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ATMOSPHERIC CARBON DIOXIDE AND THE GLOBAL CARBON CYCLE

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MASTER

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FOREWORD

Over the ages, human communities have had little or no effect on the Earth's global climate. Humans have accounted for only a small part of all of the species on the planet, and their activities have been essentially benign relative to the global atmosphere. Historically, communities were small and distant from each other and transportation was slow and difficult. Very little energy was consumed, with the burden of work being carried by humans and their domesticated animals with the assistance of elementary machinery.

Science and technology paved the way for the rapid societal changes of the 20th century. With the development of transportation and communication systems plus the machinery for industrial and agricultural production, global energy consumption grew by more than 10-fold from 770 million metric tons (10^{15} grams) of coal equivalent (mmtce) in 1900 to more than 9000 mmtce in 1984. Most of this energy was produced by burning fossil fuels. The world's population increased by threefold during the same period from 1.6 billion to 4.8 billion. (The American billion, 10^9, is the same as the British milliard, 10^9.) Urbanization, which resulted in a major increase in demand for energy, recorded an almost 20-fold increase in 84 years, as measured by the number of urban areas with populations greater than 1 million, expanding from 13 to 247 worldwide.

This immense increase in energy use is changing the layer of gases that constitutes the Earth's atmosphere, which in turn controls global climate. So, for the first time in the planet's history, humans are truly involved in a change of their environment.

Carbon dioxide (CO_2), a naturally occurring constituent of the atmosphere, is also the product of human activities—the burning of fossil fuels for energy and the clearing of land for agriculture and urbanization. Toward the end of the 19th century, Arrhenius calculated that a doubling of the CO_2 concentration would raise the average temperature by 5° or 6°C. Chamberlain later developed a hypothesis that related causes of glacial periods and the depletion of the atmospheric CO_2 concentration. Tolman, a student of Chamberlain, described the basic roles of the ocean in absorbing atmospheric CO_2 and in moving and storing it globally.

Another important role of CO_2 was also recognized more than a century ago. Von Liebig demonstrated that plants get their carbon for photosynthesis and growth from the air. By this process, a relative atmospheric constancy is maintained where assimilation by plants is roughly balanced by CO_2 exhalation by animals. This notion of constancy became dogma until modern measurements clearly showed a change in the atmospheric concentration of CO_2 due to human intervention.

It is now known that the atmospheric CO_2 concentration in 1900 was approximately 300 parts per million by volume (ppm) (indicated by recent measurements of glacial ice cores by Oeschger). However, it wasn't until 1938

that Callendar presented the first substantive data showing that the concentration of CO_2 in the atmosphere was increasing and suggested that the increase might affect the Earth's climate. After another 20 years Keeling began to monitor the atmospheric CO_2 concentration at the Mauna Loa Observatory in Hawaii. The measurements of atmospheric CO_2 in 1958 showed the annual average concentration was 316 ppm; it was approximately 345 ppm in 1985. Plass outlined theories to explain the relationship between atmospheric CO_2 and climate in 1956, and, soon after, Revelle and Suess described the relationship between CO_2 in the atmosphere and in the oceans, and Kaplan enlarged upon the role of CO_2 in the atmosphere in terms of the global heat balance.

In 1977 leading scientists assembled in Miami Beach, Florida, to discuss the current understanding of the carbon cycle, that is, the dynamics of carbon exchanges within the Earth's atmosphere, land, and oceans that determine the atmospheric CO_2 concentration. They also reviewed possible consequences of increases in atmospheric CO_2. In addition, they identified significant gaps in the knowledge base and made recommendations for research.

Since then, significant research has been carried out by the international scientific community. The Department of Energy (DOE), the lead United States agency in the study of CO_2, and other agencies including the National Science Foundation, National Oceanic and Atmospheric Administration, National Aeronautics and Space Administration, United States Geological Survey, United States Department of Agriculture, and Environmental Protection Agency, following the recommendations of the science community, conducted and have supported research activities in universities, national laboratories, industry, and other institutions.

Looking forward to the 21st century, the DOE believed it was important to "take an accounting" to see how far this considerable effort had come in 8 years in answering the questions that were previously posed and in determining future research directions. Accordingly, the Carbon Dioxide Research Division, Office of Basic Energy Sciences, of the DOE is publishing this series of four State-of-the-Art (SOA) volumes:

- *Atmospheric Carbon Dioxide and the Global Carbon Cycle* — to understand the mechanics of and quantify the sources, sinks, and exchanges of carbon between all elements of the global carbon system — the atmosphere, the biosphere, the oceans — including anthropogenic effects.
- *Projecting the Climatic Effects of Increasing Carbon Dioxide* — to project the magnitude and rate of the potential climate changes that could result from the increasing atmospheric CO_2 concentration.
- *Detecting the Climatic Effects of Increasing Carbon Dioxide* — to detect the changes in climate resulting from the increasing atmospheric CO_2 concentration and to isolate the climate changes from those caused by other contributing factors (natural or anthropogenic).
- *Direct Effects of Increasing Carbon Dioxide on Vegetation* — to determine the plant response to increased atmospheric CO_2 and develop the capability to predict crop and ecosystem responses to CO_2 enrichment.

An index and cross-reference volume accompanies the set of volumes. Two companion reports are also being published:

- *Characterization of Information Requirements for Studies of CO_2 Effects: Water Resources, Agriculture, Fisheries, Forests, and Human Health.*

- *Glaciers, Ice Sheets, and Sea Level: Effects of a CO_2-Induced Climatic Change*, from the National Research Council's (NRC) Committee on Glaciology of the Polar Research Board.

These complementary reports aid in ensuring that "the accounting" of CO_2 research activities for the past years encompasses the entire spectrum of research.

The SOAs document what is known, unknown, and uncertain about CO_2 data, analyses, and modeling capabilities. They outline potential avenues of research for reducing critical unknowns and uncertainties. More than 70 scientists from five nations have participated in the preparation of these volumes. Each chapter and each complete SOA volume has gone through extensive peer review by the American Association for the Advancement of Science (AAAS); this review, however, does not imply that AAAS endorses the statements or recommendations presented in these volumes.

These technical reports provide the basis for a Statement-of-Findings (SOF) report. While studies over the last several years have clearly shown that increasing CO_2 concentrations have the potential for significant impacts on our physical environment, these studies have not yet provided an adequate basis for addressing questions about the fundamental relationships between the benefits and impacts of various energy systems on society's activities. The SOF will summarize what we know and do not know and the degree of certainty of our knowledge. It will also present the rationale for further studies. These studies will be needed to provide an accurate scientific basis for assessments of the potential impacts of energy-related activities.

The citizens of today's nations have the responsibility for the stewardship of all the Earth, including their actions which may affect its climate. Exercising this responsibility requires an understanding of atmospheric CO_2 and its effects. Once understood, stewardship then becomes nurturing rather than unrecognized neglect.

Scientists have created the building blocks for this understanding, and the scientific community has recognized its responsibility to more fully understand CO_2-induced effects on our global environment. Through research, as we look towards the 21st century, the application of science will ensure that the additional understanding required for nurturing our planet Earth will be developed.

Sincere thanks go to everyone who has participated in developing the SOAs and companion reports. Special thanks goes to the coordinator/editors, Jennifer D. Cure, Frederick M. Luther, Michael C. MacCracken, Boyd R. Strain, John R. Trabalka, Margaret R. White, and the NRC Committee Chairman Mark Meier; their respective chapter authors; and to the AAAS, Roger Revelle, Chairman of the Climate Committee, and David M. Burns of the AAAS staff.

We hope these definitive, scientific statements will motivate scientists to recommend explicit approaches for reducing the critical uncertainties that now exist in order to permit decision making within the next decade that is based on data, learning, understanding, and wisdom.

Frederick A. Koomanoff, Director
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EDITOR'S PREFACE

The State-of-the-Art (SOA) reports represent the culmination of a major effort to review the state of scientific understanding concerning the potential effects and impacts of increasing carbon dioxide (CO_2) concentrations. Preparation of these SOA reports has taken several years and would not have been possible without the cooperation and substantial effort provided by many participants, both within and beyond the DOE-funded research program.

All sections of this volume (including the outline) have been subjected to an extensive peer review process. In addition to guidance provided by the editor's panel of reviewers (see listing in this volume), the Department of Energy contracted with the American Association for the Advancement of Science (AAAS) to arrange for external peer review of the outline, individual chapters, and the assembled volume following chapter revisions. The reviewers provided many important comments and suggestions that have led to significant improvement of the volume. Their assistance has been greatly appreciated.

I wish to gratefully acknowledge the dedicated efforts of all of the chapter authors, particularly the lead authors, in responding to numerous reviews and to my questions and inquiries. Their patient and dedicated response to over 20 reviews for each chapter, to extensive editorial suggestions, and to requests to aid in review and preparation of the executive summary and synthesis/recommendation chapters has been greatly appreciated.

I also want to express special gratitude to the many secretaries and word processing staff who have labored on these chapters at the various institutions. At Oak Ridge National Laboratory (ORNL), I would like to single out Ms. Donna Rhew and Mrs. Linda Littleton, Environmental Sciences Division Word Processing Center, and Mrs. Marvel Burtis and Ms. Joanne Sanford of my own staff for their patience, perseverance, and tolerance. I would also like to acknowledge gratefully the contributions of our technical editor, Ms. Natalie Tarr and our graphic arts coordinator, Mr. John Holbrook, both from ORNL's Information Division.

Editorial support and handling of the manuscripts for the AAAS review process have been directed by Mr. Jon Findley and co-workers of The MAXIMA Corporation in Bethesda, Maryland. The index was prepared by Mr. Fred O'Hara through a subcontract with the Oak Ridge National Laboratory.

To everyone concerned, from our review panel members to our word processing staff, a grateful thank you for a very difficult, often trying, job well done.

John R. Trabalka
Editor

AAAS REVIEW

In commissioning and publishing a summary of the state of current knowledge about the secular increase of atmospheric carbon dioxide and its effects, the Department of Energy has performed a valuable service. These volumes may prove to be the most comprehensive assembly to date of scientific results about this issue.

The Committee on Climate of the American Association for the Advancement of Science was asked to organize peer reviews of the component chapters of the books. These reviews were conducted in the traditional manner of refereeing scientific papers. We identified experts whom we knew to be well qualified, and we invited them to review anonymously an individual chapter in their field. In transmitting draft manuscripts to the reviewers, we alerted them to the ambitious nature of the "state-of-the-art" project, and particularly to the difficulties of adequately treating the many uncertainties.

The careful attention reviewers devoted to their tasks was gratifying, and indicates the importance of this issue to the world scientific community. More than 300 specialists from 23 countries gave the draft papers a careful and thorough reading and offered detailed suggestions for revision and improvement. The authors and editors thus had available a significant input from their professional colleagues as they sought to improve their drafts. But the decision as to how to use the reviewers' suggestions was the responsibility of the author(s) of the paper and the editor(s) of the books.

These volumes make clear that investigating the causes and effects of alterations to the atmosphere is an exceedingly complex undertaking, touching a wide gamut of scientific disciplines. It hardly is surprising that there were (and are) differences of interpretation.

I am grateful to the many anonymous reviewers and to my colleagues on the Committee. I hope that we have been helpful to the authors and editors in their very challenging task.

Roger Revelle, Chairman
AAAS Committee on Climate

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EXECUTIVE SUMMARY

Carbon dioxide (CO_2) in the atmosphere appears to play a major role in determining the Earth's climate and habitability through regulation of the solar radiation balance. Atmospheric CO_2 levels have increased approximately 25% from 1800 to 1985, attributable mainly to human influences, first from deforestation during the massive global expansion of agriculture, and now primarily from fossil fuel burning. A significant and disproportionate fraction of the overall increase has occurred, since the beginning of systematic atmospheric CO_2 measurements in 1958, during a period of unprecedented fossil fuel use. A doubling of the atmospheric CO_2 level of 1800 may produce a global average temperature warmer than during any period in the last 100,000 years or more, and the CO_2-induced warming ("greenhouse effect") could be augmented significantly by increasing levels of other atmospheric trace constituents.

This volume focuses on the global cycle of carbon, the dynamic balance among global atmospheric CO_2 sources and sinks, which determines the rate of increase in the atmospheric CO_2 concentration. The observed increase in CO_2 content of the atmosphere is less than the estimated release from fossil fuel consumption and deforestation because of interactions between the atmosphere and other global carbon reservoirs. These interactions must be understood to provide a basis for developing models that will project future changes in atmospheric CO_2 concentration as fossil fuel use continues. The carbon reservoirs that are known to be important in the response to anthropogenic CO_2 are the atmosphere, the ocean, and the terrestrial biosphere including soils.

Predictions of future atmospheric CO_2 concentrations must be based on carbon cycle models that accurately depict quantitative rates of carbon exchange among the major global reservoirs. As a first approximation in the validation of models, it should be possible to compute a balanced global carbon budget for the contemporary period; to date this has not been achievable and the reasons are still uncertain. Terrestrial land-use changes, principally deforestation, have contributed to the rise in atmospheric CO_2 levels, but current mathematical models of the global carbon cycle cannot accommodate contemporary biospheric releases in addition to fossil fuel releases. Thus, these models produce estimates of past atmospheric CO_2 levels that are inconsistent with the historical atmospheric CO_2 increase. This inconsistency implies that significant errors in projections are possible using current carbon cycle models.

Detailed histories of atmospheric CO_2 obtained from ice cores indicate that significant nonfossil CO_2 sources existed in the 19th century. Strong evidence exists that deforestation in the 20th century has been as intense as that in the 19th century, even though the focus of clearing has shifted from temperate to tropical regions. Thus, there appear to be several possible, perhaps complementary, explanations for the observed inconsistency in modeling: (1) undiscovered errors in the time sequence of terrestrial land-use changes; (2) existence of additional

contemporary sinks not represented in current models; and (3) masking of existing sinks by oversimplified terrestrial and ocean models. All three could be valid, each partially obscured by the others, as one simple answer.

Although the carbon cycle has been controlled by a system of checks and balances, small, naturally occurring differences in the annual sums of sources and sinks led to significant changes in atmospheric CO_2, long before land-use changes and fossil fuel burning by humans began. Over the past 100 million years, atmospheric CO_2 levels are thought to have been regulated primarily by the balance between (1) CO_2 releases from biogenic carbonate precipitation in the ocean and from tectonic processes and (2) CO_2 uptake by organic carbon burial in sediments and by mineral weathering processes. Natural changes in the rates of these processes or in the resultant flux differences (respective sums of sources and sinks) are generally thought to be too slow to significantly affect atmospheric CO_2 levels over the next few centuries. However, it is not yet clear that we have a satisfactory understanding of what drives this slow-breathing system upon which shorter term human perturbations have been imposed, this flywheel that has kept the Earth habitable for billions of years. Human effects on atmospheric composition and on the size and operations of the terrestrial ecosystems represent major perturbations to the life-support system crafted in nature. The opportunity to provide more evidence of its self-regulation (e.g., through natural feedbacks involving the biota) should be welcomed, especially if that were to moderate the human cost of regulation or adaptation to change.

Understanding the carbon cycle of the globe requires very different approaches than does the solution of more traditional, experimental scientific problems. Industrialization and changes in the landscape have disturbed natural biogeochemical cycles of carbon and nitrogen, another element of climatic interest, for several centuries. Evaluating what happened and what may happen in the future requires understanding of previous conditions that existed when there was less fuel burning, forest harvesting, or farming, reaching back past millennia of human population expansion, past the glaciations of the last million years, and through many millions of years while fossil carbon was being banked in the stratigraphic column. Development of a knowledge base for the carbon cycle requires that (1) a variety of recorded historical and contemporary changes in atmospheric CO_2 concentration needs to be understood; (2) the fossil fuel, biospheric, and oceanic interactions with the atmosphere, expressed as fluxes, need to be quantified; and (3) models that can accurately represent the major fluxes over time need to be developed for interpreting the past and predicting the future.

