural undisturbed soils and land cover types.

Source	method	N_2O emission* ($10^{-4} gNm^{-2}h^{-1}$)	cover type
CAST(1976)[c]	?	0.02	uncropped land
Demead et al (1979b)	o-	0.002	grassland (dry)
		0.63	grassland (after rain)
Mosier et al (1981)	c-	0.1	native prairie(summer)
Smith et al (1982)	c-	0.06	undrained marsh (average)
Smith et al (1983)	c-	0.04 -0.06	undrained marsh (average)
		0.01 -0.04	open water (average)
Kaplan (1984)	?	1.21[c]	tropical rain forest
Goodroad and Keeney (1984)	c-	0.02[a]-0.02[b]	burned tall gr. prairie
		0.03[a]-0.02[b]	unb. tall gr. prairie
		0.05[a]-o.15[b]	deciduous forest
		0.28[a]-0.36[b]	coniferous forest
		0.65[a]-1.49[b]	drained marsh
		0.01[a]-0.01[b]	undrained marsh
		0.31[a]-0.31[b]	wet meadow
Cates and Keeney(1987)	c-	0.004-0.022	native prairie

* In all experiments the "chamber technique" was used for collecting the emitted gases; o= open chamber with continuous flow of ambient air; c=closed chamber;1= with/without C_2H_2 inhibition method; 2= ^{15}N labelling method;-= measurement of N_2O emission exclusively. [a] summer-autumn 1979;[b] March-November 1980; [c] $2x10^{10}$ molecules cm^{-2} sec^{-1}; [c] quoted in:J.R. Simpson and K.W. Steele (1983)

4.2.1.7 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.

4.2.1.8 Natural undisturbed soils

Table 4.13 lists published data for N_2O flux from undisturbed natural soils in temperate regions. Little data could be obtained for tropical soils. More data of N_2O fluxes from tropical rain forests will be available in the near future (current NASA research in Amazonia). Kaplan (1984) states, that undisturbed tropical moist forest areas may be a significant source of N_2O. Nitrous oxide released from soils near Manaus (Brasil) is 10 to 30 times the global average, reflecting the rapidity of the N-cycle in tropical areas.

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. Forests produce more N_2O than prairies. The fluxes from coniferous forests are possibly higher than those from coniferous forests. Corresponding figures for tropical rain forests are still uncertain.

4.2.1.9 Cultivated soils

Table 4.14 shows a number of emission rates from cultivated soils by a number of researchers. 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.

Cultivated fields show a wide flux range of about 0.02×10^{-4} to 35×10^{-4} g N m^{-2} h^{-1} considering the figures in table 4.14. However, a range of 0.02×10^{-4} to 2×10^{-4} g N m^{-2} h^{-1} seems to be more realistic, since only few estimates are higher than 2×10^{-4} and most of the estimates range between 0 and 1×10^{-4} gN m^{-2} h^{-1}.

Table 4.14. N_2O emission rates from cultivated soils

source	method*	N_2O emission ($10^{-4} gNm^{-2}h^{-1}$)	remarks
Ryden et al (1978)	o-	9.4	celery
Rolston et al (1978)	c2	35.4	rye gr;wet soil
		18.8	uncropped;wet soil
Ryden et al (1979)	o1	2.1-3.1	celery; 76 h exp.
Ryden and Lund (1980)	o1	2.3-4.8	vegetables
Breitenbeck et al (1980)	c-	0.25	max.flux, NO_3^-fert.
Burford et al (1981)	c-	0.26-1.47	dir.drilled
		0.09-0.96	ploughed
Seiler and Conrad(1981)	c-	0.42	max.flux, fertilized
		0.02-0.16	routinely fertilized
Duxbury et al (1982)	c-	0.26-0.48	alfalfa;mineral soil
		0.27-0.43	corn;manure,miner.soil
		0.25-0.33	corn;N fert,miner.soil
		0.10-0.19	weeds;unfert,miner.soil
		9.7 -8.2	onions,organic soil
		8.6-17.4	sw.corn,mineral soil
		1.8- 5.5	grass,mineral soil
		6.7-18.8	fallow,mineral soil
Webster and Dowdell	c-	0.7-0.9	400 kg N as NO_3;clay soil
(1982)		0.5-0.7	400 kg N as NO_3;silt loam
		0.09-0.1	unfert. grass sward
Mosier et al (1981)	c-	0.33	urea;grassland
Ryden (1983)	o1	0.4	250kg $N(NH_4NO_3)$,ryegr
		8.3	500 kg $N(NH_4NO_3)$;ryegr.
Colbourn et al (1984a)	c1	0.07	grassland
		0.19-0.23	winterwheat
Colbourn et al (1984b)	c2	2.5-4.2	max.flux;winterwheat
		1.67	winterwheat
Eggington and Smith (1986b)	o1	0.06-0.28	2 yrs,$Ca(NO_3)_2$ fert.
		0.04-0.61	cattle slurry 2 yrs.
		0.02-0.66	controls
Cates and Keeney(1987b)	c-	0.36-0.52	manured+min.fert.
		0.03	unfert.grazing land
Folorunso and Rolston (1987)	c1	0.06-0.07	unmanured/manured

* In all experiments the "chamber technique" was used for collecting the emitted gases;c= closed chamber; o= open chamber;1= with/without C_2H_2 inhibition method;2= ^{15}N labelling method;-= measurement of N_2O emission exclusively.

4.2.1.10 Global estimates

In table 4.15 an attempt has been made to give estimated N_2O fluxes for the different land cover types. On the basis of these estimates and the extent of the cover types as presented in table 2.1 the total global emission has been calculated. The sizeable contribution of the tropical rain and wet forests and cultivated areas may have been overestimated.
Bolle et al (1986) estimated the global soil N_2O emission at $7.8x10^{12}$ to $11.9x10^{12}$ gN y^{-1}. For this study a very wide spread in the global figures is maintained. If more detailed information becomes available experts may

attempt reduce the spread.
The estimates for cultivated soils include the N_2O loss from fertilizer N. In 4.2.1.11 a separate estimate of the fraction of fertilizer N lost as N_2O gas will be presented.

Table 4.15. emission of N_2O for major land cover types.

land cover type	Emission rate (10^{-4} gN m^{-2} h^{-1})	global emission (10^{12}g N)
tropical rainforest	1.0- 2.0	6.2-12.4
tropical seasonal forest	0.1- 0.4	0.6- 2.5
teperate deciduous	0.05-0.2	0.3- 1.2
temperate coniferous	0.1- 0.4	0.6- 2.4
boreal	0.1- 0.4	0.6- 2.4
woodland	0.05-0.2	0.3-1.3
savanna	0.05-0.15	}
tundra/grassland	0.05-0.15	}1.3-4.0
cultivated land	0.02-2.0	0.3-27.6
marsh/swamp	0.5*	0.9
		---------
total		11.1-54.7

* assuming flooded conditions during half of the year.

