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8 Carbon cycle and ecosystem productivity on a global scale

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Summary

The annual rate of increase of atmospheric CO_2 by fossil fuel combustion is only about 60% of the annual CO_2 emission rate. Most of the remaining CO_2 is absorbed by the ocean, but a smaller part also by the terrestrial biosphere. The use of carbon isotopes as a tracer has shown that annually almost one tenth of the atmospheric carbon dioxide reservoir is exchanged with the ocean. Another one tenth is exchanged with the terrestrial biosphere. The rate of absorption of CO_2 in the ocean is limited by sluggish exchange of surface water with the deep sea. Increasing atmospheric CO_2 stimulates plant growth worldwide, both by increased photosynthesis and improved water use efficiency. This biotic CO_2 enrichment effect has delayed the rising atmospheric CO_2 curve by about ten years, and probably also explains the fate of CO_2 released by deforestation, the so-called "missing carbon". On a time scale of hundreds of years ocean and biosphere together will be able to absorb much more than they do now, about 85% of the total CO_2 emission. Climatic change can alter this picture in ways that are necessarily strongly diverse for different climatic zones.

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8.1 Introduction

The concentration of greenhouse gases in the atmosphere is unmistakably increasing at a fast rate, and will double at the timescale of a few human generations. This increase may induce climatic changes that will exceed the adaptive capacity of nature, and indeed of the human society. The mere existence of something like a greenhouse effect is a good thing: without it we would be frozen, but its intensification is a cause of concern, and may lead to temperature increases of the order of a few degrees centigrade. Other climatic consequences such as for precipitation or storm events are hardly known.

The most abundant greenhouse gas is not CO_2, but water vapour. Yet, CO_2 has a central role because it is less dependent on climate than water vapour. Water vapour acts more like a passive companion, enhancing a warming up by CO_2 once begun. There is usually a higher water vapour content in warmer air and therefore also a larger greenhouse effect. This sounds like a circular argument, but in fact it is the expression of a positive feed-back mechanism that destabilizes the climate system and makes prediction much more difficult. Surely, negative feed-backs also exist, such as increased cloud formation, that might weaken an initial warming. The uncertainty of the net result of the combined effects is considerable. For this reason most of the attention in this article will be given to CO_2 and the carbon cycle.

8.2 The global carbon cycle

Atmospheric CO_2 increases at a rather steady rate, at least on a decadal time scale. During the last 50 years its rate of increase has grown exponentially at a rate of about 2.4% per year, which has been about 60 % of the rate of emission of CO_2 from fossil fuel burning. The size of this fraction is not a natural constant, on the contrary, it depends on the activity and absorptive capacity of the sinks for CO_2, the ocean in the first place and the biosphere in the second place. On longer time scales, there has been a rather stable level at about 280 ppm for the last thousands of years until the effects of the industrial revolution became noticeable. Further back in time, during the last glacial period, atmospheric CO_2 was much lower and even dropped to about 200 ppm. This alternation of about 200 ppm during a glacial period and 280-300 ppm during an interglacial period has probably been repeated several times during the last one million years, when glacial periods returned about every 100 000 years. The causes of these glacial periods are still debated, and probably consist of many cooperating factors. Glacial periods are triggered by the Milankovich cycle, that is caused by precession in the planetal orbit of the earth around the sun. This cycle itself already existed during the Tertiary period but it did not cause glacials yet. Probably a slowly declining CO_2 concentration, and accompanying global cooling, passed a critical threshold so that the positive feedback mechanism of increased snow cover and glacial cap formation could take over. Only then could the Milankovich cycle trigger a run-away "ice-house" condition, signifying the beginning of the Pleistocene period.