HISTORY OF ATMOSPHERIC CO_2

From 100 million years ago until the end of the last ice age 10,000 years ago, very large and as yet poorly understood changes in the atmospheric CO_2 level are thought to have occurred. From an initial concentration perhaps as high as several thousand parts per million (ppm) suggested by long-term geochemical modeling exercises, the atmospheric CO_2 level fell gradually toward much lower values of 200-300 ppm characteristic of the glacial-interglacial cycles of the past million years. Analyses of sedimentary carbon isotope records indicate that the CO_2 level varied in a regular, periodic fashion, cycling at the ice age frequency of about 100,000 years between a low near 200 ppm in cold, glacial phases and a high near 270 ppm (but up to 350 ppm) in warm, interglacial phases, several times during the last million years.

The anthropogenic increase in atmospheric CO_2 concentration from within a range of 260-285 ppm in 1800 to 345 ppm (est.) in 1984 is approaching, but has not yet exceeded, the apparent limits of natural variation characteristic of the glacial-interglacial cycles of the past million years. Past relationships between atmospheric CO_2 levels, reservoirs of the global carbon cycle, and climate change have been inferred from a variety of methods, including ice core records for the period from the last major glaciation to the present day. An accurate characterization of such relationships may be critical for understanding some future responses of the carbon cycle and climate to human perturbations. The ice core records appear to provide the best source of information on atmospheric CO_2 (and trace gas) fluctuations over the 50,000-year period prior to the Mauna Loa record. Currently inexplicable, but provocative, rapid fluctuations ($\sim$70 ppm over a century), strongly correlated with oxygen isotope indices of climate change, are recorded in a Greenland ice core formed during the last full glacial period.

On time scales of recent decades to centuries, for example, from 1800 to the present, human impact on atmospheric CO_2 and the global carbon cycle becomes clearly measurable in ice cores, in tree rings (carbon-14 depletion), in the atmosphere, and in the surface ocean. A much-expanded, international CO_2 measurements network is now documenting a global concentration increase, at an annual rate averaging 1.5 $\pm$0.2 ppm (3.2×10^{15} g C). The atmospheric increase since 1958 represents an amount equivalent to 58 $\pm$5% of the total fossil fuel CO_2 released. This atmospheric record shows potentially important long-term changes, for example, the increasing amplitude of the seasonal cycle and year-to-year variations whose significance is not yet well understood. A synthesis of the atmospheric measurements from around the world currently represents the most precise and integrated record available of the workings of the global carbon cycle, of the fluctuating regional CO_2 exchanges of the atmosphere with both the ocean and the terrestrial biosphere, and of the response of this complex system to strong perturbations of both natural and human origins.

The available data on past fluctuations in atmospheric CO_2 and climate suggest that our current carbon cycle models, which emphasize human perturbations, may be missing natural feedback components involving both terrestrial and marine systems, perhaps even climate-induced "mode switches" in ocean circulation patterns, which could be very important in understanding changes in both climate and the carbon cycle over the next century.

Recent ice core research has narrowed the uncertainties in the historical record of atmospheric CO_2 variations over the past 200 years, but current uncertainties still do not permit an adequate definition of the magnitude and timing of either nonfossil CO_2 sources or carbon sinks during that period. The contemporary atmospheric measurements represent the principal test data for validation of carbon cycle models. The relative importance of enhanced photosynthetic uptake, enhanced respiration, wildfires, and human disturbance to the terrestrial biosphere either today or in a potentially warmer world of higher greenhouse gases in the future cannot yet be predicted with confidence, and may only be determinable from analysis of the detailed global CO_2 record itself.

FOSSIL FUEL CO_2 EMISSIONS AND RESOURCES

Global emissions of CO_2 from fossil fuels increased regularly at a 4.4% annual rate from an estimated release of 1.6×10^{15} g C/year in 1950 up to the time of the 1973 oil embargo. Since then the rate of growth has been inconsistent. Total emissions have been relatively stable ($\sim 5 \times 10^{15}$ g C) for the past 5 years; annual changes have remained within the estimated 6-10% uncertainties. A variety of

factors, including price motivation, has driven energy consumers to conservation, technological innovations, and fuel substitutions (e.g., coal) that require less oil and natural gas. Coal consumption has increased steadily since 1973. The slowed growth rate in emissions has lengthened the time until any specified atmospheric concentration of concern (e.g., twice the year-1800 level) is reached. However, the known recoverable resources of fossil fuels are so vast ($\sim$4000 x 10^{15} g C) that it is highly likely that large quantities will be utilized and that much higher atmospheric CO_2 concentrations will result in time. Refining current estimates of these resources is deemed unlikely to have an appreciable effect on uncertainties in atmospheric CO_2 over the next century.

The fossil emissions time series differentiated on a geographical and seasonal basis, and revised and updated annually, are key reference sources for both emissions and carbon cycle modeling.

UPTAKE OF CO_2 BY THE OCEANS

The oceans, particularly the areas of open ocean, have been the most important sink for anthropogenic CO_2. Knowledge of oceanic processes that determine CO_2 uptake has come mainly from extensive ocean surveys, conducted mostly in the last decade. These surveys have ascertained in varying detail the distribution of chemicals and radionuclides (tracers) that can be used to determine ocean circulation and atmosphere-ocean exchange patterns and chemical and biochemical equilibria needed to estimate CO_2 uptake. These include naturally occurring tracers and transient tracers such as chlorofluorocarbons, tritium, bomb-^{14}C, and excess carbon.

Given the intricacies of ocean circulation, chemistry, and biology, credible estimates of CO_2 uptake can come only from models. The relatively simple, one-dimensional (depth) models initially developed do not take up enough anthropogenic CO_2 to accurately account for known releases and, thus, produce higher concentrations of CO_2 in the atmosphere than was historically the case. Models of intermediate complexity, initial two-dimensional formulations, have generally taken up no more CO_2, and sometimes less, than the simpler models. This has provided further impetus for preliminary developmental work on more elaborate models, which treat chemical and biological processes in three-dimensional representations of ocean circulation. Such models can better utilize existing tracer data for their validation and can begin to incorporate potentially important climate feedback effects on CO_2 uptake. Enhanced carbon storage on continental shelves (0.2 x 10^{15} to 0.8 x 10^{15} g/year), attributed to increasing carbon and nutrient content of river outflows and to carbonate mineral reactions in sediments, could be enhancing contemporary ocean uptake but has not yet been included in models because uncertainties are very large.

Uncertainties about the adequacy of treatment of the oceans in existing models are responsible for a large portion of the uncertainties in predictions of both atmospheric CO_2 increases and climate responses over the period of the next 100 to 200 years. In addition to making fuller use of the expected increase in transient tracer data, more sophisticated models now being formulated should benefit from future direct observations of oceanic carbon uptake and changes in alkalinity (from dissolving calcium carbonate in sediments). The spatial and temporal resolution needed in data to support future modeling is not yet clear but should be clarified as the development of models proceeds. The detailed information gained from oceanic surveys cannot be interpreted without models, and there is little hope of developing more sophisticated models without much more tracer data.

CO_2 EXCHANGE BETWEEN THE ATMOSPHERE AND TERRESTRIAL SYSTEMS

The net release of carbon to the atmosphere from deforestation in 1980 is currently estimated at 0.6 x 10^{15} to 2.6 x 10^{15} g. Longer term calculations indicate a net release of 90 x 10^{15} to 180 x 10^{15} g C between 1800 and 1980. Both the ranges and the high values are smaller than earlier estimates but are still incompatible with ocean model CO_2 uptake. The total mass of organic carbon in the world's terrestrial biota and soils ($\sim$2000 $\pm$400 x 10^{15} g) is so large and heterogeneously distributed that losses at the estimated rate of 0.03-0.1% per year (circa 1980) are currently undetectable. These losses are, therefore, calculated indirectly through models, from the estimated transformation rates of natural ecosystems to human ecosystems (e.g., agricultural lands), on the basis of changes in the relative stocks of carbon in these two types of systems.

The stated uncertainties in these loss estimates result from different assumptions about types of vegetation converted to cropland or pasture, extent of historical degradation of forest carbon stocks, and current carbon stocks and deforestation rates in tropical regions. Some residual variation will remain in the estimates of both current and long-term fluxes of carbon because of the difficulty in documenting past changes in ecosystem carbon stocks and areal extents. Additional uncertainties are contributed by the modeling assumptions currently employed to calculate terrestrial carbon flux estimates. Significant questions exist about the adequacy of the "steady-state" paradigm for undisturbed ecosystems and existing concepts of succession and carbon dynamics in disturbed ecosystems. Climate change could significantly alter the role of the biota in the global carbon cycle, and the biotic flux from deforestation could become secondary to that produced by future global warming.

> *There is considerable doubt (and speculation) about whether the current terrestrial CO_2 release estimates represent the net flux of carbon between terrestrial systems and the atmosphere or whether there are other sources or sinks from changes in currently undisturbed areas, for example, from climate or CO_2 fertilization effects. Currently, there is no direct evidence for any such changes, and although no consensus exists on estimates of likely future changes, it is recognized that these uncertainties represent major unresolved issues in carbon cycle research.*

THE NATURAL CARBON CYCLE

Return of carbon from the atmosphere into the rock cycle during the last 100 million years is currently estimated to be a small fraction of the 0.2 x 10^{15} to 1 x 10^{15} g/year (or more) that has been deposited recently in peats, alluvia/colluvia, and other unconsolidated sediments. Such "subfossil" deposits may have been out of balance for several millennia, and this imbalance may account for a significant part of the mismatch that has emerged from modeling attempts to reconcile trends of carbon in the atmosphere, terrestrial biosphere, and oceans. Uncertainties about the releases of methane and CO_2 from the subfossil pools and the linkages between their respective cycles (even in the natural condition), the effects of varied human manipulations of rivers and their watersheds, and the transience of such phenomena raise very difficult, yet important, questions. Our existing data bases and perspectives are very poor, particularly for phenomena linking inland catchments and river floodplains with estuaries and coastal deltas.

Natural, localized fluctuations in CO_2 exchanges obscure global changes that may already have occurred as a by-product of human intervention. For example,

changes in the frequency of wildfires, conditioned by climatic alternations of global or zonal scope (e.g., El Niño/Southern Oscillation events), could have contributed to an important nonfossil fuel complication of the recent atmospheric CO_2 record. But stochastic contributions from fires and changes in natural burning rates through human control also complicate the estimation of both the global biospheric respiration CO_2 source and the net biotic flux with the atmosphere. The role of animals and decomposers in modifying the biospheric respiration flux of over 50 x 10^{15} g C/year remains a basic, unresolved issue—one which must be addressed to understand the impacts of increasing CO_2 and climate change on ecosystems. These examples serve to highlight the fundamental, unresolved problems involved in relating a global synthesis to the local scale at which most carbon cycle measurements or observations are made.

Ultimate reduction of the contemporary imbalance and of carbon cycle model contributions to atmospheric CO_2 prediction errors may not be possible without more basic understanding of fundamental ecosystem patterns and processes (e.g., natural controls on terrestrial carbon storage, role of subfossil carbon pools) and of relationships among ocean CO_2 uptake, climate, and ocean circulation. Lack of relevant field data is the principal obstacle to this understanding. Rethinking of current atmospheric CO_2 reconstructions and of the omission of a land-ocean interface in models is also indicated.

MODELING FUTURE GLOBAL ENERGY AND CO_2 EMISSIONS

The principal uncertainty governing atmospheric CO_2 levels a century from now appears to be the future rate of fossil fuel CO_2 emissions. This, in turn, is related to uncertainties in projected energy use and its determinants: the spectrum of potentially available technologies and fuels as well as a host of difficult-to-predict geopolitical, economic, social, and demographic factors. These currently include the nature of any global response to a climate change only dimly perceived at present. Because all aspects of this complex "human equation" cannot be foreseen, as graphically illustrated by the aftermath of the 1973 oil embargo, much of this uncertainty appears to be irreducible. The modeling of emissions nevertheless provides a necessary tool for treating these uncertainties quantitatively, even if the potential for accurate foresight is slim.

Investigations into uncertainty in patterns of future global energy production and use over the next 50 to 100 years have taken the form of either forecast studies that provide estimates of future CO_2 emissions likely in the absence of explicit control policies or feasibility studies that assess patterns of energy and economic growth consistent with explicit CO_2 emissions constraints. Differences in structure and basic assumptions among current emissions models, however, are such that the key factors in CO_2 releases, while they can be isolated, cannot now be ranked on a quantitative basis.

The rate of global economic growth will be one key factor and will depend heavily on the rate of economic development of currently less-developed countries and the energy demands and choices of energy technologies in that development. End-use energy efficiency improvements also rank high on many studies' lists of important factors. Potential changes in the technologies of both energy end use and supply and the possibilities in interfuel substitutions also must be considered, as such factors add to forecasting difficulties. Continued trends toward use of electricity as a preferred form of end-use energy offer opportunities to meet world needs and ensure economic growth with lower CO_2 emissions, whereas a heavy reliance on coal- or shale-derived synthetic fuels would have the opposite effect. However, future energy supply systems with lower emissions per unit of energy use

seem probable as technologies, including nonfossil-based ones, emerge or evolve and as human society awakens to the expected demonstration of measurable CO_2-climate change.

Forecast studies have undergone a dramatic evolution. The dominant techniques of analysis have shifted from time-trend and logistic extrapolations to economic and systems analysis. There has been a concurrent decline in rates of growth of forecasted CO_2 emissions. Prior to 1980, published studies reported annual rates of growth in CO_2 emissions between 2.0 and 4.5% per year. By 1984 the bounds of discussion had shifted to between 0 and 2.0% per year. This trend has occurred concurrently with the introduction of formal techniques of economic analysis; hence a more explicit treatment of key factors. Feasibility studies have consistently concluded that low or declining growth rates in CO_2 emissions are possible if energy conservation and shifts to CO_2-benign modes of production are adopted. Economic costs are generally found to be a small portion of the global sum of gross national products. Although conceptually quite different from forecast studies, the operational distinction between feasibility studies and forecasts is imprecise, and the quantitative differences in projected emissions growth rates between these two classes of studies has diminished. Techniques for describing and analyzing uncertainty have evolved from simple alternative scenario construction to formal Monte Carlo simulations.

Despite an apparent recent convergence in expert opinion about expected future emissions growth rates, a large band of uncertainty remains. The growth rate of global future CO_2 emissions is expected to average approximately 1% per year over the next 75 years. In the absence of explicit policies to control emissions, the 90% confidence bounds are 3.5% per year to −1.0% per year. This excludes a return to the 1950 to 1973 time trend (4.4% per year). Further, low rates of growth (<2.0% per year) in long-term global CO_2 emissions now appear to be possible even without explicit policy intervention. The field of policy analysis has remained largely unexplored.

PREDICTING FUTURE ATMOSPHERIC CO_2 LEVELS

The appearance of high uncertainty in *modeled* fossil fuel emissions and the seemingly high contributions to uncertainty in future atmospheric CO_2 levels are both artificial and potentially misleading. This is because, for a variety of reasons, society will not be free to choose any future path within the current range of potentials. Describing a given emissions scenario as a real possibility in a modeling exercise does not necessarily mean that society could achieve or would want to achieve that scenario, particularly given the implications of CO_2-climate responses for high-CO_2-release scenarios. Decisions and events already have limited the future development of some energy sources; recent trends are toward increased energy conservation and improvements in end-use energy efficiency.