4.2.1.11 N_2O emission caused by N fertilizer application

Loss of fertilizer N as N_2O usually occurs within a few weeks after fertilization. Several authors found that N_2O release in aerobic conditions occurs mainly in the uppermost soil layer and that there is a strong temperature dependence (Denmead et al, 1979b; Conrad et al, 1983). This evolution is stimulated by soil moisture (Ryden and Lund, 1980; Bremner and Blackmer, 1981; Conrad et al, 1983) and the emission rate is higher for ammonia based mineral fertilizers than for nitrate (Conrad et al, 1983; Breitenbeck et al, 1980; Bremner and Blackmer, 1981). This latter effect appears to be independent of the type of counter ion. Conrad et al (1983) concluded, that the level of fertilizer application does not influence the percentage N-loss. 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.

Table 4.16 shows the percentage of applied fertilizer N lost as N_2O reported by a number of authors. Bolle et al (1986) assumed average N_2O loss rates of 0.04% for nitrate, 0.15-0.19% for ammonium and urea, and 5 % for anhydrous ammonia. These values seem to be independent of climate and may thus be globally representative. Based on the total production rates of mineral fertilizers, the global loss of mineral fertilizers in the form of N_2O is estimated at 0.5 to 2.0%.

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 2×10^{-4} to 11.9×10^{-4} 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).

Table 4.16. Loss of N as N_2O as % of fertilizer N added for different types of fertilizer as reported by various authors.

reference	method*	%loss as N_2O	fertilizer gift
urea			
Breitenbeck et al (1980)	c-	0.05-0.14	125-250 kg N ha^{-1}
Mosier et al (1981)	c-	0.6	450 kg N ha^{-1}
Smith et al (1982)	c-	0.01-0.05	90-180 kg N ha^{-1};wet rice
NH_3			
Hutchinson and Mosier (1979)	c-	1.3	200 kg N ha^{-1}
Breitenbeck et al (1980)	c-	0.08-0.18	125-250kg N ha^{-1}
Bremner and Blackmer (1981)	1	0.11-0.18	
Mosier and Hutchinson (1981)	c-	1.3	200 kg N ha^{-1}
Seiler and Conrad (1981)	c-	0.03-0.09	100kg N ha^{-1} as NH_4Cl
Conrad et al (1983)	c-	0.025-0.376	100kg N ha^{-1} as NH_4Cl
Mosier et al (1986)	c2	0.4	200 kg N ha^{-1}; barley
		1.5	200 kg N ha^{-1}; corn
Minami(1987)	c-	0.06-0.34	80-200 kg N ha^{-1}; upland
		0.33-0.55	90-100 kg N ha^{-1};wet rice
NH_4NO_3			
Ryden (1981)	o1	1.3	250 kg N ha^{-1}
Ryden (1983)	o1	1.6	500 kg N ha^{-1}
		1.4	250 kg N ha^{-1}
NO_3			
Ryden and Lund (1980)	o1	3.9-6.7	680-430 kg N ha^{-1}
Breitenbeck et al (1980)	c-	0.01-0.04	125-250 kg N ha^{-1}
Seiler and Conrad (1981)	c-	0.01-0.05	100kg N ha^{-1} as $NaNO_3$
Conrad et al (1983)	c-	0.001-0.073	100kg N ha^{-1} as $NaNO_3$
Armstrong (1983)	c-	0.11-0.15	200 kg N ha^{-1}; grassland
Colbourn et al (1984b)	c2	2.1	100 kg N ha^{-1} as $Ca(NO_3)_2$
Eggington and Smith (1986a)	o1	3-11	100 kg N ha^{-1}
Eggington and Smith (1986b)	o1	1.9	700 kg N ha^{-1}
organic manures			
Eggington and Smith (1986a)	o1	0.0-0.4	>1200 kg N ha^{-1}; slurry
Eggington and Smith (1986b)	o1	0.5	700 kg N ha^{-1}; slurry
Cates and Keeney(1987b)	c-	2	168 kg N ha^{-1} + 13-69 kg min.fert.

* In all experiments (except Bremner and Blackmer, 1981) the "chamber technique" was used for collecting the emitted gases; c= closed; o= open; l= laboratory;1= with/without the C_2H_2 inhibition method;2= with the ^{15}Nlabelling method;-= measurement of N_2O exclusively.

With a total nitrogen fertilizer consumption of about $74x10^{12}$ g in 1985 (FAO, 1985), the total N_2O emission due to application of nitrogen fertilizers amounted to $0.7x10^{12}$ to $3.0x10^{12}$ g N y^{-1}. This estimate shows less variability than the figure for cultivated fields as presented in table

4.15. However, conditions under which N_2O is lost form fertilizers are as variable as the conditions in cultivated fields. Nevertheless, the fertilizer N_2O loss fits well within the rage given for cultivated fields.

4.2.1.12 N_2O emission from other 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. N_2O emission due to biomass burning is estimated at $1x10^{12}$ to $2x10^{12}$ g N y^{-1} (Crutzen ,1983).

4.2.1.13 Global sources and sinks of nitrous oxide

Most studies presented in section 4.2 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.

Nevertheless, 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 soils ($11.1x10^{12}$ to $54.7x10^{12}$ g N y^{-1}) and emissions from other sources ($1x10^{12}$ to $2x10^{12}$ g N y^{-1}) yield a likely total global emission of 12.6 $x10^{12}$ to $57.2x10^{12}$ g N y^{-1}. In this estimate it is assumed that the emission induced by N-fertilization ($0.7x10^{12}$ to $3.0x10^{12}$ g N y^{-1}) is included in the flux rate given for cultivated fields (see table 4.15). The estimate for the global emission given by Bolle et al (1986) is $7.8x10^{12}$ to $11.9x10^{12}$ g N y^{-1}. Lipschultz et al (1981) estimated the global N_2O source associated with nitrification only at $5\text{-}10x10^{12}$ g N y^{-1} and the source of NO at $15x10^{12}$ g N y^{-1}. The broad range expresses the large uncertainties in the estimates of individual sources, particularly the nitrification / denitrification losses.

The total global emission from all possible sources including oceans and freshwater, fossil fuel burning and lightning estimated by Bolle et al (1986) is $12x10^{12}$ to $15x10^{12}$ g N y^{-1}.

4.2.3 Nitric oxide and nitrous oxide (NO_x)

4.2.3.1 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.

4.2.3.2 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 sources are listed in table 4.17.

Table 4.17 Global tropospheric sources of NO_x (10^{12} g N y^{-1})

	mean	range
surface sources		
1. fossil guel combustion	21	14-28
2. biomass burning	12	4-24
3. nitrification/denitrification in soils	8	4-16[1]
atmospheric sources		
1. lightning	8	2-20[2]
2. NH_3 oxidation		1-10[3]
3. from stratosphere	0.5	
4. high flying aircraft	___	
total	50	25-99

All data are from Logan (1983);[1] Levine et al (1984) estimated biogenic production at $10x10^{12}$ g N y^{-1}; [2] Levine et al (1984) gave a source of $1.8x10^{12}$ to $18x10^{12}$ g N y^{-1}. [3] Crutzen (1983) estimated that 10% of all atmospheric NH_3 is oxidated, i.e. $12\text{-}15x10^{12}$ g N y^{-1} (see section 4.2.5).