Surely, on different time scales, different processes are important. On a shorter time scale than the decadal one, seasonal or even diurnal fluctuations dominate. The seasonal effect of carbon uptake by plant growth is quite outspoken in atmospheric CO_2 records that have been obtained during the past 30 years, such as the well-known record of Mauna Loa. Most remarkably, this seasonal fluctuation is almost absent on the southern hemisphere. The best explanation is that carbon uptake by algal growth in the oceans is buffered by the large carbonate content of seawater. The alternative explanation that carbon uptake in the ocean is simply much smaller than that on land, has recently been disproven by simultaneous measurements of atmospheric oxygen. Until recently the corresponding seasonal cycle of atmospheric oxygen could not be measured, due to the very high back

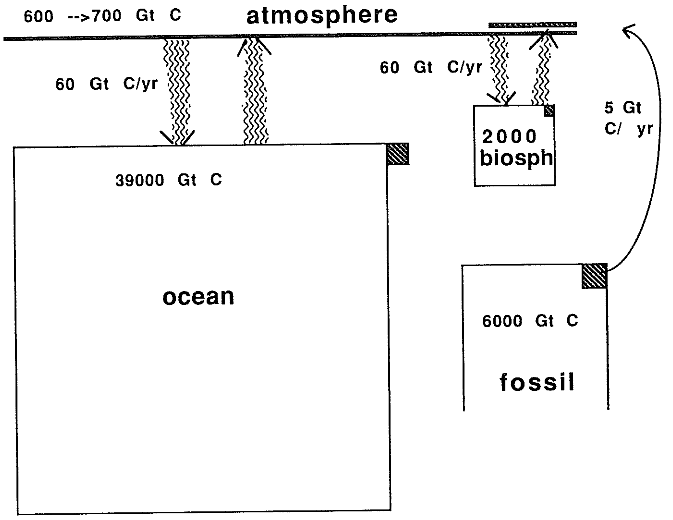

Figure 8.1 Major global reservoirs and fluxes of carbon, expressed in Pg (or Gt).

ground concentration, but now it has been shown that the cycle is present both above the continents and the oceans. Oxygen is not buffered in seawater the same way as CO_2 is, and so it does follow the annual cycle.

On the whole, the terrestrial and marine annual fluxes of biotic carbon uptake and release are of the same order of magnitude and equal to about 60 Pg of carbon ($1\ Pg = 10^{15}\ g = 1\ Gt$) each (Figure 8.1). The total flux of about 120 Pg is drawn from the atmosphere that contains about 700 Pg of C in total. Therefore the whole CO_2 content of the atmosphere is recycled once per six years. This is such a short period of time that it seems as if any disturbance by injecting CO_2 from fossil fuel should be absorbed without difficulty. This impression is even reinforced when one considers the large size of the ocean pool of carbon: almost 40 000 Pg of C, but it is a false impression as we will see.

8.3 The limited absorption capacity for CO_2 in the carbon cycle

On a geological time scale the largest carbon reservoir on earth is rock carbonate. This huge reservoir does participate in the global carbon cycle, but very slowly. During a process of granite rock weathering, carbonate and SiO_2 are formed from silicate and CO_2 drawing down atmospheric CO_2. Eventually this process will absorb practically all CO_2 from fossil fuel burning, but it will only do so at a time scale of tens of thousands of years (Berner and Lasaga, 1989). The largest next reservoir of carbon in the carbon cycle is the dissolved bicarbonate in the ocean, and this reservoir is active on a decadal time scale. In comparison to carbon in the ocean, fresh water of lakes and rivers can be totally ignored. The amount of CO_2 that can be dis-

solved in water is proportional to the partial pressure in the air above it. This holds for both fresh water and seawater, but in seawater bicarbonate ions have a much higher concentration than dissolved CO_2, because of the high pH (about 8.1). In fact, a chemical equilibrium exists in the chain $CO_2 <-> HCO_3^- \leftrightarrow (CO_3)^{2-}$, which shifts to the right with increasing pH (Butler, 1982). When the CO_2 concentration in the air increases, more CO_2 gets dissolved and an equilibrium shift occurs towards HCO_3^-, which will release H^+ ions. Therefore the pH goes down, offering a strong negative feed-back to further absorption. The net result is that in seawater the incremental absorption is not proportional to partial pressure of CO2 and total amount (about 25 g m^{-3}) of carbon dissolved, but to only about 1/10 of this product. Therefore, effectively the pool size of ocean carbon is only about 4000 Pg , and 36 000 Pg can be considered as being chemically locked up.
Still, 4000 Pg is much more than what the biosphere contains. However, the second reason for limited absorptive capacity is the slowness of mixing of deep ocean water and surface water. The upper few hundred meters of the ocean are in good contact with the atmosphere, but the deeper layers are separated by a thermal inversion and the exchange rate with them is slow. The average residence time of water in the deep Pacific is several hundreds of years. The water of the Atlantic is recirculated at a much faster rate, in less than 100 years time. The overall effect is that out of the 4000 Pg chemically accessible, only about one tenth is immediately in contact with the atmosphere. The other nine tenth will absorb its portion in due time, but only after hundred of years.
These two facts tell a lot about the global carbon cycle. In the long run (a thousand years) redistribution of emitted CO_2 will occur between the atmospheric reservoir of 700 Pg and the oceanic effective reservoir of 4000 Pg. A fraction of about 700/(700 + 4000) or about 15 % will remain in the atmosphere. On the short decadal time scale, this remanent fraction is much higher due to lack of time for mixing. A similar calculation gives a remanent fraction of about 700/(700 + 400) or about 64 %. This value is very close to the observed remanent fraction, and can serve as a first guideline for scenario studies.

