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Atmospheric Carbon Dioxide[1] and the Global Carbon Cycle: The Key Uncertainites

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ABSTRACT

The biogeochemical cycling of carbon between its sources and sinks determines the rate of increase in atmospheric CO_2 concentrations. The observed increase in atmospheric CO2 content is less than the estimated release from fossil fuel consumption and deforestation. This discrepancy can be explained by interactions between the atmosphere and other global carbon reservoirs such as the oceans, and the terrestrial biosphere including soils.

Over the last ten years, research on global carbon cycle processes has significantly improved our understanding of the increase in atmospheric CO2 concentrations. Nevertheless, key uncertainties remain that must be resolved before atmospheric CO_2 concentrations can be predicted with sufficient accuracy to predict CO_2-climate interactions and to determine the effects of increasing CO_2 concentrations and climate change on vegetation and renewable resources.

Results from studies of past fluctuations in atmospheric CO_2 and climate suggest that current carbon cycle models that emphasize human disturbances may be doing so at the expense of natural feedback components involving both terrestrial and marine systems. Because of current uncertainties in atmospheric CO_2 fluctuations over the last 200 years, the contribution of nonfossil carbon sources cannot be satisfactorily estimated.

Undoubtedly, the oceans have been the most important sinks for CO_2 produced by man. But, the physical, chemical, and biological processes of oceans are complex and, therefore, credible estimates of CO_2 uptake can probably only come from mathematical models. Unfortunately, one- and two-dimensional ocean models do not allow for enough CO_2 uptake to accurately account for known releases. Thus, they produce higher concentrations of atmospheric CO_2 than was historically the case. More complex three-dimensional models, while currently being developed, may make better use of existing tracer data than do one-and two-dimensional models and will also incorporate climate feedback effects to provide a more realistic view of ocean dynamics and CO_2 fluxes. The inability of current models to estimate accurately oceanic uptake of CO_2 creates one of the key uncertainties in predictions of atmospheric CO_2 increases and climate responses over the next 100 to 200 years.

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DISCLAIMER

In describing the various CO_2 exchanges between the atmosphere and terrestrial systems, it is apparent that estimates of the net release of carbon to the atmosphere from deforestation are incompatible with ocean-model CO_2-uptake. Furthermore, the estimated loss rate terrestrial organic carbon from biota and soils (0.03 to 0.1%) is so small compared to the total mass of organic carbon that these losses are largely undetectable and can only be estimated indirectly from models. Much of the uncertainty results from inadequately described deforestation rates, forest carbon stock estimates, and documented land-use changes. In addition, there is the problem of how terrestrial ecosystems respond to changes in climate. Will they release or store carbon? Besides providing allowances for natural annual variability and uncertainty in measuring atmospheric CO_2 levels, a carbon cycle model should reproduce the contemporary CO_2 record while balancing past fluctuations from fossil fuels with terrestrial and ocean components. This is not the case. While there are suggestions as to the cause of this uncertainty, it is apparent that carbon cycle models must take care of existing omissions by including large geologically active pools of living and dead material in vegetation and soils and subfossil carbon pools. Approaches must be adopted that emphasize both data and model improvements, keying in on critical uncertainties, and addressing areas where no easy solutions exist or where knowledge is needed. Finally, ocean CO_2 models must be developed that allow for enough absorption of carbon from the atmosphere to account for past fossil fuel burning and contemporary atmospheric sources.

INTRODUCTION

Measurements of atmospheric CO_2 concentration at Mauna Loa Observatory have shown a steady increase since 1958 (Bacastow and Keeling, 1981; and Gammon et al., 1985). These findings led to an awareness of the need to understand patterns and processes of the global carbon cycle because of the implications for future climatic warming resulting from potential fossil fuel burning.

During the past decade, significant progress has been made in measuring and understanding the global carbon cycle and in developing methods for projecting future changes in the atmospheric CO_2 concentration. Major studies have estimated the sizes of carbon reservoirs (Fig. 1), examined the components of the ocean carbon cycle, estimated the fluxes between the terrestrial biosphere and the atmosphere, and integrated components of the carbon cycle into global models. Progress in each study has not been independent, but has been stimulated by developments in the others. Although some of the estimates of the magnitudes of the fluxes and reservoirs have become more refined over the last decade, the uncertainties are still high (Tables 1 and 2).

