

# The Carbon Cycle

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## CO<sub>2</sub> and the Carbon Cycle

(Extracted from the Intergovernmental Panel on Climate Change (IPCC) Report, “Climate Change, 1994”)

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### Summary

Interest in the carbon cycle has increased because of the observed increase in levels of atmospheric CO<sub>2</sub> (from ~280 ppmv in 1800 to ~315 ppmv in 1957 to ~356 ppmv in 1993) and because the signing of the UN Framework Convention on Climate Change has forced nations to assess their contributions to sources and sinks of CO<sub>2</sub>, and to evaluate the processes that control CO<sub>2</sub> accumulation in the atmosphere. Over the last few years, our knowledge of the carbon cycle has increased, particularly in the quantification and identification of mechanisms for terrestrial exchanges, and in the preliminary quantification of feedbacks.

### The Increase in Atmospheric CO<sub>2</sub> Concentration Since Pre-Industrial Times

Atmospheric levels of CO<sub>2</sub> have been measured directly since 1957. The concentration and isotope records prior to that time consist of evidence from ice cores, moss cores, packrat middens, tree rings, and the isotopic measurements of planktonic and benthic foraminifera. Ice cores serve as the primary data source because they provide a fairly direct and continuous record of past atmospheric composition. The ice cores indicate that an increase in CO<sub>2</sub> level of about 80 ppmv paralleled the last interglacial warming. There is uncertainty over whether changes in CO<sub>2</sub> levels as rapid as those of the 20th century have occurred in the past. However, there is essentially no uncertainty that for approximately the last 18,000 years, CO<sub>2</sub> concentrations in the atmosphere have fluctuated around 280 ppmv, and that the recent increase to a concentration of ~356 ppmv, with a current rate of increase of ~1.5 ppmv/yr, is due to combustion of fossil fuel, cement production, and land use conversion.

### The Carbon Budget

The major components of the anthropogenic perturbation to the atmospheric carbon budget are anthropogenic emissions, the atmospheric increase, ocean exchanges, and terrestrial exchanges. Emissions from fossil fuels and cement production averaged  $5.5 \pm 0.5$  GtC/yr over the decade of the 1980s (estimated statistically). The measured average annual rate of atmospheric increase in the 1980s was  $3.2 \pm 0.2$  GtC/yr. Average ocean uptake during the decade has been estimated by a combination of modelling and isotopic measurements to be  $2.0 \pm 0.8$  GtC/yr.

Averaged over the 1980s, terrestrial exchanges include a tropical source of  $1.6 \pm 1.0$  GtC/yr from ongoing changes in land use, based on land clearing rates, biomass inventories, and modelled forest regrowth. Recent satellite data have reduced uncertainties in the rate of deforestation for the Amazon, but rates for the rest of the tropics remain poorly quantified. For the tropics as a whole, there is incomplete information on initial biomass and rates of regrowth. Potential terrestrial sinks may be the result of several processes, including the regrowth of mid-latitude and high latitude Northern Hemisphere forests ( $0.5 \pm 0.5$  GtC/yr), enhanced forest growth due to CO<sub>2</sub> fertilisation ( $0.5$ – $2.0$  GtC/yr) and nitrogen deposition ( $0.2$ – $1.0$  GtC/yr),

and, possibly, response to climatic anomalies ( $0$ – $1.0$  GtC/yr). Partitioning the sink among these processes is difficult, but it is likely that all components are involved. While the CO<sub>2</sub> fertilisation effect is the most commonly cited terrestrial uptake mechanism, existing model studies indicate that the magnitude of contributions from each process is comparable, within large ranges of uncertainty. For example, some model-based evidence suggests that the magnitude of the CO<sub>2</sub> fertilisation effect is limited by interactions with nutrients and other ecological processes. Experimental confirmation from ecosystem-level studies, however, is lacking. As a result, the role of the terrestrial biosphere in controlling future atmospheric CO<sub>2</sub> concentrations is difficult to predict.

### Future Atmospheric CO<sub>2</sub> Concentrations

Modelling groups from many countries were asked to use published carbon cycle models to evaluate the degree to which CO<sub>2</sub> concentrations in the atmosphere might be expected to change over the next several centuries, given a standard set of emission scenarios (including changes in land use). Models were constrained to balance the carbon budget and match the atmospheric record of the 1980s via CO<sub>2</sub> fertilisation of the terrestrial biosphere.

### Stabilisation of Atmospheric CO<sub>2</sub> Concentrations

Modelling groups carried out stabilisation analyses to explore the relationships between anthropogenic emissions and atmospheric concentrations. The analyses assumed arbitrary concentration profiles (i.e., routes to stabilisation) and final stable CO<sub>2</sub> concentration; the models were then used to perform a series of inverse calculations (i.e., to derive CO<sub>2</sub> emissions given CO<sub>2</sub> concentrations). These calculations did three things: (1) assessed the total amount of fossil carbon that has been released (because land use was prescribed), (2) determined the partitioning of this carbon between the ocean and the terrestrial biosphere, and (3) ascertained what the time course of carbon emissions from fossil fuel combustion must have been to arrive at the selected arbitrary atmospheric CO<sub>2</sub> concentrations while still matching the atmospheric record through the 1980s.

Results suggest that in order for atmospheric concentrations to stabilise below 750 ppmv, anthropogenic emissions must eventually decline relative to today's levels. All emissions curves derived from the inverse calculations show periods of increasing anthropogenic emissions, followed by reductions to about a third of today's levels (i.e., to ~2 GtC/yr) for stabilisation at 450 ppmv by the year 2100, and to about half of current levels (i.e., ~3 GtC/yr) for stabilisation at 650 ppmv by the year 2200. Additionally, the models indicated that if anthropogenic emissions are held constant at 1990 levels, modelled atmospheric concentrations of CO<sub>2</sub> will continue to increase over the next century.

Among models the range of emission levels that were estimated to result in the hypothesised stabilisation levels is about 30%. In addition, the range of uncertainty associated with the parametrization of CO<sub>2</sub> fertilisation (evaluated with one of the models) varied between  $\pm 10\%$  for low stabilisation values and  $\pm 15\%$  for higher stabilisation values. The use of CO<sub>2</sub> fertilisa-

tion to control terrestrial carbon storage, when in fact other ecological mechanisms are likely involved, results in an underestimate of concentrations (for given emissions) of 5 to 10% or an overestimate of emissions by a similar amount (for given concentrations).

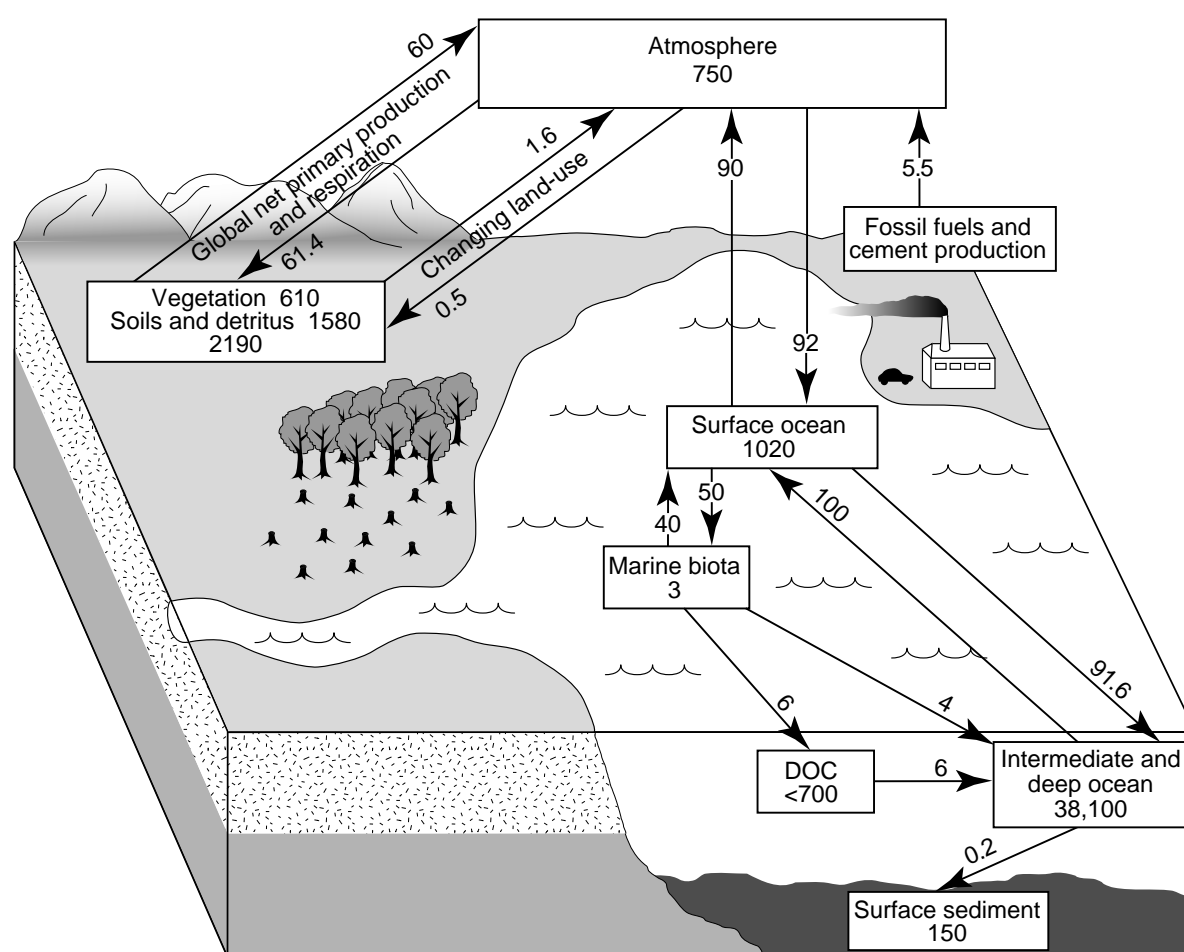
### Feedbacks to the Carbon Cycle

Climate and other feedbacks via the oceans and terrestrial biosphere have the potential to be significant in the future. The effects of temperature on chemical and biological processes in the ocean are thought to be small (tens of ppmv changes in the atmosphere), but the effects of climate on ocean circulation could be larger, with possible repercussions for atmospheric concentration of  $\pm 100$ – $200$  ppmv. Effects of changing precipitation, temperature, and atmospheric CO<sub>2</sub> can also have effects on the terrestrial biosphere, resulting in feedbacks to the atmosphere. Models suggest transient losses of about 200 GtC from terres-

trial ecosystems as temperatures warm, with a potential for long-term increases in carbon storage above present levels by a few hundred gigatons. Patterns of changing land use will have a substantial effect on terrestrial carbon storage and decrease the potential of terrestrial systems to store carbon in response to CO<sub>2</sub> and climate. Representation of feedbacks on the carbon cycle through oceanic and terrestrial mechanisms needs to be improved in subsequent analyses of future changes to CO<sub>2</sub>.

### 1.1 Description of the Carbon Cycle

Atmospheric CO<sub>2</sub> provides a link between biological, physical, and anthropogenic processes. Carbon is exchanged between the atmosphere, the oceans, the terrestrial biosphere, and, more slowly, with sediments and sedimentary rocks. The faster components of the cycle are shown in Figure 1.1. In the absence of anthropogenic CO<sub>2</sub> inputs, the carbon cycle had pe-



**Figure 1.1:** The global carbon cycle, showing the reservoirs (in GtC) and fluxes (GtC/yr) relevant to the anthropogenic perturbation as annual averages over the period 1980 to 1989 (Eswaran et al., 1993; Potter et al., 1993; Siegenthaler and Sarmiento, 1993). The component cycles are simplified and subject to considerable uncertainty. In addition, this figure presents average values. The riverine flux, particularly the anthropogenic portion, is currently very poorly quantified and so is not shown here (see text). While the surface sediment storage is approximately 150 Gt, the amount of sediment in the bioturbated and potentially active layer is of order 400 Gt. Evidence is accumulating that many of the key fluxes can fluctuate significantly from year to year (terrestrial sinks and sources: INPE, 1992; Ciais et al., submitted; export from the marine biota: Wong et al., 1993). In contrast to the static view conveyed by figures such as this one, the carbon system is clearly dynamic and coupled to the climate system on seasonal, interannual and decadal time-scales (Schimel and Sulzman, 1994; Keeling and Whorf, 1994).

### RECENT ANOMALIES

The last few decades have been characterised by a number of observed changes in the carbon cycle:

- The early 1980s were characterised by a period of relatively constant or slightly declining fossil carbon emissions. After 1985, emissions again exceeded the 1979 level; each year's release during the latter half of the 1980s was 0.1 to 0.2 GtC above that for the previous year (Boden *et al.*, 1991).
- Direct measurements and ice core data have revealed a general decrease in atmospheric levels of <sup>13</sup>C relative to <sup>12</sup>C by about 1‰ over the last century (Friedli *et al.*, 1986; Keeling *et al.*, 1989a; Leuenberger *et al.*, 1992). This decrease is expected from the addition of fossil and/or terrestrial biospheric carbon, both of which are poor in <sup>13</sup>C relative to the atmosphere. In contrast, the atmospheric <sup>13</sup>C/<sup>12</sup>C ratio remained nearly constant from 1988 to 1993. This constant ratio must reflect changes in the fluxes between the ocean, terrestrial biosphere, and atmosphere which currently are not quantified. The atmospheric record of <sup>13</sup>C shows a decrease of ~0.4‰ between the first measurements in 1978 and the present time. The ice core record provides information on the changing <sup>13</sup>C levels over the past few hundred years (Friedli *et al.*, 1986; Leuenberger *et al.*, 1992).
- Relative to the long-term average rate of atmospheric CO<sub>2</sub> concentration increase (~1.5 ppmv/yr), the years 1988 to 1989 had relatively high CO<sub>2</sub> concentration growth rates (~2.0 ppmv/yr) while subsequent years (1991 to 1992) had very low growth rates (0.5 ppmv/yr) (Boden *et al.*, 1991). The magnitude of the 1988 to 1989 anomaly depends on what is defined as "normal" for the long-term trend. At Mauna Loa, Hawaii, the 1988 to 1989 increase was similar to a variation which occurred in 1973 to 1974 (see Figure 1.2). The subsequent decrease exceeds any previous anomaly since the Mauna Loa record began in 1958. Data for 1993 indicate a higher growth rate than that for 1991 to 1992.
- The CO<sub>2</sub> record exhibits a seasonal cycle, with small peak-to-peak amplitude (about 1 ppmv) in the Southern Hemisphere but increasing northward to about 15 ppmv in the boreal forest zone (55–65°N). This cycle is mainly caused by the seasonal uptake and release of atmospheric CO<sub>2</sub> by terrestrial ecosystems. Part of the seasonal signal is driven by oceanic processes (Heimann *et al.*, 1989). The amplitude of the seasonal atmospheric CO<sub>2</sub> cycle varies with time. For instance, at Mauna Loa (see Figure 1.3; Keeling *et al.*, 1989a), it was roughly constant at 5.2 ppmv (peak-to-trough) from the beginning of the Mauna Loa measurements in 1958 until the mid-1970s. It then increased over the late 1970s to reach 5.8 ppmv for most of the 1980s. The most recent data indicate a further increase. Because the trend is not well-correlated with the CO<sub>2</sub> concentration increase (Thompson *et al.*, 1986; Enting, 1987; Manning, 1993), it provides at best weak evidence for CO<sub>2</sub> fertilisation of terrestrial vegetation, in contrast with interpretations that claimed strong evidence (e.g., Idso and Kimball, 1993). The variation in amplitude does indicate changes in terrestrial metabolism, but not necessarily increased photosynthesis or storage.