The problems inherent in modeling energy/economic/CO_2 emissions futures on century time scales and, to a lesser extent, current uncertainties in modeling the global carbon cycle make it difficult to accurately project atmospheric CO_2 levels. Despite this, atmospheric CO_2 levels are expected to increase through the year 2075. Although the range of current estimates of the projected increase by 2075 varies from less than 100 ppm to over 1000 ppm above the 1984 level (345 ppm), more likely estimates of this increase are on the order of 150 to 300 ppm. Consequently, doubling of the year-1800-atmospheric CO_2 level before 2025 now seems highly unlikely, and doubling may not even occur within the next century.

Projections of future atmospheric CO_2 levels are based on current understanding of the operations of the carbon cycle and associated geochemical

and climatic systems over the past few centuries. It is important to recognize that human activities or even natural phenomena have the potential for altering the dynamics of these systems enough to invalidate *some* current CO_2 projections. Prerequisites for more confidence in the underlying science and assumptions about both balanced and unbalanced conditions of the global carbon system are more widely understood perspectives on the natural biogeochemical cycles of other critical elements (particularly nitrogen and phosphorus, but also oxygen, silicon, and sulfur) and the naturally varying cycles of climate. Because these kinds of cycles interact, studies of them cannot remain as isolated as they typically have been up to now.

Changes in atmospheric CO_2 and climate could be expressed in the form of feedbacks—both positive (unbalancing) and negative (stabilizing)—on the carbon cycle, energy systems, and human ecology. The magnitudes of these feedbacks could be very important. Yet basic understanding is lacking, even in the short term, of several critical processes. These limitations mean that particular outcomes cannot be defined with confidence. Thus, contingency planning for technological and social adjustments will continue to be argued over a range of possibilities that only can be narrowed or elaborated on the basis of new research or insight. Nevertheless, the exploration of a variety of policy options for dealing with the complex issue of global climate change, including the cooperative international development and implementation effort that may be necessary, is still a worthwhile initiative, one that must be undertaken far in advance of demonstrable climatic effects.

TASKS FOR THE FUTURE

No simple guidelines exist for evaluating current uncertainties or for developing the agenda for future inquiries into the workings of the global carbon system. Given the complexity of the CO_2-climate issue in all its varied aspects, uncertainties are to some extent irreducible. Policy decisions, therefore, need to be made intelligently in the face of this uncertainty, both now and in the future. An attempt at a better understanding of the global carbon system seems essential to provide better information for a timely and reasoned response to the effects of CO_2-climate changes. However, projections of CO_2 levels provide only a portion of the atmospheric trace constituent information needed to model future climate changes. Thus, the scope of global biogeochemical research should also be broadened to address the sources and sinks of climatically significant atmospheric constituents other than CO_2.

Significant improvements in the ability to understand and treat (i.e., detect and model) critical processes, including feedbacks, and to attempt a resolution of existing uncertainties are projected to entail a major research effort. Future carbon cycle research should be keyed to removing early, tentative oversimplifications in models as well as to pursuing obvious areas of uncertainty. Without a better perception of the causes for the contemporary imbalance in the global carbon cycle, however, it is not currently feasible to provide a satisfactory ranking of the key sources of uncertainty—important contributions may result from a variety of factors. This ranking is critical to the analysis of policy options, an important precursor to policy formation. In order to attempt to reduce uncertainties over the next decade, the following is required:

1. *Development of a quantitative ranking and evaluation of key determinants of future global energy use and CO_2 emissions using research tools such as historical, engineering, and uncertainty analyses.*

2. *Development of modeling approaches and methods that make more explicit statements of basic biological, chemical, and physical processes responsible for carbon cycling and make better use of available data (particularly for analyzing historical changes in land-use patterns and multiple tracer distributions in the oceans).*
3. *Continuation of detailed reconstructions of land-use changes from primary historical records and from correlates such as population patterns, topography, and so forth, to improve estimates of the terrestrial biospheric CO_2 flux generated during the expansion of farming, forestry, and population over the past 250 years.*
4. *Expansion of process-oriented field research in areas critical to carbon cycle model improvement. Highest priority should be given to the following:*
    - *Completion of planned global synoptic ocean tracer sampling effort (e.g., Transient Tracers in the Ocean Program) in coordination with planned satellite remote-sensing programs (e.g., World Ocean Circulation Experiment) to clarify the relationships among ocean circulation, climate, and geochemical cycles of carbon and critical nutrient elements.*
    - *Continuation of global atmospheric CO_2 monitoring network and its extension to major land biomes in both tropical and high latitudes, coupled with an intensified effort in satellite remote-sensing and ground-support studies to quantify the carbon dynamics of terrestrial ecosystems.*
    - *Intensification of geochemical sampling and modeling (e.g., based on records of elemental and isotopic content in air, ice cores, ocean waters, sediments, and soils) to understand the meaning of atmospheric CO_2 fluctuations, CO_2-climate relationships, and critical links between cycles of carbon and other elements in the recent and distant geologic past. Imbalances over the last 100 to 100,000 years seem in particular need of better quantification as background for modeling and predicting changes in the next century.*

1. THE GLOBAL CYCLE OF CARBON

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1.1 THE CARBON DIOXIDE QUESTION

This chapter outlines the nature of the carbon dioxide (CO_2) question and the general role of the global carbon cycle in regulating the amount of CO_2 in the atmosphere. The scientific and societal issues that result from interactions between changing CO_2 emissions and the global carbon cycle processes are then described as perspective for subsequent chapters on specific and detailed aspects of the global carbon cycle.

Carbon dioxide, the fourth most common gaseous compound in the atmosphere, is transparent to short wavelengths that comprise the sunlight intercepted by clouds or the Earth, but not to the much longer infrared wavelengths subsequently reradiated back toward space. The "greenhouse effect" is produced when some of this infrared radiation is absorbed by CO_2 and reradiated back to Earth, rather than being transmitted into space. CO_2 appears to be the most significant of the atmospheric gases in producing climatic effects. However, other greenhouse or radiatively active gases (e.g., chlorofluorocarbons, methane, nitrous oxide, water vapor) may produce similar effects.

In the 1970s, recognition of a growing world population, the rising per capita use of energy, and the concomitant growth in the rates of fossil fuel CO_2 emissions focused attention on the atmospheric CO_2 increases being documented by Keeling and others (Keeling et al. 1982; Chapters 3 and 4, this volume; henceforth all chapter references apply to this volume only). At the heart of the concern is an anthropogenic climate change that may be too gradual to detect until it is well advanced, too advanced to stop by the time it is detected, and capable of inducing profound changes in the delicate environmental balances that determine whether global breadbaskets will produce adequate food supplies and whether coastal areas will be affected by rising sea levels (National Research Council [NRC] 1977, 1979, 1982, 1983; Seidel and Keyes 1983).

1.1.1 Climatic Effects

Most of the concern over atmospheric CO_2 is focused on its climatic effects. By coincidence, the recent documentation of rising CO_2 concentrations was paralleled by the development of general circulation models (GCMs) of the atmosphere. The original GCMs were designed to use the power of rapidly expanding computer technology to simulate the interactions among climate's most important energy exchanges and processes. The GCM simulations of the CO_2 and climate relationship (e.g., Manabe and Wetherald 1975, 1980) revealed the quantitative climatic implications of our current atmospheric knowledge and documented the importance of these implications to valid prediction of specific gaps in that knowledge. Current GCM simulations suggest that a doubling of atmospheric CO_2 would increase global average temperature from 1.5° to 4.5°C, with high-latitude temperatures rising as much as 9°C and with most of the warming in winter (NRC 1979, 1982). The specific values depend on differing assumptions and data sets used in the GCMs.

Many uncertainties remain. For example, the actual importance of CO_2 in producing climate changes is unclear. The atmospheric warming from 1880 to about 1940, which led Callendar (1938) to hypothesize a relationship between temperature and fossil fuel use, reversed thereafter until the early 1960s, despite continued and greatly enhanced fossil fuel use. Although Hansen et al. (1981) document the likely reasons for the reversal and for the continued warming since the 1960s, the relationship between CO_2 and climate is obviously very complex and nonlinear.

General circulation models of the atmosphere used to quantify the relationship between radiatively active gases and climate generate large uncertainties themselves. For example, GCMs are unable to reproduce features of today's climate when today's boundary conditions are used to constrain the models. Most of the divergence between reality and model output is caused by important atmospheric processes (cloud generation, weather disturbances and fronts, water vapor dynamics, atmosphere-ocean interactions) that are not yet treated adequately because of insufficiencies in basic knowledge, in computing power, or in other resources. Instead, empirical relationships, assumptions, or voids substitute in the models for the missing processes and knowledge (e.g., Manabe 1983; Mitchell 1983). In fact, the model simulations are remarkably realistic when one considers that GCMs were developed as a theoretical aid to understanding fundamental interrelationships between atmospheric processes, rather than as a means for solving applied climatological problems. The research on climate response to atmospheric CO_2 is treated in detail in another volume in the state-of-the-art series (MacCracken and Luther 1985).

1.1.2 Indirect Effects

Human concerns center on the CO_2-induced changes in climate. For example, the world agricultural harvest is tuned to the local and regional climatic status quo, and particularly to a narrow range of average, rather than extreme, seasonal temperature and rainfall values. However, identifying gradual shifts away from the status quo requires many years of climatic data, simply because of the high inherent variability in climate from year to year. Therefore, changes in the averages and in the

frequency of extremes could significantly reduce global agricultural production, while both those deprived or favored by the changes take no action, awaiting proof that the climate has indeed changed. In fact, climate controls the current geographic distribution of many interrelated biospheric features of direct interest to humans (including regional hydrologic cycles) and abundances of forest resources, wildlife and fish populations, disease and pest populations, endangered species, and so on. These biospheric climate dependencies and their potential responses to climate change are addressed in a companion report (White 1985). The expected CO_2-induced effects of climate change are based on knowledge of ecological relationships or merely on the geographic coincidence of today's climatic, landscape, and biotic variables. These effects translate into changes in carbon storage capacity and in carbon fluxes between the major global reservoirs and must be included in any accounting of future carbon cycle dynamics (Chapter 10).

1.1.3 Vegetation Effects

The potential for direct enhancement of plant growth by increasing levels of atmospheric CO_2 may be significant in carbon cycle research. Agricultural scientists have long known that increased CO_2 concentrations in greenhouse atmospheres increase the amounts and the rates at which CO_2 is fixed (transformed with water into simple sugars through photosynthesis) in plants not otherwise limited by nutrients, water, temperature, light, and so forth. This effect could be further enhanced by parallel climate warming, producing longer growing seasons and more rapid growth.

However, large uncertainties exist in translating knowledge derived from controlled greenhouse environments into the outside world (Chapter 10). For example, temperature increases would increase rates of respiration (breakdown of simple sugars into CO_2 and water, the reverse of photosynthesis) during both light and dark periods, possibly balancing the enhanced photosynthesis during light periods. Enhanced photosynthesis requires increased nutrients and water that may not often be available in natural soils. Many ecologists believe that most unmanaged vegetation is currently limited by factors other than CO_2 and therefore could not respond to enhanced CO_2 levels. Agriculture may be similarly unable to use the increased atmospheric CO_2 because added fixed carbon in crops implies added fertilizer and water, both of which potentially require higher fossil fuel energy inputs.

Although of obvious interest to a hungry world, enhanced CO_2 fixation by global vegetation (increased density of the terrestrial carbon reservoir) could also imply a concomitant decline in atmospheric CO_2 concentrations, thereby reducing climatic effects. To date, such CO_2-mediated interactions between the carbon cycle and climate have not been seriously incorporated into models, because there are no firm data to support the existence of such an effect, much less to quantify it.

Future climatic effects may be quite different than expected if, for example, enhanced photosynthesis from atmospheric CO_2 enrichment becomes commonplace. Biota could respond to climate differently, with plants surviving greater drought and temperature extremes, forest products being generated more rapidly, and grazing animals and insect pests being less limited by the quantities of plant foods (e.g., Shands in press). Hydrologic regimes could also be altered because, for example, less precipitation runoff results because plants take up more moisture. However, at the present time, these are only hypotheses waiting to be tested.

The existence of a CO_2 and vegetation link could merely result in greater annual turnover of organic carbon (increased seasonal atmospheric CO_2 amplitude) because decomposer organisms are provided with more litter and warmer temperatures in which to metabolize the litter. Such increased annual turnover may have been detected, particularly in high latitudes of the Northern Hemisphere where land masses are largest and temperature increases from CO_2-climate change would be maximum (Chapter 3). However, there is no direct evidence for the source of the increased amplitude of the annual CO_2 signal (Chapter 3) cited as the manifestation of increased turnover. These topics are evaluated in detail in another volume in the state-of-the-art report series (Strain and Cure 1985).

Although uncertainties abound, it is clear that the concentration of atmospheric CO_2 is increasing. Thus, the amount in the atmosphere now and in the future is of paramount concern, whether the topic is climate response, plant growth enhancement, indirect climatic effects, or societal responses, or interactions among them all.

1.2 THE GLOBAL CARBON CYCLE

Before the industrial and agricultural revolutions, humans may have had relatively little impact on the global cycling of carbon (Chapters 6 and 8). But with increased industrialization and agricultural development, human activities began to play a more significant role in some of the natural cycles. For example, the current amount of CO_2 injected into the atmosphere annually from fossil fuel combustion is equivalent to approximately 2.5% of the combined amount exchanged between the atmosphere and the terrestrial biosphere (including soils) and between the

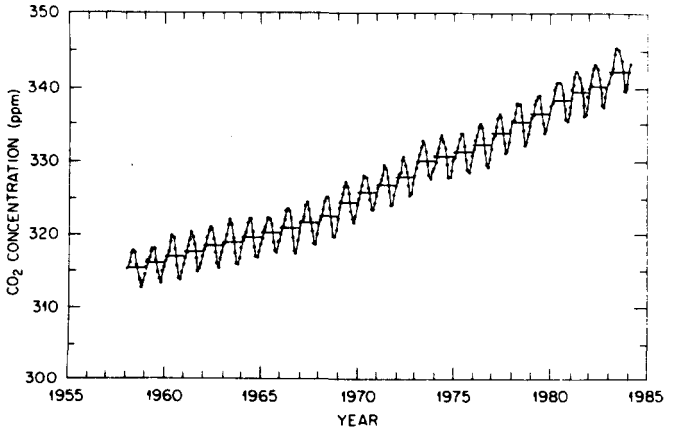


Figure 1.1. Concentration of atmospheric CO_2 at Mauna Loa Observatory, Hawaii, expressed as a mole fraction in parts per million of dry air. The dots depict monthly averages of visually selected data adjusted to the center of each month. The horizontal bars represent annual averages. Data were obtained by C. D. Keeling, Scripps Institution of Oceanography, University of California, La Jolla, California, and are from files in the Carbon Dioxide Information Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee.

atmosphere and the oceans. The cumulative effect of this annual anthropogenic flux superimposed on the natural exchanges, and the role of anthropogenic sources in increasing atmospheric CO_2 concentrations, must be understood before the effect of CO_2 on climate can be evaluated.

Originally, in an atmosphere- and ocean-free Earth, all the terrestrial carbon existed in meteorites and igneous rocks. Thus far only 1 to 5% of this carbon has been released, primarily as volatile CO_2 but also as CO, COS, and so on. The CO_2 has been converted largely to marine carbonates and hence to limestone (82%), with the remainder converted to various forms of reduced carbon. At any given time, the amount of carbon that resides in and circulates between the atmosphere, biosphere, and oceans is a very small portion (less than 0.1%) of the total carbon in the sedimentary rocks of the lithosphere (Chapter 8).