THE ATMOSPHERE

The size of the atmospheric carbon reservoir has been quantified since 1958 when Keeling started continuous measurements of atmospheric CO_2 at Mauna Loa Observatory. As of 1980, the atmosphere contained 712 Pg (1 Pg = 10^{15} g) of

carbon (Bacastow and Keeling, 1981). Although an upward trend in atmospheric CO_2 is evident from measurements at all sites over recent decades, the observed increase accounts for only about 58% (with a range of 30 to 80%) of CO_2 released by fossil fuel burning (Trabalka, 1985). For this reason, a better understanding of long-term atmospheric CO_2 trends and the role of the oceans and terrestrial biosphere has been sought.

The long-term record of atmospheric CO_2 concentrations is not well known. Measurements of atmospheric CO_2 levels before 1958 were usually rather inaccurate, because samples were not collected with adequate consideration of sampling methods and locations.

Measurements of atmospheric concentrations of CO_2 have been extended about 30,000 years back in time by the analysis of air trapped in polar ice (Delmas et al., 1980; and Neftel et al., 1982). The mean concentration during the Holocene was about 270 ppmv. Measurements have recently been extended to the period 1750 to 1960 (Neftel et al., 1985). The values determined for samples from the 1950s agree well with the Mauna Loa data, while estimated concentrations during 1750 to 1800 are about 280 ppmv.

THE OCEANS

Because the global ocean is a major carbon storage reservoir, the ocean plays a major role in determining the atmospheric CO_2 concentration. An understanding of processes that regulate ocean-atmosphere exchange of carbon has developed over the past 20 years. Specifically, it was found that ocean vertical gradients of dissolved inorganic carbon (DIC) and alkalinity are crucial to atmospheric CO_2 concentration, and that the carbonate chemistry of the surface ocean determines the rate of uptake of CO_2 by the surface ocean. However, high precision measurements of these properties were not introduced until the GEOSECS program in 1973 to 1978. These data became the foundation for a series of advances elucidating the role of ocean chemistry and circulation in carbon cycling (Broecker and Peng, 1982). Results of systematic measurements of isotope tracers such as ^{14}C, ^{3}H, ^{3}He, and ^{85}Kr during the GEOSECS and TTO expeditions have led to the development of improved ocean models that use these tracers measured at different times to parameterize models of the chemical and mixing processes (Broecker et al., 1985a).

Models of the ocean-atmosphere system have been very useful tools in carbon cycle research. These models evolve and improve as understanding of the ocean-atmosphere dynamic system increases as new field data are acquired. The simplest model is the two-box model consisting of a well-mixed surface layer and a well-mixed deep ocean (Craig, 1957; Bolin and Eriksson, 1959; and Keeling, 1973); however, because of its simplicity, this model is too unrealistic to be used for quantitative projections (Bacastow and Bjorkstrom, 1981; and Bolin et al., 1983).

A major improvement in ocean modeling was the development of the box-diffusion model, which describes the world ocean in terms of a well-mixed surface layer (75 m) and a deep sea reservoir with eddy diffusion vertical transfer (Oeschger et al., 1975). Peng et al. (1983) further improved this model by adding biological cycling, upwelling, and deep water formation, and by increasing the eddy diffusivity in the thermocline. However, results of this model did not show a significant increase in ocean CO_2 uptake as compared with the simpler box-diffusion model. Siegenthaler (1983) extended the box-diffusion model to include the role of intermediate and deep water formation at high latitudes. An upper limit for CO_2 uptake by this type of model is estimated to be 47% of fossil fuel release when model parameters are calibrated with bomb-^{14}C data. Broecker et al. (1985b) developed a lateral transport model of the global oceans, using the distribution of bomb-produced radiocarbon in the oceans as observed during the GEOSECS and TTO programs. This model reproduces a reasonable global tritium distribution in the ocean (Broecker et al., 1986), but it can only take up 35% of the fossil fuel released in the period 1958 to 1980 (Peng, 1986), which is not significantly more than predictions by box-diffusion models.