riods of millennia in which large carbon exchanges were in near balance, implying nearly constant reservoir contents. Human activities have disturbed this balance through the use of fossil carbon and disruption of terrestrial ecosystems. The consequent accumulation of CO<sub>2</sub> in the atmosphere has caused a number of carbon cycle exchanges to become unbalanced. Fossil fuel burning and cement manufacture, together with forest harvest and other changes of land use, all transfer carbon (mainly as CO<sub>2</sub>) to the atmosphere. This anthropogenic carbon then cycles between the atmosphere, oceans, and the terrestrial biosphere. Because the cycling of carbon in the terrestrial and ocean biosphere occurs slowly, on time-scales of decades to millennia, the effect of additional fossil and biomass carbon injected into the atmosphere is a long-lasting disturbance of the carbon cycle. The relationships between concentration changes and emissions of CO<sub>2</sub> are examined through use of models that simulate, in a simplified manner, the major processes of the carbon cycle. The terrestrial and oceanic components of carbon cycle models vary in complexity from a few key equations to spatially explicit, detailed descriptions of ocean and terrestrial biology, chemistry, and transport processes. The simpler models, in general, are designed to reproduce observed behaviour while the more complex models

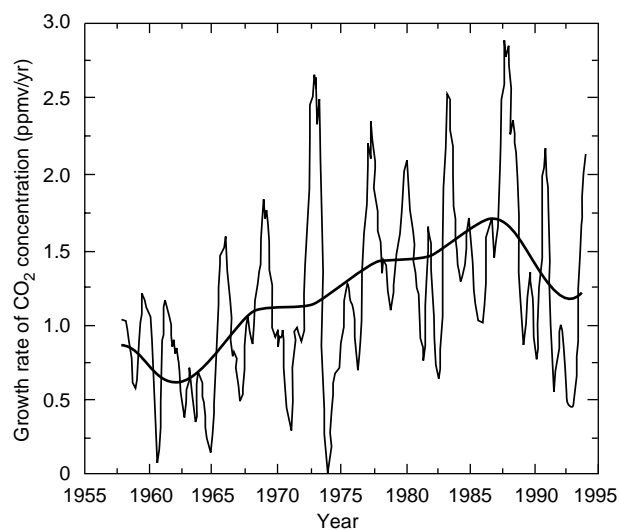
are aimed at incorporating the processes that cause the observed behaviour. The latter are thus potentially more likely to yield realistic projections of changes in storage under conditions different from the present (e.g., changing climate).

### 1.2 Time-Scales

Because CO<sub>2</sub> added to the atmosphere by anthropogenic processes is exchanged between reservoirs having a range of turnover times, it is not possible to define a single atmospheric "lifetime." This is in contrast with other anthropogenic compounds such as N<sub>2</sub>O and the halogens that are destroyed chemically in the atmosphere. Because the time-scales involved in the CO<sub>2</sub> exchanges range from annual to millennial (thousands of years), the consequences of anthropogenic perturbations will be long-lived. In this regard, the "turnover time" of about 5 years for atmospheric CO<sub>2</sub> deduced from the rate of bomb <sup>14</sup>CO<sub>2</sub> removal is relevant to the initial response of the carbon system, but it does not characterise the much slower, long-term response of atmospheric concentrations to the anthropogenic perturbation.

The remainder of this chapter reviews what is known about the carbon cycle as a basis for understanding past changes and





**Figure 1.2:** The growth rate of CO<sub>2</sub> concentrations since 1958 (from the Mauna Loa record). The high growth rates of the late 1980s, the extremely low growth rates of the early 1990s, and the recent increase in the growth rate are all evident. The smooth curve shows the same data but filtered to suppress any variations on time-scales less than approximately 10 years.

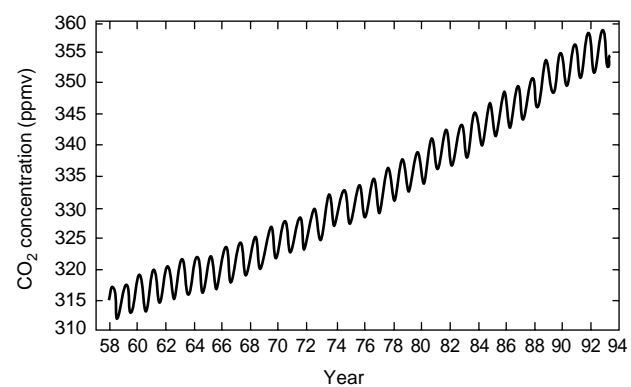
relationships between future emissions and concentrations. We do not attempt to make specific predictions of likely future changes in CO<sub>2</sub> concentration – rather, we assess the sensitivity of the system to particular scenarios of future emissions and concentrations. We also analyse key areas in which quantitative understanding is deficient. The final section of the chapter presents the results of a set of calculations relating future CO<sub>2</sub> concentrations and future CO<sub>2</sub> emissions. These calculations, produced by modelling groups from many countries following an agreed set of specifications, explored various aspects of uncertainty.

### 1.3 Past Record of Atmospheric CO<sub>2</sub>

#### 1.3.1 Atmospheric Measurements Since 1958

Precise, direct measurements of atmospheric CO<sub>2</sub> started in 1957 at the South Pole, and in 1958 at Mauna Loa, Hawaii. At this time the atmospheric concentration was about 315 ppmv and the rate of increase was ~0.6 ppmv/yr. The growth rate of atmospheric concentrations at Mauna Loa has generally been increasing since 1958. It averaged 0.83 ppmv/yr during the 1960s, 1.28 ppmv/yr during the 1970s, and 1.53 ppmv/yr during the 1980s. In 1992, the atmospheric level of CO<sub>2</sub> at Mauna Loa was 355 ppmv (Figure 1.3) and the growth rate fell to 0.5 ppmv/yr (see “Recent Anomalies” list). Data from the Mauna Loa station are close to, but not the same as, the global mean.

Atmospheric concentrations of CO<sub>2</sub> have been monitored for shorter periods at a large number of atmospheric stations around the world (e.g., Boden *et al.*, 1991). Measurement sites are distributed globally and include sites in Antarctica, Australia, several maritime islands, and high northern latitude sites,

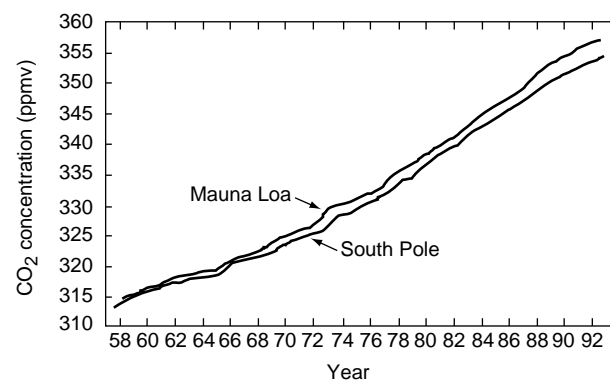


**Figure 1.3:** CO<sub>2</sub> concentrations measured at Mauna Loa, Hawaii since 1958 showing trends and seasonal cycle.

but, at present, nowhere on the continents of Africa or South America. The reliability and high precision of the post-1957 record is guaranteed by comparing the measured concentration of CO<sub>2</sub> in air with the concentration of reference gas mixtures calibrated by a constant volume column manometer. The increase shown by the atmospheric record since 1957 can be attributed largely to anthropogenic emissions of CO<sub>2</sub>, although considerable uncertainty exists as to the mechanisms involved. The record itself provides important insights that support anthropogenic emissions as a source of the observed increase. For example, when seasonal and short-term interannual variations in concentrations are neglected, the rise in atmospheric CO<sub>2</sub> is about 50% of anthropogenic emissions (Keeling *et al.*, 1989b) with the inter-hemispheric difference growing in parallel to the growth of fossil emissions (Keeling *et al.*, 1989a; Siegenthaler and Sarmiento, 1993; Figure 1.4).

#### 1.3.2 Pre-1958 Atmospheric Measurements and CO<sub>2</sub>-Ice Core Record over the Last Millennium

While several sets of relatively precise atmospheric measurements of CO<sub>2</sub> were carried out as early as the 1870s (e.g., Brown and Escombe, 1905), they did not allow assessment of concentration trends, as they were neither adequately calibrated nor temporally continuous.



**Figure 1.4:** Trends in CO<sub>2</sub> concentration and the growing difference in concentration between the Northern and Southern Hemispheres.

Measurements of CO<sub>2</sub> concentration from air extracted from polar ice cores are presently the best means to extend the CO<sub>2</sub> record through the geologically recent past. The transformation of snow into ice traps air bubbles which are used to determine the CO<sub>2</sub> concentration. Providing certain conditions are met, which include no fracturing of the ice samples, absence of seasonal melting at the surface, no chemical alteration of the initial concentrations, and appropriate gas extraction methods, the ice record provides reliable information on past atmospheric CO<sub>2</sub> concentrations (Raynaud *et al.*, 1993). It has been suggested that, under certain meteorological circumstances, the CO<sub>2</sub> data from Greenland ice cores may be contaminated, apparently influenced by varying levels of carbonate dust interacting with acid (Delmas, 1993) or organic matter deposition onto the ice sheet. Antarctic ice, however, has been uniformly acidic throughout the complete range of climate regimes of the last climatic cycle, and the available evidence suggests that these data are reliable throughout the entire record. Data from appropriate sites with unfractured ice are reliable to within  $\pm 3$ -5 ppmv (Raynaud *et al.*, 1993). The recent ice core record is validated by comparison with direct atmospheric measurements (Nefel *et al.*, 1985; Friedli *et al.*, 1986; Keeling *et al.*, 1989a).

Several high resolution Antarctic ice cores have recently become available in addition to the Siple core (Nefel *et al.*, 1985; Friedli *et al.*, 1986) for documenting both the "industrial era" CO<sub>2</sub> levels and the pre-industrial levels over the last millennium (Figure 1.5). The main results are:

- The ice core record can be used in combination with the direct atmospheric record to estimate, in conjunction with an oceanic model, the net changes in CO<sub>2</sub> flux between terrestrial ecosystems and the atmosphere (Siegenthaler and Oeschger, 1987).

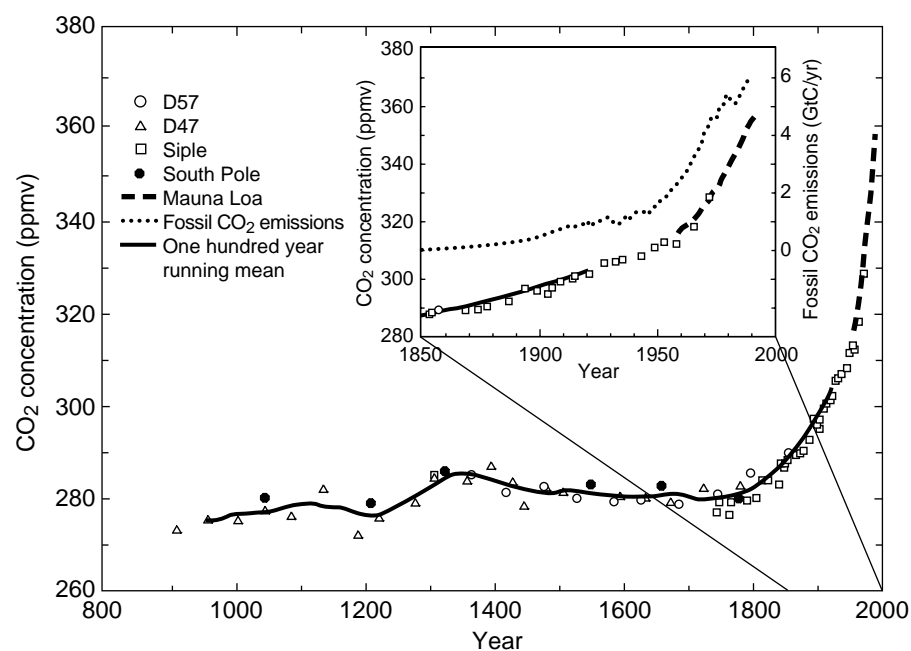
- The pre-industrial level over the last 1000 years shows fluctuations up to 10 ppmv around an average value of 280 ppmv. The largest of these, which occurred roughly between AD 1200 and 1400, was small compared to the 75 ppmv increase during the industrial era (Barnola *et al.*, in press, and Figure 1.5). Short-term climatic variability is believed to have caused the pre-industrial fluctuations through effects on oceanic and/or terrestrial ecosystems.

Finally, an important indicator of anthropogenically induced atmospheric change is provided by the <sup>14</sup>C levels preserved in materials such as tree rings and corals. The <sup>14</sup>C concentration measured in tree rings decreased by about 2% during the period 1800 to 1950. This isotopic decrease, known as the Suess effect (Suess, 1955), provides one of the most clear demonstrations that the increase in atmospheric CO<sub>2</sub> is due to fossil inputs.

### 1.3.3 The CO<sub>2</sub> Record over the Last Climatic Cycle

Although the magnitude and rate of climate changes observed in the palaeo-record covering the last climatic cycle may differ from those involved in any future greenhouse warming, these records provide an important perspective for recent and potential future changes. The glacial-interglacial amplitudes of temperature change are of similar order to the high estimate of equilibrium temperature shifts predicted for a doubling of CO<sub>2</sub> levels (Mitchell *et al.*, 1990), although the shifts of the past took thousands of years.