1.2.1 Atmosphere

The present atmospheric total of 720×10^{15} g (342 ppm) of elemental carbon represents a significant increase from about 590×10^{15} g C, largely due to the accelerated burning of fossil fuels for energy (Table 1.1). Sensitive measurements of atmospheric CO_2 levels, begun at Mauna Loa Observatory in 1958, show a regular annual oscillation and a continual annual increase (Figure 1.1 and Table 1.1). Mauna Loa data are not alone in documenting the rise and the seasonal variations of atmospheric CO_2. Measurements made at other locations all over the globe demonstrate similar changes, with seasonal cycles being more apparent in the Northern Hemisphere and most apparent at high latitudes (Chapter 3). Any theories that seek to predict future carbon cycle dynamics must account for the size of the annual CO_2 increment as well as the slowly increasing amplitude of the seasonal signal (see Chapters 3 and 6). Remarkably, the atmospheric measurements clearly reflect effects on CO_2 concentration of El Niño/Southern Oscillation (ENSO) climatic events, but do not yet clearly reflect the shift from 4.4% annual growth in carbon emissions before the 1973 oil embargo to the essentially zero annual growth since 1979 (Figure 1.2; Chapters 3 and 4).

At all sites where measurements have covered multiple years, the upward trend of annual CO_2 concentration values is strongly evident. Yet, when these data are geographically integrated for the globe and compared with atmospheric CO_2 emissions, only about 58% (the observed airborne fraction) of the CO_2 produced can be found in the atmosphere (Table 1.1). The remaining CO_2 cannot be satisfactorily accounted for in calculated uptake by ocean and terrestrial carbon reservoirs. Indeed, a missing fraction has either been taken up by the oceans at a much more rapid rate than geochemists believe can occur (Chen and Millero 1979), or been incorporated into the terrestrial biosphere, in

Table 1.1
Estimated Values and Uncertainties of Key Parameters in the Global Carbon Cycle

Parameter	Value	Uncertainty Range	Section
Total Fossil Fuel Release (1860-1982)	170×10^{15} g C	$150\text{-}190 \times 10^{15}$ g C	4.2
Current Fossil Fuel Release (1982)	5.0×10^{15} g C	$4.5\text{-}5.5 \times 10^{15}$ g C	4.2
Annual Increase in Fossil Fuel CO_2 Emissions	0.00 (1982)	Highly variable in past and future[a]	4.2; 4.4; 9.5
Recoverable Conventional Fossil Fuel Resources	3.93×10^{18} g C	$3.68\text{-}4.25 \times 10^{18}$ g C	4.3
Total Carbon in the Atmosphere (1982)	720×10^{15} g C[b]	$715\text{-}725 \times 10^{15}$ g C	3.4
CO_2 Concentration in the Troposphere (1982)[c]	341 ppm	339-343 ppm	3.4
Airborne Fraction (Fossil Fuels, 1958-1982)	0.58	(0.54-0.63)[d] (0.50-0.80)[e]	10.2.8 10.2.8
Annual Atmospheric CO_2 Cycle[c]	7 ppm	0-15 ppm	3.4
Annual Increase in Atmospheric CO_2	1.5 ppm	1.3-1.7 ppm	3.4
Historic Atmospheric CO_2 Concentration (1800)	280 ppm	260-285 ppm	3.3
Surface Area of Ocean	3.61×10^{14} m^2[f]	- -	- -
Carbon in Ocean Surface Layer (0-75m)	630×10^{15} g C	$590\text{-}720 \times 10^{15}$ g C	8.2
Carbon in Intermediate and Deep Ocean	38×10^{18} g C	$32\text{-}42 \times 10^{18}$ g C	8.2
Carbon in Ocean Sediments	10^{23} g C	?	8.2
Marine Particulate Carbon Flux	4×10^{15} g C	$2\text{-}6 \times 10^{15}$ g C	5.2.2
Gross Annual Ocean CO_2 Uptake	107×10^{15} g C[g]	$106\text{-}108 \times 10^{15}$ g C	
Net Ocean CO_2 Uptake[h]	2.4×10^{15} g C	$1.5\text{-}3.3 \times 10^{15}$ g C	10.2.6
River Export of Carbon to Oceans	0.7×10^{15} g C	$0.3\text{-}1.3 \times 10^{15}$ g C	5.8.3
Historic Terrestrial Biomass	900×10^{15} g C	$700\text{-}1100 \times 10^{15}$ g C	8.2
Contemporary Biomass	560×10^{15} g C[i]	$420\text{-}660 \times 10^{15}$ g C	6.3
Historic Soil Carbon	1.7×10^{18} g C	$1.7\text{-}2.0 \times 10^{18}$ g C	8.2
Contemporary Soil Carbon	1.5×10^{18} g C	$1.2\text{-}1.8 \times 10^{18}$ g C	6.3; 8.2
Net Carbon Flux From Land Biosphere Since 1800	120×10^{15} g C	$90\text{-}180 \times 10^{15}$ g C	6.5 10.2.5
Gross Annual Terrestrial Plant CO_2 Uptake	120×10^{15} g C	$90\text{-}120 \times 10^{15}$ g C	6.4
Net Primary Production	60×10^{15} g C[i]	$45\text{-}62 \times 10^{15}$ g C	6.4
Annual Tropical Forest Area Conversion (1970-1980)	0.004	0.003-0.006	6.5.1
Annual Net Carbon Flux From Land Conversion (1970-1980)	1.3×10^{15} g C	$0.0\text{-}2.6 \times 10^{15}$ g C	6.5.1;10.2.5

[a] A key variable of the CO_2 issue determined by energy demand and associated economic, technological, social, demographic, environmental, and political factors.
[b] Fraser et al. 1983.
[c] Value is for Mauna Loa; latitude dependent (e.g., smaller in Southern Hemisphere and larger in high northern latitudes).
[d] Based only on atmospheric CO_2 and fossil fuel emissions record errors.
[e] Based on combined carbon cycle uncertainties.
[f] Broecker et al. (1979).
[g] Bolin et al. (1979).
[h] Includes enhanced coastal wetland sink ($0.5[\pm 0.3] \times 10^{15}$ g C; Table 5.2).
[i] Olson et al. (1983).

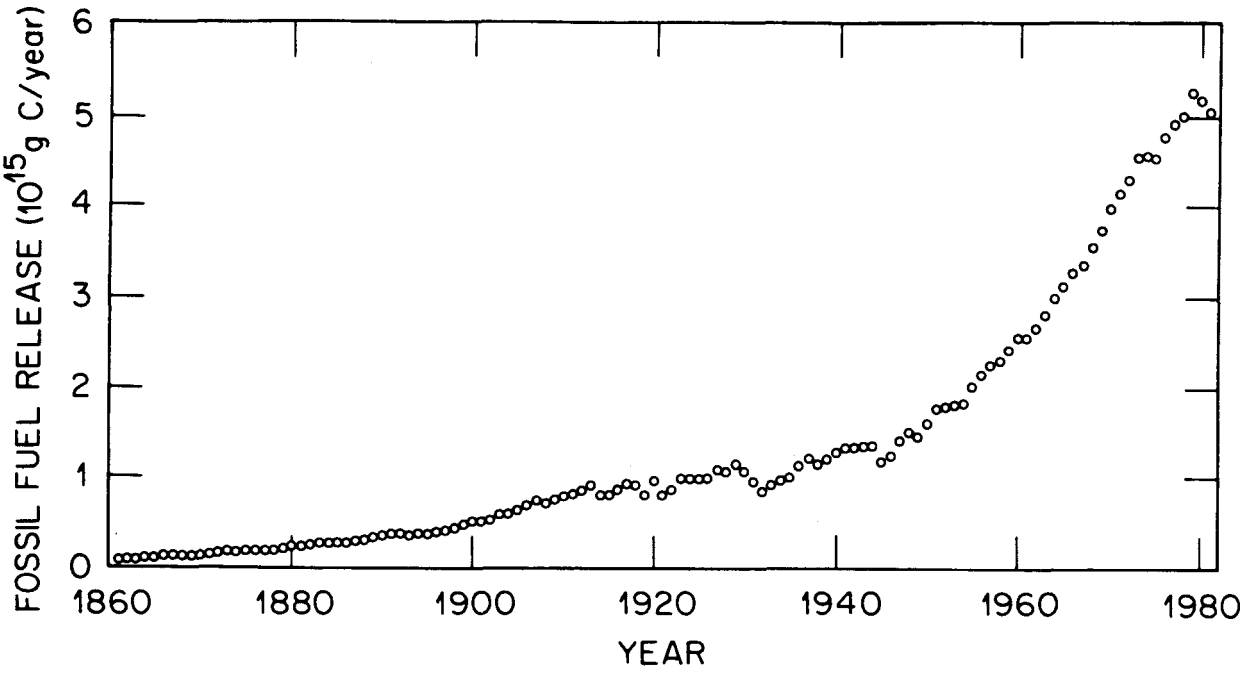


Figure 1.2. Fossil fuel CO_2 emissions: 1860-1982 (Marland and Rotty 1983).

the face of strong evidence to the contrary (Chapter 6). Resolution of this question is high on the carbon cycle research agenda (Chapters 3 to 11).

1.2.2 Oceans

For examining CO_2 exchanges between atmosphere and oceans, the ocean is divided into functional layers. The surface layer extends on average to about 75 m; it is heated by the sun and agitated by the wind (Figure 1.3; Table 1.1). It is relatively well mixed and more or less at equilibrium with atmospheric CO_2. Beneath this layer is a relatively stagnant region (the thermocline) that is stabilized by decreasing temperature and increasing density from the mixed layer to about 1000 m. The deep ocean is isolated from the surface waters by the thermocline. When the surface waters are sufficiently cold and/or saline, however, the thermocline weakens and the surface waters sink (or mix) to various depths and spread horizontally. This produces a worldwide circulation pattern that is quite complex and not yet adequately understood. It involves descending waters mostly in polar regions and upwelling of deep water from various depths elsewhere (Chapter 5).

The most abundant forms of carbon in the ocean are inorganic products of the reaction of CO_2 and water with minerals. The total amount of inorganic carbon in the well-mixed surface waters (about 580 $\times$ 10^{15} g) is close to that of the atmosphere. The total amount in the entire ocean reservoir is 38,000 $\times$ 10^{15} g, or about 50 times the atmospheric content (Chapters 5 and 8).

Organic carbon ($[CH_2O]_n$ in Figure 1.3) is much less abundant (dominated by dead organic matter), existing in a continuum of particle sizes down to small organic molecules. The concentration is variable, with a total organic carbon pool of 50 $\times$ 10^{15} g for the surface region and about 1725 $\times$ 10^{15} g for the entire ocean (Chapter 8). Living carbon, about 0.36 $\times$ 10^{15} g, is mainly represented by the photosynthesizing organisms (the phytoplankton) confined to the surface region; carbon in zooplankton and larger animals is much less (0.02 $\times$ 10^{15} g). When allowances are made for the more productive areas of coastal zones, reefs, and upwelling, there may be 4 $\times$ 10^{15} g C in the total living marine biomass (Chapters 5 and 8).

The ocean reservoir also contains carbonate solids (formed by sea organisms) found both suspended in the water column and incorporated in surface sediments. This material is largely calcium carbonate produced in surface waters and redissolved in deep water. The deposits are generally confined to ocean depths less than 4 km. The top ~10 cm of these deposits is mixed sufficiently by the foraging activities of bottom animals that it may be reasonable to include it as a reactive part of the ocean system. The amount of carbon present as calcium carbonate in these mobile deposits is about 410 $\times$ 10^{15} g (Baes et al. 1976). The deeper sediments contain an amount of carbon (7 $\times$ 10^{22} g) that dwarfs

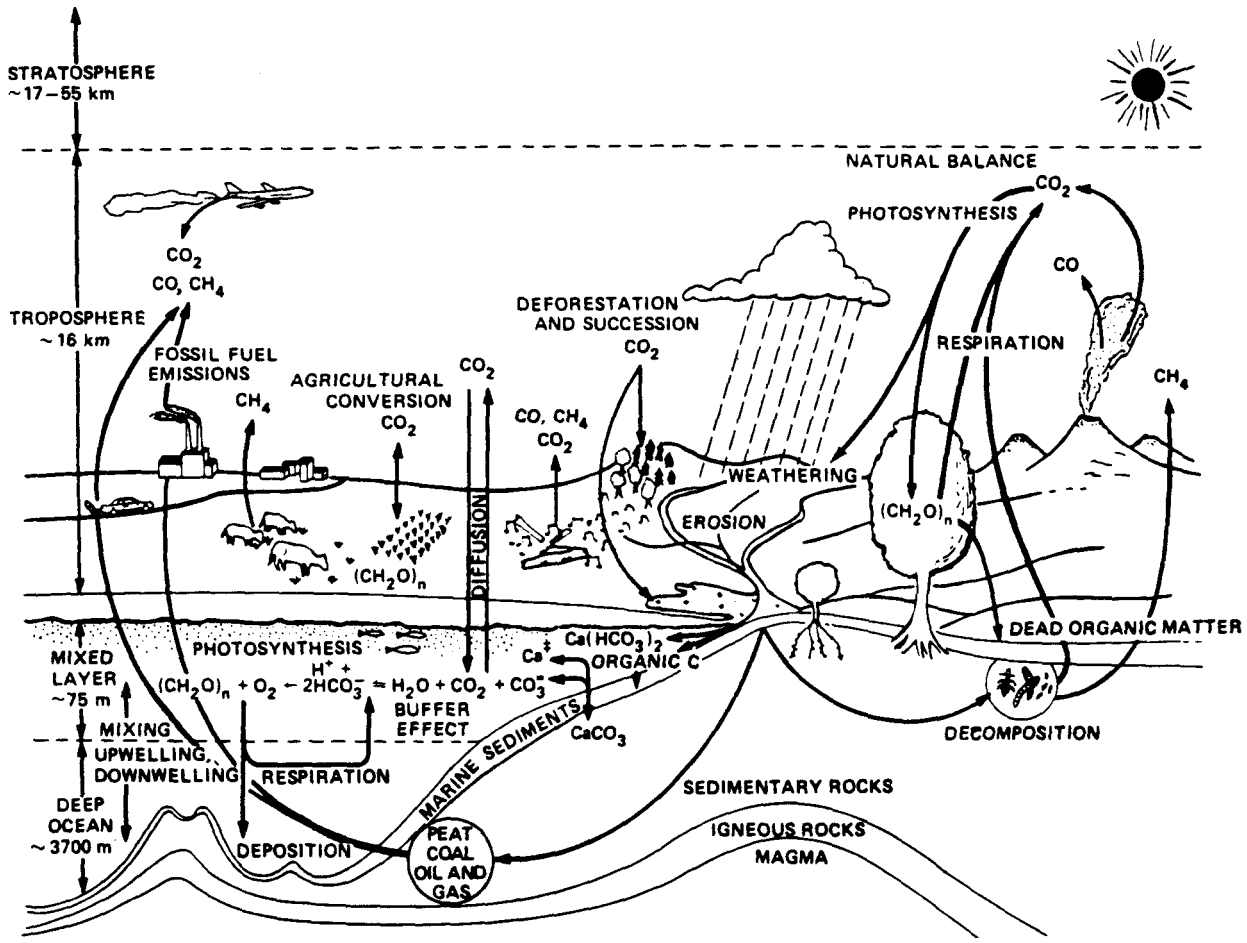


Figure 1.3. Schematic diagram of the global carbon cycle. Current work in carbon cycle research seeks to quantify the CO_2 exchanges depicted; an additional effort is needed to quantify CH_4 and CO.

the carbon content of ocean water, but it cannot react at an appreciable rate with the ocean and atmosphere over time scales shorter than millennia (Table 1.1; Chapter 8).