To better represent ocean dynamics, two approaches have been pursued recently to develop more detailed three-dimensional models. First, a general circulation model (GCM) of the ocean developed on the basis of dynamical principles (Hasselmann, 1982) has been used for a first simulation of the transfer of carbon and ^{14}C (Maier-Reimer, 1984). More improvement is necessary, however, both to ascertain that the ocean circulation models describe the ocean adequately and to include biological processes. Second, Bolin et al. (1983) have approached the problem by working backwards from data to obtain parameters for a 12-box model of the oceans. Attempts at validation of the model with field data have shown that improvement in the spatial resolution of the model is needed. At this time, it is not known how much more CO_2 uptake will be accounted for by these three-dimensional models.

Recent data have indicated that other ocean processes, such as seasonal variation and biological pump, must be considered in understanding the exchange between the ocean and atmosphere. Seasonal variations of the distribution of CO_2 sources and sinks in the global oceans have been observed, and strong latitudinal variations with season have been discovered (Takahashi et al., 1986). Production, decomposition, and dissolution of biogenic material in the ocean also play important roles. Studies have been directed at determining the role of biological pumping, particularly in the major upwelling areas of the Antarctic Ocean (Knox and McElroy, 1984; Sarmiento and Toggweiler, 1984; and Siegenthaler and Wenk, 1984).

In predictions of future atmospheric CO_2 concentrations, it is commonly assumed that the general circulation of the oceans will not change. However, the ice core CO_2 record indicates that the atmospheric CO_2 concentration increased about 80 ppmv at the end of last glacial period (Neftel et al., 1982; Stauffer et al., 1984; and Barnola et al., 1983), which could be the result of an ocean circulation change (Broecker et al., 1985a; and Broecker and Peng, 1986). Results of some model experiments indicate that the possible implications for the global carbon cycle may be significant (Bolin, 1981; and Broecker, 1981).