The close association between CO<sub>2</sub> and temperature changes during glacial-interglacial transitions was first revealed by data from the ice core record. Samples from Greenland and Antarctica representing the last glacial maximum



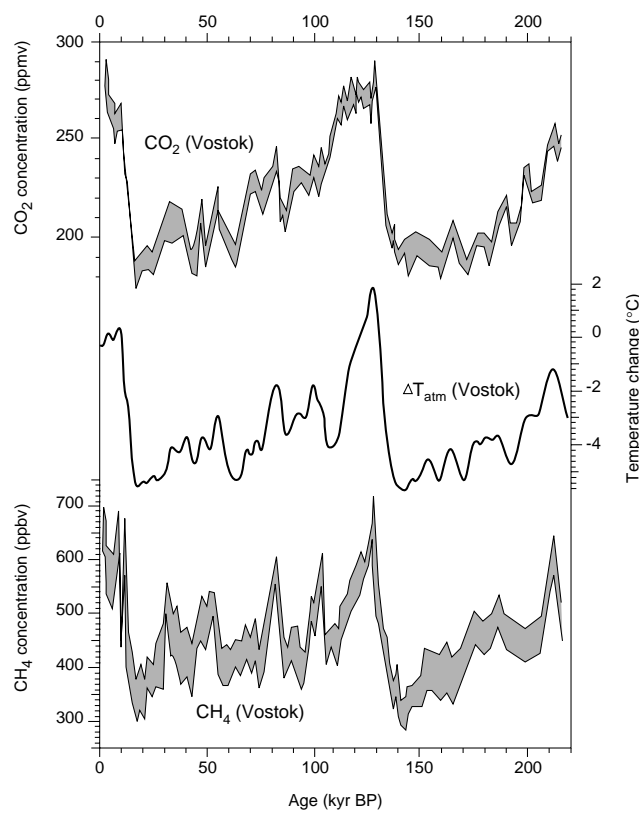
**Figure 1.5:** CO<sub>2</sub> concentrations over the past 1000 years from the recent ice core record and (since 1958) from the Mauna Loa measurement site. The inset shows the period from 1850 in more detail including CO<sub>2</sub> emissions from fossil fuel. Data sources: D47 and D57 (Barnola *et al.*, in press); Siple (Nefel *et al.*, 1985 and Friedli *et al.*, 1986) and South Pole (Siegenthaler *et al.*, 1988). The smooth curve is based on a 100yr running mean. All ice core measurements were taken in Antarctica.

(about 18,000 years before present) indicate that CO<sub>2</sub> concentrations at that time were 190–200 ppmv, i.e., about 70% of the pre-industrial level (Delmas *et al.*, 1980; Neftel *et al.*, 1982). Thus, an increase of about 80 ppmv occurred in parallel with the warming starting at the end of the glacial period, when the estimated glacial-interglacial rise of the mean surface temperature of the Earth rose by 4°C over about 10,000 years (Crowley and North, 1991). These discoveries have since been confirmed by detailed measurements of the Antarctic Byrd core for the 8,000 to 50,000 year BP period (Neftel *et al.*, 1988). Analyses of ice cores from Vostok, Antarctica, have provided new data on natural variations of CO<sub>2</sub> levels over the last 220,000 years (Barnola *et al.*, 1987, 1991; Jouzel *et al.*, 1993). The record shows a marked correlation between Antarctic temperature, as deduced from the isotopic composition of the ice, and the CO<sub>2</sub> profile (Figure 1.6).

Clear correlations between CO<sub>2</sub> and global mean temperature are evident in much of the glacial-interglacial palaeorecord. This relationship of CO<sub>2</sub> concentration and temperature may carry forward into the future, possibly causing a significant positive climatic feedback on CO<sub>2</sub> fluxes. Information about leads and lags between climatic variations and changes in radiatively active trace gas concentrations is contained in polar ice and deep sea sediment records (Raynaud and Siegenthaler, 1993). There is no evidence that CO<sub>2</sub> changes ever significantly (> 1 kyr) preceded the Antarctic

temperature signal. In contrast, CO<sub>2</sub> changes clearly lag behind the Antarctic cooling at the end of the last interglacial. As temperature changes in the South generally preceded temperature changes in the North (CLIMAP Project Members, 1984), it cannot be assumed that CO<sub>2</sub> changes never led northern temperature changes. Comparison of atmospheric CO<sub>2</sub> concentration and continental ice volume suggests that CO<sub>2</sub> started to change ahead of any significant melting of continental ice. It is possible that CO<sub>2</sub> changes may have been caused by changes in climate, and that CO<sub>2</sub> and other trace gases acted to amplify palaeoclimatic changes.

Changes in climate on time-scales of decades to centuries have occurred in the past. The question remains whether these changes have been accompanied by changes in greenhouse trace gas concentrations. The Greenland ice cores (Johnsen *et al.*, 1992; Grootes *et al.*, 1993) show that during the last ice age and the last glacial-interglacial transition, there was a series of rapid (over decades to a century) and apparently large climatic changes in the North Atlantic region (~5 to 7°C in Central Greenland: Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993). These changes may have been global in scale: the methane record suggests the potential for parallel changes in the tropics (Chappellaz *et al.*, 1993). Evidence for rapid climate oscillations during the last interglacial has also recently been reported (GRIP Project Members, 1993). However, because the details were not confirmed by a second core retrieved from the same area (Grootes *et al.*, 1993), the possibility that these features were caused by ice-flow perturbations cannot be discounted. The Dye 3 ice core from Greenland indicates that CO<sub>2</sub> concentration shifts of ~50 ppmv occurred within less than 100 years during the last glacial period (Staufner *et al.*, 1984). These changes in CO<sub>2</sub> were paralleled by abrupt and drastic climatic events in this region. Such large and rapid CO<sub>2</sub> changes have not been identified in Antarctica, even after accounting for the lower time resolution of the Antarctic core (Neftel *et al.*, 1988). It is possible that impurities have introduced artefacts into the Greenland CO<sub>2</sub> record (Delmas, 1993; Barnola *et al.*, in press), and that these features do not represent real atmospheric CO<sub>2</sub> changes. Nevertheless, taken together with independent support from isotopic studies of mosses (White *et al.*, 1994), rapid CO<sub>2</sub> events recorded during the past cannot be disregarded.



**Figure 1.6:** Temperature anomalies and methane and CO<sub>2</sub> concentrations over the past 220,000 years as derived from the ice core record at Vostok, Antarctica.

## 1.4 The Anthropogenic Carbon Budget

### 1.4.1 Introduction

The phrase “carbon budget” refers to the balance between sources and sinks of CO<sub>2</sub> in the atmosphere, expressed in terms of anthropogenic emissions and fluxes between the main reservoirs – the oceans, the atmosphere, the terrestrial carbon pool – and the build-up of CO<sub>2</sub> in the atmosphere. Because of the relative stability of atmospheric CO<sub>2</sub> concentrations over several thousand years prior to AD 1800, it is assumed that the net fluxes among carbon reservoirs were close to zero prior to anthropogenic disturbance. The data described in Section 1.3

provide the essential background for understanding the carbon budget's changes over time. Several approaches are used to quantify the components of this budget (other than atmospheric mass build-up, which can be measured directly), often in combination.

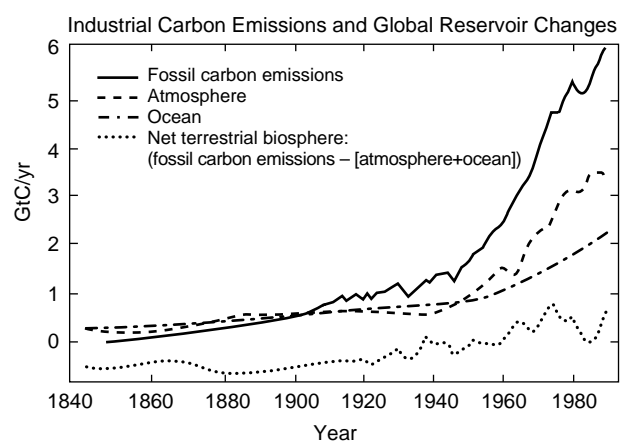
- direct determination of rates of change of the carbon content in atmospheric, oceanic, and terrestrial carbon pools, either by observations of local inventory changes or by local flux measurements, extrapolated globally;
- indirect assessment of the atmosphere-ocean and atmosphere-terrestrial biosphere fluxes by means of carbon cycle model simulations, either calibrated or partially validated using analogue tracers of CO<sub>2</sub>, such as bomb radiocarbon or tritium, or with use of chlorofluorocarbons;
- interpretation of tracers or other substances that are coupled to the carbon cycle (<sup>14</sup>C/<sup>12</sup>C and <sup>13</sup>C/<sup>12</sup>C ratios and atmospheric oxygen).

The heterogeneity of some aspects of the oceanic and terrestrial carbon systems makes reliable extrapolation of flux measurements to the entire globe dependent on high resolution geographic information and accurate modelling of processes. Because of this, method (a) is used only for atmospheric carbon and for estimating effects of land use, and estimates of other carbon reservoirs are based primarily on methods (b) and (c).

#### 1.4.2 Methods for Calculating the Carbon Budget

##### 1.4.2.1 Classical Approaches

The calculated ocean uptake rate, together with the estimated fossil emissions and the observed atmospheric inventory change, allows inference of the net terrestrial biospheric balance. Figure 1.7 shows the results of a time-series rate of change calculation for the atmospheric carbon mass, the oceanic com-



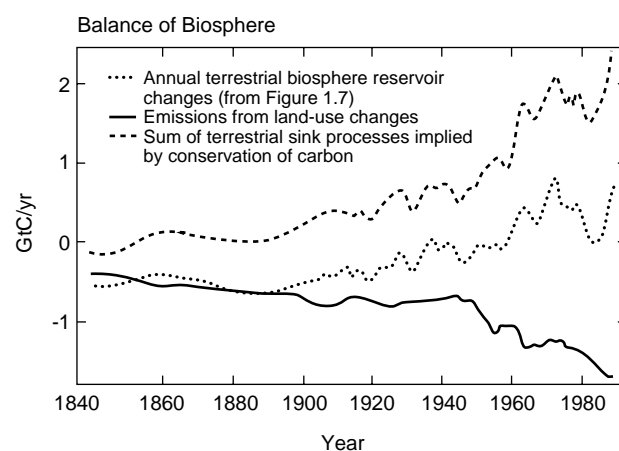
**Figure 1.7:** Fossil carbon emissions (based on statistics of fossil fuel and cement production), and representative calculations of global reservoir changes: atmosphere (deduced from direct observations and ice core measurements), ocean (calculated with the GFDL ocean carbon model), and net terrestrial biosphere (calculated as remaining imbalance). The calculation implies that the terrestrial biosphere represented a net source to the atmosphere prior to 1940 (negative values) and a net sink since about 1960.

ponent, fossil emissions, and the residual of the first three terms. In this calculation, the time history of fossil plus cement emissions was deduced from statistics (Keeling, 1973; Marland and Rotty, 1984; WEC, 1993; Andres *et al.*, 1994), atmospheric accumulation was determined from the observational and ice core record (e.g., Barnola *et al.*, 1991; Boden *et al.*, 1991), and ocean uptake was modelled with the GFDL ocean general circulation model (Sarmiento *et al.*, 1992). Ocean carbon uptake was determined by forcing the ocean chemistry with the time history of atmospheric concentrations to obtain uptake as a function of the non-linear chemistry of carbon dioxide in the ocean. This calculation illustrates the classic means by which the carbon budget is calculated. Figure 1.8 shows the balance of the terrestrial biosphere, now introducing the time history of land use effects together with a calculation of the inferred terrestrial sink over time. This calculation illustrates the classic estimation of a “missing sink,” in which a residual sink arises because the sum of anthropogenic emissions (fossil, cement, changing land use) is greater than the sum of ocean uptake and atmospheric accumulation. Note that this approach is quite different from the estimation of sources and sinks from the spatial distribution of atmospheric concentrations and isotopic composition, discussed in Section 1.4.2.3 (Keeling *et al.*, 1989b; Tans *et al.*, 1990; Enting and Mansbridge, 1991).

##### 1.4.2.2 New Approaches: Budget Assessment Based on Observations of <sup>13</sup>C/<sup>12</sup>C and O<sub>2</sub>/N<sub>2</sub> Ratios

Several promising approaches have recently been proposed to assess the current global carbon budget with less dependence on models. Included are observations of <sup>13</sup>C/<sup>12</sup>C and O<sub>2</sub>/N<sub>2</sub>, both of which are strongly influenced by the anthropogenic perturbation of the carbon cycle.

The methods based on <sup>13</sup>C exploit the fact that the <sup>13</sup>C/<sup>12</sup>C ratio in fossil fuels and terrestrial biomass is less than that in the atmosphere. There has been a decline in the <sup>13</sup>C/<sup>12</sup>C ratio of



**Figure 1.8:** The carbon balance of the terrestrial biosphere. Annual terrestrial biosphere reservoir changes (from Figure 1.7), land use flux (plotted negative because it represents a loss of biospheric carbon) and the sum of the terrestrial sink processes (e.g., Northern Hemisphere regrowth, CO<sub>2</sub> and nitrogen fertilization, climate effects) as implied by conservation of carbon mass.

atmospheric CO<sub>2</sub> over the last century (Keeling *et al.*, 1989a). This atmospheric <sup>13</sup>C/<sup>12</sup>C ratio change propagates through the global carbon cycle, causing isotopic ratio changes in the ocean and the terrestrial carbon reservoirs.

Quay *et al.* (1992) proposed a method to determine the global budget from repeated measurements of vertical profiles of the <sup>13</sup>C/<sup>12</sup>C ratio in oceanic dissolved inorganic carbon. In principle, these observations allow us to determine the rate of change of the oceanic <sup>13</sup>C/<sup>12</sup>C ratio which, together with the observed changes in the atmospheric <sup>13</sup>C/<sup>12</sup>C ratio, provide another constraint on the global carbon balance. The ratio permits discrimination between oceanic and terrestrial biospheric sinks because terrestrial uptake (photosynthesis) discriminates against the heavy isotope (<sup>13</sup>C) much more strongly than does ocean uptake. A preliminary analysis based on a data set from seven stations in the Pacific Ocean (sampled in the early 1970s and again in 1990) yielded a mean oceanic sink of  $2.1 \pm 0.8$  GtC/yr (Quay *et al.*, 1992).

Additional information for constraining the oceanic carbon budget may be provided by the <sup>13</sup>C/<sup>12</sup>C isotopic disequilibrium between the air and the sea (Tans *et al.*, 1993). The disequilibrium reflects the isotopic adjustment of the ocean to the atmospheric perturbation and can be used to assess the atmospheric balances of <sup>13</sup>C and CO<sub>2</sub>, thereby discriminating between oceanic and biospheric components. Tans *et al.* (1993) used this method to estimate the net air-sea flux of CO<sub>2</sub> for the period 1970 to 1990, which they found to be less than 0.4 GtC/yr, with unspecified error ranges.