1.2.3 Terrestrial Biosphere

Part of the excess carbon from fossil fuels, unaccounted for in the atmosphere, may be accumulating in terrestrial organic matter (Figure 1.3). Nonfossil organic carbon pools are dominated by materials with slowly exchanging carbon. Approximately 1500×10^{15} g C are in dead organic matter (i.e., humus and recent peat) (Table 1.1) that may require 100 to 1000 years to turn over. The more rapidly exchanging carbon in live organic material (turnover time 10 to 100 years) may total 560×10^{15} g (Table 1.1). The smaller pools of rapidly exchanging organic carbon (turnover time 1 to 2 years; plant leaves, small branches, and roots; litter; invertebrates; decomposers), when combined, are equal to about 20% of the total living pool (Chapter 8). Only very small fractions of total carbon in terrestrial carbon pools ($\sim 4 \times 10^{15}$ g) are represented by animals, short-lived and long-lived, including humans (Chapter 8).

The prehistoric terrestrial biosphere was often a carbon sink, otherwise there would be no fossil fuel carbon problem (e.g., no vast high-latitude peatlands where only continental ice sheets occurred 15,000 years ago). Yet, clearly, today's terrestrial carbon reservoir, particularly the high-carbon-density forests, occupies significantly less land each year (Chapter 6). Although changes in the rates of forest clearing over time are still a major research issue, land clearance today is concentrated in the tropical forests of highest carbon density (Table 1.1; Chapter 6). About one-half of

the area and carbon remains in forests today compared with that existing in the middle Holocene (Chapter 8). Furthermore, geochemical models constrained by measured $^{13}C/^{12}C$ isotopic ratios in tree rings suggest that biota has been a carbon source throughout most of the industrial era (Chapters 3, 6, 7, and 10).

1.2.4 Fluxes Between the Global Reservoirs

The observed annual increase in carbon content of the atmosphere is less than the cumulative input from fossil fuel combustion and other sources because of net CO_2 uptake by other reservoirs. The dynamics of these exchanges will determine the future CO_2 concentrations that result from additional fossil fuel combustion. The major components and exchanges of the global carbon cycle that are significant in the response of the cycle to fossil fuel CO_2 are illustrated in Figure 1.3. On a scale of 1 or 2 centuries, dissolved inorganic carbon in the oceans and terrestrial organic carbon appear to be the major reservoirs for CO_2 exchange with the atmosphere.

Net carbon exchange between the atmosphere and oceans is controlled by the temperature-dependent equilibrium between carbonates and bicarbonates, by biological processes, by turbulent mixing, and by circulation of water masses that transport carbon from the surface into deeper waters (Figure 1.2; Chapter 5). Some vertical redistribution of carbon in the oceans results from settling of organic and carbonate particles produced by organisms in the surface waters. Many uncertainties in the atmosphere-ocean flux remain, despite the great strides in data collection and modeling of the past decade (Table 1.1; Chapter 5). Yet, even the most sophisticated analyses indicate that only 30 to 40% of the carbon emitted to the atmosphere has been taken up by the oceans (Chapter 5). However, potentially as much as 40% of the carbon emitted to the atmosphere is unaccounted for (the "missing carbon" problem; Chapter 10).

The net exchange of carbon with terrestrial ecosystems depends on the balance between carbon assimilation by photosynthesis and carbon release from both living and dead material (Figure 1.2). However, the net balance is quite uncertain at present (Table 1.1; Chapters 6 and 10). It should be recognized that biotic sources may occur as a result of either land-use practices, such as forest clearing and soil cultivation, or natural processes.

Forest clearing and other disturbances to the terrestrial carbon system (Figure 1.2) lead to net releases of CO_2 to the atmosphere, which could have been more significant in the past 200 years than the fossil fuel contribution to the carbon cycle (Table 1.1; Chapter 6). However, these terrestrial fluxes could soon diminish in relative importance, and, over longer time periods, fossil fuel release and rate of exchange with the oceans may become the major factors influencing atmospheric CO_2 concentrations. How much of the observed past increases in atmospheric CO_2 have been significantly affected by biotic sources has yet to be determined (Chapters 7 and 10). Feedbacks in the carbon cycle from future climate change and from elevated CO_2 concentrations in the atmosphere could shift the values of both reservoirs and fluxes between the reservoirs.

Besides these major elements of the carbon cycle, several other reservoirs and exchanges may be important in determining the response of the cycle to the CO_2 released from fossil fuel combustion. Examples are the transfer of terrestrial carbon to coastal wetlands and continental shelves by rivers, formation of subfossil carbon (e.g., peats) in wetlands, enhancement of marine productivity by increased nutrient availability (from coastal eutrophication), and formation and dissolution of mineral carbonates in the sea and in arid lands. Many of these fluxes are partially or poorly understood, and, although quantitatively each may not be large, collectively they may be significant in balancing the carbon budget on a global scale (Chapters 5, 7, and 8).

Finally, the dynamics of the carbon cycle are linked to global cycles of certain other elements (particularly nitrogen and phosphorus); that is, fluxes or pool sizes in the carbon cycle depend to some degree on fluxes or pool sizes of the other elements. Carbon uptake by terrestrial biota depends on the availability of fixed nitrogen and phosphorus. Phosphorus and nitrogen as fertilizers washed from the land into rivers and then to the sea should increase riverine and ocean surface productivity (Garrels et al. 1975), reducing CO_2 content of surface waters and CO_2 concentration in the atmosphere. Indeed, Walsh (Walsh et al. 1981; Walsh 1984) believes that anthropogenic nitrogen in oceans could have produced enough organic matter to explain the current imbalance in the carbon cycle, although others argue logically to the contrary (Chapter 5; see also Peng and Broecker 1984). All agree that changing climate could change not only the individual element cycles but their linkages as well.

1.3 SCIENTIFIC ISSUES

The central scientific question that ultimately must be addressed concerns the environmental consequences of the increasing concentrations of CO_2 in the atmosphere. The related aim of carbon cycle research is to provide information for estimating the climatic, indirect, and direct vegetative effects. For example, we cannot yet provide an accurate answer to the fundamental question of how the concentration of CO_2 in the

atmosphere will change as more and more enters the atmosphere. Before consequences can be estimated, the potential future rates and amounts of CO_2 fluxes must be documented. Yet, our scientific understanding of global carbon sources and sinks, such as the uptake of CO_2 by the ocean and the fundamental exchanges with the biosphere, is inadequate for documenting today's fluxes, much less those in the future (Chapters 10 and 11).

Without better knowledge of the carbon cycling behavior of the terrestrial biosphere, its unequivocal identification as a current or future source or sink of CO_2 is impossible. If the biosphere releases more CO_2 than it absorbs, then behavior of the oceans must be different from that currently represented in models (see Chapters 5 and 7). The ability of the ocean to act as a CO_2 sink is great, but the rate of carbon uptake is important because it determines how rapidly the fossil fuel carbon can be permanently removed from the atmosphere. Before confidently estimating what fraction of the CO_2 from fossil fuels will remain in the atmosphere, we must better understand the roles of both the biosphere and the oceans in the carbon cycle.

A primary objective of carbon cycle research is to estimate the change in atmospheric CO_2 concentration expected to accompany additional combustion of fossil fuels. Related objectives are to understand how human intervention in biogeochemical processes (such as clearing forests and forcing climate change) may alter the global carbon cycle, as well as to estimate mitigation capacities of natural components of the global carbon system. Answers to the following sets of questions are needed to achieve these objectives:

1. What future releases of CO_2 to the atmosphere can be expected from fossil fuel combustion? What socioeconomic and technological factors must be considered to project future releases? What is the history of such releases and of atmospheric responses in the past?
2. What is the temporal trend, annual cycle, and geographic variability of atmospheric CO_2? How has atmospheric CO_2 content changed over the past centuries and millennia? Did steady-state conditions persist in the carbon cycle before significant human intervention?
3. Is the terrestrial biosphere in equilibrium with atmospheric CO_2, or is it a carbon source or sink? How variable is the carbon exchange between the atmosphere and different terrestrial ecosystems, and what has been the influence of the disturbance of forests and soils? Has the globally averaged flux between the atmosphere and terrestrial ecosystems changed substantially in the past 100 years?
4. What is the temporal and geographic variability of atmosphere-ocean CO_2 exchange? What is the rate of ocean uptake of atmospheric CO_2? How rapidly is carbon removed from surface water to the deep ocean?
5. Are there significant sources or sinks of atmospheric CO_2 not now considered in carbon cycle models? How likely are significant future changes in CO_2 fluxes?

The usual scientific approach to such questions is to divide the world's oceans and air and land masses into separate boxes, with exchanges between the boxes representing the separate parts of the carbon cycle. Early models consisted of three boxes: the atmosphere, the surface ocean, and the deep sea sediments. Atmospheric CO_2 exchanges occurred at the ocean surface, whereas the sediments served as sinks for long-term deposition of carbon. To incorporate current knowledge of the carbon cycle into the models, modelers have increased the number of boxes, first from 3 to 12 and now to many score, while including more complex mathematical functions to describe the many separate biogeochemical processes. Without qualifications, none of these models duplicates the present well enough to instill confidence that anything more than generalities can be reliably predicted about the future.

Future carbon cycling dynamics may be influenced by processes that have not yet been included in even the most sophisticated of carbon cycle models. Certainly no models include adequate process descriptions to explain the recently observed depression of atmospheric CO_2 values during ENSO events (Chapter 3) or the significant changes in the temperature and salinity of the deep North Atlantic occurring over less than a decade (Brewer et al. 1983). If the atmospheric CO_2 concentrations are so tightly linked to surface-ocean conditions, and the deep sea can change with such surprising swiftness, presumably in response to atmospheric changes, how will the ocean's capacity to remove CO_2 from the atmosphere respond to climatic warming?

Glacier ice, containing evidence of rapid changes in atmospheric CO_2 in tandem with climate changes, is another example of atmosphere-ocean linkages not yet modeled. Recent analyses of ice by several laboratories have shown that atmospheric CO_2 rose from about 190 to almost 300 ppm, closely paralleling warming at the end of the last glacial period 10,000 B.P. (Oeschger and Stauffer 1986). Detailed analysis of ice formed in the middle of the last glacial period (40,000 B.P.) has now revealed comparable changes requiring little more than 100 years (Oeschger and Stauffer 1986).

Another process hypothesized to be important in

future carbon cycling involves potential saturation of ocean water by carbonates, blocking further uptake of atmospheric CO_2 by the oceans (Chapter 5). Today, anthropogenic CO_2 that reaches deep sea sediments dissolves the carbonate in them and is removed from the cycle. Thus, sediments are currently a sink for CO_2. But if the carbon in available fossil fuels is injected into the atmosphere within the next 100 years, the neutralization of CO_2 by ocean sediments could cease. It is possible that carbonate at the sediment surface would dissolve faster than bottom-dwelling animals could churn it up; a protective layer of noncarbonate sediment would then form; and neutralization on the ocean bottom (and hence oceanic CO_2 uptake) would be limited by the slow fallout of carbonate from the ocean surface (e.g., see Sundquist 1986). Many other hypotheses that may involve significant perturbations to the carbon cycle remain to be evaluated by scientific research.

These scientific concerns are currently overshadowed by the magnitude of the potential uncertainty in future fossil fuel emissions (Chapters 9 and 10). Projections of global fossil fuel emissions appear to be the greatest source of error in predicting future atmospheric CO_2, but it may not be reasonable to assume that a significant reduction in this uncertainty is possible if projections on century time scales are needed. There exist obvious limits to our ability to predict economic/energy-use futures, as very recent history, including the aftermath of the 1973 oil embargo, has taught us (see Chapters 4 and 9). Uncertainties in emissions sources may always remain dominant in CO_2 projections. The real challenge will be to make policy decisions intelligently in the face of continuing uncertainty.

1.4 SOCIAL ISSUES

The fundamental issue of concern is the societal consequences of the environmental changes brought about by increasing CO_2 concentrations. Most of the answer lies with the resolution of the scientific issues, but their resolution in turn depends on accurate projections of rates of fossil fuel use by society. If all of the fossil fuel resources were to be burned at a steady rate over a 10,000-year period, the atmospheric CO_2 concentrations would rise too slowly to cause immediate concern. On the other hand, if large increases in the combustion of fossil fuels occur (e.g., up to 3.5% per year over the next century; Chapter 9), much of this resource could be consumed in the next 100 years and the resultant atmospheric concentration of CO_2 could reach several times its historic levels (Chapter 10). Estimates of the consequences from just a CO_2 doubling indicate increases in the average surface temperature of the Earth that range from possibly acceptable (NRC 1983) to potentially catastrophic (Seidel and Keyes 1983). A better understanding of the future energy requirements of various regions (e.g., Rotty et al. 1984) and the alternatives for meeting these requirements are essential in determining and possibly controlling rates of warming.

Clearly it is also necessary to understand more accurately the consequences of the continued use of fossil fuels by learning (1) more about the carbon cycle (particularly ocean circulation, the effect of CO_2 on photosynthesis, and the effect of changing patterns of land use on biosphere exchanges of CO_2) to judge how much of the fossil carbon CO_2 will remain in the atmosphere in the future (Chapter 2); (2) more about the climatic (particularly regional) effects of increased atmospheric CO_2 through a greatly increased and integrated climate research effort; and (3) how to predict the full impact of a given change in regional climate on humans, their environment, and their society. This knowledge would allow a judgment of how rapidly global fossil fuel reserves could be safely used.

Future energy demands of the world will probably require extensive further exploitation of fossil fuels. Without major development of nonfossil fuel alternatives, the atmospheric concentrations of CO_2 could double during the 21st century (Chapter 10). A resultant general global warming would force society to face a historic dilemma. Survival of the growing global human population may require burning fossil fuels and clearing forests, thereby warming the globe during the next century. Such changes could have significant impacts on global ecological systems, agricultural production, human health and social patterns, and the economics associated with these factors (NRC 1983; Seidel and Keyes 1983). Increasing temperatures could also affect vegetation distribution and growth and might alter geographical patterns by enlarging the world's arid zones or melting the polar ice (NRC 1983; Clark 1982).

The dilemma is that action to mitigate the climate changes could also produce destabilizing effects. Obvious corrective actions would be major reductions in the consumption of fossil fuels and lowering the rate of forest clearance (i.e., reducing land expansion for agriculture and grazing, land toxification, and other consequences of industrial development). Efforts to change the patterns and growth of industrial, mining, agricultural, and other land uses, coupled with restrictions on the burning of fossil fuels, could disrupt the social and economic order as much as changes caused by a significant warming of the globe (Clark 1982).

This debate highlights the human questions of the CO_2 issue. We must determine what levels of atmospheric CO_2 increase can occur without significant

negative or irreversible impacts on environmental, agricultural, economic, and societal systems. Further, we must determine what mitigative actions can be implemented to control or ameliorate the effects of anthropogenic CO_2 emissions. Although these fundamentally complex questions are far easier to pose than to answer, their solutions lie in interpreting the technical knowledge forthcoming from a broad array of scientific disciplines.

This volume does not (and cannot) address all of these questions, but rather is focused on the global cycle of carbon, the dynamic balance between atmospheric CO_2 sources and sinks that determine the rate of increase in the atmospheric CO_2 concentration. The emphasis is on the global biogeochemical cycle, the focus of most past research. Much less is known (*or can be known*) about projected energy usage and its determinants–the palette of available technologies and fuels as well as demographic, economic, social, and geopolitical factors–which represent the largest potential uncertainty in anthropogenic CO_2 emissions and atmospheric CO_2 levels over the next century. Human ecology and future CO_2 emissions are also linked through the effects of CO_2-climate change. Because all aspects of this complex "human equation" cannot be foreseen, including the global response function for a climate change only dimly perceived at present, much of this uncertainty appears irreducible. This implies that *international* emissions control and CO_2 mitigation strategies be developed along with better models of global carbon biogeochemistry (linked to other critical element cycles and to climate) in order to respond in a timely and reasoned manner to avoid or lessen the deleterious effects of CO_2-climate change.