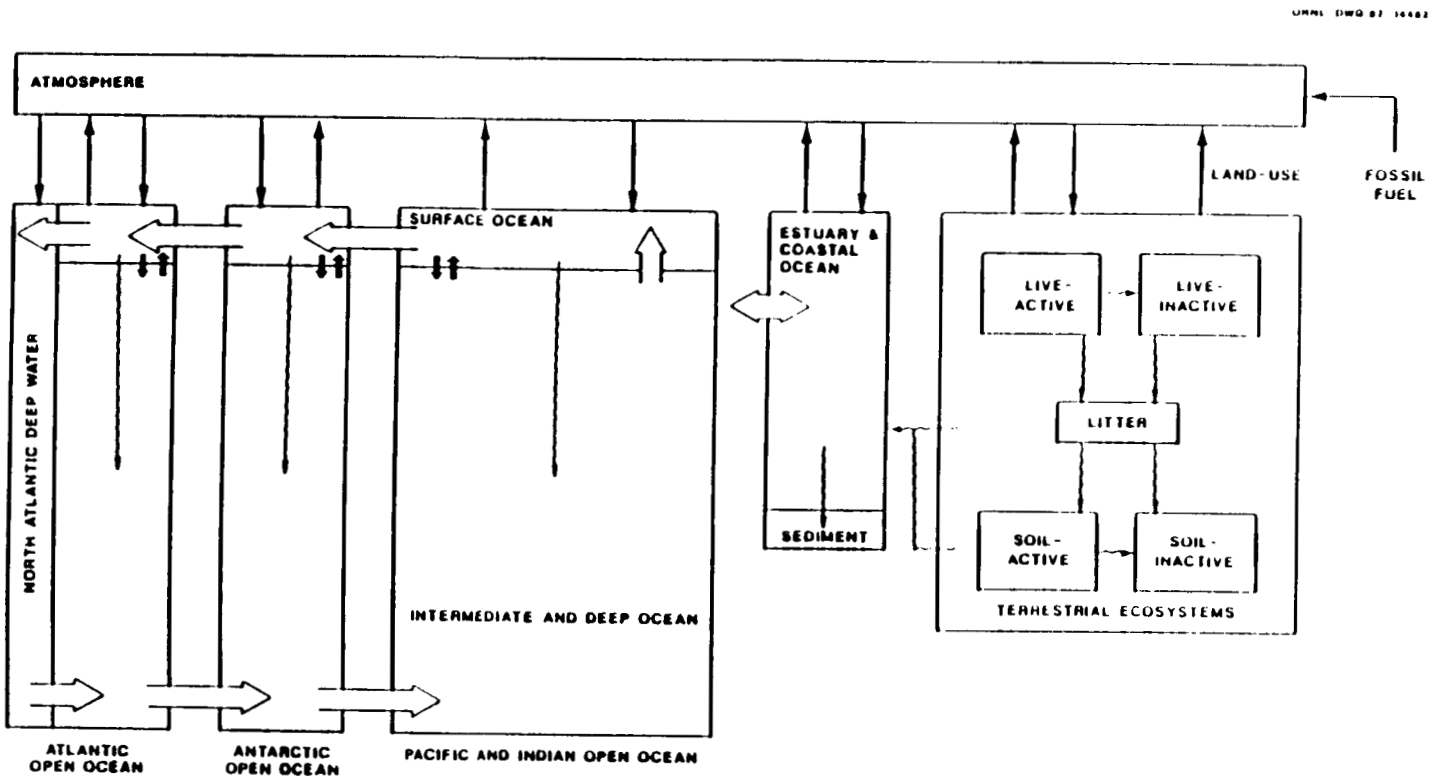

Figure 1. The atmosphere, ocean, and land are three major reservoirs of carbon.

Table 1. Magnitudes of Fluxes (Pg C/yr) in Figure 1.

Flux Type	Reservoirs Involved	Current (1985) Estimate	References	Previous (<1975) Estimate	References
--->	Ocean-Atmosphere				
	North Atlantic deep water uptake Atlantic uptake Antarctic uptake Pacific & Indian uptake	90-104	Peng et al (1979) Solomon et al. (1985) Moore & Bolin (1986) Olson et al. (1985) Bolin (1986)	57-141	Revelle & Suess (1957) Backastow & Keeling (1973) Craig (1957)
	Atlantic release Antarctic release Pacific & Indian release	90-102	Same	57 - 141	same
--->	Estuary & Coastal Ocean - Atmosphere	not available		not available	
-->	Terrestrial-Atmosphere				
	Gross fixation	90 - 130	Houghton et al. (1985)		
	Autotrophic respiration	60		50 - 172(net)	Rodin et al. (1975)
	Current release from land clearing	.9 to 2.9	Detwiller et al. (1985) Melillo et al. (in press) Houghton et al. (1983)	not available	
	Historical release from land clearing (since 1860)	135-228	Houghton et al. (1981)	72	Revelle and Munk (1977)
	Organic matter decomposition				
	Litter	54-60	Schlesinger (1979)	not available	
	Active soil	2-5		not available	
	Inactive soil	< 1		not available	

Table 1. (continued)