The extent of biogenic NO_x production is highly uncertain. Lipschultz et al (1981) estimated the biogenic NO source at $15x10^{12}$ 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 $x10^{12}$ g N y^{-1}, the estimated NO emission is $10x10^{12}$ 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 ration of NO:N_2O is a highly uncertain factor in all such calculations.

4.2.4 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.

4.2.4.1 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 $74x10^{12}$ 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 $3000x10^{12}$ g, the anual emission is 4 - $12x10^{12}$ 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 $60x10^{12}$ to $70x10^{12}$ 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 $3.7x10^{12}$ 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 $20x10^{12}$ to $35x10^{12}$ 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 $117x10^{12}$ to $150x10^{12}$.

4.2.4.2 Sinks of NH_3

Crutzen (1983) estimated, that about 10% of all atmospheric NH_3 ($12x10^{12}$ to $15x10^{12}$ g N) reacts with OH. to from NO or NO_2. The annual dry deposition is $72x10^{12}$ to $151x10^{12}$ and wet deposition amounts to $30x10^{12}$ to $60x10^{12}$ g N.

4.2.5 Conclusions

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.

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 <u>Change of runoff due to deforestation</u>

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) conclude, 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.
Increased runoff enhances soil detachment, splash erosion and sediment transport.

A review of the literature on increased runoff and increased erosion hazard due to forest clearing is given by Lal (1987).

5.3 <u>Change in total evapotranspiration due to deforestation</u>

5.3.1 <u>General</u>

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):

$$LE = \frac{s(R_n - G) + \check{S} * C_p * (e(T_z) - e_z)/r_A}{s + \ddot{Y} * (1 + r_C/r_A)} \qquad (W\ m^{-2}) \qquad (5.2)$$

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 sat. vapour pressure curve (mbar K^{-1})
$\check{S}$ =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})
$\ddot{Y}$ =psychometric constant (mbar K^{-1})
z =reference height (m)
T_z=temperature (^oK) at reference height z.

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.

<u>The Priestley-Taylor equation</u>
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:

$$LE = \tilde{a} \frac{s}{s + \ddot{Y}} (R_n - G) \qquad (5.3)$$

where: ã = constant

In conditions of non-limiting water supply the actual evapotranspiration equals the potential rate calculated according to equation 5.3 with ã 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 $s/s+\ddot{Y}$. 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:

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

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:

$$LE= \text{ã} \frac{s}{s + \text{Ÿ}} (R_n-G) \qquad (5.5)$$

Inputs into equation 5.5 are estimates of the surface fluxes E and H. The parameter ã is dependent on many variables such as the vapour deficit, H and LE. De Bruin (1983) showed that ã =1.3 if r_s=0; ã=1 if r_s=60-90sm^{-1}; ã<1 if r_s>100sm^{-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_s causes the evapotranspiration to decrease by only 10-20%.

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):

$$LE= \frac{s(R_n-G) + \text{Š}*C_p*(e(T_z) - e_z)/r_A}{s + \text{Ÿ}} \qquad (W\ m^{-2}) \qquad (5.6)$$

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, $LE/(R_n-G-S)^2$, 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)[1]	0.2(0.9)	0.6(1.4)	0.1(0.6)
$LE/(R_n-G)$	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[2] (s m^{-1})	5-10	5-10	20-200	20-200

[1] maxima in parenthesis; [2] 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 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 deforest.	complete change	albedo 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 (1987) 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 tropical deforestation of the South-American Amazon. The major changes imposed by the deforestation were:

- decreasing vegetation cover and decreasing roughness length of the land cover;
- deceasing soil depth (through erosion) and increase in run-off;
- increasae of the relative density of roots in the soil's surface layer;
- increase of the albedo of the land cover;
- 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, but they duplicate well with other climatic models. 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 responsible 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 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. 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 are 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 (Sellers, 1987).

The albedo is one of the factors in the energy balance which determine the partitioning of the sun's net radiation. Albedo in this respect is the ratio of reflected to total incident radiation. The longwave terms in the radiation balance are the key factors in the greenhouse effect since infrared 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:

$$R_n + H + LE + G_E = 0 \qquad (6.1)$$

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, radiation (the sum of incoming and outgoing radiation) 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 6.2).

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

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

where: $\bar{a}$ = 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 um. The earth, a body with approximate temperature of about 300 K, emits radiation with wavelengths of 4 to 40 m.

Global radiation 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, wherby some of the required parameters such as cloudiness, albedo and planetary reflectance are obtained from satellites (see 6.4.4).

Long wave radiation emitted or reflected by the surface can in principle be obtained by measuring the emittance in the 8 to 14 um spectral range using radiometers installed in airplane or satellite. The radiation terms in equation (6.2) will be discussed in more detail in the following sections.

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.

6.3 Radiation

6.3.1 Shortwave radiation terms

- Global radiation R_{sw}

Global radiation 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 2 Sr.

- 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. Usually only the fraction of reflected radiation in the visible and near infrared part of the spectrum are considered in the determination of albedo (Schanda, 1986).

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 absorbed energy equals the amount emitted (Schanda, 1986). The following formulas describe the above relations:

$$r = 1 - e \tag{6.3}$$

$$e = a \text{ (Kirchhoff's law)} \tag{6.4}$$

where: r = reflectivity, e = emissivity, a = absorptivity (r,a and e are dimensionless)

It is obvious, that the emissivity of any rough or plane surface is related to the reflectivity viewed from the same direction (i.e. angular and polarization dependency). The above relations hold for any hard rough or plane surface, but also for transition layers many wavelengths thick, where volume scattering dominates the reflection properties. The term roughness is understood as to be measured in wavelengths. A surface may appear rough optically and at the same time be smooth at microwave scale.

Reflectivity depends on the wavelength of the incoming radiation and in general increases with wavelengths up to 1.2 um (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 m is emitted by the earth. The quantity of emitted energy depends on the surface's 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:

- Sky radiation R_{sky}

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):

$$R_{sky} = \hat{E}'_{sky} * k * T_a^4 \tag{6.5}$$

where: $\hat{E}'_{sky}$ = 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).

- Surface emittance R_s

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

$$R_s = \hat{E} * k * T_s^4 \tag{6.7}$$

where: $\hat{E}$ = 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 0.9 (dry quartz sand) to approximately 1.0. 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 E 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:

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

where: $\hat{E}$ = emissivity

The emissivity equals the absorptivity (Kirchhoff's law, equations 6.3 and 6.4). 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 7a and 7b can be combined into:

$$R_s = \hat{E}*k*T_s^4 + (1-\hat{E})*R_{sky} \tag{6.9}$$

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

$$R_s = f_c\hat{E}_cT_c^4 + (1-f_c)\hat{E}_sT_s^4 + f_c(1-\hat{E}_c)R_{sky} + (1-f_c)(1-\hat{E}_s)R_{sky} \qquad (6.10)$$

where: f_c = vegetation coverage expressed as a fraction
$\hat{E}_s$ = soil emissivity
$\hat{E}_c$ = 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.