Relative to achievable measurement precision, the anthropogenic perturbation has a greater effect on isotopic composition of dissolved organic carbon than on its concentration, because the former is not affected by chemical buffering reactions. The required analytical quality of the isotopic measurements is still high. The uncertainty range of the current oceanic sink, as estimated by the methods employing isotope measurements, may be reduced substantially if an extended measurement programme is vigorously pursued (Heimann and Maier-Reimer, submitted).

The trend in atmospheric oxygen, as revealed by measurements of the oxygen to nitrogen ratio, can also be used to assess the global carbon budget (Keeling and Shertz, 1992; Bender *et al.*, 1994). Oxygen in many respects is complementary to carbon. It is consumed during combustion and respiration and is released during biotic carbon uptake via photosynthesis. The crucial difference compared to carbon is that, owing to low solubility of O<sub>2</sub> in water, the magnitude of the oceanic oxygen pool is negligible and can be ignored in the global oxygen balance. Therefore, measurements of the temporal trend in atmospheric oxygen together with the known O:C stoichiometric relations during combustion, respiration, and photosynthesis, permit establishment of global oxygen and carbon balances, although interannual variations in marine photosynthesis and respiration may complicate the interpretation of O:C data (Keeling and Severinghaus, 1994). Data are limited, as measurements of the oxygen/nitrogen ratio have been reported for only two years. Within 10 years it is possible that an accurate, model-independent estimate of the oceanic carbon budget will be achieved by this method. The atmospheric oxygen trend during

the last several decades can also be reconstructed from measurements of CO<sub>2</sub> and the O<sub>2</sub>/N<sub>2</sub> ratio in glacial surface layers (firn) as a function of depth. Preliminary results indicate past ocean CO<sub>2</sub> uptake rates were large (> 3 GtC/yr), but associated uncertainty ranges are also large (Bender *et al.*, 1994).

Despite high uncertainty, these new approaches provide a different, model-independent means of assessing the global carbon budget. They provide only estimates of the oceanic uptake rate, however. The information gained can help reduce the uncertainty in the net terrestrial sink quantified by differencing, but to partition the uptake into land use fluxes and terrestrial sinks requires other approaches.

#### 1.4.2.3 Constraints from Spatial Distributions

The different spatial patterns of sources and sinks of atmospheric CO<sub>2</sub> create gradients in the concentration that vary inversely with the strength of atmospheric transport. These spatial distributions of concentration can be interpreted quantitatively using models of atmospheric transport to match observed concentration distributions. This process is known as "inversion."

The problem of deducing sources and sinks of trace gases from observations of surface concentrations presents a considerable technical challenge. This is partly because the small-scale features of the spatial distribution of sources and sinks are blurred by atmospheric mixing. The "reconstruction" process amplifies the small-scale details, but the errors and uncertainties introduced along the way are similarly amplified. Hence, there is a trade-off between the spatial resolution sought and the accuracy of the estimates obtained. Consequently, only a small number of independent source/sink components can be reliably determined from the data, and, in general, only the largest spatial scales can be resolved (Enting, 1993). This explains, in part, the range of interpretations obtained in different studies. The uncertainty analysis of Enting *et al.* (1994a) indicates that such analyses cannot, on their own, estimate global totals of net fluxes more accurately than the classical approach to carbon budget analysis. Thus, at present, the constraints from the spatial distribution of CO<sub>2</sub> act mainly as consistency checks with budgets derived from analyses of individual budget components.

Synthesis inversions include additional information beyond atmospheric concentration and transport, and can suppress unwanted amplification of small-scale variations either by restricting the number of source components considered, or through the use of additional constraints such as incorporation of independent estimates of source strengths. Inversions using two-dimensional atmospheric transport models have been presented by Tans *et al.* (1989) and Enting and Mansbridge (1989, 1991). Three-dimensional modelling studies of source and sink distributions have been presented by Keeling *et al.* (1989b), Tans *et al.* (1990), and Enting *et al.* (1994a).

The main results that have been obtained from inverse calculations using atmospheric transport modelling are:

- *Northern Hemisphere source.* Inversion calculations reveal a strong Northern Hemisphere release, as expected from increased use of fossil fuels. This source is so

strong that it tends to obscure other details of the source-sink distribution and is additional confirmation of the role of fossil fuel emissions in atmospheric concentration change.

- *Northern Hemisphere sink.* After subtracting the fossil source, there is a strong net sink in the Northern Hemisphere which may involve both marine and terrestrial components (e.g., Tans *et al.*, 1990). The partitioning remains controversial (Keeling *et al.*, 1989b; Ciais *et al.*, submitted), but recent isotopic analyses suggest an appreciable terrestrial component (Ciais *et al.*, submitted). A Northern Hemisphere sink, as estimated by inverse calculations, is consistent with results from terrestrial studies that suggest sinks are the result of changing land use (e.g., Dixon *et al.*, 1994) and nitrogen deposition (largely confined to the industrialised Northern Hemisphere) (e.g., Schindler and Bayley, 1993).
- *Indirect evidence of the existence of a tropical biotic sink.* The net equatorial source from oceanic outgassing and changes in tropical land use is smaller than expected based on other estimates (see Sections 1.4.3.3 and 1.4.3.4), consistent with a role for nutrient and CO<sub>2</sub> fertilisation and forest regrowth.
- *Southern Ocean source.* Most studies indicate a source (probably oceanic) at high southern latitudes.
- *Southern Hemisphere sink.* The net Southern Hemisphere sink (which is presumably oceanic, given the small proportion of Southern Hemisphere land) is weaker than expected in comparison with northern ocean sinks.

### 1.4.3 Sources and Sinks of Anthropogenic CO<sub>2</sub>

#### 1.4.3.1 Fossil Carbon Emissions

The dominant anthropogenic CO<sub>2</sub> source is that generated by the use of fossil fuels (coal, oil, natural gas, etc.) and production of cement. The total emissions of CO<sub>2</sub> from the use of fossil carbon can be estimated based on documented statistics of fossil fuel and cement production (Keeling, 1973; Marland and Rotty, 1984; WEC, 1993; Andres *et al.*, 1994). Average (1980 to 1989) fossil emissions were estimated to be 5.5 GtC/yr (Andres *et al.*, 1994). During 1991, the reported emissions totalled 6.2 GtC (Andres *et al.*, 1994). The cumulative input since the beginning of the industrial revolution (1751 to 1991) is estimated to be approximately 230 GtC (Andres *et al.*, 1994). Uncertainties associated with these estimates are less than 10% for the decade of the 1980s (at the 90% confidence level) based on the methods presented in Marland and Rotty (1984).

#### 1.4.3.2 Atmospheric Increase

The globally averaged CO<sub>2</sub> concentration, as determined through analysis of NOAA/CMDL data (Boden *et al.*, 1991; Conway *et al.*, 1994), increased by  $1.53 \pm 0.1$  ppmv/yr over the period 1980 to 1989. This corresponds to an annual aver-

age rate of change in atmospheric carbon of  $3.2 \pm 0.2$  GtC/yr. Other carbon-containing compounds like methane, carbon monoxide, and larger hydrocarbons contain ~1% of the carbon stored in the atmosphere and can therefore be neglected in the atmospheric carbon budget.

#### 1.4.3.3 Ocean Exchanges

The ocean contains more than 50 times as much carbon as the atmosphere. Over 95% of the oceanic carbon is in the form of inorganic dissolved carbon (bicarbonate and carbonate ions); the remainder is composed of various forms of organic carbon (living organic matter, particulate and dissolved organic carbon) (Druffel *et al.*, 1992).

The role of the oceans in the global carbon cycle is twofold: first, it represents a passive reservoir which absorbs excess atmospheric CO<sub>2</sub>. It is this role that is discussed in this section. Second, changes in the physical state of the ocean (temperature, circulation) and the marine biota may affect the rate of air-sea exchange, and thus future atmospheric CO<sub>2</sub>. This second role, subsumed under “feedbacks,” is addressed in Section 1.5.3.

The oceanic uptake of excess CO<sub>2</sub> proceeds by (1) transfer of the CO<sub>2</sub> gas through the air-sea interface, (2) chemical interactions with the oceanic dissolved inorganic carbon, and (3) transport into the thermocline and deep waters by means of water mass transport and mixing processes. Though there are large geographical and seasonal variations of the surface ocean partial pressure of CO<sub>2</sub>, averaged globally and annually, the surface water value is close to that for equilibrium with the atmosphere. Therefore, processes (2) and (3) are the main factors limiting the capacity of the ocean to serve as a sink on decadal and centennial time-scales. The chemical buffering reactions between dissolved CO<sub>2</sub> and the HCO<sub>3</sub><sup>-</sup> and CO<sub>3</sub><sup>2-</sup> ions reduce the rate of oceanic CO<sub>2</sub> uptake. At equilibrium, an atmospheric increase in CO<sub>2</sub> concentration of 10% is associated with an oceanic increase of dissolved inorganic carbon of merely 1%. This potential uptake occurs only in those parts of the ocean that are mixed with surface waters on decadal time-scales. Therefore, on time-scales of decades to centuries, the ocean is not as large a sink for excess CO<sub>2</sub> as it might seem from comparison of the relative sizes of the main carbon reservoirs (Figure 1.1). While carbon chemistry is known in sufficient detail to perform accurate calculations, oceanic transport and mixing processes remain the primary uncertainties in the determination of the oceanic uptake of excess CO<sub>2</sub>.

The marine biota, if in steady state, are believed to play a minor role, if any at all, in the uptake of excess anthropogenic CO<sub>2</sub>. The marine biota, however, play a crucial role in maintaining the steady-state level of atmospheric CO<sub>2</sub>. About three-quarters of the vertical gradient in dissolved inorganic carbon is generated by the export of newly produced carbon from the surface ocean and its regeneration at depth (a process referred to as “biological pump”). In the open ocean, however, this process is believed to be limited by the availability of nutrients, light, or by phytoplankton population control via grazing, and not by the abundance of carbon (Falkowski and Wilson, 1992).

Therefore, a direct effect of increased dissolved inorganic carbon (less than 2.5% since pre-industrial times) on carbon fixation and export is unlikely, although a recent study by Riebesell *et al.* (1993) suggested that under particular conditions the rate of photosynthesis and hence phytoplankton growth might indeed be limited by the availability of CO<sub>2</sub> as a dissolved gas. The global significance of this effect, however, remains to be assessed. (See Section 1.5.3 for a discussion of the potential indirect effects of increased dissolved inorganic carbon.)

Flux of carbon from the terrestrial biosphere to the oceans takes place via river transport. Global river discharge of carbon in organic and inorganic forms may be ~1.2–1.4 GtC/yr (Schlesinger and Melack, 1981; Degens *et al.*, 1991; Maybeck, 1993). A substantial fraction of this transport (up to 0.8 GtC/yr), however, reflects the natural geochemical cycling of carbon and thus does not affect the global budget of the anthropogenic CO<sub>2</sub> perturbation (Sarmiento and Sundquist, 1992). Furthermore, the anthropogenically induced river carbon fluxes reflect, to a large extent, increased soil erosion and not a removal of excess atmospheric CO<sub>2</sub>.

The role of coastal seas in the global carbon budget is poorly understood. Up to 30% of total ocean productivity is attributed to marine productivity in the coastal seas, which compose only ~8% of the oceanic surface area. Here, discharge of excess nutrients by rivers might have significantly stimulated carbon fixation (up to 0.5–1.0 GtC/yr). At present, however, it is not known how much of this excess organic carbon is simply reoxidised, and how much is permanently sequestered by export to the deep ocean, or in sediments on the shelves and shallow seas. Because of the limited surface area, a burial rate significantly exceeding 0.5 GtC/yr is not very likely, as it would require all coastal seas to be under-saturated in partial pressure of CO<sub>2</sub> by more than 50  $\mu$ atm on annual average, in order to supply the carbon from the atmosphere. Such under-saturations have been documented, e.g., in the North Sea (Kempe and Pegler, 1991), but these measurements are unlikely to be representative of all coastal oceans. Based on the above considerations, the role of the coastal ocean is judged most likely to be small, but, at present, cannot be accurately assessed and so is neglected in the budget presented in this chapter (Table 1.3).

Exchanges of carbon between the atmosphere and the oceans (net air-sea fluxes) can be deduced from measurements of the partial pressure difference of CO<sub>2</sub> between the air and surface waters. This calculation also requires knowledge of the local gas-exchange coefficient, which is a relatively poorly known quantity (Watson, 1993). Furthermore, while the globally and seasonally averaged partial pressure of surface waters is close to the value for equilibrium with the atmosphere, large geographical and seasonal variations exist, induced both by physical processes (upwelling, vertical mixing, sea surface temperature fluctuations) and by the activity of the marine biota. Representative estimates of the seasonally and regionally averaged net air-sea carbon transfer thus necessitates sampling with high spatial and temporal resolution (Garçon *et al.*, 1992). Estimates of the regional net air-sea carbon fluxes have been obtained, albeit with considerable error margins. However, the global oceanic carbon balance is more difficult to deduce by this method, as it represents a small residual computed

from summing up the relatively large emissions from super-saturated and uptake in under-saturated regions (Tans *et al.*, 1990; Takahashi *et al.*, 1993; Wong *et al.*, 1993; Fung and Takahashi, 1994). The World Ocean Circulation Experiment (WOCE) survey has nevertheless demonstrated the feasibility of direct measurement programmes, and the importance of measurements as a cross-check for other approaches should not be underrated.

The oceanic contribution to the global carbon budget can also be assessed by direct observations of changes in the oceanic carbon content. This approach also suffers from the problem of determining a very small signal against large spatial and temporal background variability. Model estimates of the rate of change induced by the anthropogenic perturbation are of the order of 1 part in 2000/yr. Therefore, on a 10-year time-scale, variations in dissolved inorganic carbon would have to be measured with an accuracy of better than 1% in order to determine the carbon balance in a particular oceanic region. While such accuracy can be obtained, a substantial sampling effort is required (Keeling, 1993). Determination of the global oceanic budget by this approach does not appear feasible in the near future. However, repeated observational surveys might reveal regional carbon inventory changes, and thus might provide a cross-check for other approaches.