8.4 The role of the terrestrial biosphere

Deforestation as an additional source of CO_2 release to the atmosphere is estimated to amount to a rate between 1 and 2 Pg of C per year. When this rate is added to the known emission rate of fossil fuel burning, atmospheric CO_2 should rise faster than it actually does. This in a nutshell is the problem of the so-called "missing carbon". In Figure 8.1 the total biosphere reservoir is given, as well as the annual exchange flux of carbon with the atmosphere. The size of this exchange flux is of the order of 50 Pg of C per year, but it is worthwhile to consider this flux and its definition in more detail.
This flux can be identified with the Net Primary Productivity, the annual sum of the daily net growth rates of plant dry matter (expressed in carbon). This definition means that the own respiration of the vegetation has been taken into account , and is not included any more. The flux *before* subtraction of respiration is the Gross Primary Productivity (*GPP*), which can be considered as the sum of gross CO_2 assimilation rates. However, the calculation of this gross flux is even more imprecise

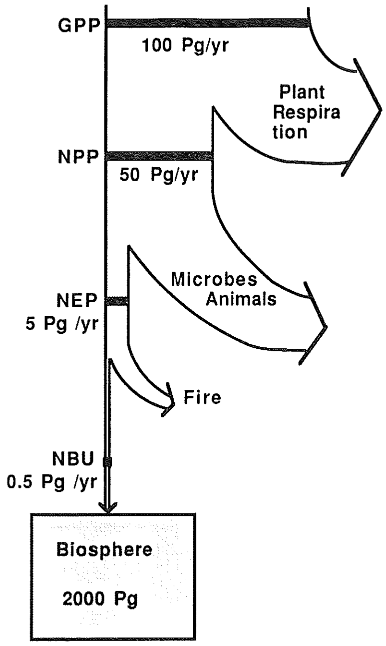

Figure 8.2 Major terrestrial carbon fluxes, as consumed by the plants themselves, by heterotrophic organisms and by fire. Net biospheric uptake is the remainder of large positive and negative fluxes.

than that of the *NPP*, and based on rough guesses of the rate of respiration. The estimate of *NPP* is based on field observations of production rates of new leaves, twigs, branches and roots and is therefore somewhat more reliable. The order of magnitude of global *NPP* is 50 Pg of C per year, whereas that of *GPP* is about twice as much (Figure 8.2). The Net Primary Productivity is largely consumed by grazers, such as insects and various kinds of mammals, or left over to be decomposed by fungi and microbes. Live biomass is estimated to be about 500 Pg , mainly in the form of wood. After death it turns to litter and eventually to soil organic matter (about 1500 Pg of C). The magnitudes of the pools of organic matter in various forms and of the fluxes of formation and consumption are related by residence times and partitioning coefficients. Their basic relationship is given in Figure 8.3 as a scheme driven by *NPP*. The outflow of each state variable in this scheme is calculated as content divided by longevity, so that its equilibrium content is simply the product of longevity and inflow. The outflows cascade down to litter, humus, and resistent soil carbon. From biomass to humus a considerable fraction of carbon is lost by respiratory processes, and also from humus to resistant soil carbon. The complement of this fraction returns to the atmosphere as respiratory CO_2. More complex approaches to decomposition of organic matter (Janssen, 1992) may be needed, but were not considered here.