Flux Type	Reservoirs Involved	Current (1985) Estimate	References	Previous (<1975) Estimate	References
>	Society-Atmosphere				
	total 1860-1985	140 190	Rotty (1987) Edmonds et al. (1986)	98 126 (up to 1970)	Keeling (1973)
->	Ocean Ocean				
	Particulate (Org. + $CaCO_3$) sinking surface to deep				
	Atlantic Antarctic Pacific & Indian	2 - 4 15 - 22	Moore & Bolin (1986) Baes et al. (1985), Sundquist (1985) Solomon et al. (1985) Chavez and Barber (1987)	1 - 7	Broecker (1971, 1974) Keeling & Bolin (1968) Broecker & Li (1970)
==>	Ocean Ocean				
	Diffusive flux - surface & deep				
	Atlantic Antarctic Pacific & Indian	11	Peng & Broecker (1984)	2.8 - 12	Broecker (1966) Oeschger et al. (1975)
-=-=>	Ocean Ocean				
	Advective flux A_{surf}-NA NA-A_{deep} A_{deep}-AA_{deep} AA_{deep}-PI_{deep} PI_{deep}-PI_{surf} PI_{surf}-AA_{surf} AA_{surf}-A_{surf}	36 - 40	calculated from Peng et al. (1983) Moore & Bolin (1986)	not available	
	Terrestrial-Terrestrial				
- - ->	live-active to live-inactive carbon	15	Bolin et al. (1983), (1986) Emanuel et al. (1984)	not available	These numbers are global aggregates that do not exist outside the framework of a global carbon model. No global carbon model before Emanuel et al. (1984) included more than 2 terrestrial pools. Site specific data was available from IBP in 1975, but not used.
->	litter production from live-active	40	Bolin et al. (1983), (1986) Emanuel et al. (1984)		
- - ->	litter production from live-inactive	15			
- - ->	conversion litter to active soil OM	2-5	Bolin et al. (1983), (1986) Emanuel et al. (1984)		
- - ->	conversion litter to inactive soil OM	< 1	Bolin et al. (1983), (1986)		
- ->	conversion active soil OM to inactive soil OM	< 1	Olson et al. (1985)		

Table 1. (continued)

Flux Type	Reservoirs involved	Current (1985) Estimate	References	Previous (<1975) Estimate	References
	Terrestrial-Estuary & Coastal Ocean				
------>	Dissolved & particulate OM in rivers from:				
	Litter Active soil Inactive soil	1-2	Sundquist (1985) Degens et al. (1981), Solomon et al. (1985), Garrels & McKenzie (1972)	not available	
----->	Estuary & Coastal Ocean to Sediment	1-5	Peterson & Melillo (1984) Dawes (1981), Walsh (1984) Baes et al. (1985) Wollast & Billen (1981)	not available	

Table 2. **Magnitudes of Compartments (Pg C) in Figure 1.**

	Current Estimate	References	Previous Estimates	References
North Atlantic	570	Sundquist (1985)	not available	
Surface Ocean				
Atlantic, Antarctic, Pacific & Indian	inorg. 700-900 org. 25-45 biota 3 (annual increase 0.3)	Olson et al. (1985), Moore and Bolin (1987) Bolin (1986), Bolin et al. (1983) Sundquist (1985)	inorg. 700-900 organic 25-45 biota .4 - 5.0	Broecker (1974) Keeling & Bolin (1968) Bowen (1966), Bogorov (1967), Whittaker and Likens (1973)
Intermediate & Deep Ocean				
Atlantic, Antarctic, Pacific & Indian	inorg. 34,000-36,700 org. 975 (annual increase 2.5)	Olson et al. 1985, Bolin (1986), Solomon et al. (1986)	inorg. 34,000-36,700 organic 975	Broecker (1974) Keeling & Bolin (1968)
Terrestrial photosynthetic	100 - 110	Olson et al. (1978, 1983) Bazilevich et al. (1971)	1850 - 2400	Whittaker and Likens (1973), Rodin et al. (1975)
Terrestrial structural	450			
Litter	60	Ajtay et al. (1979) Olson et al. (1985) Schlesinger (1979)	not available	
Active soil	1200-1400			
Inactive soil	160-1000	Zinke et al. (1984), Post et al. (1982)		
Atmosphere (1860)	270-280	Neftel et al. (1985)	290	Keeling (1973), Bray (1959)
Atmosphere (past variation)	180-300	Neftel et al. (1982)	assumed constant at 290	Keeling (1973), Backastow & Keeling (1973)

THE TERRESTRIAL ECOSYSTEMS

Because storage in terrestrial organic matter may be responsible for part of the excess CO_2 unaccounted for in the atmosphere, there have been several attempts to estimate the storage of carbon in the world's terrestrial ecosystems. Most of these estimates fall within the range of 1200 to 1600 Pg C (Bolin, 1986; and Houghton et al., 1985). The soil carbon pool is believed to be 1400 ± 200 Pg C (Schlesinger, 1977; and Post et al., 1982) Estimates of the amount of carbon in terrestrial biota range from 421 to 827 Pg C (Olson et al., 1983; and Houghton et al., 1985). The differences among estimates are due to methodological differences in (1) classifying ecosystems into types, (2) determining areas of each type, and (3) determining carbon stocks of each type (see Houghton et al., 1985; and Bolin et al., 1986).