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2. APPROACHES TO STUDYING THE GLOBAL CARBON CYCLE

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2.1 INTRODUCTION

This chapter describes the approaches taken in formulating research on the global carbon cycle, and is designed principally to orient the nonspecialist. Understanding the global carbon cycle requires very different approaches than does the solution of more traditional and experimental scientific problems. Development of a knowledge base for the carbon cycle requires that we understand all aspects of measured changes in atmospheric carbon dioxide (CO_2) concentration, that we quantify fossil fuel, biospheric, and oceanic fluxes in the global carbon cycle, and that we develop accurate models for interpreting the past and predicting the future. For discussion, these approaches can be categorized as follows.

1. Information exchange: To promote the optimum use of existing information, data and models must be readily available to all interested researchers.
2. Extant data analysis: Much potentially useful information exists in historical records and in scientific publications that must be located, deciphered, evaluated, and interpreted.
3. Monitoring: Repeated long-term measurements will be needed to determine the status of several reservoirs and sources in the global cycle and to validate cause-effect relationships for observed changes.
4. Experimentation: To study the dynamics of the global carbon cycle, experimentation will be necessary to reveal the scientific basis for cause-effect relationships so fluxes can be quantified.
5. Modeling: Mathematical representations of the biogeochemical processes are necessary to integrate a diversity of data and to permit projections of future conditions. The following sections present more detailed discussions of these categories.

2.2 INFORMATION EXCHANGE

The rate of progress regarding the CO_2 question depends in large part on the speed and accuracy with which research results are exchanged among workers. The Carbon Dioxide Information Center (CDIC), supported by the Department of Energy (DOE) and located at Oak Ridge National Laboratory (ORNL), Oak Ridge, Tennessee, helps coordinate and promote the international exchange of information on CO_2 research (Watts 1984). The CDIC serves as a repository for data from monitoring, as well as for storing and distributing research reports.

Many other organizations play a role in coordinating CO_2 information exchange. They include (but are not limited to) the World Meteorological Organization (WMO), the Commission of the European Communities (CEC), the International Energy Agency (IEA), the International Institute for Applied Systems Analysis (IIASA), the International Atomic Energy Agency (IAEA), the United Nations Environment Programme (UNEP), the International Council of Scientific Unions (ICSU), and the Scientific Committee on Problems of the Environment (SCOPE), a member committee of ISCU (Dougher 1983). There are other international organizations that participate in CO_2-related activities, such as some of the other specialized agencies of the United Nations and committees of ICSU, but their efforts usually only involve support or funding for specific conferences or meetings sponsored by one of the eight major organizations.

One important example of institutional cooperation is the development of CO_2 standard reference materials essential for long-term measurement of atmospheric CO_2 and for interstation calibration. Responsibility for standards originated at the Scripps Institution of Oceanography. To ensure long-term continuity and availability of standard reference materials, the National Bureau of Standards (NBS) is developing and will distribute CO_2-in-air standard gases. The NBS will develop protocols for distributing and maintaining continuous calibration of the CO_2 reference materials. Systematic convergence of Scripps and NBS standards will ensure that the CO_2 researchers will lose no reference materials (U.S. Department of Energy [U.S.DOE] 1984). These plans are being coordinated with recommendations for standardization proposed by the WMO in September 1981 at the meeting on standards in Geneva, Switzerland. The interlaboratory comparisons of standards currently supervised by the WMO Central CO_2 Laboratory at the Scripps Institution of Oceanography will also continue.

In addition to the above-mentioned international organizations, many national academies, governmental ministries, universities, and agencies throughout the world are involved in CO_2 research and the exchange of information. International conferences, projects, and programs on CO_2-related research have been arranged and carried out by the individual countries, with the major international organizations having broad promotional and coordinating functions. There is also close collaboration and cooperation between the international organizations, particularly the UNEP, WMO, and ICSU.

2.3 EXTANT DATA ANALYSIS

Understanding the response of the global carbon cycle to fossil fuel CO_2 requires reasonable estimates of the

natural fluctuations in major biological and geological processes affecting carbon, that is, those that occurred before the onset of substantial fossil fuel use (Chapter 8, this volume; henceforth all chapter callouts are to this volume). One critical benchmark for estimating the effects of natural phenomena on the global carbon cycle would be records of past atmospheric CO_2 levels–even to within 5 parts per million by volume (ppm)–at key dates in the 18th, 19th, and early 20th centuries (Chapter 3). However, reliable direct observations of atmospheric CO_2 concentration exist only since 1958 (Chapter 3). Older records based on chemical measurements (e.g., World Meteorological Organization [WMO] 1983; Elliott 1984; Fraser et al. 1986), while suggesting values lower than those of 1958, may not be precise enough to establish benchmark values (Chapter 3).

We must understand the historical impact of anthropogenic carbon emissions on the atmospheric concentration of CO_2 before we can evaluate future impacts. The past production of fossil fuels has been estimated from data of the United Nations Statistical Office and primary historical reference sources (Chapter 4). The amount of CO_2 released also has been estimated from historical and current production data on fossil fuels, the carbon content of each fuel type, and the patterns of fuel use. These estimates provide a necessary chronological record of atmospheric CO_2 emissions which is now revised and updated annually (Chapter 4).

The maximum recoverable fossil fuel resource, which will determine future CO_2 releases, has been estimated using data from the U.S. Geological Survey, the Bureau of Mines, and the United Nations (see Chapter 4). These estimates, with data on the carbon contents and emissions characteristics of fossil fuels, can then be used in energy-economic models to predict future CO_2 emissions (Chapter 9). Information from the models can be used to place bounds on future CO_2 emissions to the atmosphere and to estimate the potential for the success of mitigation strategies.

Information on changes in the biosphere, such as forest clearing, together with knowledge of historical fossil emissions and the ocean's past CO_2 uptake, are also important for understanding both the natural carbon cycle and human-induced changes. Considerable data bases already exist on the nature of terrestrial ecosystems and soils prior to major human disturbance. Global maps of world vegetation and soil carbon storage (e.g., contemporary, Zinke et al. 1984; Olson et al. 1983; and for natural conditions, Post et al. 1982; Emanuel et al. 1985), with spatial resolution of one-half degree latitude by one-half degree longitude, have been prepared from available data. Estimates of natural carbon storage, together with information on regional changes in land-use and rates of population growth, have been combined with carbon storage response functions

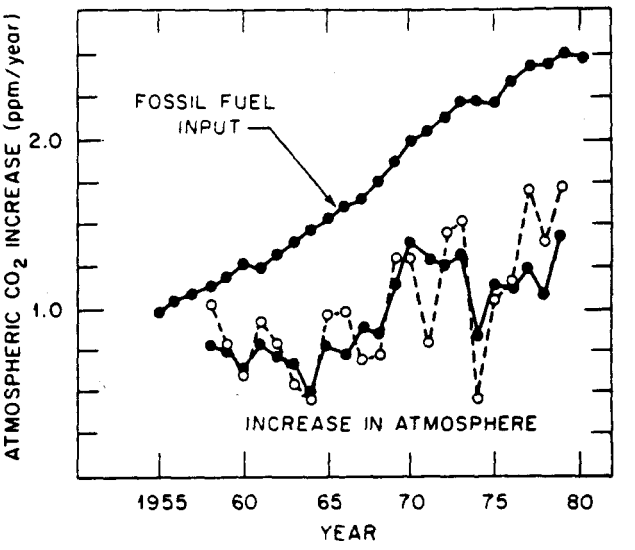


Figure 2.1. Upper plot: Theoretical annual increase in atmospheric CO_2 if CO_2 source was only fossil fuels, neglecting any possible removal. Lower plot: Closed circles and solid lines represent measured annual increase of atmospheric CO_2, adjusted to reduce the influence of ENSO events. Open circles and dashed lines represent the same data, unadjusted for ENSO events. Atmospheric CO_2 data are based on averages from Mauna Loa Observatory, Hawaii, and the South Pole. (Source: Keeling 1983).

(Houghton et al. 1983; Chapter 7) to estimate the history of terrestrial carbon fluxes to the atmosphere (Chapter 6). Yet, this research is very incomplete and in great need of upgrading, as discussed in the conclusions and recommendations of Chapters 6, 7, and 8.

2.4 MONITORING

The basis for the energy-CO_2 link with the CO_2-climate issue is that atmospheric CO_2 concentrations have increased over the past two decades concomitant with increased fossil fuel combustion (Figure 2.1; see Chapters 3 and 10). The "link" is obviously far from perfect even when some suspected sources of variance in nonfossil fuel carbon are removed (Figure 2.1). Continued tracking of atmospheric CO_2 concentrations will be required to quantify this relationship and to predict the responses of the atmosphere and oceans to fossil fuel CO_2, as well as to predict the effects of increased CO_2 on climate and on changes in biospheric productivity.

Some carbon cycle responses to fossil fuel CO_2 can be deduced from the atmospheric records by careful analysis of temporal, geographic, and seasonal variations of CO_2 concentrations. Research thus far has identified many causes of the observed

atmospheric variations (e.g., trends and patterns in fossil fuel combustion, equatorial upwelling in the oceans, and seasonal carbon exchanges in terrestrial ecosystems [Chapter 3]). However, existing analytical approaches do not allow unequivocal resolution, from atmospheric CO_2 records, of the magnitudes of the fluxes producing these changes (Chapter 7). Seasonal variability exhibited by the Mauna Loa record, for example, is attributed to the annual cycle of biospheric photosynthesis and respiration on land, but other processes also contribute to the annual and seasonal variations. Annually varying air-sea exchange of CO_2 is of unknown importance during El Niño Southern Oscillation (ENSO) events. Reductions in fossil fuel CO_2 inputs appear to have had measurable effects since the 1973 oil embargo (Chapter 3), but minor climate change could be as important to the global carbon cycle as any of the other factors. As examples, CO_2 released from the biosphere may increase if warmer temperature accelerates oxidation of organic materials at higher latitudes, and ocean currents, mixing processes, and CO_2 uptake could be affected if the temperature changes (Chapters 5 to 9).

Data from the National Oceanic and Atmospheric Administration's Geophysical Monitoring for Climate Change Program and the Scripps Institution of Oceanography have provided the most comprehensive evidence of atmospheric CO_2 changes. The data also clearly show the existence of spatial and seasonal gradients (e.g., Chapter 3). Temporal trends in atmospheric CO_2 are used to constrain global carbon cycle models used to estimate future atmospheric CO_2 concentrations (Chapters 7 and 10). Precise seasonal records may also reflect regional carbon exchanges (e.g., of respiration and photosynthesis), and annual data may indicate variations in emissions sources and in global CO_2 circulation (Chapters 3 and 7). Strategically located atmospheric measurements, including individual carbon isotopes ($^{13}CO_2$ in addition to $^{12}CO_2$), may thus help to identify and to quantify sources or sinks of CO_2 (Chapters 3, 6, and 7).

The net releases of CO_2 to the atmosphere from land-use change at some points in the past may have been as large as, or larger than, those from fossil fuel emissions (e.g., Chapters 3, 6, 8, and 10). Therefore, rates of some anthropogenic disturbances other than fossil fuel combustion, particularly forest clearing, need to be monitored to define the global carbon balance. Annually updated records from the Food and Agricultural Organization (FAO) of the United Nations, which has monitored land-use change since 1947, have been analyzed to estimate contemporary changes in terrestrial carbon storage (Houghton et al. 1983; Brown and Lugo 1984). Satellite photographic records obtained annually since 1973 provide another potentially powerful tool for estimating the areal extent and characteristics of terrestrial ecosystems and rates of recent disturbance from land-use change (Chapters 3 and 6).

Research programs, such as the Geochemical Ocean Sections (GEOSECS) and Transient Tracers in the Ocean (TTO) Programs, were designed to provide synoptic information on ocean circulation and carbon chemistry from a series of measurements repeated over time. The programs have generated a sequential series of data sets from several ocean regions which are critical to carbon cycle research (Chapters 5 and 7). Repeated, sensitive chemical measurements on oceanic waters may provide direct evidence of carbon buildup in the oceans from increased atmospheric CO_2 (e.g., either from dissolved CO_2 or inorganic carbon). These data are potentially very useful for calibrating carbon cycle models (Figure 2.2; Chapter 5).

2.5 EXPERIMENTATION

The ocean is the ultimate sink for fossil fuel CO_2. Two main fluxes control this uptake. One is the relatively rapid exchange between the atmosphere and surface water, creating near equilibrium conditions for CO_2, HCO_3^-, and CO_3^{2-}. This exchange is influenced by inorganic carbon chemistry, biological productivity,

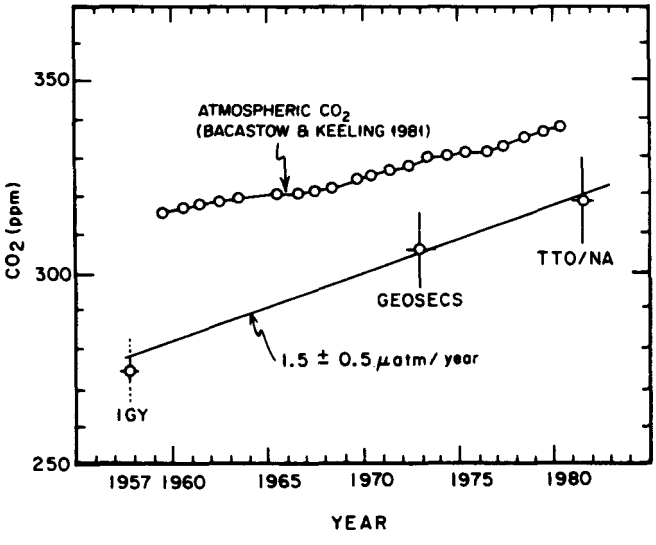


Figure 2.2. Mean Sargasso Sea gyre surface water pCO_2 versus time. The atmospheric CO_2 concentration is expressed in mole fraction CO_2 (fCO_2) in dry air, and was obtained at Mauna Loa, Hawaii. IGY = International Geophysical Year. (Source: Takahashi et al. 1983).

physical resistances of thin-film diffusion of CO_2, and wind stress. Reaction rates affecting air-sea exchanges for quiescent surface water in warm seasons are reasonably well understood, but additional information is critically needed for regions of high CO_2 invasion and for other seasons (Chapter 5). The second and the most critical flux is the transfer of carbon from surface waters into the deep ocean by vertical mixing and deep water injection. This flux limits long-term ocean removal of atmospheric CO_2 from the atmosphere, and our uncertainties about this transport are linked to our limited understanding of ocean circulation (see Chapters 5 and 7).

Particular regions may play a dominant role in the carbon dynamics of the oceans. For example, the formation of North Atlantic deep water is believed to be a critical process for transport of carbon from surface waters. Measurements of partial pressure of CO_2 (pCO_2) in arctic waters, such as the Norwegian and Greenland seas, imply a large net influx of CO_2 from the atmosphere. The reasons for this influx, as well as the patterns of carbon circulation in the oceans, have been inferred from tracer data. Chemical (salinity, temperature, alkalinity, density, oxygen, total carbon, major nutrients, and chlorofluorocarbons) and isotopic (^{14}C, ^{3}H, ^{3}He, ^{85}Kr, ^{39}Ar, and radon) tracers have been used in a variety of ways to describe air-sea exchange, ocean mixing, and vertical transport of carbon to the deep ocean (Chapters 5 and 7).

The increase in atmospheric CO_2 during the past 25 years is often attributed to combustion of fossil fuel alone. But some of it may also be caused by human-induced changes in the biosphere (e.g., oxidation of organic carbon from forest conversion and soil cultivation and/or climatic effects on ocean circulation). The uncertainty in biospheric contributions must be resolved so we can accurately determine the relative importance of other carbon cycle subsystems and the exact contribution of fossil emissions to changes in atmospheric CO_2 concentrations (Chapters 6, 7, and 10).