### Model Results

The present-day oceanic uptake of excess CO<sub>2</sub> is estimated using ocean carbon model simulation experiments, with observed atmospheric CO<sub>2</sub> concentration as a prescribed boundary condition. It is necessary to model the transport and mixing of the excess carbon perturbation only if the marine biota are assumed to remain on an annual average in a steady state, as excess CO<sub>2</sub> will disperse within the ocean like a passive tracer, independent of the background natural distribution of dissolved inorganic carbon. In particular, the perturbation excludes a natural cycle of carbon transported by rivers, out-gassed by the oceans and then taken up by the terrestrial biota.

Until recently, most models of the oceanic uptake of anthropogenic CO<sub>2</sub> consisted of a series of well-mixed or diffusive reservoirs (“boxes”) representing the major oceanic water masses, connected by exchange of water (Oeschger *et al.*, 1975). Global transport characteristics of these models were obtained from simulation studies of the oceanic penetration of bomb radiocarbon (<sup>14</sup>C) validated by comparison with observations (Broecker *et al.*, 1985). Bomb radiocarbon, however, does not accurately track the oceanic invasion of anthropogenic carbon, because oceanic concentration changes of carbon isotopes depend on the full time history of those inputs and <sup>14</sup>C entered the ocean with an atmospheric history different from that of natural CO<sub>2</sub>. Furthermore, CO<sub>2</sub> uptake depends on chemical interactions with dissolved inorganic C, while uptake of <sup>14</sup>C does not. Nevertheless, bomb radiocarbon provides a powerful constraint on ocean carbon models, a constraint which may be supplemented by analyses of other steady-state and transient tracers, such as halocarbons, tritium from nuclear weapon testing, and possibly <sup>13</sup>C.

Recently, three-dimensional oceanic general circulation models (OGCMs) have been used for modelling CO<sub>2</sub> uptake by the oceans (Maier-Reimer and Hasselmann, 1987; Sarmiento *et al.*, 1992; Orr, 1993). These models calculate oceanic circulation on the basis of the physics of fluid dynamics, and the few adjustable model parameters are primarily tuned to reproduce the relatively well-known large-scale patterns of ocean temperature and salinity. It is known that the OGCMs used in published global carbon cycle studies show significant and similar deficiencies (e.g., too weak a surface circulation, inaccurate deep convection, absence of high resolution features). Simulation of transient tracers, bomb radiocarbon in particular, provides an important validation test.

The average of modelled rates of oceanic carbon uptake is 2.0 GtC/yr over the decade 1980 to 1989 (Orr, 1993; Siegenthaler and Sarmiento, 1993). This value is corroborated by model experiments projecting future concentrations of atmospheric CO<sub>2</sub> (see Section 1.6, and Enting *et al.*, 1994b) and by a simulation with a newly-developed two-dimensional global ocean model (Table 1.1; Stocker *et al.*, 1994). The spread of the modelled uptake rates corresponds to a statistical uncertainty of only about  $\pm 0.5$  GtC/yr (at the 90% confidence level). However, recent assessments of the global radiocarbon balance (Broecker and Peng, 1994; Heshaimer *et al.*, 1994) suggest that the oceanic carbon uptake estimates of box models tuned by bomb radiocarbon might have to be revised downwards by up to 25%. Furthermore, the spread of the OGCM results listed in Table 1.1 might be fortuitously small in view of the similar deficiencies in these models. Based on these considerations we do not change the estimate of the uncertainty of the ocean uptake from the value of  $\pm 0.8$  GtC/yr as given by IPCC in 1990 (Watson *et al.*, 1990) and in 1992 (Watson *et al.*, 1992).

#### 1.4.3.4 Terrestrial Exchanges

In previous assessments (IPCC, 1990, 1992), the remaining intact terrestrial biosphere has often been assumed to be a significant sink for carbon dioxide, balancing or exceeding emissions derived from changing land use. The calculated imbalance, which has been a persistent feature of global carbon

cycle calculations (Broecker *et al.*, 1979), arises from the difference between measured atmospheric changes, statistically derived fossil fuel and cement emissions, modelled ocean uptake, and estimated emissions from changing land use. In this assessment, part of the previously calculated imbalance is accounted for by effects of changing land use in the middle and high latitudes ( $0.5 \pm 0.5$  GtC/yr: Table 1.3). Several processes may contribute to increased terrestrial carbon storage (Table 1.2), but the problem of detecting these increases is troublesome.

Difficulty in quantifying the role of the terrestrial biosphere in the global carbon cycle arises because of the complex biology underlying carbon storage, the great heterogeneity of vegetation and soils, and the effects of human land use and land management. We consider the following issues:

- *Deforestation and change in land use.* CO<sub>2</sub> is emitted to the atmosphere as a result of land use changes such as biomass burning and forest harvest through the oxidation of vegetation and soil carbon. These “emissions from changing land use” are currently largest in the tropics, but prior to the 1950s, the middle latitudes were a larger source than were the tropics.
- *Forest regrowth.* Carbon is absorbed by regrowing forests following harvest. This absorption is included as a factor in the calculation of net emissions from disturbed regions. In the tropics, emissions are thought to exceed the uptake by secondary growth following forest harvest, but uptake from regrowth in the middle and high latitudes apparently results in a net sink for atmospheric CO<sub>2</sub> (see below).
- *Fertilisation by carbon dioxide.* As atmospheric CO<sub>2</sub> increases, plants increase their uptake of carbon, potentially increasing carbon storage in the terrestrial biosphere. This mechanism has often been hypothesised to account for the calculated imbalance in the global carbon budget (the “missing sink”). There is strong physiological evidence for photosynthetic increase with CO<sub>2</sub> fertilisation (e.g., Woodward, 1992), but interspecific differences in the magnitude of effects on photosynthesis and growth range from marked enhancement to negli-

Table 1.1. Excess CO<sub>2</sub> uptake rates (average 1980 to 1989) calculated by various ocean carbon cycle models.

Model	Ocean uptake (GtC/yr)	Reference
<i>Bomb radiocarbon-based box models</i>		
Box-diffusion model	2.32	Siegenthaler & Sarmiento (1993)
HILDA model (“Bern model”)	2.15	Siegenthaler & Joos (1992)
<i>Three-dimensional ocean general circulation models</i>		
Hamburg Ocean Carbon Cycle Model (HAMOCC-3)	1.47	Maier-Reimer (1993)
GFDL Ocean General Circulation Model	1.81	Sarmiento <i>et al.</i> (1992)
LODYC Ocean General Circulation Model	2.10	Orr (1993)
<i>Two-dimensional ocean circulation models</i>		
<b>AVERAGE of all models</b>	<b>2.0</b>	Stocker <i>et al.</i> (1994)



Table 1.2. Processes leading to increased terrestrial storage and their magnitudes (GtC/yr): average for the 1980s.

Processes	
Mid-/high latitude forest regrowth	0.5 ± 0.5
CO <sub>2</sub> fertilisation	0.5-2.0
Nitrogen deposition	0.2-1.0
Climatic effects	0-1.0

gible and even negative responses. Considerable uncertainty exists as to how to translate laboratory and field results into a global estimate of changing biospheric carbon storage, as ecosystem feedbacks and constraints are numerous (see Section 1.5.2).

- **Nitrogen fertilisation.** Most terrestrial ecosystems are limited by nitrogen, as evidenced by increased carbon storage after nitrogen fertilisation. Additions of nitrogen to the terrestrial biosphere through both intentional fertilisation of agricultural land and deposition of nitrogen arising from fossil fuel combustion and other anthropogenic processes can result in increased terrestrial storage of carbon.
- **Climate.** The processes of terrestrial carbon uptake (photosynthesis) and release (respiration of vegetation and soils) are influenced by climate. Decadal time-scale variations in climate may have caused natural changes in carbon storage by the terrestrial biosphere, acting in conjunction with or even counter to the anthropogenic effects described above. This effect on storage is separate from the potential future changes in carbon storage which might arise from greenhouse gas-induced changes of climate.
- **Interactions.** The processes mentioned above are not independent, and their magnitudes are not necessarily additive. For example, much of the anthropogenic nitrogen has been deposited by the atmosphere on regrowing forests of the middle and high latitudes, thereby potentially enhancing carbon storage in these regions. Recent measurements of mid-latitude and high latitude forest regrowth may therefore reflect the combined effects of nitrogen deposition, elevated CO<sub>2</sub> concentrations, and climate variability. Also, CO<sub>2</sub> fertilisation can affect the nitrogen cycle and vice versa (see Section 1.5.2). It is possible that nitrogen deposition will enhance the effectiveness of carbon storage by offsetting the increased plant demand for nitrogen caused by increased concentrations of CO<sub>2</sub>. Interactive effects with air pollutants such as tropospheric ozone may also be important.

#### 1.4.3.4.1 Emissions from Changing Land Use

Deforestation and other changes in land use (including land management) cause significant exchanges of CO<sub>2</sub> between the land and atmosphere. Changes in land use from 1850 to 1990

Table 1.3. Average annual budget of CO<sub>2</sub> perturbations for 1980 to 1989. Fluxes and reservoir changes of carbon are expressed in GtC/yr, error limits correspond to an estimated 90% confidence interval.

CO <sub>2</sub> sources	
(1) Emissions from fossil fuel combustion and cement production	5.5 ± 0.5
(2) Net emissions from changes in tropical land use	1.6 ± 1.0
(3) Total anthropogenic emissions (1)+(2)	7.1 ± 1.1
<i>Partitioning among reservoirs</i>	
(4) Storage in the atmosphere	3.2 ± 0.2
(5) Oceanic uptake	2.0 ± 0.8
(6) Uptake by Northern Hemisphere forest regrowth	0.5 ± 0.5*
(7) Additional terrestrial sinks (CO <sub>2</sub> fertilisation, nitrogen fertilisation, climatic effects) [(1) + (2)] - [(4) + (5) + (6)]	1.4 ± 1.5

\* from Table 1.2

resulted in cumulative emissions of 122 ± 40 GtC (Houghton, 1994a). From the last century through the 1940s, expansion of agriculture and forestry in the middle and high latitudes dominated carbon emissions from the terrestrial biosphere (Houghton and Skole, 1990). Since then, conversion of temperate forests for agricultural use has diminished and forests are regrowing in previously logged forests and on abandoned agricultural lands (Melillo *et al.*, 1988; Birdsey *et al.*, 1993; Dixon *et al.*, 1994). Emissions from the tropics have generally been increasing since the 1950s. Estimates for 1980 range from 0.4 to 2.5 GtC (Houghton *et al.*, 1987; Detwiler and Hall, 1988; Watson *et al.*, 1990).

While uncertainties in the estimated rate of tropical deforestation are large (FAO, 1993), estimation of deforestation rates and area of regrowing forests have improved recently through the use of remote sensing. For example, satellite imagery has reduced estimated rates of deforestation in the Brazilian Amazon (INPE, 1992; Skole and Tucker, 1993). The lack of satellite analyses for other tropical areas may mean that the flux from the tropics as a whole has been overestimated.

While there is considerable uncertainty in estimated emissions from changing land use, the most recent compilations (Dixon *et al.*, 1994; Houghton, 1994b) agree that tropical emissions averaged 1.6 ± 1.0 GtC/yr in the 1980s. These analyses take into account the changing estimates of deforestation rates (INPE, 1992; Skole and Tucker, 1993) as well as other new data. The uncertainty estimate reflects the controversy over quantity and distribution of biomass (e.g., Brown and Lugo, 1992; Fearnside, 1992) as well as the errors associated within and between individual studies. At this writing, considerable work is in progress; it is possible that the estimate

of emissions from changing land use in the tropics may be modified in the near future. Of particular concern is the currently poor evaluation of effects of changing land use in the drier, open canopy forests of the seasonal tropics (FAO, 1993), and poor knowledge of carbon accumulation in regrowing forests.

#### 1.4.3.4.2 Uptake of CO<sub>2</sub> by Changing Land Use

Several recent analyses suggest a sink of carbon in regrowing Northern Hemisphere forests. Estimates for the magnitude of this term range from essentially zero to 0.74 GtC/yr (Melillo *et al.*, 1988; Houghton, 1993; Dixon *et al.*, 1994). It is difficult to assess the magnitude of this sink for two reasons. The first is the wide disparity in published regional estimates (e.g., a Russian sink of 0.01–0.06 GtC/yr (Melillo *et al.*, 1988; Krankina and Dixon, 1994) versus 0.3–0.5 GtC/yr (Dixon *et al.*, 1994).

A more fundamental problem is the confusion of carbon uptake via forest growth versus forest *regrowth*. As a basis for understanding how fluxes might change in the future, we seek in this report to separate increases in forest carbon storage that result from the regrowth of previously harvested forests versus storage due to processes such as CO<sub>2</sub> fertilisation, nitrogen deposition, and changes in fire frequency, management practices, or climate, all of which can affect forest growth. Inventory-based estimates of forest carbon storage are made by multiplying the age distribution of forest stands (rates of carbon accumulation generally decrease as stands age) by age-specific rates of carbon accumulation, estimated from observations. However, as forest carbon accumulation has been measured during the period of changing CO<sub>2</sub> and nitrogen deposition, and possibly climate, forest inventories cannot distinguish the amount of accumulation attributable to individual processes.

The confusion of mechanisms is not important when inventory data are used to corroborate the results of inverse modelling (e.g., Tans *et al.*, 1990; Enting and Mansbridge, 1991; Ciais *et al.*, in press), but separation of carbon storage mechanisms is essential for projections of future changes in carbon storage. Estimates of carbon storage due to nitrogen deposition or CO<sub>2</sub> fertilisation (Peterson and Melillo, 1985; Schindler and Bayley, 1993; Gifford, 1994) cannot be added to inventory-based land use sinks (Dixon *et al.*, 1994; Houghton, 1994b). Moreover, inventory data are ill-suited for the task of projecting future carbon uptake because the inventory-based method does not distinguish between demographic effects such as forest harvest, agricultural abandonment, and natural disturbance such as fire and wind (Kolchugina and Vinson, 1993). The problem of quantifying the contribution of individual mechanisms to total carbon storage is even greater in the tropics, where few demographic data exist. Thus, it is inappropriate to rely entirely on inventory-based measurements for initialisation of terrestrial carbon models.

In addition to changes on forested lands, the decline in soil carbon storage in agricultural lands may have been reversed in the middle latitudes. Best management practices may even lead to storage increases in the future (Metherell, 1992). While the world-wide extent of management changes and their effect on carbon balance is poorly known, a recent assessment of car-

bon storage in U.S. croplands suggests possible increases over the coming few decades of 0.02–0.05 GtC/yr; the authors note that similar changes are likely to have occurred over the past few decades (Donigian *et al.*, 1994). In terms of the global C budget, increases in soil storage are small (Schlesinger, 1990), although they may be significant at the regional, or even national, scale.