The net flux of carbon between the atmosphere and terrestrial ecosystems has been impossible to measure directly, but several methods of indirect measurement have been developed. One approach, deconvolution, has been the use of global models in conjunction with atmospheric, oceanographic, and tree-ring or ice-core data to deduce the terrestrial-atmosphere exchange rate. A second approach, reconstruction, has been to use land-use change data together with consideration of the resultant changes in carbon stored in vegetation and soil.

To understand the historic terrestrial carbon flux, Peng et al. (1983), Peng and Freyer (1986), and Emanuel et al. (1984) combined ocean models and the ^{13}C tree-ring records assembled by Freyer and Belacy (1983) and Freyer (1986). The results of these deconvolution approaches differ dramatically from each other, as well as from the pattern inferred from historical reconstruction (Trabalka et al., 1985). Furthermore, these models must assume a pre-historical atmospheric CO_2 concentration of around 240 ppmv to achieve even partial consistency with the Mauna Loa data. This value is substantially lower than the current estimate of 280 ppmv (Neftel et al., 1985).

There are many technical problems associated with using tree-ring ^{13}C data as a surrogate for direct atmospheric measurements. However, direct atmospheric measurements of historical ^{13}C and CO_2 levels are now becoming available from ice-core analyses. Depending on the ocean models used, deconvolution of pCO_2 record gives a cumulative terrestrial release of about 90 to 150 Pg C as of 1980, of which more than 50% was released before 1900 (Siegenthaler and Oeschger, 1987).

Studies using the historical reconstruction approach have analyzed net carbon released to the atmosphere from land clearing for specific regions of the earth (Detwiler et al., 1985; Molofsky et al., 1984; Houghton et al., 1985; and Melillo et al., in press). The net flux of carbon from the terrestrial ecosystems to the atmosphere from land-use is estimated to fall

within the range of 0.9 to 2.9 Pg C in 1980 (Houghton et al., 1985; Detwiler et al., 1985; and Melillo et al., in press), almost all of which is from the tropics. However, two important processes were not considered in these analyses: (1) increases in carbon storage within existing forests and forest expansion into woodlands and grasslands in the northern hemisphere due to fire suppression, and (2) gradual degradation and the associated reduction of carbon stocks of vegetation and soil from logging, harvesting of fuel wood, deliberate burning, and grazing in tropical open forests.

Four studies have attempted to estimate the net biotic flux based on land-use history since 1860 (Revelle and Munk, 1977; Moore et al., 1981; Houghton et al., 1983; and Richards et al., 1983). In particular, Houghton et al. (1983) built upon the work of Moore et al. (1981) and developed a model that reconstructs yearly changes of carbon in terrestrial ecosystems from 1860 to 1980. That model considers ten geographic regions with up to 14 types of ecosystems, including agroecosystems. The predicted pattern of carbon release to the atmosphere is significantly different than that arrived at by deconvolution. Whether or not closer agreement between estimates from deconvolution and historical reconstruction can be achieved by further refinements of techniques and data is unknown.

Thus, there are great uncertainties in the amount of flux from the terrestrial biosphere to the atmosphere (Table 1). Although current estimates of 0.9 to 2.9 Pg $C \cdot yr^{-1}$ are lower than some earlier estimates, they are still incompatible with estimated ocean model CO_2 uptake.

Because terrestrial ecosystems do not mix or exchange carbon directly, as do different volumes of inter-mixing ocean water, spatial heterogeneity is an important consideration in aggregating the many site-specific measurements of ecosystem carbon dynamics into global models with appropriate dynamics. Recent work has attempted to incorporate spatial detail into analyses. Several global maps of contemporary terrestrial ecosystems at 0.5° latitude by 0.5° longitude resolution have been constructed (Matthews, 1983; and Olson et al., 1983). The relationship between these maps and the reconstruction of land-use based on tabular data by country has not been determined. Satellite imagery also has been used for assessing current vegetation and land-use, although, so far, only for selected areas (Tucker et al., 1985; and Woodwell et al., 1986). Remote sensing has been used to classify current vegetation cover over large areas (Tucker et al., 1985; and Townshend et al., 1986), relate seasonal phenology observed by satellite to atmospheric CO_2 concentration (Tucker et al., 1985), and monitor changes in productivity (Warrick et al., 1986).