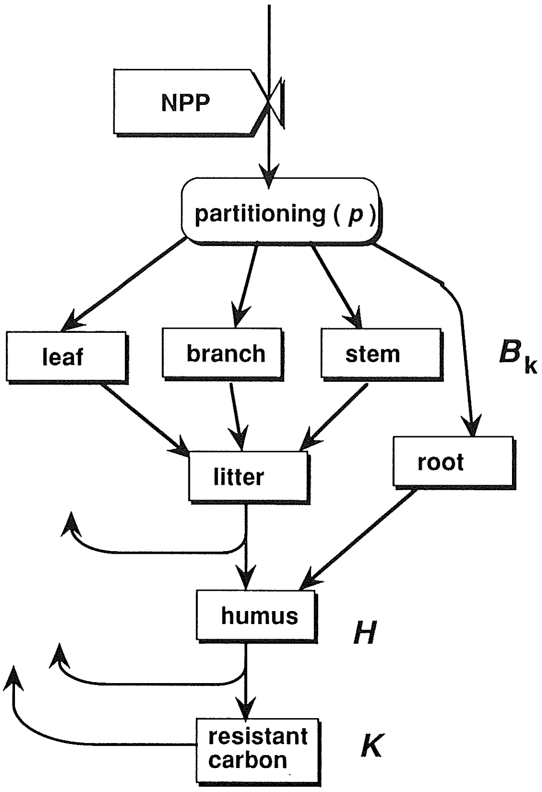

Figure 8.3 Simplified model structure for an ecosystem. Each box itself is described by first order decay. Characteristics such as *NPP*, its allocation, reservoir longevities and transfer (respiratory) losses depend on ecosystem type.

Above-ground biomass consists of leaves, stemwood, roots, branches and litter. Combination of the driving forces and residence times leads to a steady state distribution of surface densities of carbon for a biosphere model consisting of six different ecosystem types as presented in Figure 8.4 (Goudriaan 1990). In this figure the width of each bar represents the area of the vegetation type, and the height of each bar the carbon areal density. The high areal density of soil carbon in temperate forest and in grasslands as compared with agricultural land and tropical forests is noteworthy.

Many ecosystems in the world, however, are not in a steady state, but they accumulate dry matter during a number of years and are then disrupted by fire or other drastic events, so that they go through a saw tooth-like life cycle. Fire is often a natural process, and necessary for rejuvenation of ecosystems. Locally, disruption by fire is a discontinuous event, but summed over the entire globe there is a con-

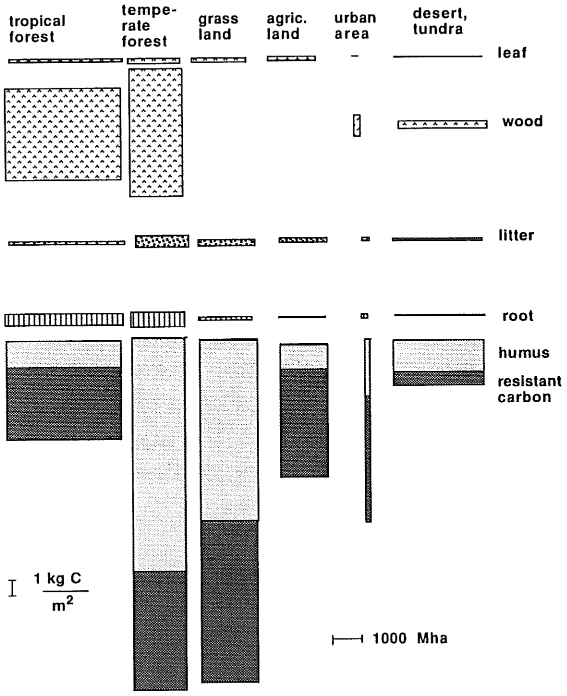

Figure 8.4 Simulated carbon areal density (heights of the columns) and land area for each vegetation type (widths of the columns). In this representation the areas of the columns are proportional to the simulated amounts of carbon.

tinuous release of CO_2 by fire, which is practically compensated by regrowth at theother places. The total carbon release flux due to fire is of the order of 4 - 7 Pg of carbon per year. The largest portion in this flux is on account of periodic burning of tropical grasslands (Hall and Scurlock 1991). Although this flux is almost as large as the rate of fossil fuel consumption, grassland burning does not contribute to the increase of atmospheric CO_2. Within a time span of one year most of the biomass turns to CO_2 anyway, if not by fire then by decomposition. Moreover, fire residues such as charcoal are added to resistant soil carbon.