#### 1.4.3.4.3 Other Terrestrial Sink Processes

While little has been done to increase understanding of potential sinks in tropical forests, analyses based on models of forest growth, age distribution data, and the age-growth relationship suggest carbon accumulation in Northern Hemisphere forests may be as high as ~1 GtC/yr (Dixon *et al.*, 1994). Analyses based on observed atmospheric CO<sub>2</sub> and <sup>13</sup>CO<sub>2</sub> suggest a strong Northern Hemisphere terrestrial sink. While various pieces of evidence support a substantial terrestrial sink in the Northern Hemisphere, direct observations to confirm the hypothesis and to establish the processes responsible for increasing carbon storage are lacking. For example, tree ring studies of contemporary and pre-industrial forest growth rates are contradictory. These studies have revealed enhanced growth in subalpine conifers (LaMarche *et al.*, 1984), no growth enhancement beyond that explained by climatic variability (Graumlich, 1991; D'Arrigo and Jacoby, 1993), and both positive and negative changes in growth rate dependent upon species and location (Briffa *et al.*, 1990). Models and observational data to confirm tropical sinks are even more deficient (e.g., Brown and Lugo, 1984, 1992; Fearnside, 1992; Skole and Tucker, 1993; Dixon *et al.*, 1994). Improved observations for identifying terrestrial sinks is necessary for the improvement of models capable of projecting future states of the carbon cycle, and also for quantifying regional contributions to terrestrial sources and sinks. Processes that may contribute to a terrestrial sink are explored below (see Table 1.2).

### CO<sub>2</sub> Fertilization

Experimental studies of agricultural and wild plant species have shown growth responses of typically 20–40% higher growth under doubled CO<sub>2</sub> conditions (ranging from negligible or negative responses in some wild plants to responses of 100% in some crop species (Körner and Arnone, 1992; Rochefort and Bazzaz, 1992; Coleman *et al.*, 1993; Idso and Kimball, 1993; Owensby *et al.*, 1993; Polley *et al.*, 1993; Idso and Idso, 1994)). The majority of these studies were short-term and were conducted with potted plants (Idso and Idso, 1994). The effect of increased growth under elevated CO<sub>2</sub> conditions, known as the CO<sub>2</sub> fertilisation effect, is often assumed to be the primary mechanism underlying the imbalance in the global carbon budget. This assumption implies a terrestrial sink that increases as CO<sub>2</sub> increases, acting as a strong negative feedback. However, while the potential effect of CO<sub>2</sub> in experimental and some field studies is relatively strong (e.g., Drake, 1992; Idso and Kimball, 1993; see also review by Idso and Idso, 1994), natural ecosystems may be less responsive to increased levels of atmospheric CO<sub>2</sub>. Evidence that the effects

of CO<sub>2</sub> on long-term carbon storage may be less than is suggested by short-term pot studies of photosynthesis or plant growth includes:

- (1) Field studies showing reduced responses over time (Oechel *et al.*, 1993) and zero, small, or statistically insignificant responses (Norby *et al.*, 1992; Jenkinson *et al.*, 1994);
- (2) Evidence that plants with low intrinsic growth rates, a common trait in native plants, are less responsive to CO<sub>2</sub> increases than are rapidly growing plants, such as most crop species (Poorter, 1993);
- (3) Results of model simulations which suggest that increases in carbon storage eventually become nutrient-limited (Comins and McMurtrie, 1993; Melillo *et al.*, 1994; reviewed in Schimel, 1995). (Long-term nutrient limitation is different from the short-term nutrient limitation of photosynthesis and plant growth observed in experimental studies; it reflects the need for nutrients in addition to carbon to increase organic matter production.)

There is also evidence that while water- or temperature-stressed plants are more responsive to CO<sub>2</sub> increase than are unstressed plants (because CO<sub>2</sub> increases water-use efficiency (Polley *et al.*, 1993)), nitrogen-limited plants are less sensitive to CO<sub>2</sub> level (Bazzaz and Fajer, 1992; Comins and McMurtrie, 1993; Díaz *et al.*, 1993; Melillo *et al.*, 1993; Ojima *et al.*, 1993; Idso and Idso, 1994). The reduction in response of short-term photosynthesis and plant growth to CO<sub>2</sub> caused by nitrogen limitation is highly variable and may on the whole be small (Idso and Idso, 1994), but as all organic matter contains nitrogen, carbon storage increases should eventually become limited by the stoichiometric relationships of carbon and other nutrients in organic matter (Comins and McMurtrie, 1993; Díaz *et al.*, 1993; Schimel, 1995).

While current models assume that nitrogen inputs are not affected by CO<sub>2</sub> fertilisation, this may not be true (Thomas *et al.*, 1991; Gifford, 1993; Idso and Idso, 1994). CO<sub>2</sub> fertilisation could stimulate biological nitrogen fixation, because nitrogen-fixing organisms have high energy (organic carbon) requirements. If nitrogen inputs are stimulated by increasing CO<sub>2</sub>, acclimation of carbon storage to higher CO<sub>2</sub> levels could be temporarily offset. However, evidence from native ecosystems suggests that other nutrients, such as phosphorus, may be more limiting to nitrogen fixation than is energy (Cole and Heil, 1981; Eisele *et al.*, 1989; Vitousek and Howarth, 1991).

As a sensitivity analysis, the effects of CO<sub>2</sub> enrichment was varied in one model over the plausible range of values (by the equivalent of 10, 25, and 40% increases in plant growth at doubled CO<sub>2</sub>). The results indicate terrestrial carbon storage rates (increased net ecosystem production: NEP) due to CO<sub>2</sub> fertilisation were 0.5, 2.0, and 4.0 GtC/yr during the 1980s (Gifford, 1993). A modelling experiment by Rotmans and den Elzen (1993) indicated the strength of CO<sub>2</sub> fertilisation might be ~1.2 GtC/yr. Overall, it is likely that CO<sub>2</sub> fertilisation plays a role in the current terrestrial carbon budget, and may have amounted to storage of 0.5 to 2.0 GtC/yr during the 1980s.

### Nitrogen Fertilization

Many terrestrial ecosystems are nitrogen limited: added fertiliser will produce a growth response and additional carbon storage (e.g., Vitousek and Howarth, 1991; Schimel, 1995). Nitrogen deposition from fertilisers and oxides of nitrogen released from the burning of fossil fuel during the 1980s is estimated to amount to a global total, but spatially concentrated, 0.05–0.06 GtN/yr (Peterson and Melillo, 1985; Duce *et al.*, 1991). The carbon sequestration which results from this added nitrogen is estimated to be of the order 0.2–1.0 GtC/yr (Peterson and Melillo, 1985; Schindler and Bayley, 1993; Schlesinger, 1993), depending on assumptions about the proportion of nitrogen that remains in ecosystems. Estimates significantly higher than 1 GtC/yr are unrealistic because they assume that all of the N would be stored in forms with high carbon to nitrogen ratios, while much atmospheric nitrogen is in reality deposited on grasslands and agricultural lands where storage occurs in soils with low average carbon to nitrogen ratios. Uptake of carbon due to long-term increases in nitrogen deposition could increase as nitrogen pollution increases; but possibly to a threshold, after which additional nitrogen may result in ecosystem degradation (e.g., Aber *et al.*, 1989; Schulze *et al.*, 1989). Because most deposition of anthropogenic nitrogen occurs in the middle latitudes, some of the effect of added nitrogen may already be accounted for in measurement-based estimates of mid-latitude carbon accumulation (discussed above). However, existing analyses do not allow identification of the fraction of measured forest growth that is due to nitrogen addition or effects of CO<sub>2</sub>.

### Climate Effects

Climate affects carbon storage in terrestrial ecosystems because temperature, moisture, and radiation influence both ecosystem carbon gain (photosynthesis) and loss (respiration) (Houghton and Woodwell, 1989; Schimel *et al.*, 1994). While a number of compensatory processes are possible, warming is thought to reduce carbon storage by increasing respiration, especially of soils (Houghton and Woodwell, 1989; Shaver *et al.*, 1992; Townsend *et al.*, 1992; Oechel *et al.*, 1993; Schimel *et al.*, 1994). Conversely, cooling would be expected to increase carbon storage. In nutrient-limited forests, however, warming may increase carbon storage by “mineralising” soil organic nutrients, which are generally stored in nutrient-rich, or low carbon-to-nutrient ratio forms, thereby allowing increased uptake by trees, which store nutrients in high carbon-to-nutrient forms (Shaver *et al.*, 1992). This is not true in nutrient-limited tundra ecosystems, where recent warming has resulted in a local CO<sub>2</sub> source (Oechel *et al.*, 1993). Increases in precipitation will generally increase carbon storage by increasing plant growth, although in some ecosystems compensatory increases in soil decomposition may reduce or offset this effect (Ojima *et al.*, 1993). While these effects have been discussed as components of future responses of ecosystems to climate change, climate variations during the past century may have influenced the terrestrial carbon budget. In a provocative paper, Dai and

Fung (1993) suggested that climate variations over the past decades could have resulted in a substantial sink. Ciais *et al.* (in press) suggested that cooling arising from the effects of Mt. Pinatubo may have increased terrestrial carbon storage and contributed to the observed reduction in the atmospheric growth rate during the 1991 to 1992 period. Palaeoclimate modelling studies likewise suggest major changes in terrestrial carbon storage with climate (Prentice and Fung, 1990; Friedlingstein *et al.*, 1992). Modelling and observations of ecosystem responses to climate and climate anomalies will be important tools for validating predictions of future changes; meanwhile, investigations of the effects of climate on carbon storage remain suggestive rather than definitive.

#### 1.4.4 Budget Summary

In Table 1.3 we present an estimated budget of carbon perturbations for the 1980s, shown as annual average values.

The budgetary inclusion of a Northern Hemisphere sink in forest regrowth reduces the unaccounted-for sink compared with earlier budgets (IPCC, 1999, 1992) by assigning a portion of this sink to forest regrowth. The remaining imbalance in this budget implies additional net terrestrial sinks of  $1.4 \pm 1.5$  GtC/yr.

### 1.5 The Influence of Climate and Other Feedbacks on the Carbon Cycle

#### 1.5.1 Introduction

The description of the carbon cycle in the previous sections addressed the budget of anthropogenic CO<sub>2</sub>, without emphasising the possibility of more complex interactions within the system. There are, however, a number of processes that can produce feedback loops. One important distinction (see Enting, 1994) is between carbon cycle feedbacks and CO<sub>2</sub>-climate feedbacks. The only direct carbon cycle feedback is the CO<sub>2</sub> fertilisation process described in Section 1.4.3.4.3. In contrast, CO<sub>2</sub>-climate feedbacks involve the effect of climate change (potentially induced by CO<sub>2</sub> concentration changes) on the components of the carbon cycle. For terrestrial components, the most important effects are likely to be those involving temperature, precipitation, and radiation changes (through changes in cloudiness) on net primary production and decomposition (including effects resulting from changes in species composition). For marine systems, the primary effects to be expected arise through climatic influences on ocean circulation and chemistry. Such changes would affect the physical and biological aspects of carbon distribution in the oceans including physical fluxes of inorganic carbon within the ocean and changes in nutrient cycling (Manabe and Stouffer, 1993). The feedbacks may also include changes in the species composition of the ocean biota, which determines the location and magnitude of oceanic CO<sub>2</sub> uptake.

Factors with a strong influence on the global carbon cycle that are similar to and/or modify feedback loops also exist. Among such factors are the effects of increased UV on terrestrial

and marine ecosystems and the anthropogenic toxification and eutrophication of these ecosystems. While significant effects of possible changes in mid-latitude and high latitude UV radiation, tropospheric ozone and other pollutants have been documented in experimental studies, there is little basis for credible global extrapolation of these studies (Chameides *et al.*, 1994). They will not be discussed further, but should be addressed in subsequent assessments as more data become available.

#### 1.5.2 Feedbacks to Terrestrial Carbon Storage

The responses of terrestrial carbon to climate are complex, with rates of biological activity generally increasing with warmer temperatures and increasing moisture. Because photosynthesis and plant growth increase with warmer temperatures, longer growing seasons and more available water, storage of carbon in living vegetation generally increases as well (Melillo *et al.*, 1993). Storage of carbon in soils generally increases along a gradient from low to high latitudes, reflecting slower decomposition of dead plant material in colder environments (Post *et al.*, 1985; Schimel *et al.*, 1994). Flooded soils, where oxygen becomes depleted, have extremely low rates of decomposition and may accumulate large amounts of organic matter as peat. Global ecosystem models based on an understanding of underlying mechanisms are designed to capture these patterns, and have been used to simulate the responses of terrestrial carbon storage to changing climate. Some models also include the effect of changes in land use. Model results suggest that future effects of changing climate, atmospheric CO<sub>2</sub>, and changing land use on carbon storage may be quite large (Vloedbeld and Leemans, 1993).

#### Effects of Temperature and CO<sub>2</sub> Concentration

Ecosystem models have been used to simulate the response of the terrestrial biosphere to changes in climate, rate of change of climate ("transient" changes), effects of changing CO<sub>2</sub>, and the effects of changing land use. Models may consider only the response of plant growth and decomposition, or they may also allow movement of vegetation "types" such as forests and grasslands. The simplest case, involving a general circulation model- (GCM-) simulated climate change and stationary vegetation patterns projected losses of terrestrial carbon of about 200 Gt over an implicit time period of a few hundred years (Melillo *et al.*, 1993). When vegetation types were allowed to migrate, models projected a long-term increase in terrestrial carbon storage: 60–90 GtC over 100–200 years (Cramer and Solomon, 1993; Smith and Shugart, 1993). In Smith and Shugart's (1993) model, climate change and vegetation redistribution led to a transient release of CO<sub>2</sub> from die back before regrowth (~200 GtC over ~100 years), followed by an eventual accumulation of ~90 GtC. This study, like those mentioned above, did not incorporate the effects of CO<sub>2</sub> fertilisation. The TEM model of Melillo *et al.* (1993) was also used to assess the response of ecosystem carbon storage to CO<sub>2</sub> fertilisation under a scenario of climate change resulting from an in-

stantaneous doubling of CO<sub>2</sub>. The results of this experiment showed that, over the period of CO<sub>2</sub> doubling, uptake of carbon occurred only when the CO<sub>2</sub> fertilisation effect was included (loss of ~200 GtC without the fertilisation effect, and a gain of ~250 GtC when the effect was included: Melillo *et al.*, 1994). Rotmans and den Elzen (1993) used the IMAGE model to assess the effects of including CO<sub>2</sub> fertilisation and temperature feedbacks on their modelled carbon budget. Estimated biospheric uptake increased by 1.2 GtC/yr over the decade of the 1980s when CO<sub>2</sub> and temperature feedbacks were included. The most complex assessment of the interactive effects on carbon storage are carried out with models of transient changes in climate that include CO<sub>2</sub> fertilisation. Esser (1990) simulated an increase of 170 GtC over 200 years with transient changes of climate and CO<sub>2</sub> fertilisation (this model does not take vegetation redistribution into account). Alcamo *et al.* (1994b) performed a similar experiment, but included redistribution of vegetation. Their results revealed eventual increased carbon storage of ~200–250 Gt, depending on assumptions of future land use. Results from ecosystem models suggest that changes in climate associated with CO<sub>2</sub> doubling are likely to lead to a significant transient release of carbon (1–4 GtC/yr) over a period of decades to more than a century (Smith and Shugart, 1993; Dixon *et al.*, 1994).