The degree of reflectivity and refraction/absorption of incident radiation by condensed matter (soil, vegetation) is determined by the polarizability of the medium. The polarization by the presence of an electric field (electromagnetic waves) is the basis of the dielectric properties of condensed matter. The dielectric properties determine the permittivity of the medium (or in other words the impedance). The permittivity (E) and the index of refraction (n) describe macroscopically the behaviour of a propagating wave. In formula:

$$E = E_r * E_0 \qquad (6.11)$$
$$n = E_r^{1/2} \qquad (6.12)$$

where: E = electric permittivity
E_o= permittivity in vacuum
E_r =dielectric constant

Permittivities of materials are interrelated with the frequencies of the incident radiation. This phenomenon causes the wavelength dependency of the reflection by a medium. Ferro-electric material, for example, is characterized by a very high permittivity (E_r -->oo). Knowledge of the dielectric properties is therefore, a prerequisite for the interpretation of remotely sensed data.

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 um and 83% in the range 0.3-1.1 um. Hence an instrument sensing fluxes in the interval 0.3-2 um 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 (ã) will decrease in the morning and increase in the late afternoon. Furthermore ã is dependent on wavelength of incident radiation. ã in the 0.4 to 1.1 um spectral range (as in LANDSAT) is representative of a much wider spectral range of 0.2 to 3.5 um. Knowledge of ã-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 um) as carried in satellites allow a good determination of ã. When dealing with multispectral scanners, one should account for the percentage of solar energy in each band to calculate ã from the various single band reflectances.

The dielectric properties of soils are strongly related to soil moisture content. (Internal) Reflection takes place at the surface of waterfilms surrounding soil particles, i.e. radiation reflected on the surface of soil particles is re-reflected at the surface of the water film. This process breaks down at some soil moisture content, where the water film gets thinner and the interaction radiation-soil particles becomes dominant. Thus, 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 um (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), etc. etc.

6.4.3 Spectral reflectance characterisitics of vegetation

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

- 0.4-0.7 um: intense absorption of incident radiation by pigments in the plant leaves with major absorption bands of about 0.43-0.45 um and 0.66-0.64 um (Schanda, 1986).
- 0.7-1.3 um: low absorption, high reflection.
- 1.3-2.6 um: 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 um).

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.

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 independantly. Recent developments show the ability to distinguish between vegetation species and to identify water stress (Schanda, 1986).

Table 6.1. Description of surface classes and corresponding mean values and ranges of the surface albedo.

surface class	characteristics	surface albedo mean	range
0 swampland, oceans	coastal swamps, rivers, smooth oceans >50% water	0.09	0.0 -0.10
1 dense forest	uniform dark vegetation, 10% light vegetation or soils	0.15	0.10-0.16
2 moderate forest	mostly dark vegetation, 30% light vegetation or soils	0.17	0.16-0.21
3 mixed vegetation	evenly mixed surface; 50% dark vegetation, 50% light vegetation/soils	0.22	0.21-0.26
4 savanna	mostly low light grasses, cult. land some dark scrub; 70% light dry grass 30% dark scrub or rock	0.28	0.26-0.31
5 mixed desert	dry light surface, coloured soils + rock outcrops; 50% light soils, 50% scrub, light grasses	0.33	0.31-0.36
6 moderate desert	sparse vegetation, mostly light sands, scrub; 30% low scrub, light grasses, rock	0.39	0.36-0.42
7 desert	uniform sand surface, little variation in surface over large areas; 10% low scrub or coloured rock	0.43	0.42-1.0

Source: Rockwood and Cox (1978)

6.4.4 Surface albedo values

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).

To illustrate the possible variations in surface albedo influenced by various factors such as zenith angle and vegetation type, the determination of the surface albedo according to Rockwood and Cox (1978) will be discussed briefly. These authors calculated the surface albedo for the Sahel Region in NW Africa from SMS-1 (Synchronous Meteorological Satellite, launched in 1974) taking the planetary albedo at the top of the atmosphere and accounting for the atmospheric absorptivity and transmittance. The same method was applied on 3 different dates (July 1974, August 1974 and September 1974). It was observed, that there was a change in the surface albedo north-south gradient from July-September. At about 15°N there was a change in albedo from 25 to 19%; at 14 and 17°N relative changes of 15 and 5% were observed; North of 18°N changes from 0 to 15% were inferred. Surface class 7 (see table 6.1) with albedo values greater than 42% was nearly unaffected by the seasonal variation of moisture. Surface class 3, dominant at 15°N in July, had a range of 21-26% and appeared to change to class 2 (moderate forest). The Sahara zone North of 10°N as a whole was found to be comparatively stable with respect to changes in albedo.

Table 6.2. 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)	ā (%)	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)

Baumgartner (1965, 1970) demonstrated the interrelations between net radiation, albedo, sensible heat and latent heat. Although his figures as presented in table 6.2 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 6.2.

A very useful albedo model was developed by Hummel and Reck (1979), which was already mentioned in chapter 3. Their data as presented in table 6.3 illustrate the seasonal variation of albedo. Another seasonal global albedo dataset is the one produced by Matthews (1985) with a 1x1° grid. The latter data set is based on the theoretical natural vegetation type and does not include snow and ice cover.

Pinker et al (1980) found a mean albedo of 10% (0.13) 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% 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 6.3. land cover types, extent and seasonal (Northern Hemisphere)

albedo values.

albedo (%) 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

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 6.2). 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 postive 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, since 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) used a deforestation rate of 0.6% yr^{-1}, whereby the extent of desertification could be of greater importance. They arrived at maximum planetary albedo increase of 0.00033 to 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.

6.5 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

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.

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).

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.

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.

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 um 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 developed 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.

The amount of reflected power gathered by the sensor is proportional to the square of the field of view, the sensor aperture, the irradiance, irradiance angles, sensor view angle and the scattering properties of the target. The angular relations of the source, target and sensor can greatly influence the measured reflectance. For example, the reflectance measured by a sensor vertically viewing soil in a plowed field with deep furrows will be higher when the solar azimuth angle is parallel to the furrows (no shadows) than when the solar azimuth is perpendicular to the furrow direction (shadows present). For smooth targets the azimuthal orientation is irrelevant. The power received by the sensor is proportional to the square of the field of view of the sensor. However, more factors determine the choice of the field of view. A wide enough field of view is necessary to represent the geometric features of the target adequately.

In field situations an appropriate field of view is 15^{o}, and in laboratory situations a narrow field of view of 1^{o} may be properly used if the surface roughness can be controlled (Baumgardner, 1984).

In relation to monitoring of vegetation at a global scale various remote sensing techniques are applicable depending on the purpose and desired detail. These remote sensing techniques are discussed extensively in Woodwell (1984). 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).

In all remote sensing techniques the objects in the field are projected on the two-dimensional plain of the image. Aerial photography offers a certain advantage in this respect through their relief displacement enabling stereoscopic view of pairs of photos (stereopairs). However, also SPOT satellite imagery has stereoscopic coverage.

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 um) 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 that 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 air- 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 can be 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	a 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	a 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 um (green); 0.6-0.7 um (red); 0.7-0.9 um and 0.8-1.1 um (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 7.3.