On a global scale, grasslands occupy about 20% of the well vegetated land area, contributing about 23% to both annual carbon flux and total carbon storage. The surprisingly large share in terms of carbon storage is entirely due to the large amount of organic matter that is maintained in grassland soils, owing to a combination of large partitioning of dry matter to below-ground parts and to a relatively slow rate of decay of soil organic matter. When a forest is turned into land used for agriculture, the standing biomass is removed (often partly burned), and the disturbance of the soil will give rise to enhanced decomposition of soil organic matter and so to a strong efflux of CO_2. In arable land, the disturbance is repeated annually by ploughing and after a few decades the soil carbon content has declined to a much lower level than originally present in the virgin soil. In grassland, however, the situation is different. The soil is much less disturbed, the crop is perennial, and the remnants of leaves, stolons, roots, stubble and of cattle manure accumulate and give rise to a soil organic matter level as high as in the original forest soil. At present grasslands are extended far beyond their natural range. In the Holdridge diagram (Holdridge,1967; Monserud and Leemans, 1992) the term "cool temperate steppe" is used to indicate the climatic zone where grasslands naturally occur. This zone is a rather dry one, with annual potential evapotranspiration exceeding annual precipitation. These truly natural grasslands occur in the prairie zone of North America, the chernozem zone of the Ukraine and in the pampas of Argentina. Their combined area is about 900 Mha, containing a total amount of soil carbon of 120 Pg (Post et al. 1985) In the actual man-made situation, the most productive grasslands are found in the zone with a larger moisture supply, indicated by the term "moist temperate forests". These natural forests have long ago been replaced by agricultural land, and indeed also by pasture land. The area of these man-made grasslands is also about 900 Mha, but their soil carbon content is higher than that of natural grasslands, probably around 200 Pg.

8.5 The use of carbon isotopes as a tracer

Ammerlaan and de Visser (1993) demonstrate the usefulness of carbon isotopes in plant physiological research. On a global scale as well, time courses of carbon isotopes are utilized to yield important information. The largest global tracer "experiment" ever came from the hydrogen bomb experiments that peaked around 1962. Due to these experiments the concentration of ^{14}C in the atmosphere was approximately doubled. After the international ban in 1965 this peak gradually declined by dilution of ^{14}C in the ocean and biosphere. The rate of this decline offers a method to estimate the exchange rate. In early studies the exchange rate with the biosphere was ignored, and the estimated rate was totally ascribed to the ocean. This yielded exchange rates of the order of 100 Pg C per year. In fact, part of the exchange is with the biosphere, and only the remainder is with the ocean. Therefore the best method is to use a model that allows for both exchange rates, to impose the exchange rate of the biosphere (*NPP*) and then to find the exchange rate at the ocean surface by curve fitting. This method yields a lower rate of exchange (50 Pg per year) than the one normally assumed, but it describes very well the measured decline rate of ^{14}C in the atmosphere (Figure 8.5).

The second man-made global tracer experiment is the emission of fossil carbon. No ^{14}C remains in fossil fuel at all, and therefore the atmospheric ^{14}C content before 1960 gradually declined ("Suess effect"). Later the hydrogen bomb experiments completely swamped this effect, and made it impossible to follow it any further. However, the other carbon isotope, ^{13}C, was not disturbed by the nuclear experiments and remained available as a tracer. Naturally, about 1% of carbon is in the form of ^{13}C, but in several processes a slight discrimination occurs. The process of photosynthetic CO_2 assimilation (in C_3 plants at least) discriminates against the heavier ^{13}C , so that the ^{13}C concentration in plant biomass is about 25 per thousand lower than in the atmospheric source. This has been the case ever since photosynthesis started, and so the depletion is equally present in fossil carbon. Additional discrimination in methanogenesis has further depleted ^{13}C content in