#### Effects of Land Use

Several of these studies also included the effects of changing land use. For example, Esser (1990) simulated a gain of 170 GtC due to the effects of climate and CO<sub>2</sub> change, but losses of 322 GtC when he assumed areal reduction of global forests by 50% by the year 2300, emphasising the importance of maintaining forest ecosystems for terrestrial carbon storage. In the Cramer and Solomon (1993) study, inclusion of dense land clearing also substantially reduced simulated C storage (by up to 152 GtC). In the Alcamo *et al.* (1994b) study, global scenarios that required additional land for agriculture and biomass energy production resulted in lower carbon storage.

While the results from global terrestrial models range widely, all suggest that the terrestrial biosphere could eventually take up 100–300 GtC in response to warming, albeit after a significant transient loss (e.g., Smith and Shugart, 1993). While existing simulations have a number of shortcomings, including the lack of agreed-upon climate and land use scenarios that would allow rigorous comparison of results, they suggest possible exchanges between the atmosphere and the terrestrial biosphere of the order of 100s of GtC over decades to a few centuries (e.g., Alcamo *et al.*, 1994a).

#### Soil Feedbacks

Because soil carbon is released with increasing temperature, it has been suggested that global warming would result in a large positive feedback (Houghton and Woodwell, 1989; Townsend *et al.*, 1992; Oechel *et al.*, 1993). In order to evaluate this, the

sensitivity of carbon storage to temperature was assessed using a number of models (Schimel *et al.*, 1994). Inter-comparison of model results revealed rates of global soil carbon loss of 11–34 GtC per degree warming. Vegetation growth and C storage are stimulated in many of these models because as soil carbon is lost, soil nitrogen is made available to the modelled vegetation. This fertilisation effect can ameliorate or even reverse the overall loss of carbon (Shaver *et al.*, 1992; Gifford, 1994; Schimel *et al.*, 1994). In the Century ecosystem model, the nitrogen feedback reduces the effect of warming on soil carbon loss by 50% (Schimel *et al.*, 1994).

#### Nutrient Limitation of CO<sub>2</sub> Fertilization

The temperature feedback on nitrogen availability may interact with CO<sub>2</sub> fertilisation. CO<sub>2</sub>-fertilised foliage is typically lower in nitrogen than foliage grown under current concentrations of CO<sub>2</sub> (e.g., Coleman and Bazzaz, 1992). As dead foliage from high-CO<sub>2</sub> conditions works its way through the decomposition process, soil decomposer organisms require additional nitrogen (Díaz *et al.*, 1993), and, as a result, vegetation may become more nitrogen limited (attenuating but not eliminating the effects of CO<sub>2</sub> fertilisation (Comins and McMurtrie, 1993; Schimel, 1995)). Additional nitrogen released by warming, as described above, can alleviate the nitrogen stress induced by high-CO<sub>2</sub> foliage: this interaction between warming and CO<sub>2</sub> fertilisation is responsible for the large estimated effect of CO<sub>2</sub> on carbon storage in Melillo *et al.* (1993). Deposition of atmospheric nitrogen could also influence the effectiveness of CO<sub>2</sub> fertilisation (Gifford, 1994), although some modelling studies suggest the effect may be modest (Rastetter *et al.*, 1992; Comins and McMurtrie, 1993). Empirical studies are few, but a recent analysis by Jenkinson *et al.* (1994) revealed no measurable change in hay yield over the past 100 years despite substantial increases in nitrogen inputs via precipitation and a 21% increase in atmospheric CO<sub>2</sub> concentration during the period of study. Nutrient feedbacks clearly influence the response of terrestrial ecosystems to the interactive effects of climate and CO<sub>2</sub>.

#### 1.5.3 Feedbacks on Oceanic Carbon Storage

Climate feedbacks influence the storage of carbon in the ocean through physical, chemical, and biological processes. Changes in sea surface temperature affect oceanic CO<sub>2</sub> solubility and carbon chemistry. At equilibrium, a global increase in sea surface temperature of 1°C is associated with an increase of the partial pressure of atmospheric CO<sub>2</sub> of approximately 10 ppmv (Heinze *et al.*, 1991), serving as a weak positive feedback between temperature and atmospheric CO<sub>2</sub>. During a transient climate change on time-scales of decades to centuries this temperature feedback would be even weaker (MacIntyre, 1978).

Changes in temperature might also affect the remineralisation of dissolved organic carbon. Climatically driven changes in its turnover time are not expected to result in significant variations in atmospheric CO<sub>2</sub>, as the quantity of dissolved organic

carbon in the surface ocean is less than 700–750 Gt (Figure 1.1). Earlier suggestions (Sugimura and Suzuki, 1988) that this pool might be much larger than previously thought (i.e., about twice the value shown in Figure 1.1) are now generally discounted (Suzuki, 1993).

Changes in the oceanic circulation and their effects on the oceanic carbon cycle are more difficult to assess. Simulation studies of the transient behaviour of the ocean-atmosphere system using coupled GCMs (Cubasch *et al.*, 1992; Manabe and Stouffer, 1993) indicate a strong reduction of the deep thermohaline circulation, which is driven by cooling and sinking of surface water at high latitudes, especially in the North Atlantic. Smaller effects are expected on the wind-driven features of the oceanic general circulation: the shallow upwelling cells at the equator and the eastern boundaries of the warm sub-tropical gyres and in the cyclonic areas in higher latitudes. A reduction of the vertical water exchange processes would impact the oceanic carbon cycle in several ways:

First, the downward transport of surface waters in contact with the atmosphere and thus enriched with excess CO<sub>2</sub> would be reduced, thereby leading to a smaller oceanic excess CO<sub>2</sub> uptake capacity.

Second, changes in the vertical water mass transports would affect the marine biological pump. This would reduce the flux of nutrients transported to the surface and, in nutrient-limited regions, result in reduced marine production. This effect would constitute a weakening of the marine biological pump (i.e., less export of carbon to depth), and could potentially lead to an increase in dissolved inorganic carbon and partial pressure of CO<sub>2</sub> at the surface. Conversely, smaller vertical water mass transports imply a reduced upward transport of deeper waters enriched in dissolved inorganic carbon. The two effects result from the same process, but affect the surface concentration of dissolved inorganic carbon, and hence the partial pressure and the air-sea flux of CO<sub>2</sub>, in opposite directions. The latter of the two effects dominates in models that use constant carbon to nutrient ratios to describe the marine biosphere (Bacastow and Maier-Reimer, 1990; Keir, 1994). These models project a small increase in oceanic carbon storage as a result of the reduction of oceanic circulation and vertical mixing.

Initial model results suggest that the effects of predicted changes in circulation on the ocean carbon cycle are not large (10s rather than 100s of ppmv in the atmosphere). However, exploration of the long-term impacts of warming on circulation patterns has just begun; hence, analyses of impacts on the carbon cycle must be viewed as preliminary. For example, changes in the oceanic environment (temperature and circulation) have the potential to change the composition of marine ecosystems in ways not yet included in global models. This could lead to changes in carbon to nutrient ratios, which are assumed constant in present models, or to changes in the relationship between organic and inorganic carbon fixation, and/or change the efficiency by which marine organisms utilise available nutrients. Furthermore, the remineralisation depths of nutrients and carbon might be affected differently (Evans and Fasham, 1993). An extreme lower bound may be estimated by assuming complete utilisation of the oceanic surface nutrients

which would reduce atmospheric CO<sub>2</sub> by 120 ppmv. Similarly, an upper bound is given by assuming extinction of all marine life, increasing atmospheric CO<sub>2</sub> by almost 170 ppmv (Bacastow and Maier-Reimer, 1990; Shaffer, 1993; Keir, 1994). Such large effects, however, are very unlikely to occur. Indeed, the oceanic carbon system appears to be rather stable as evidenced by the very small fluctuations of the atmospheric CO<sub>2</sub> concentration prior to anthropogenic perturbation. It has also proven very difficult to account for the lower atmospheric CO<sub>2</sub> concentration in glacial times (i.e., by 80 ppmv) merely by changing biospheric parameters within current three-dimensional ocean carbon models (Heinze *et al.*, 1991; Archer and Maier-Reimer, 1994).

In areas of excess nutrients (e.g., in the equatorial and sub-arctic Pacific and in some parts of the Southern Ocean) the micronutrient iron appears to limit marine primary production (Martin, 1990). Changes in atmospheric iron loading thus potentially could affect the biological pump and impact atmospheric carbon dioxide levels. Modelling studies have shown, however, that even excessive “iron fertilisation” of these oceanic areas would have a relatively small impact on atmospheric CO<sub>2</sub> levels (Joos *et al.*, 1991a; Peng and Broecker, 1991; Sarmiento and Orr, 1991; Kurz and Maier-Reimer, 1993). Therefore, at least during the past 200 years, it is very unlikely that iron loading had a significant impact on the present day carbon balance.

External impacts on the oceanic carbon system not directly related to global warming must also be considered. Increased ultraviolet radiation (UV-B, corresponding to light wavelengths of 280–320 nm) resulting from the depletion of stratospheric ozone could affect marine life and thus influence marine carbon storage. However, model simulation studies show that even a complete cessation of marine productivity in high latitudes, where increases in UV-B are expected to occur, would result in an atmospheric CO<sub>2</sub> increase of less than 40 ppmv (Sarmiento and Siegenthaler, 1992). Increased input of anthropogenic nitrogen or other limiting nutrients might also affect the oceanic biota. The global effects on atmospheric CO<sub>2</sub> are most likely much smaller than the extreme bounds discussed above.

## 1.6 Modelling Future Concentrations of Atmospheric CO<sub>2</sub>

### 1.6.1 Introduction

The clear historical relationship between CO<sub>2</sub> emissions and changing atmospheric concentrations implies that continuing fossil fuel, cement, and land-use-related emissions of CO<sub>2</sub> at or above current rates will result in increasing atmospheric concentrations of this greenhouse gas. Understanding how CO<sub>2</sub> concentrations will change in the future requires quantification of the relationship between CO<sub>2</sub> emissions and atmospheric concentration using models of the carbon cycle. This section presents the results of a set of standardised calculations, carried out by modelling groups from many countries, that analyse the relationships between emissions and concen-

trations in a number of ways (see Enting *et al.*, 1994b, for full documentation of the modelling exercise).

Two questions are considered:

- For a given CO<sub>2</sub> emission scenario, how might CO<sub>2</sub> concentrations change in the future?
- For a given CO<sub>2</sub> concentration profile leading to stabilisation, what anthropogenic emissions are implied?

As an initial condition, it was required that all models have a balanced carbon budget in which the components matched, within satisfactory limits, a prescribed 1980s-mean budget based on Watson *et al.* (1992). Modelling groups were provided with prescribed future land use fluxes, and a variety of time-series of concentrations and fossil-plus-cement emissions. Because the model intercomparison was conducted simultaneously with the rest of the assessment, the prescribed budget differs from the budget presented in this chapter (Table 1.3) in that:

- (i) the 1980s-mean concentration growth rate used (1.59 ppmv/yr or 3.4 GtC/yr) was higher than the current estimate (1.53 ppmv/yr or 3.2 GtC/yr);
- (ii) the net flux from changing land use was set at 1.6 GtC/yr rather than the current estimate of 1.1 GtC/yr;
- (iii) the only mechanism used in the models to simulate terrestrial uptake was CO<sub>2</sub> fertilisation, whereas we suggest that several other mechanisms may be important (Table 1.2).

Modellers were required to continue whatever processes were used to balance the 1980s budget into the future.

The approach used in balancing the budget probably biases the modelling exercise for this chapter towards lower concentrations when emission profiles were employed and higher anthropogenic emissions when concentrations were prescribed. There are two reasons for this. First, the effect of forest regrowth and nitrogen deposition result in changing terrestrial carbon storage which need not increase with increasing CO<sub>2</sub> concentrations as is required by the CO<sub>2</sub> fertilisation effect. Second, as noted in previous IPCC reports (Watson *et al.*, 1990, 1992) climate-related feedbacks may result in additional transient releases of CO<sub>2</sub> to the atmosphere. In contrast, most of the calculations employed in the 1990 IPCC assessment (Watson *et al.*, 1990) assumed constant terrestrial carbon content after 1990 and probably resulted in biases in the other direction. Revisions to the 1980s budget used in model initialisation for this report, however, are in a direction that would lead to higher concentrations by 5 to 10% (for given emissions) and lower emissions by a similar amount (for given concentrations).