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 delineation of major cover types and also disturbed forest lands can be reliably defined, although 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) series of polar orbiting sun synchronous meteorological satellites (Townshend and Justice, 1986). The spectal range of NOAA-9 (launched in 1984) is 0.58-0.69 um (band 1), 0.725-1.1 um (band 2), 3.5-3.93 um (band 3), 10.3-11.3 um (band 4) and 11.5-12.5 um (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. 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. For a given scan line 4 pixels (picture elements) are averaged, then one pixel is ignored, the next 4 pixels averaged, and so on. Once completed a scan line, the next 2 scan lines are ignored and the third is sampled as the first. Thus each 5 by 3 pixel subarea is represented by 4 pixels in a row. 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 a resampling procedure to about 20 km resolution is applied. 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 scale 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:

- when the leaf area index exceeds 2 or 3;
- when there are patches of 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 sen-

sors. These dimensions, called brightness and greenness, are related to target 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:

$$B = A * DHC \qquad (5.1)$$

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 (5.1) 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). Shortwave reflectance values measured by METEOSAT, which usually range between 0.16 and 0.45 (Rockwood and Cox, 1978), indicate the presence of clouds or vegetation. 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:

$$LE = p\ C_p \frac{T_a - T_c}{r_{ah}} + (1-a_0)R_{sw} + R_{sky} - EkT_c^4 - G_E \qquad (7.2)$$

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 express 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.

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

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:

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

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). 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. Soil heat flux estimated as a fraction of net radiation was proposed by Clothier et al (1986).

The basic equation used by Asrar et al (1985) is:

$$M = \sum_{i=1}^{n} e_c{}^*e_i{}^*e_s{}^*S{}^*C \tag{5.5}$$

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 evapotranspira tion)

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 4 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).

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.

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.

7.3.3 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 m). 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.

7.4 Conclusion

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, 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.

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APPENDIX I

Conversion to agricultural land for the various ecosystems in 10^{10} m^2. Data are averaged over the period 1965-1982.

Region	Ecosystem*									
	1	2	3	4	5	6	7	8	9	10
N. America										
Europe			-.2	-.2	-.2			-.2		
USSR										
Pacific Dev.	.03					.05		.12		
China	.07								.59	
Latin America	.96		.39	.19				.39		
N.Africa-Middle East	.15							.73		.59
Trop.Africa	1.60	.16	.95					.48		
S. Asia	.10	1.0				1.90	.16			
S.E. Asia	.25	.04							.35	
World	3.15	1.20	1.14	-.01	-.2	1.95	0.00	1.88	.94	.59

adapted from: Houghton et al(1983);
* acc. to Whittaker and Likens (1975) see table 2.1, chapter 2.

APPENDIX II

The chemical composition of the atmosphere measured in 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
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)	0.01-0.04	40ppb	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 III.1

Area harvested of paddy rice in $10^7 m^2$

Source: FAO (1979,1986) Production Yearbooks 1978, 1985.

Countries of the tropical regions

	70	85
Africa		
Angola	21	20
Benin	3	8
Botswana	?	?
Burkina Faso	40	22
Burundi	3	6
Cameroon	16	27
Central Afr.Rep.	14	16
Chad	44	16
Congo	3	4
Equatorial Guinea	?	?
Ethiopia	?	?
Gabon	--	--
Gambia	28	20
Ghana	55	87
Guinea	411	561
Guinea-Bissau	30	140
Ivory Coast	286	470
Kenya	6	9
Liberia	154	210
Madagascar	931	1201
Malawi	23	45
Mali	158	160
Mozambique	76	70
Niger	16	20
Nigeria	272	700
Rwanda	1	1
Senegal	91	78
Sierra Leone	331	400
Somalia	--	1
Sudan	4	3
Tanzania	144	350
Togo	25	15
Uganda	16	25
Zaire	245	325
Zambia	1	9
Zimbabwe	3	1
North America		
Barbados	?	?
Costa Rica	44	69
Cuba	164	155
Dominican rep.	80	110
El Salvador	12	17
Guatemala	11	22
Haiti	38	60
Honduras	9	39
Jamaica	--	2
Mexico	152	276
Nicaragua	26	41
Panama	105	105
Trinidad and Tobago	1	1

	70	85
South America		
Bolivia	59	120
Brasil	4788	4752
Colombia	260	328
Ecuador	78	120
Guyana	109	92
Paraguay	19	92
Peru	130	218
Surinam	37	70
Venezuela	120	148
Asia		
Bangladesh	9842	10430
Bhutan	178	31
Burma	4748	4800
India	37677	42000
Indonesia	8158	9540
Kampuchea	2074	1750
Lao People's D.Rep.	665	600
Malaysia	697	675
Nepal	1185	1400
Pakistan	1527	1400
Philippines	3157	2000
Sri Lanka	579	859
Thailand	6919	9585
Viet Nam	4916	5700
Oceania		
Papua New Guinea	1	--

APPENDIX III.2
Countries of the temperate regions

	70	85
Africa		
Algeria	1	--
Cape Verde	?	?
Comoros	9	14
Djibouti	?	?
Egypt	487	422
Lesotho	?	?
Libya	?	?
Mauritania	1	6
Mauritius	--	--
Morocco	6	2
South Africa	1	1
Swaziland	3	--
Tunisia	?	?
North America		
Canada	--	--
United States	777	1013
South America		
Argentina	89	117
Chile	23	39
Uruguay	34	85
Asia		
Afghanistan	203	215
Bahrain	?	?
China	34622	32075
Cyprus	?	?
Iran	362	390
Iraq	97	40
Israel	?	?
Japan	2966	2342
Jordan	?	?
Korea D.People'sRep.	445	840
Korea, rep.	1204	1237
Kuwait	?	?
Lebanon	?	?
Mongolia	?	?
Oman	?	?
Qatar	?	?
Saudi Arabia	--	--
Singapore	?	?
Syrian Arab rep.	--	--
Turkey	63	64
Un. Arab Emirates	?	?
Yemen	?	?
Yemen, Dem.	?	?

	70	85
USSR	356	667
Europe		
Albania	4	4
Austria	--	--
Belgium	--	--
Bulgaria	17	16
Czechoslovakia	--	--
Denmark	--	--
Finland	--	--
France	22	11
German Dem. Rep.	--	--
Germany, Fed. Rep.	--	--
Greece	17	16
Hungary	24	11
Iceland	--	--
Ireland	--	--
Italy	172	186
Luxemburg	--	--
Malta	--	--
The Netherlands	--	--
Norway	--	--
Poland	--	--
Portugal	40	30
Romania	28	31
Spain	63	74
Sweden	--	--
Switzerland	--	--
United Kingdom	--	--
Yugoslavia	7	9
Oceania		
Australia	38	126
Fiji	10	11
New Zealand	?	?
Solomon Islands	1	3

APPENDIX III.3
Irrigated rice area, deepwater rice area and total area for rice (10^7 m^2).
Source: Huke (1982); data are for 1980.