Variations in atmospheric CO_2 levels contain information about interactions between the atmosphere and other carbon reservoirs. The amplitude of the seasonal cycle varies from place to place on the surface of the earth. In particular, there is a strong latitudinal pattern, with an amplitude that is very small at the South Pole but that increases northward to a maximum at Pt. Barrow, Alaska (Harris and Bodhaine, 1983; and Gammon et al., 1985). These oscillations are largely due to seasonal patterns of terrestrial vegetation growth and decomposition.

Recent measurements made at several Northern Hemisphere locations show significant growth rates in the amplitude of the seasonal cycle. At Mauna Loa, the mean rate of increase was 0.66% yr^{-1} from 1958 to 1981, resulting in a nearly 1 ppm (from 5.5 to 6.4 ppm) increase in the CO_2 seasonality over this interval. Keeling (1983) suggested that an increasing CO_2 seasonal amplitude indicates increased plant activity, but not necessarily increased net carbon storage. Seasonality of the growing fossil fuel use is insufficient to explain the increase (Pearman and Hyson, 1980).

Modeling terrestrial fluxes is being directed at using the information of CO_2 geographical distribution sources and sinks represented in the seasonal cycle (Fung, 1986; and Fung et al., 1983). Using a three-dimensional atmospheric GCM to advect CO_2 with specified sources and sinks of CO_2 at the earth's surface, it is possible to determine the sensitivity of global CO_2 distribution to various assumptions about CO_2 sources and sinks.

CONCLUSIONS

The above discussion is only an outline of the voluminous research that has gone into global CO_2 research in recent years. Major questions remain. However, the knowledge of major factors affecting the carbon cycle has increased over the last decade (Tables 1 and 2) and will help in directing future efforts towards critical research areas. New measurement methodologies should increase the spatial and temporal resolution and the accuracies of estimates. Continued increases in sophistication of mathematical models will integrate this information into a comprehensive picture of the global CO_2 cycle.

To conclude, several significant improvements in the understanding of the carbon cycle over the last ten years can be listed:

1. A better picture of ^{3}H and ^{14}C distributions in the global ocean through GEOSECS and TTO data has increased the understanding of how the ocean has transferred these tracers, which have great relevance to anthropogenic carbon distribution in the ocean.

2. There is an increasing data base of carbon cycle-related gas tracers (such as Freon, ^{85}Kr, and ^{39}Ar) in the ocean, due to improved measuring techniques.

3. The importance of high-latitude deep-water formation processes in controlling atmospheric CO_2 levels has been recognized. Based on outcrop-box-diffusion model, a maximum of 43% of fossil fuel CO_2 taken up by the ocean could be attributed to deep water formation (Siegenthaler, 1983).

4. Realization has emerged of the role of biological pumping in controlling atmospheric CO_2 concentrations. Based on an 11-box geochemical model of global ocean, the CO_2 content of the atmosphere is increased by 6.7% for each 10% that the ocean content of limiting nutrient is reduced in the absence of other effects (Broecker and Peng, 1986).

5. Based on GEOSECS and TTO chemical data, the Redfield ratio (P:N:C:O_2) has been revised from 1:16:106:138 to 1:16:140:172, thus improving the understanding of the effects of biogenic fluxes on carbon cycle.

6. Ice-core CO_2 data have guided the development of new ideas on the role of paleoocean changes in the history of atmospheric CO_2 changes and in the contribution of the terrestrial biosphere to the increase in atmospheric CO_2. The atmospheric CO_2 concentration increased from 200 ppmv during the last glacial period to 280 ppmv in the present interglacial period before human disturbances.