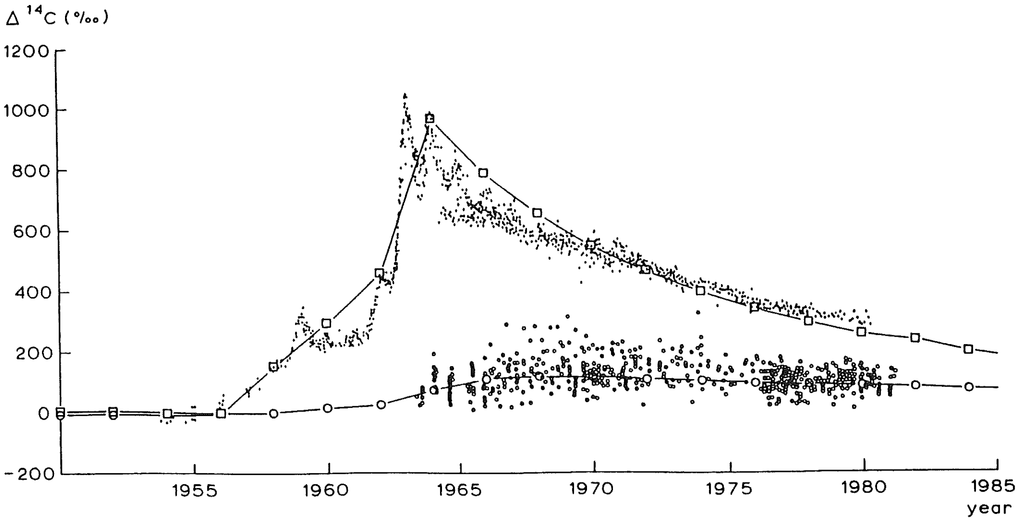

Figure 8.5 Simulated and observed time courses of $\Delta^{14}C$ in the atmosphere (squares) and in the sea surface water at low latitudes (circles), for the period around the nuclear tests. Data from Bolin (1986) in his Figure 3.6.

natural gas reserves, down to about -40 per thousand. The release of isotopically depleted CO_2 by fossil fuel burning is visible in various records of atmospheric CO_2 and also in tree rings. This decline serves as a validation of model runs for the global carbon cycle (Goudriaan,1989; Goudriaan, 1992).

8.6 The biosphere as a sink for carbon

The increase of atmospheric CO_2 itself induces an increase in net CO_2 assimilation and also in plant growth. Since industrialization, this effect by itself has stimulated biospheric uptake rate by about 1 Pg of C per year, which is at present equivalent to 20 % of the global emission rate of CO_2. As another comparison, this CO_2-induced absorption is just about equal to the fluxes released in large scale deforestation. This is a coincidental compensation, but it explains why a model with just the ocean, and without biosphere at all, still works reasonably well. Such a model will obviously fail to explain effects of deforestation or afforestation.

This role of the biosphere as a sink cannot be directly measured, but the circumstantial evidence is very strong, indicating a strong stimulation of plant growth by increased atmospheric CO_2, not only in agricultural crops but also in natural vegetation (Strain and Cure, 1985). According to a summary of Kimball (1983), there is a mean 40% increase of dry matter in C_3 crops upon doubling of CO_2, and of 15% for C_4 crops. This effect of CO_2 not only occurs in crop plants, but is a general phenomenon (Lemon 1984). The primary physiological effect of increased CO_2-concentration is a stimulation of the rate of CO_2 assimilation, and also the respiration rate might be reduced (Amthor 1991). The CO_2 enrichment effect is maintained when growth is limited by water (Gifford 1979), which can be explained by control of both water loss and CO_2 assimilation by stomatal resistance. Any increase in ambient CO_2 will then stimulate CO_2 uptake without raising water loss. Nutrient shortage, especially of phosphorus and of potassium, tends to impose a more absolute limitation to crop growth without leaving much room for stimulation by CO_2. Nitrogen, however,differs from other nutrients in that it permits a small positive CO_2 effect, even under rather severe nitrogen shortage (see also Arp and Berendse, 1993).

Arp (1991) showed that part of the conflicting evidence in literature on adaptation of plants during growth can be explained by differences in pot size used in the reported experiments, the disappearance of the CO_2 effect on growth being associated with small pot size. In the field, a primary CO_2 stimulus can lead to better exploration of the soil (Rosenberg 1981), so that nutrient uptake may go up even under nutrient limited circumstances. Indeed, in an experimental study during 4 years of continued exposure of a natural salt marsh vegetation to high CO_2 , Arp et al.(1991) found no decline in increased photosynthesis and no decline in water use efficiency .