	irrigated area			rainfed[1] wetland	deep-water	total(wet, rainfed, deepwater+ dryland) harvested[2]
	wet season	dry season	total irrigated			
Southeast Asia						
Burma	780	115	895	3456	173	5317
Indonesia	3274	1920	5194	1618	258	8204
Kampuchea	214	---	214	883	435	1020[3]
Laos	67	9	76	277	---	695
Malaysia	266	220	486	92	---	735
Philippines	892	622	1514	1586	---	3515
Sabah	8	4	12	9	---	42
Sarawak	6	4	10	57	---	127
Thailand	866	320	1186	6130	400	8677
Vietnam	1326	894	2220	2526	420	5573
Total SE.Asia	7685	4100	11785	16568	1686	33905
South Asia						
Bangladesh	170	987	1157	6880	1117	10006
Bhutan	---	---	---	161	---	189
India	11134	2344	13478	17147	2434	38941
Nepal	261	---	261	908	53	1262
Pakistan	1710	---	1710	---	---	1710
Sri Lanka	294	182	476	232	---	760
Total S.Asia	13569	3513	17082	25328	3604	52868
South Korea	1120	---	1120	99	---	1282
North Korea	500	---	500	150	---	600
China	23986	9690	33676	7746	---	36162
Total Asia	46860	17303	64163	49891	5290	124817

[1] rainfed shallow+intermediate wetland; this includes rainfed areas with up to 30 cm (shallow) and 100 cm (intermediate) of water in bounded fields; fields may temporarily dry up.
[2] double cropped area counted twice.
[3] average for the period 1964-1972.

APPENDIX III.4

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

Table 1. The distribution of the soils and ecosystems (a) of North America (FAO/UNESCO soil map of the world VOL.II)
Ecosystem (1000000 ha) (c)

Soil type (b)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols			21.1	19.9	5.0	2.0			25.0	9.0		28.1			110.1
Camoisols			46.2	40.4	5.0	9.1			2.6			7.3		15.0	125.6
Chernozems					5.0	7.5			10.1			12.0			34.6
Podzoluvisols			3.6	1.2								1.7			6.5
Rendzinas									1.3						1.3
Ferralsols															0.0
Gleysols			28.4	24.0	5.0	5.0			15.6			30.0	5.0	53.5	166.5
Phaeozems			7.5	7.5		2.0			25.4			30.7			73.1
Lithosols			33.0	80.0										100.9	213.9
Fluvisols			9.5									3.6			13.1
Kastanozems					5.0				63.2			72.6		12.0	152.8
Luvisols			66.4	20.9	1.0	2.0			38.4			39.3			168.0
Greyzems									0.1			5.2			5.3
Nitosols															0.0
Histosols			38.5	7.9	19.0								31.4	5.0	101.8
Podzols			120.0	23.3	5.0							15.4	2.0		165.7
Arenosols															0.0
Regosols			21.5	9.4	2.0				27.9	186.9		3.3			251.0
Solonetz									3.2	2.0		5.5			10.7
Andosols			14.5						3.1						17.6
Rankers															0.0
Vertisols									4.7	1.0		3.5			9.2
Planosols						2.0						4.5			6.5
Xerosols										23.4		1.3			24.7
Yermosols							10.0			7.0	70.4	1.0			88.4
Solonchaks												0.1			0.1
Miscellaneous												-			
Total	0.0	0.0	410.2	234.5	52.0	29.6	10.0	0.0	220.6	229.3	70.4	265.1	38.4	186.4	1746.5

Table 2. The distribution of the soils and ecosystems (a) of Mexico and Central America (FAO/UNESCO soil map of the world VOL.III)
Ecosystem (1000000 ha) (c)

Soil type (b)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols			6.5	1.0		3.0			2.9			2.5			15.9
Cambisols			7.1	1.0		2.0	3.0		2.0			9.0			24.1
Chernozems															0.0
Podzoluvisols															0.0
Rendzinas				8.5								1.0		2.0	11.5
Ferralsols		0.8													0.8
Glevsols		3.0						4.0				1.6			8.6
Phaeozems								0.5				0.1			0.6
Lithosols		4.0	4.0				5.6			10.0	20.0				43.6
Fluvisols						3.0						2.4			5.4
Kastanozems			9.8	3.7		1.0			6.4			3.5			24.4
Luvisols		4.0	4.0	2.0		1.0		10.9	8.0			3.0			32.9
Greyzems															0.0
Nitosols		3.0				1.0		3.0				4.6			11.6
Histosols													2.2		2.2
Podzols															0.0
Arenosols											0.2	0.2			0.4
Regosols			4.0	3.0		2.0			5.0			1.7			15.7
Solonetz															0.0
Andosols									4.0			7.0		1.2	12.2
Rankers															0.0
Vertisols						0.1	4.0	2.0		5.0		0.1			11.2
Planosols								0.7				0.2			0.9
Xerosols										8.2	3.0				11.2
Yermosols							4.0			5.0	15.0				24.0
Solonchaks											0.4				0.4
Miscellaneous															8.7
Total	0.0	14.8	35.4	19.2	0.0	13.1	16.6	21.1	28.3	28.2	38.6	36.9	2.2	3.2	266.3

[a] acc. to FAO/UNESCO soil map of the world (FAO, 1971–1978) VOL II–X.

[b] The presented acreages are estimated values of the extents of the soils, whether occurring as dominant soil, associated soil or inclusions, according to the FAO/UNESCO soil map of the world (VOL.II–X).

[c] ecosystems according to Whittaker and Likens (1975).
(1=tropical rainforest 6=woodland/shrubl. 11=extreme desert
2=tropical seasonal forest 7=savanna 12=cropland
3=temperate evergreen forest 8=trop.grassland 13=swamp/marsh
4=temperate deciduous forest 9=temp.grassland 14=tundra/alpine
5=boreal (taiga) 10=desert/semi-desert scrub)

APPENDIX IV.2

Table 3. The distribution of the soils and ecosystems (a) of South America (FAO/UNESCO soil map of the world VOL.IV)

Ecosystem (1000000 ha) (c)

Soil type (b)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols	112.8	90.9				38.6	30.2					13.5	6.8		292.8
Cambisols		10.6	5.2	11.5			7.2					11.0		3.0	48.5
Chernozems															0.0
Podzoluvisols															0.0
Rendzinas															0.0
Ferralsols	210.0	121.2					58.3	9.6				34.6			433.7
Gleysols	16.1						9.6					24.0	51.9		101.6
Phaeozems		9.5				3.8	34.2		8.8						56.3
Lithosols	8.0	26.6	8.0	23.0		26.1	2.0			4.0	13.7	9.0		65.3	185.7
Fluvisols		23.8				3.2	17.5		1.6			10.0	8.5		64.6
Kastanozems				22.2		12.6	9.5		0.3			6.8			51.4
Luvisols		28.5		6.6		3.5				30.2		12.0			80.8
Greyzems															0.0
Nitosols		17.3						10.0	1.9			7.5			36.7
Histosols													3.6		3.6
Podzols				1.5											1.5
Arenosols		21.0					28.0			41.5					90.5
Regosols						12.0	2.1		10.0	5.9	5.1	6.7			41.8
Solonetz						3.7			5.6						9.3
Andosols			0.5	7.7		1.6						3.5		6.5	19.8
Rankers				1.5					0.4						1.9
Vertisols						1.1	11.5		9.3	0.2		2.9			25.0
Planosols		17.2					21.5		5.8			3.0	1.0		48.5
Xerosols										36.0		1.7			37.7
Yermosols											58.6				58.6
Solonchaks									9.9		11.8				21.7
Miscellaneous															
Total	346.9	366.6	13.7	74.0	0.0	106.2	231.6	19.6	53.6	117.8	89.2	146.2	71.8	74.8	1712.0