Indirect experimental techniques have been used as one means to estimate the contribution of CO_2 from terrestrial ecosystems. For example, organic matter in vegetation and humus is depleted in the heavy carbon isotopes (^{13}C and ^{14}C) because photosynthesis discriminates against the heavier nuclides. When a large amount of ^{13}C-depleted biospheric carbon residue is oxidized, the ^{13}C-depleted CO_2 can cause the $^{13}C/^{12}C$ ratio of atmospheric CO_2 to become more negative. This isotopic change of atmospheric CO_2 could then be registered in plants (e.g., tree-ring cellulose), thereby providing a proxy record of historical biospheric effects on atmospheric CO_2.

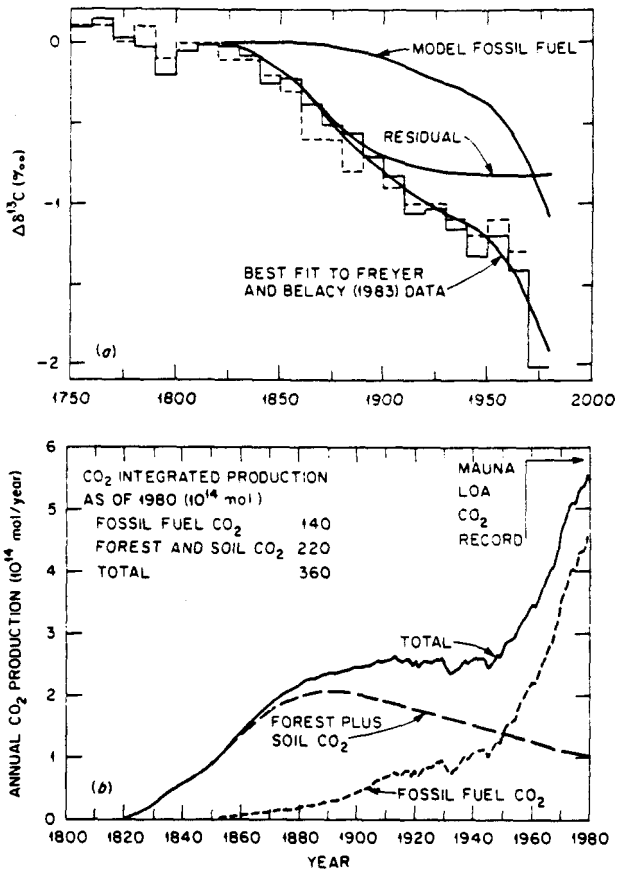


Figure 2.3. (A) Forest and soil CO_2 contribution (solid curve marked residual) to the decline in atmospheric $^{13}C/^{12}C$ ratio as obtained by subtracting the model-derived contribution of fossil fuel CO_2 (solid curve marked model fossil fuel) from the actual atmospheric $^{13}C/^{12}C$ decline as reconstructed by Freyer and Belacy (1983) (solid step curve). (B) Forest and soil CO_2 input time history based on the deconvolution of Freyer and Belacy's tree-ring-based atmospheric $^{13}C/^{12}C$ curve. Shown for comparison are fossil fuel CO_2 input time history and the sum of these two contributions. (Source: Peng et al. 1983).

Fossil fuel carbon is also depleted in ^{13}C relative to the atmosphere. Deforestation and other losses of terrestrial biomass have been estimated by subtracting the expected ^{13}C decrease attributable to fossil fuel combustion from the decrease, either measured in the atmosphere or inferred from proxy $^{13}C/^{12}C$ time series such as tree rings (Figure 2.3; Chapters 3, 6, 7, and 10).

The carbon isotopic record of atmospheric CO_2 in tree rings is difficult to interpret because of interferences associated with physiological effects of isotopic fractionation and because an accurate global carbon cycle model is needed for interpretation (Chapters 3, 7, 10, and 11). Several other types of potentially useful records for the past history of atmospheric CO_2, which required new analytical techniques for their interpretation, are also being evaluated. For example, these records

include a series of spectrographic plates exposed at the Earth's surface from 1898 to the present, which may register the CO_2 content as changes in absorption of solar radiation, and CO_2 in "old air" (from 20 to 100,000 years B.P.) extracted from gas bubbles trapped in ice from Greenland and Antarctica (Chapter 3).

2.6 MODELING

Mathematical models are the primary analytical tools for addressing questions on the past and future responses of the global carbon cycle to CO_2 emissions. Systems analysis and mathematical models provide a rigorous quantitative basis for examining the relationships within and among the temporal and spatial hierarchies of processes that comprise the global carbon cycle. However, faulty data and/or incompletely modeled processes can be very difficult to detect, both because the nonlinear and interacting portions of many models defy intuitive understanding and because modern technical journals are increasingly reluctant to publish the large amounts of information required to analyze complex models.

Systems analysis and modeling of the carbon cycle address three major research needs: (1) determination of the characteristics of the carbon cycle from direct analysis of data; (2) establishment of the significance of various processes and the validity of the models through development, calibration, and testing of models in historical contexts; and (3) assessment of the likely consequences of alternative energy policies and land-use practices to future atmospheric levels of CO_2, based on the application of valid models. Model approaches to meeting these research needs are discussed below.

The primary factors determining the oceanic uptake of CO_2 from the atmosphere are complex and geographically, annually, and seasonally variable. Mathematical models permit integration of these complex biogeochemical processes, leading to calculation of the global exchanges of carbon from experimental data and knowledge about the global ocean system.

Models require structural organization and functional interconnections to represent the global dynamics of the biogeochemical cycle (Figure 2.4). Mathematical models of the ocean system not only integrate and synthesize these discrete sets of information, but also allow researchers to simulate the behavior of the chemical processes under changed atmospheric conditions and to evaluate the relative importance of different processes in regulating the capacity of oceanic uptake of CO_2 (Chapter 7).

Annual cycles and fluxes of CO_2 between the biosphere and the atmosphere are highly ecosystem-dependent and, therefore, highly variable geographically, as are air-sea CO_2 exchange, ocean current structure, and vertical mixing phenomena. The response of terrestrial ecosystems to a disturbance (e.g., forest clearing) is known to vary significantly among different types of ecosystems and to depend on the history of previous disturbance (e.g., the cycle of disturbance and regrowth, or other shifts in land use). In some regions, disturbance may have affected the carbon balance up to several hundred years before the start of the industrial era. Potentially important climatic feedbacks or CO_2 fertilization effects in the global carbon cycle are also expected to be highly dependent on ecosystem type and geographic location. Treatment of these regional differences cannot be achieved without the use of models (Chapters 7, 10, and 11).

The annual cycle of photosynthesis and respiration by land plants is one of the most pronounced features of the terrestrial component of the carbon cycle and may be treated in models to analyze the annual and the longer variations apparent in CO_2 records. Spatially differentiated atmospheric, oceanic, and terrestrial models could allow simulation of the annual carbon cycle. Differences in the simulated and observed values at any location would suggest inconsistencies in the models (Chapter 7).

Particularly important to the biosphere is the nonlinear photosynthesis-respiration response of ecosystems to temperature. Direct CO_2 stimulation of the biosphere ("fertilization") could also affect the global carbon cycle, even without climate-induced changes. However, caution to avoid inappropriate application of theory to the real world is required (see also Chapters 1 and 10). For example, forests, which contain 80% of the world's organic carbon, are of great importance among terrestrial carbon sinks. Global plant growth stimulation because of enhanced atmospheric CO_2 has been extrapolated from results of greenhouse experiments on herbaceous crops (Goudriaan and deRuiter 1983; Goudriaan and Ketner 1984; Esser 1984). The enhancement, in turn, has been used to balance present-day global geochemical carbon budgets (e.g., Gifford 1980; Goudriaan and Ketner 1984). In contrast to the models, tropical forests are actually disappearing at 0.3% per year because of a growing human population (Chapter 6). Dead or senescent forest trees are obvious today over large areas of the Northern Hemisphere (e.g., McLaughlin et al. 1984; Plochmann 1984), whereas growth enhancement remains a speculative issue, suggested only in a few individual trees growing at high elevations (LaMarche et al. 1984).

At the very least, models that require positive vegetation growth enhancements from atmospheric CO_2 are currently premature. Assessments based on incorrect or inadequate information are likely to yield misleading predictions of future ocean and terrestrial

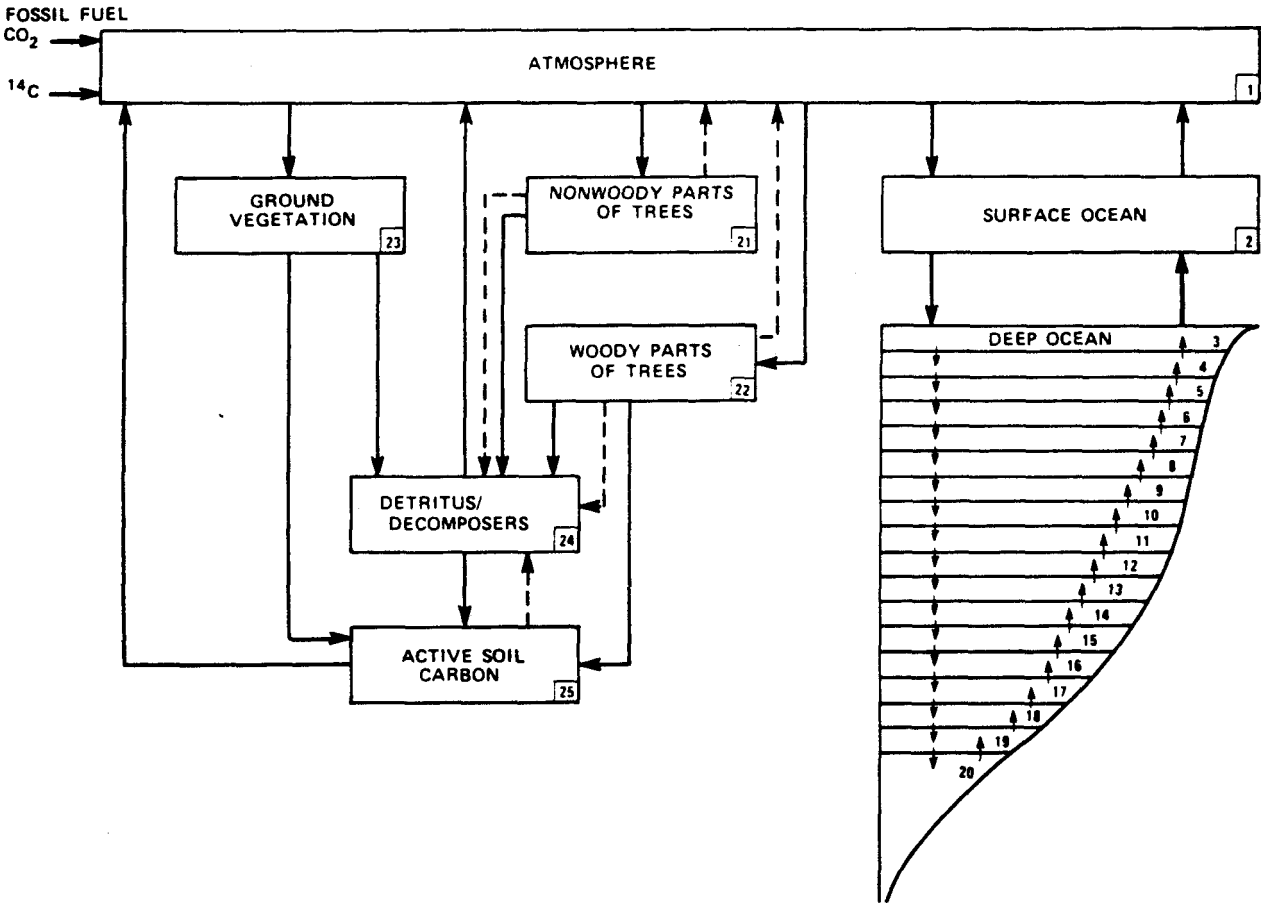


Figure 2.4. Compartment diagram of a model designed to represent the global carbon cycle. The atmosphere, which is treated as a single well-mixed reservoir, exchanges carbon with the inorganic pool in the oceans and with vegetation and soils on land. The *surface ocean* corresponds to waters above 75 m. Carbon turnover in the *deep ocean* is by exchanges between adjacent layers. Assimilation of carbon from the atmosphere by plants (primary production) is divided among *nonwoody parts of trees, woody parts of trees,* and *ground vegetation,* each of which decomposes (releases carbon) at a different rate. Carbon in dead organic matter is divided between *detritus/decomposers,* which correspond to litter at the soil surface, and *active soil carbon,* which corresponds to organic debris in the soil column, undergoing decomposition. Solid arrows indicate naturally occurring carbon fluxes; dashed arrows are associated with forest clearing. There are two inputs of carbon to the model system, release of CO_2 by fossil fuel combustion and natural or bomb-^{14}C, indicated by double arrows. Indexes in each compartment indicate the variable assignment in model equations (Chapter 7). (Source: Emanuel et al. 1984).

ecosystem responses. Models that include the other sources and sinks discussed above can provide no better information than the data on which they are based. At present, new basic research on many physical, chemical, physiological, and ecological mechanisms is required before adequate models can be developed.

Carbon cycle models with globally averaged inputs and outputs are currently the principal tools for projecting future atmospheric CO_2 concentrations. These models are used to calculate CO_2 concentrations for, and the response of other variables to, different fossil fuel and land-use scenarios. Fossil fuel scenarios are now generated with quite complex regionally and technologically differentiated energy-economic models, although the impact of many potentially important human factors remains unknown (e.g., governmental policies, feedback effects of climate change itself, and the perception of change). The field of policy analysis remains largely unexplored territory (Chapter 9). Best current assumptions on model formulation and parameters are adapted to generate base-case projections. Alternative models and parameter values must also be considered because of current uncertainties (Chapters 7, 9, and 10). In instances where alternative model formulations or parameter sets yield substantially different results from a base-case formulation, the causes or modeling assumptions leading to the differences can then be isolated and analyzed. The

sensitivity of calculated CO_2 values to such changes can indicate areas where uncertainties require particular attention in research and analysis.

Unfortunately, current global averages of critical CO_2 fluxes used in models (e.g., terrestrial CO_2 exchange and atmosphere-ocean carbon flux) do not represent measured values. Rather, they result from the judgmental weighting and averaging of assumed values derived, in turn, from mathematical representations that are only surrogates of the real processes (e.g., atmosphere-ocean CO_2 exchange and vertical mixing, Chapters 5 and 7). The processes themselves operate at very small spatial scales and are highly variable temporally (Chapters 3 and 5–8). Thus, development of a global model requires very careful assembly of hierarchical sets of data and assumptions to compensate for spatial and temporal variability and for uncertain knowledge about key processes. The process of assembly is a highly judgmental one at present and, in itself, is a key area of inquiry (Chapters 8 and 11).

The impact of some of these assumptions can be estimated by propagating uncertainties in parameter values through the models. This can be done to develop ranges of uncertainty on predicted atmospheric CO_2 concentrations. Monte Carlo simulations of model response are one means to conduct an uncertainty analysis (e.g., National Research Council [NRC] 1983; Chapters 9 and 10). Where uncertainty in projected CO_2 concentrations is unacceptably large, the contributing uncertainties in model parameters can be analyzed and, where possible, reduced as data and model refinements permit (Chapter 11).

2.7 SUMMARY AND CONCLUSIONS

How does one investigate a problem like the issue of increasing global CO_2? The causes and consequences of the problem are visible and are intertwined at all spatial and temporal scales, providing a single issue of unrivaled complexity. In addition, there is no aid to be gleaned from written protocols nor from previous studies on other similar problems. In contrast to the solution of more traditional scientific problems, resolution of the global-scale CO_2 issue requires very different research methods, new measurement procedures, and a multiplicity of coordinated approaches (Chapter 11). These include enhanced information exchange, creative use of extant data to describe the past behavior of the global carbon cycle, monitoring of key components to understand the contemporary system, experimentation to uncover key processes and major fluxes, and modeling to integrate the diverse components and processes.