8.7 Deforestation and CO_2 fertilization

The CO_2-induced sink effect must be separated from the effect of land-use changes, because the latter is independent of atmospheric CO_2. The precise definition of emission due to deforestation is full of pitfalls: next to direct removal and burning of wood, it contains also a stimulated oxidation of soil carbon. This process takes many years to reach a new steady state, and during that period the soil is still releasing a flux of CO_2 larger than it used to do before. This continued flux increase was represented in the carbon cycle model by decreased longevity for humus after land is turned into arable land. Another factor included in the model was incomplete biomass burning, leading to partial charcoal formation (10 - 20% of the above ground biomass). The net effect of these factors is expressed in the total amount of simulated biospheric carbon. Comparison of total biospheric carbon for model runs with and without land use changes (deforestation in particular) resulted in a difference, due to deforestation, of 90 Pg C over the period 1780 - 1980 (Table 8.1, first column for $\beta = 0$). The correct order of magnitude of this number is confirmed by a comparison with the amount of biomass that was present on the total forest area lost during this period. For tropical forests almost 800 Mha was lost with a mean biomass of 8 kg C m^{-2} (64 Pg C) and for temperate forests 200 Mha with a mean biomass of 11 kg C m^{-2} (22 Pg C).

However, deforestation and CO_2 fertilization occur at the same time. Therefore the true effect of land use changes must be determined from model results *without* this CO_2 enrichment effect (left hand column in Table 8.1). As the next step, the CO_2 enrichment effect must be estimated *with* the land use changes going on (lower row in Table 8.1).

The sequence is important, as there appeared to be a strong interaction between the effects of deforestation and CO_2 enrichment. The positive CO_2 enrichment effect was larger when deforestation was included, and on the other hand the biomass removal by deforestation became smaller when CO_2 fertilization occurred.

Table 8.1 Simulated total amount of carbon in Pg C in the terrestrial biosphere for the year 1980. Without any land use changes and without any CO_2 effect ($\beta = 0$), the biosphere would have remained identical to that in the starting year 1780. The separate effects of these two factors and of their combination showed a strong interaction. The arrows indicate the correct sequence for determining the effects of land use changes and of CO_2 enrichment respectively (bold). Their combined effect gave a net loss of carbon of 39 Pg C.

	$\beta = 0$		$\beta = 0.5$	Enrichment effect
No land use changes	1894 Pg$^{(1780AC)}$		1926 Pg	32 Pg
	↓			
With land use changes	1804 Pg	⟶	1855 Pg $^{(1980AC)}$	**51** Pg
				↓
Land use effect	– **90** Pg		– 71 Pg	– **39** Pg

This interaction can be understood as follows: By deforestation the dynamics in the biosphere is enhanced, and the CO_2 enrichment effect is more effectively transferred into the slow compartments.

The other interaction is more straighforward: CO_2 fertilization will reduce the net effect of deforestation, simply by negative feedback. The baseline of deforestation emission was 90 Pg C. The CO_2 fertilization effect itself will recycle about 20% of this emission into the biosphere, which leaves about 71 Pg C as effective emission. This number is however not a good estimate for the size of the source itself, since the negative feedback is already included. The combined effect of CO_2 fertilization and deforestation is a simulated biospheric carbon loss of 39 Pg between 1780 and 1980 (diagonal transition in Table 8.1).

Summarizing, the total emission of 249 Pg C (159 fossil and 90 biospheric) was partitioned to the three major reservoirs as 88 Pg to the ocean, 51 Pg to the biosphere and 110 Pg remaining in the atmosphere. For the biosphere, the net result was a loss of 39 Pg of carbon, which was added to the release of 159 Pg of fossil carbon. The time course of total emission followed an approximately exponential increase at a relative growth rate of about 2.4% a^{-1}.

8.8 Redistribution of current fluxes

The almost exponential increase of fluxes is still going on. The rate of deforestation in the year 1980 was about 12 Mha per year, and with a loss or carbon (including soil carbon) of about 14 kg C m^{-2}, a flux of release of carbon due to deforestation can be expected of about 1.7 Pg C a^{-1} (including soil carbon losses). This figure is well within the range of a recent estimate of Houghton(1991).