Table 4. The distribution of the soils and ecosystems (a) of Europe (FAO/UNESCO soil map of the world VOL.V)

Ecosystem (1000000 ha) (c)

Soil type (b)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols				2.5								0.6			3.
Cambisols			40.0	38.8					28.1			44.7		15.0	166.
Chernozems			7.0	10.0					17.0			58.2			92.
Podzoluvisols			18.5	18.5	18.5				21.0			70.2			146.
Rendzinas			3.2	3.2					4.9			4.4		2.5	18.
Ferralsols															0.
Gleysols			2.2	2.1					10.7			15.0		14.0	44.
Phaeozems									2.0			7.4			9.
Lithosols			7.1		3.0	11.0			15.3					15.0	51.
Fluvisols			8.4	8.4					1.8			12.2			30.
Kastanozems			7.1	7.1					19.7			22.5			56.
Luvisols			14.5	14.5					40.9			41.9			111.
Greyzems			2.3	2.3					4.0			4.8			13.
Nitosols															0.
Histosols			5.0	5.0					9.0			0.9	22.5	8.2	50.
Podzols			43.8	30.0	50.0				30.0			26.7		20.0	200.
Arenosols			1.5									1.1			2.
Regosols			0.5		0.5				8.0			3.3		9.8	22.
Solonetz									8.9			1.0			9.
Andosols			0.5						2.4			1.2			4.
Rankers			4.2	4.2					1.3						9.
Vertisols									1.2			4.6			5.
Planosols			0.5									1.3			1.
Xerosols										15.6		14.5			30.
Yermosols															0.
Solonchaks									1.9			0.5			2.
Miscellaneous															8.
Total	0.0	0.0	166.3	146.6	72.0	11.0	0.0	0.0	228.1	15.6	0.0	337.0	22.5	84.5	1092.

[a] acc. to FAO/UNESCO soil map of the world (FAO, 1971-1978) VOL II-X.

[b] The presented acreages are estimated values of the extents of the soils, whether occurring as dominant soil, associated soil or inclusions, according to the FAO/UNESCO soil map of the world (VOL.II-X).

[c] ecosystems according to Whittaker and Likens (1975).
(1=tropical rainforest 2=tropical seasonal forest 3=temperate evergreen forest 4=temperate deciduous forest 5=boreal (taiga) 6=woodland/shrubl. 7=savanna 8=trop.grassland 9=temp.grassland 10=desert/semi-desert scrub 11=extreme desert 12=cropland 13=swamp/marsh 14=tundra/alpine)

APPENDIX IV.3

Table 5. The distribution of the soils and ecosystems (a) of Africa (FAO/UNESCO soil map of the world VOL.VI) (1000000 ha) (c)

Soil type (b)	Ecosystem 1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols	28.7	37.5				0.8	21.6					16.4			105.0
Cambisols		9.6	10.0	5.0		17.0	34.0	8.0		3.6		12.5			99.7
Chernozems															0.0
Podzoluvisols															0.0
Rendzinas			8.8			0.2	0.3								9.3
Ferralsols	169.0	24.8				50.0	20.9	0.4				35.0	2.1		302.2
Gleysols	29.8	0.4				0.1	25.1	2.5				14.5	8.0		80.4
Phaeozems						0.4	0.3								0.7
Lithosols		19.7	3.5	3.3		25.2	78.9	4.6		102.5	135.2				372.9
Fluvisols	9.5	1.5				2.6	38.7					39.3	12.7		104.3
Kastanozems			0.2			2.4									2.6
Luvisols		12.6	5.0	4.9		60.9	74.9	6.9		6.5		59.5			231.2
Greyzems															0.0
Nitosols		27.8	18.9			4.9	22.0	0.8				24.0			98.4
Histosols	0.3						1.3						1.0		2.6
Podzols						1.1	1.2								2.3
Arenosols		15.9				45.5	135.3	14.6		63.5	30.0				304.8
Regosols		2.0	3.1			5.0	123.3			77.9	43.9	8.0			263.2
Solonetz							7.5	9.3							16.8
Andosols	0.2	1.5				0.4	2.1					0.5			4.7
Rankers															0.0
Vertisols						4.1	92.1			0.5		2.1			98.8
Planosols			0.5			3.0	4.8	9.3							17.6
Xerosols							23.0			54.8	7.6				85.4
Yermosols							16.5			65.2	291.9				373.6
Solonchaks							7.9			6.5	5.2				19.6
Miscellaneous															207.7
Total	237.5	153.3	50.0	13.2	0.0	223.6	731.7	56.4	0.0	381.0	513.8	211.8	23.8	0.0	2803.8

Table 6. The distribution of the soils and ecosystems (a) of South Asia (FAO UNESCO soil map of the world VOL.VII) (1000000 ha) (c)

Soil type (b)	Ecosystem 1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols	5.0	4.0				0.5						7.0			16.5
Cambisols		2.7		2.5		0.5						15.0			20.7
Chernozems															0.0
Podzoluvisols															0.0
Rendzinas			0.2												0.2
Ferralsols															0.0
Gleysols				0.6								2.0	1.0		3.6
Phaeozems		0.4										0.5			0.9
Lithosols		20.0	32.0	20.0		12.0				40.0	40.0	10.0			174.0
Fluvisols						0.7						15.0			15.7
Kastanozems															0.0
Luvisols						0.7						10.0			10.7
Greyzems															0.0
Nitosols		5.0				1.0		2.2				7.5			15.7
Histosols													1.4		1.4
Podzols												0.5	1.0		1.5
Arenosols		2.0								10.0	19.0	6.0			37.0
Regosols			7.0			4.0		9.0		2.0	2.0	65.0			89.0
Solonetz												4.0			4.0
Andosols												0.5			0.5
Rankers															0.0
Vertisols						0.7						4.0			4.7
Planosols															0.0
Xerosols							0.7			33.0		2.0			35.7
Yermosols										35.0	95.0	65.0			195.0
Solonchaks						7.0	17.0			12.5	12.5	8.0			57.0
Miscellaneous															
Total	5.0	34.1	39.2	23.1	0.0	27.1	17.7	11.2	0.0	132.5	168.5	222.0	3.4	0.0	683.8

[a] acc. to FAO/UNESCO soil map of the world (FAO, 1971–1978) VOL II–X.

[b] The presented acreages are estimated values of the extents of the soils, whether occurring as dominant soil, associated soil or inclusions, according to the FAO/UNESCO soil map of the world (VOL.II–X).

[c] ecosystems according to Whittaker and Likens (1975).