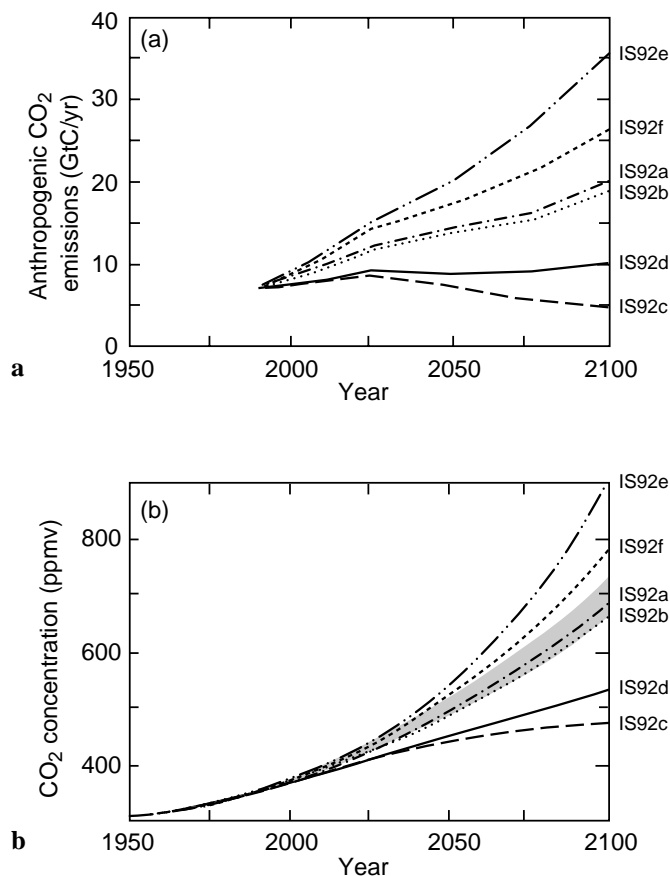
Results from a range of different carbon cycle models are considered in order to assess the sensitivity of calculated emission and concentration profiles to model formulation. The complexity of the models employed varies considerably. The most detailed was a model coupling a three-dimensional ocean circulation and chemistry model to a terrestrial biosphere model incorporating a full geographical representation of ecological processes; but, as with other complex models, only a small set of calculations could be performed. The simplest models were

designed to simulate only critical processes and represent terrestrial and oceanic carbon uptake with a minimum number of equations. Because most of the models employed incorporate some representation of terrestrial processes, they tended to have initially higher rates of CO<sub>2</sub> uptake from the atmosphere and produced lower estimates for the effective lifetime of CO<sub>2</sub> than the models used in the 1990 IPCC report (Moore and Braswell, 1994). Thus, the GWPs based on these newer carbon cycle model results are higher for other trace gases than those of earlier IPCC assessments (Shine *et al.*, 1990).

We chose one model, the “Bern model”, for a number of important illustrative calculations, because its results were generally near the mid-point of the results obtained with all models, and because complete descriptions exist in the literature (Joos *et al.*, 1991a; Siegenthaler and Joos, 1992). Selected sensitivity analyses from the model of Wigley (1993) were also used as the configuration of that model allowed for ready modification to consider certain issues. Both the Bern and Wigley models have a balanced carbon cycle consisting of a well-mixed atmosphere linked to oceanic and terrestrial biospheric compartments. In the Bern model, the ocean is represented by the HILDA model, which is a box-diffusion model with an additional advective component. It was tuned to observed values of natural and bomb radiocarbon (Joos *et al.*, 1991a, b; Siegenthaler and Joos, 1992) and validated with CFCs and Argon-39 (Joos, 1992). The Wigley model uses a representation of the ocean based on the ocean general circulation carbon cycle model of Maier-Reimer and Hasselmann (1987). Both the models have similar terrestrial components, with representations of ground vegetation, wood, detritus, and soil (Siegenthaler and Oeschger, 1987; Wigley, 1993). A possible enhancement of plant growth due to elevated CO<sub>2</sub> levels is taken into account by a logarithmic dependency between additional photosynthesis and atmospheric CO<sub>2</sub>, which, in the Bern model, probably overestimates biological storage at high CO<sub>2</sub> concentrations. The Bern model was chosen for illustrative purposes in discussions where presentation of multiple results would be confusing (and where all models produced similar patterns). This model was also used to define the reference case for the GWPs presented in Albritton *et al.* (1995). Neither model is recommended nor endorsed by the IPCC or the authors of this chapter as having higher credibility than other models.

### 1.6.2 Calculations of Concentrations for Specified Emissions

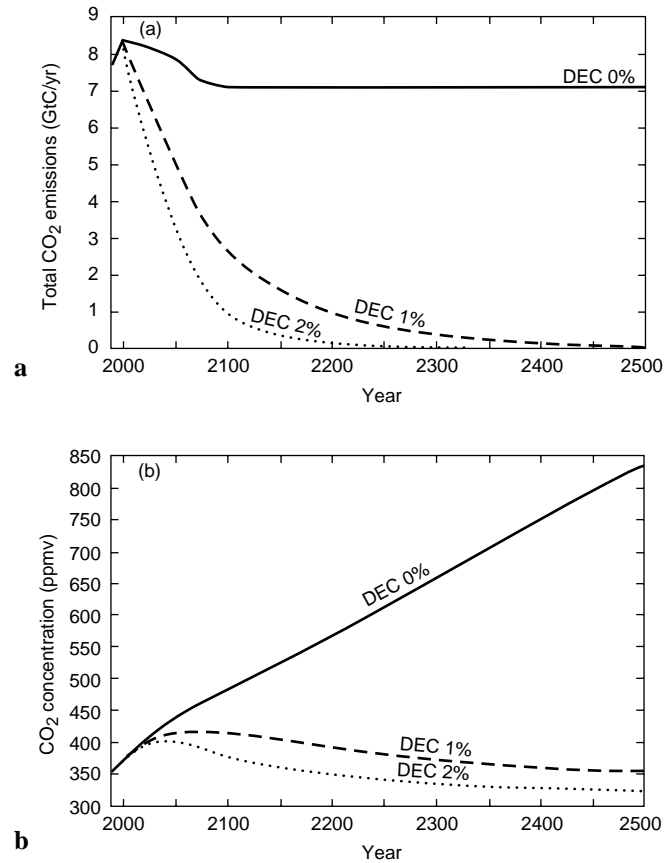
Six greenhouse gas emissions scenarios were described in the 1992 IPCC report (Leggett *et al.*, 1992), based on a wide range of assumptions regarding future economic, demographic, and policy factors. The anthropogenic CO<sub>2</sub> emissions for these scenarios are shown in Figure 1.9a. Scenario IS92c, which has the lowest CO<sub>2</sub> emissions, assumes an eventual decrease in population, low economic growth, and severe constraints on the availability of fossil fuel supplies. The highest emission scenario (IS92e) assumes moderate population growth, high economic growth, high fossil fuel availability, and a phase-out of nuclear power. Concentration estimates for these scenarios



**Figure 1.9:** (a) Anthropogenic CO<sub>2</sub> emissions for the IS92 Scenarios. (b) Atmospheric CO<sub>2</sub> concentrations calculated from the scenarios IS92a-f (Leggett et al., 1992) using the Bern model (Siegenthaler and Joos; 1992). The typical range of results from different carbon cycle models is indicated by the shaded area.

have previously been published (e.g., Wigley, 1993). Figure 1.9b shows concentration results from the Bern model that typify the responses of the wide range of models used for the full analysis. This and the other models show strong increases in concentration to well above pre-industrial levels by 2100 (75 to 220% higher). None of the six scenarios leads to stabilisation of concentration before 2100, although IS92c leads to a very slow growth in CO<sub>2</sub> concentration after 2050. IS92a, b, e, and f all produce a doubling of the pre-industrial CO<sub>2</sub> concentration before 2070, with rapid rates of concentration growth. Scenarios developed by the World Energy Council show a similar range of results (Figures 1.11a, b).

In addition to the IS92 emissions cases, three arbitrarily chosen “science” emissions profiles and the newly produced World Energy Council (WEC) Scenarios were also examined. In the former, fossil emissions followed IS92a to the year 2000 and then either stabilised (DEC0%) or decreased at 1% or 2%/yr (see Figure 1.10). For the WEC Scenarios, where only energy-related CO<sub>2</sub> emissions were originally given (WEC, 1993), estimates of gas-flaring and cement production emissions were added (M. Jefferson and G. Marland, personal communications) to ensure consistency with the IS92 Scenarios. Total fos-



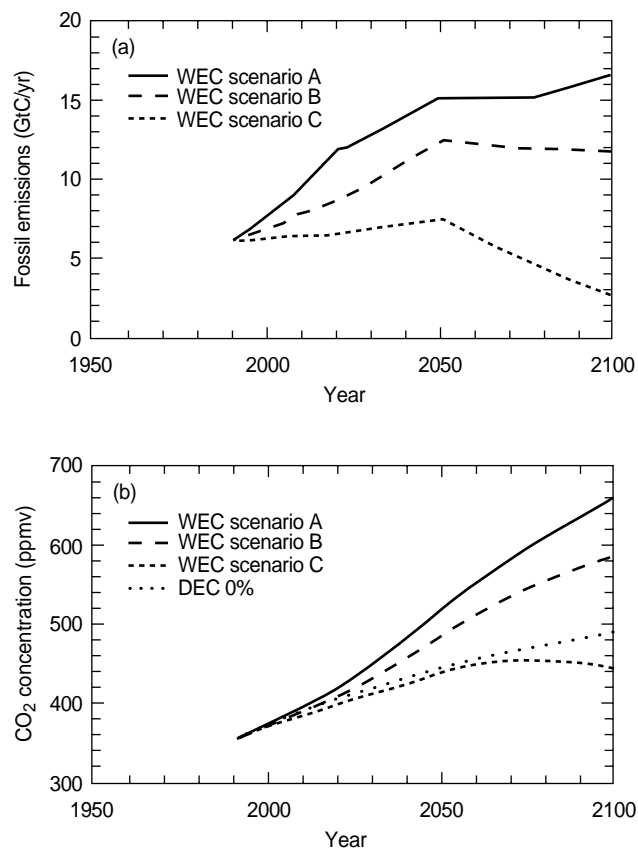
**Figure 1.10:** (a) Anthropogenic CO<sub>2</sub> emissions calculated by following the IS92a Scenario to the year 2000 and then either fixed fossil fuel emissions (DEC 0%) or emissions declining at 1%/yr (DEC 1%) or 2%/yr (DEC 2%). Land-use emissions followed the modified IS92a scenario (see Section 1.5.2). (b) Atmospheric CO<sub>2</sub> concentrations resulting from DEC 0%, DEC 1% and DEC 2% emissions. Curves are for the model of Wigley (1993).

sil emissions are given in Figure 1.11a. In all cases net land use emissions were assumed to follow IS92a to 2075. For the WEC cases, IS92a was followed to 2100. For the science scenarios, land use emissions dropped to zero in 2100 and remained zero thereafter. Concentration results are shown in Figures 1.10 and 1.11. Perhaps the most important result (which could be anticipated from the IS92c case) is that stabilisation of emissions at 2000 levels does not lead to stabilisation of CO<sub>2</sub> concentration by 2100; in fact, the calculations show that concentrations continue to increase slowly for at least several hundred years. The lowest of the WEC Scenarios, where emissions were based on policies driven by “ecological” considerations (see WEC, 1993), gives an idea of the sort of emissions profile that could lead to concentration stabilisation.

### 1.6.3 Stabilization Calculations

The calculations presented in this section illustrate additional aspects of what may be required to achieve stabilisation of atmospheric CO<sub>2</sub> concentrations. The exercise was motivated by



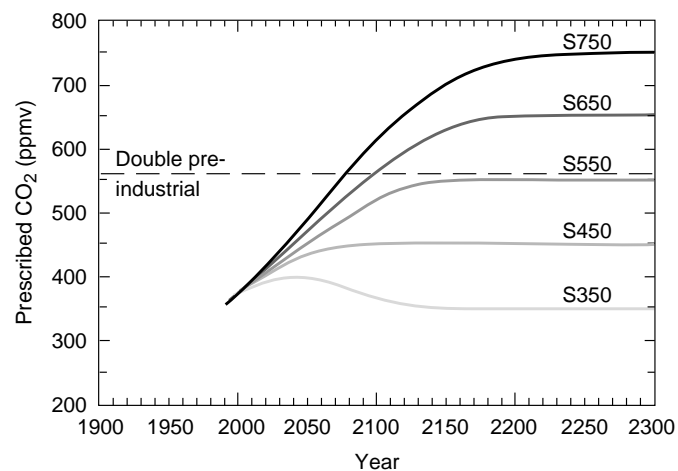


**Figure 1.11:** (a) Emission scenarios from the World Energy Council (WEC) modified by including gas flaring and cement production. (b) Atmospheric CO<sub>2</sub> concentrations calculated from the WEC Scenarios using the model of Wigley (1993). Concentrations resulting from the fixed emissions case (DEC 0%) (see Figure 1.10) using the same model are included for comparison.

the Framework Convention on Climate Change (United Nations, 1992), which states:

*The ultimate objective of this Convention and any related legal instruments that the Conference of the Parties may adopt is to achieve, in accordance with the relevant provisions of the Convention, stabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system. Such a level should be achieved within a time-frame sufficient to allow ecosystems to adapt naturally to climate change, to ensure that food production is not threatened and to enable economic development to proceed in a sustainable manner.*

In the context of this objective it is important to investigate a range of emission profiles of greenhouse gases which might lead to atmospheric stabilisation. It is not our purpose here to consider the climate response (this will be done in the 1995 IPCC Scientific Assessment report), nor to define what might constitute “dangerous interference”, nor to make any judgement about the rates of change that would meet the criteria of



**Figure 1.12:** CO<sub>2</sub> concentration profiles leading to stabilisation at 350, 450, 550, 650 and 750 ppmv. These are the profiles prescribed for carbon cycle model calculations in which the corresponding emission pathways (shown in Figure 1.13) were determined.

the objective. In this chapter only CO<sub>2</sub> is considered; the stabilisation of other greenhouse gas concentrations is discussed in Prather *et al.* (1995).

Several carbon cycle models have been used to calculate the emissions of CO<sub>2</sub> that would lead to stabilisation at a range of different concentration levels. These calculations are designed to illustrate the relationship between CO<sub>2</sub> concentration and emissions. Concentration profiles have been devised (Figure 1.12) in which CO<sub>2</sub> concentrations stabilise at levels from 350 to 750 ppmv (for comparison, the pre-industrial CO<sub>2</sub> concentration was close to 280 ppmv and the 1990 concentration was ~355 ppmv). Many different stabilisation levels and routes to stabilisation could have been chosen. Those in Figure 1.12 give a smooth transition from the 1990 rate of CO<sub>2</sub> concentration increase to stabilisation. As a result, the year of stabilisation differs with stabilisation levels from around 2150 for 350 ppmv to 2250 for 750 ppmv. Further details on the concentration profiles, and the implied emissions results are given in Enting *et al.* (1994b).

Figure 1.13 shows the model-derived profiles of total anthropogenic emissions (i.e., the sum of fossil fuel use, changes in land use, and cement production) that lead to stabilisation following the concentration profiles shown in Figure 1.12. Initially emissions rise, followed some decades later by quite rapid and large reductions. Stabilisation at any of the concentration levels studied (350–750 ppmv) is only possible if emissions are eventually reduced well below 1990 levels (Figure 1.13). For comparison, the emissions corresponding to IS92a, c, and e are also shown up to 2100 in Figure 1.13. Emissions for all the stabilisation levels studied are lower than those for IS92a and e, even in the first few decades of the 21st century. For the IS92c Scenario, emissions lie between those which in this study eventually achieve stabilisation between 450 and 550 ppmv.

The concentration profiles here are illustrative. Stabilisation at the same level, via a different route, would produce dif-