There are, however, a number of processes that reduce the rate of emission from the biosphere as a whole. First, loss of soil carbon does not occur instantaneously but it is delayed in time. A dynamic simulation model takes such a delay into account, in contrast with statistical calculations that assume immediate release. To give an impression of the effect of this delay: assuming a time coefficient of 20 years for soil carbon and a relative growth rate of deforestation of 2% per year, the rate of release is reduced by a factor 1.5. Surely the soil carbon will eventually be released, but the model deals with the *current* rates. Secondly, not all standing biomass is immediately oxidized, part of it is even conserved in the form of charcoal, which has a much longer longevity than biomass. As a third factor, regrowth of forests occurs in particular in the temperate parts of the world (Brouwer et al. 1991). The total rate of regrowth is probably of the order of one third of the rate of tropical deforestation. In the model these processes together reduced the net rate of global carbon release due to land use changes to about 0.6 Pg C a^{-1}.

This net rate of release was counteracted by a current CO_2 fertilization flux of 1.1 Pg C a^{-1}, which was derived from model output using the same method as for the accumulated amounts, given in Table 8.1. This model result can be approximately understood as follows:

The annual rate of increase of atmospheric CO_2 is about 0.4% per year, leading to an assumed CO_2 fertilization effect of 0.2% per year for *NPP*. In the simple biosphere model explained in the beginning of this text, the resulting annual increase of equilibrium should then be 4 Pg a^{-1}, based on 2000 Pg present. The effect of the slowness of the different compartments reduces this rate to about 1.1 Pg a^{-1}.

The result for the terrestrial biosphere as a whole is *a net uptake rate of carbon of 0.5 Pg* a^{-1}. The rate of emission due to fossil fuel burning was about 5.2 Pg C a^{-1}. When this flux is added to the biospheric release rate of 0.6 Pg a^{-1} , we find a total emission rate of about 5.8 Pg C a^{-1} . The rate of absorption by the ocean was simulated to be 2.0 Pg C a^{-1} , and so the remaining rate of accumulation in the atmosphere was 2.7 Pg C a^{-1}. The remanent fraction based on *current* emission rates is now found to be 0.46, very close to the figure based on accumulated amounts over the past 200 years. This small difference emphasizes that the actual increase over time for the combined fossil/biosphere emission has been very close to an exponential one. Yet, the time patterns of biospheric emission and fossil fuel emission separately have not been identical. Around the turn of the century the biospheric emission was larger than the fossil fuel emission was, whereas nowadays the situation has reversed. Over the past 200 years together the biosphere has been a net source, but somewhere during the past decades it must have turned into a net sink. This switch is the result of the steadily increasing CO_2 fertilization effect.

The dual character of the biosphere in being both a source and a sink is undeniable. The biosphere may not be quite as strong a sink as the ocean is, but it is well capable of counteracting its own role as a source (Lugo & Brown, 1986).

8.9 Final remarks

There are different ways to calculate the remanent fraction in the atmosphere. The classical method is simply to take the ratio of atmosphere increment divided by fossil fuel emission, which yields a value of 0.69. This number is too high, since the biospheric emission is not included. But which contribution should we use? Based on *total* biospheric emission (159 + 90) we find a value of 0.44 , but based on *net* biospheric emission (159 + 39) we find a value of 0.55. When biospheric emission is taken for granted and only the ocean is considered as a sink interacting with the

atmosphere, the value of the remanent fraction is 0.55. When the sink response of the biosphere is included, the remanent fraction is smaller and arrives at a value of 0.46.
Biospheric release fluxes of CO_2 that are caused by periodic fires in savannas e.g. should not be added to the base of the calculation. These fluxes do not constitute a permanent transfer of carbon from the biosphere to the atmosphere, but are part of the system itself in much the same way as decomposition and respiration are.
Climatic change *per se* was not discussed yet in this manuscript. Warming, if it occurs, will almost certainly have various effects. In some places enhanced decomposition of soil organic matter will prevail, such as is most likely the case in the tundra regions. Simultaneously, this enhanced decay in the soil compartments may be compensated by increased formation of above ground dry matter (Esser, 1987). In other places, extension of the duration of the climatic growing season will be more important and the net carbon storage will increase. Agricultural zones will shift (Parry,1990; Leemans and Solomon, 1993; Cramer and Solomon,1993).

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