(1=tropical rainforest 6=woodland/shrubl. 11=extreme desert
2=tropical seasonal forest 7=savanna 12=cropland
3=temperate evergreen forest 8=trop.grassland 13=swamp/marsh
4=temperate deciduous forest 9=temp.grassland 14=tundra/alpine
5=boreal (taiga) 10=desert/semi-desert scrub)

APPENDIX IV.4

Table 7. The distribution of the soils and ecosystems (a) of North and Central Asia (FAO/UNESCO soil map of the world VOL.VIII)
Ecosystem (1000000 ha) (c)

Soil type (b)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols		50.3		3.2		20.0			28.9			25.6			128.0
Camoisols		5.2		34.7	148.7				6.4			15.7		58.1	268.8
Chernozems				1.1	6.0				20.5			61.5		0.2	89.3
Podzoluvisols					10.5				45.3						55.8
Rendzinas															0.0
Ferralsols															0.0
Gleysols				17.1	190.4				9.1			48.0	5.0	59.8	329.4
Phaeozems					5.4				12.8						18.2
Lithosols		12.8		72.5	21.1	5.0			89.0		20.5	5.6		150.6	377.1
Fluvisols		2.5		1.4	18.1				14.1			28.0			64.1
Kastanozems									96.3			14.1		30.7	141.1
Luvisols		4.7		14.2	8.2							8.0			35.1
Greyzems					9.7				7.1						16.8
Nitosols															0.0
Histosols					94.7				12.2				20.0	10.5	137.4
Podzols					14.5								5.0		19.5
Arenosols											2.2				2.2
Regosols					38.0				1.2					27.2	66.4
Solonetz									37.3		14.8				52.1
Andosols		4.9		0.9	10.0										15.8
Rankers															0.0
Vertisols						2.5						2.5			5.0
Planosols					2.0				0.3						2.3
Xerosols										51.3	33.8	18.8			103.9
Yermosols										24.2	118.8	7.5			150.5
Solonchaks		2.4							10.4	17.0	54.0				83.8
Miscellaneous															
Total	0.0	82.8	0.0	145.1	577.3	27.5	0.0	0.0	390.9	92.5	244.1	235.3	30.0	337.1	2162.6

Table 8. The distribution of the soils and ecosystems (a) of Southeast Asia (FAO/UNESCO soil map of the world VOL.IX)
Ecosystem (1000000 ha) (c)

Soil type (b)	1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols	85.0	20.0				16.0	22.0	3.0				19.0			165.
Camoisols		8.0				7.0	11.0	6.0				8.6			40.
Chernozems															0.
Podzoluvisols															0.
Rendzinas		4.0				1.0						0.9			5.
Ferralsols	13.0	1.0					1.0	1.0				2.6			18.
Gleysols	5.0	2.0				3.0	2.1	2.0				10.0	4.0		28.
Phaeozems															0.
Lithosols	7.0	2.0				2.0	5.2			2.0			6.0		24.
Fluvisols		2.0				0.7		5.0				12.0			19.
Kastanozems															0.
Luvisols		2.0				3.0	1.1	3.0				8.0			17.
Greyzems															0.
Nitosols		3.0				2.0	4.0	2.0				6.5			17.
Histosols								5.0				3.5	7.0		15.
Podzols			2.5	2.5								1.5			6.
Arenosols		2.0					2.0			4.0		1.2			9.
Regosols						3.0	1.0					0.3			4.
Solonetz															0.
Andosols		2.0						0.8				4.0		0.8	7.
Rankers		0.4													0.
Vertisols							2.0			1.0		1.7			4.
Planosols		0.2													0.
Xerosols															0.
Yermosols															0.
Solonchaks															0.
Miscellaneous															
Total	110.0	48.6	2.5	2.5	0.0	37.7	51.4	27.8	0.0	7.0	0.0	79.8	17.0	0.8	385.

[a] acc. to FAO/UNESCO soil map of the world (FAO, 1971-1978) VOL II-X.

[b] The presented acreages are estimated values of the extents of the soils, whether occurring as dominant soil, associated soil or inclusions, according to the FAO/UNESCO soil map of the world (VOL.II-X).

[c] ecosystems according to Whittaker and Likens (1975).

(1=tropical rainforest 6=woodland/shrubl. 11=extreme desert
2=tropical seasonal forest 7=savanna 12=cropland
3=temperate evergreen forest 8=trop.grassland 13=swamp/marsh
4=temperate deciduous forest 9=temp.grassland 14=tundra/alpine
5=boreal (taiga) 10=desert/semi-desert scrub)

APPENDIX IV.5

Table 9. The distribution of the soils and ecosystems (a) of Australasia (FAO/UNESCO soil map of the world VOL.X)

Soil type (b)	Ecosystem (1000000 ha) (c) 1	2	3	4	5	6	7	8	9	10	11	12	13	14	total
Acrisols	5.7			0.2		11.4	3.0	4.7	1.2			3.0			29.2
Cambisols		7.2		9.0		0.2	0.4	0.3	56.7			8.6		0.4	82.6
Chernozems															0.0
Podzoluvisols															0.0
Rendzinas		3.0													3.0
Ferralsols	3.9					4.0		2.0				2.5	0.1		12.5
Gleysols	2.0					0.1			0.2			0.8			3.1
Phaeozems						1.6		0.5	0.1			0.9			3.1
Lithosols			5.8	4.6		53.0	0.6			0.1					64.1
Fluvisols				0.6			0.5					0.5			1.6
Kastanozems						1.7						0.1			1.8
Luvisols			3.6	3.7		32.0	6.0		6.4	0.7		11.5			63.9
Greyzems															0.0
Nitosols						2.1						2.1			4.2
Histosols													0.9		0.9
Podzols				5.0		4.3									9.3
Arenosols			2.3			10.7				74.4		1.0			88.4
Regosols				1.1		34.3		5.0	60.6	22.5		2.0			125.5
Solonetz						42.4		4.7				1.0			48.1
Andosols		0.1	1.6	1.0				1.5				1.5			5.7
Rankers						1.0									1.0
Vertisols						31.2		31.7				2.0			64.9
Planosols						11.5		25.0				5.0		7.5	49.0
Xerosols										68.0		1.0			69.0
Yermosols										30.5	132.9				163.4
Solonchaks															0.0
Miscellaneous															
Total	11.6	10.3	13.3	25.2	0.0	241.5	10.5	75.4	125.2	196.2	132.9	43.5	1.0	7.9	894.5

[a] acc. to FAO/UNESCO soil map of the world (FAO, 1971–1978) VOL II-X.

[b] The presented acreages are estimated values of the extents of the soils, whether occurring as dominant soil, associated soil or inclusions, according to the FAO/UNESCO soil map of the world (VOL.II-X).

[c] ecosystems according to Whittaker and Likens (1975).
(1=tropical rainforest, 2=tropical seasonal forest, 3=temperate evergreen forest, 4=temperate deciduous forest, 5=boreal (taiga), 6=woodland/shrubl., 7=savanna, 8=trop.grassland, 9=temp.grassland, 10=desert/semi-desert scrub, 11=extreme desert, 12=cropland, 13=swamp/marsh, 14=tundra/alpine)