7. Based on ice-core CO_2 data and ^{14}C variations in the ocean as recorded in forams in deep sea sediments, we learned that the mixing rate and the pattern of ocean circulation during the last glacial period was very different from the current ocean operation. The relationships between ocean operation and atmospheric CO_2 changes in the past provide important guidelines for predicting the future atmospheric CO_2 variation due to climate changes.

8. The concept of different modes of ocean operation being caused by past climate changes has been proposed, though not yet confirmed.

9. Development of an outcrop box-diffusion ocean model (Siegenthaler, 1983) has placed an upper limit on CO_2 uptake amounting to 47% of the fossil fuel release.

10. Seasonality has significant effects on latitudinal distribution of oceanic CO_2 sources and sinks. A strong oceanic CO_2 sink during the summer months could change to a weak CO_2 sink or even become a weak CO_2 source, especially in the high latitude regions.

11. Time history of the terrestrial contribution to atmospheric CO_2 increase since 1800 was obtained using tree-ring ^{13}C records, and ice-core ^{13}C and pCO_2 records.

12. Estimates of terrestrial biomass have been refined from 1000 to 1855 Pg C to 421 to 827 Pg C and those of soil carbon from 700 to 2200 Pg C to 1400 ± 200 Pg C.

13. Maps detailing spatial distribution of terrestrial carbon pools have been developed for geographical analysis and comparison with tabular data on land-use.

14. Historical reconstruction based on tabular data has reduced initial estimates of land-clearing flux from 20 Pg $C \cdot yr^{-1}$ to 0.9 to 2.9 Pg $C \cdot yr^{-1}$. The historical pattern of flux derived from tabular data differs from deconvolution results.

15. River fluxes of carbon to coastal zones have been estimated to average 1 to 2 Pg $C \cdot yr^{-1}$.

16. The seasonal pattern of atmospheric CO_2 has been analyzed at many latitudes for correspondence with short term terrestrial sources and sinks of CO_2 and it has been suggested that an increasing CO_2 seasonal amplitude indicates increased plant activity, but not necessarily increased net carbon storage (Keeling, 1983).

The major uncertainties remaining stem from an inadequate knowledge of the following processes:

1. Rates of vertical and horizontal mixing in the ocean.

2. Rates of upwelling and its spatial distribution in the ocean.

3. Rates of deep water formation in the North Atlantic and in the Antarctic.

4. Degree of CO_2 equilibrium between atmospheric and oceanic CO_2 in the source regions of deep water formation before it sinks and is transported in the deep ocean.

5. Wind speed dependence of CO_2 exchange rates across the sea-air interface and the spatial distribution of these rates.

6. Net biogenic carbon flux from the surface productive layer to the deep ocean.

7. The sensitivity of marine primary production to changes of nutrient availability in the surface waters.

8. Short term changes (decades) of alkalinity, and thus the buffering factor of sea water due to an increasing amount of DIC.

9. Effects of increasing DIC due to carbonate dissolution for a long term perspective (centuries) on carbonate chemistry in the ocean.

10. Effects of seasonal variations in the zonal distribution of oceanic CO_2 sources and sinks on global CO_2 uptake.

11. Effects of climate changes on ocean circulation and the carbon cycle.

12. Fertilization and the increase of biomass and organic matter in soils in terrestrial ecosystems from increasing atmospheric CO_2 concentrations and, possibly, deposition of nutrients emitted from anthropogenic sources.

13. Enhanced rate of decay of organic matter in soils, particularly during the process of forest exploitation and conversion of natural vegetation to cultivation.

14. Burying rate of organic matter in the sediments of estuaries and coastal oceans.

15. Uncertainty in the land-use flux from lack of information concerning the pre-1860 standing stocks of carbon, the rates of land clearing and abandonment, the rate of biomass recovery after disturbance, and the degree of decrease in carbon stocks due to disturbance.

16. Effects of climate change on distribution, dynamics, and storage capacity of terrestrial ecosystems, particularly carbon rich boreal and tundra systems.

17. Effects of climate change on the terrestrial hydrological cycle, which affects both the rates of carbon storage in organic soils and the flux of carbon and nutrients to the estuary and coastal oceans, thereby altering the rate of sediment formation.

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