The tendency exists to view carbon cycle research as a purely applied program (e.g., refining carbon budgets, measuring changes in ecosystem areas, monitoring atmospheric changes, measuring penetration rates or distributions of oceanic tracers, or recording fossil fuel use). Yet, our inability to balance the carbon cycle is determined in large part by insufficient basic research devoted to resolving the more fundamental biogeochemical processes that govern the global carbon cycle. The basic research and new data collections that must be accomplished to reduce carbon cycle uncertainties, and the current knowledge of the dynamics of that cycle, are detailed in the following chapters.

In no aspect of the carbon system is global inference more difficult than in modeling fossil fuel CO_2 emissions: enter the complex and essentially insoluble human equation, complete with the threat of global nuclear disaster and unstable political and socioeconomic structures but also with the latent capability for adaptation and technological wizardry. Thus, our future is unknowable, and much of the uncertainty in global estimates of energy-related CO_2 emissions is doubtlessly irreducible. The reasons for this conclusion and the resulting implications were outlined at the close of Chapter 1, and are discussed more fully in the concluding chapters in the volume.

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3. HISTORY OF CARBON DIOXIDE IN THE ATMOSPHERE

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3.1 OVERVIEW

The evolution of the concentration of carbon dioxide (CO_2) in the Earth's atmosphere during the last 100 million years may be divided into three stages (Figure 3.1). During the first stage, from 10^8 years B.P. until the end of the last ice age 10^4 years B.P., very large, and as yet poorly understood, changes in the atmospheric CO_2 level are thought to have occurred. From an initial concentration, perhaps as high as several thousand parts per million (ppm) in Cretaceous times, the atmospheric CO_2 level fell gradually toward much lower values (200-300 ppm) characteristic of the glacial-interglacial cycles of the past few million years. The CO_2 level seems to have varied in a regular, periodic fashion, cycling at the ice-age frequency ($\sim 10^5$ years) between a low value near 200 ppm (cold, glacial phase) and a high value near 270 ppm (warm, interglacial phase) several times during the last million years.

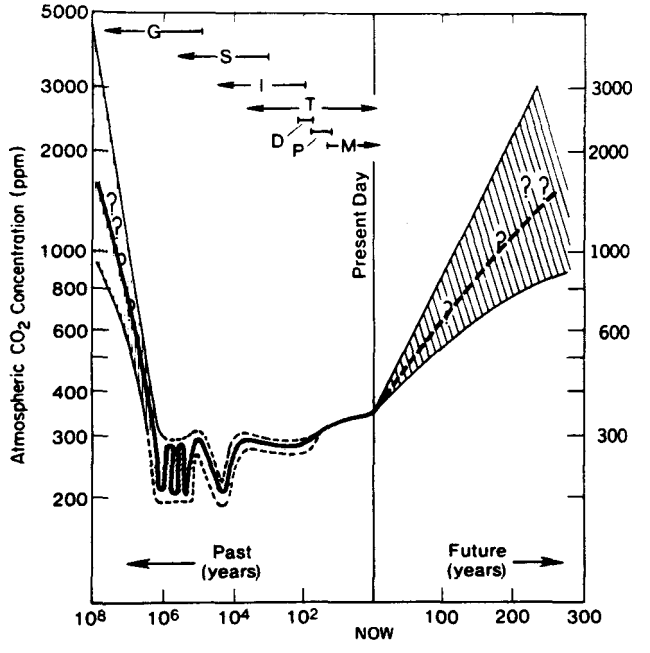


Figure 3.1. Overview of the history of atmospheric CO_2 from 10^8 years ago to the present day on a log-log scale. For comparison, the possible future CO_2 levels projected for the next few centuries are shown on a linear time scale. The analytical method appropriate to each time scale is indicated by the letters and arrows: (G) geological carbon cycle models, (S) ocean sediment cores, (I) trapped air bubbles in ice cores, (T) ^{13}C isotopic studies of tree rings, (D) direct chemical measurements of the past century, (P) spectroscopic plates from Smithsonian Solar Constant Program, (M) Mauna Loa record and subsequent precise atmospheric CO_2 measurements by nondispersive infrared spectroscopy. The hatching and dashed lines indicate the general level of uncertainty about the estimated CO_2 concentration in each time interval. The ice-age cycling is only representative of the range of CO_2, not the specific number of glaciations.

The second stage was the postglacial period of relatively steady atmospheric CO_2 concentration (260-290 ppm), which lasted from 10,000 years B.P. until the beginning of the 19th century (Figure 3.1). The small variations in CO_2 level observed or inferred during this period are considered natural, and human influences on these variations minimal until recent millennia.

The third stage, from 1800 B.P. to the present, is one in which the human impact on the global carbon cycle becomes clearly measurable in the atmosphere, in tree rings, in ice cores, and in the ocean. Initially during the 19th century through land-use practices (i.e., deforestation) to meet the needs of a growing global population, and later during the 20th century through the increasing reliance of the world on fossil fuel energy resources, human activities rapidly came to dominate all natural sources of CO_2 variability.

These three stages are addressed chronologically in Sections 3.3 to 3.5 of this report. The review of the history of atmospheric CO_2 is preceded by a brief summary (Section 3.2) of relevant analytical methods. Of these, the direct measurement of CO_2 in trapped air bubbles in polar ice cores has been emphasized as the most promising by the World Meteorological Organization (WMO 1983). Proxy CO_2 records derived from $\delta^{13}C$ isotopic measurements in tree rings are briefly reviewed here and are also discussed in the following chapters (all chapter, section, figure, and table references in the text are to this volume).

As a point of definition, we have not used the word "preindustrial" to describe atmospheric CO_2 levels measured or inferred after 1850, but have instead chosen the year 1800 as a better boundary before which time human influence on the global carbon cycle is difficult to document in the atmospheric record.

The modern period of very precise (<0.1%) time series measurements of atmospheric CO_2 began with Keeling's pioneering infrared measurements in Hawaii in 1958. Keeling recently presented an excellent review of current knowledge of the carbon cycle (Keeling 1983). In this chapter, we emphasize particular aspects of the recent atmospheric CO_2 record, with significant implications for the separation of the natural and manmade sources of variability in CO_2 and the global carbon cycle:

1. The potentially slower growth rate of atmospheric CO_2 at Mauna Loa in the past decade in parallel with the reductions in global fossil fuel production rates.
2. The unusual perturbation of the global carbon cycle caused by the very strong El Niño/Southern Oscillation (ENSO) event of 1982/83.

Table 3.1
Summary of Direct and Indirect Methods for Reconstructing the Atmospheric History of CO_2 From the mid-Cretaceous (100 million years B.P.) to the Present

Method	Time Frame (years)	Resolution (years)	Approximate Precision[a] (percent)
Geological Carbon Cycle Model	$10^8 \rightarrow 10^5$ B.P.	10^5	(?) × 2-3
Isotope in Ocean Sediment Cores	$3 \times 10^5 \rightarrow 10^3$ B.P.	10^3	10-20
Direct Measurement CO_2 in Trapped Polar Ice	50,000 B.P.→ 1980 A.D.	20-300 [lag $\sim 10^2$-10^3]	(1)
Carbon Isotopes in Tree Rings, Coral	10,000 B.P.→ present	Annual	5(2)
Ocean Chemistry	100 A.D.→present	30-150	5
Direct Wet Chemical	1860→1960 A.D.	Day/month	3
Optical Absorption Spectra (Smithsonian plates)	1900→1958 A.D.	Day/month	5(2)
Nondispersive Infrared (Mauna Loa, South Pole)	1958→present	Minute	0.1(0.03)

[a]Precision values in parentheses represent improvements anticipated for the near future.

3. The growing evidence that the annual exchange of CO_2 between the atmosphere and the Northern Hemisphere land biosphere has begun to change in recent decades.
4. Atmospheric CO_2 vertical profiling and isotopic time series in Australia.

The final section (3.6) summarizes current uncertainties and research priorities in the global carbon cycle that present and future atmospheric measurements should address.

3.2 REVIEW OF ANALYTICAL METHODS USED TO ESTABLISH THE HISTORY OF ATMOSPHERIC CO_2 LEVELS

The history of atmospheric CO_2 over the past 100 million years can be reconstructed by combining results from several distinct analytical approaches (Figure 3.1, Table 3.1). Some are less precise, are relatively indirect, and require important assumptions and carbon cycle models (e.g., geological carbon cycle models and interpretations of carbon isotopic records in ocean sediment cores, tree rings, and corals). Other methods are more direct and precise, and are not dependent on model assumptions (e.g., CO_2 determinations in trapped air bubbles in polar ice cores, the direct wet chemical measurements of the late 19th century, and the modern infrared measurements beginning in 1958). The time spans accessible by each method are shown in Figure 3.1 and in Table 3.1.

For that part of the postglacial period before human influence becomes measurable (10,000 B.P. to 1800 A.D.), the applicable data sources and associated time intervals for recovering the history of atmospheric CO_2 are (Table 3.1): sediment cores (3×10^5 to 10^3 B.P.), ice-core air bubbles (entire period), tree-ring $\delta\ ^{13}C$ (entire period, and ocean chemistry (since $\sim 10^3$ B.P.). The two most promising sources, ice cores and tree rings, will be reviewed first.

3.2.1 Direct CO_2 Measurements of Trapped Air in Polar Ice Cores

The two major research centers are at the Physics Institute of the University of Bern, Switzerland (Oeschger et al. 1984), and the Laboratory of Glaciology and Environmental Geophysics in Grenoble, France (Raynaud and Barnola, in press). Both groups use a dry extraction technique, crushing ice samples of various volumes at low temperature and measuring the CO_2 evolved from the trapped air bubbles by laser infrared spectroscopy (Bern) or gas chromatography (Grenoble). The accessible time frames and analytical resolution of the CO_2 ice-core measurement are 50,000 to 20 years B.P. and ~20 years, respectively (Schwander and Stauffer 1984; Neftel et al., 1985).

Direct CO_2 measurements in trapped polar ice are currently limited to ~50,000 years B.P. by the length of the available cores. Although below ~1300 m, the pressure of the overlying ice causes the trapped bubbles

to disappear into the ice matrix, initial tests suggest that dry extraction still yields sufficient and unfractionated trapped air to give reliable CO_2 values (Stauffer et al., in press). Deeper cores should allow direct CO_2 determinations over the last full glacial cycle (125,000 years). Improved drilling and sample handling are required to minimize contamination, and on-site analysis may be required.

The useful ice-core CO_2 record generally stops at 1850-1900 A.D. for most cores. This limit is set by the time required for accumulating snow to compact to solid ice, thus closing off air bubbles and ending contact with the atmosphere. This lag time (bubbles younger than surrounding ice) is 10^2 to 10^3 years, depending on local polar temperatures and precipitation rates (Oeschger et al. 1984). The atmospheric ventilation of the compacting snow and firn (névé) in the upper layers of the ice core smears the CO_2 signal across the annually deposited layers, the effective time resolution being one-third to one-fifth the lag time, or 20 to 300 years. To improve on this time resolution, one must find very cold (−30°C) sites with very high snowfall rates, improve our understanding of the meltwater enhancement of trapped CO_2, and use other tracers (^3H, ^{14}C, chlorofluorocarbons) to quantify the process of bubble isolation. The Bern and Grenoble groups currently disagree on the time scale of this close-off process.

With regard to current analytical precision, both laboratories currently claim 2-3% (Barnola et al. 1983). The natural noise level in the ice is unknown, but improvements in analytical precision to 1% or better are predicted. Both groups are now attempting to resolve recently observed sample volume and methodological effects, which have resulted in differences in measured 19th century (and earlier) concentrations of 10 to 20 ppm between the two laboratories (Moor and Stauffer, in press; Schwander and Stauffer 1984; cf. Neftel et al. 1985; with Raynaud and Barnola 1985).

Some very useful related measurements will greatly enhance the value of the ice-core CO_2 record. As accelerator mass spectrometric techniques for ^{14}C dating improve, the dating accuracy of the deep cores should improve dramatically. Studies of the profiles of anthropogenic tracers (^{210}Pb, ^{14}C, chlorofluorocarbons) in the tops of the cores will allow close-off processes to be better quantified. The observed close correlation of rapid variations in atmospheric CO_2 and oxygen isotopes ($^{18}O/^{16}O$) in Greenland ice (see Section 3.3.2) should be pursued in other locations (Antarctica) to establish whether such fluctuations ($\leq 10^2$ years) are truly global (Oeschger and Stauffer 1986; Stauffer et al., in press). Measurements of atmospheric $^{13}CO_2$ in trapped ice should also be attempted, despite the sample requirements, as well as measurements of other trace gases of the carbon cycle [e.g., methane (CH_4) and carbon monoxide (CO), and their isotopes], gases that are now also implicated in climate change via radiative or chemical effects. The Bern group has already begun the determination of $^{13}CO_2$ in the ice air bubbles (Friedli et al. 1984; Neftel et al. in press).

3.2.2 Indirect Determination of Atmospheric CO_2 History from Carbon Isotopes in Tree Rings

The development of a proxy atmospheric CO_2 record from carbon cycle model interpretations of the ^{13}C and ^{14}C isotopic measurements of tree rings is less direct than the ice-core method, but offers the promise of an unbroken CO_2 record, with annual resolution from 10^4 years B.P. to the present (Stuiver 1986). The controversial part of this approach is in the choice of trees with truly global, rather than just local climate and CO_2 signals (Freyer 1986; Freyer and Belacy 1983; Leavitt and Long 1983; Lorius and Raynaud 1983; Peng et al. 1983; Stuiver 1982), and in the methods by which the influence of variable plant physiology and local environment is normalized out of the $\delta^{13}C$ record to leave the residual signal, which is then interpreted as representing historical changes in global atmospheric CO_2 levels (Stuiver et al. 1984; Stuiver and Quay 1981; Francey and Farquhar 1982; Farquhar 1983; Francey 1986). The proxy CO_2 record from $\delta^{13}C$ tree-ring interpretations (Stuiver et al. 1984) shows greater apparent CO_2 variability than does the contemporaneous record from direct determinations of CO_2 in trapped air bubbles in ice cores (Oeschger and Stauffer 1986). This suggests that at least a part of the apparent CO_2 variability in the tree-ring record is, in fact, a residual signal of plant physiological effects. Recent progress in understanding the ways in which plants fractionate carbon under varying climatic conditions (water stress, temperature, nutrients, pollutants) offers the hope that future efforts to extract the history of atmospheric CO_2 concentration from ^{13}C isotopic time series in tree rings will be more realistic in accounting for that part of the variable $\delta^{13}C$ signal attributable to changes in plant physiology (Farquhar 1982; Francey and Farquhar 1982; also see Chapters 6, 10, and 11).

3.2.3 Direct Wet Chemical Measurements of Atmospheric CO_2 in the Late 19th Century

For the more recent period of the last 2 centuries in which the human impact on the global carbon cycle is first measurable (1800-1958 A.D.), sediment cores

have insufficient time resolution (10^3 years) and the ice-core record has often been limited by the dynamics of the snow compaction to dates earlier than 1900. Thus, direct wet chemical determinations first made in Western Europe in the latter half of the 19th century provide potentially useful information about the atmospheric CO_2 level in the transition period from "natural" (before 1800 A.D.) to "modern" [post-International Geophysical Year (IGY)] times. For this reason, these direct atmospheric determinations were recently reexamined in detail (WMO 1983; Stanhill 1982; Waterman 1983; Wigley 1983; Elliott 1984; Pearman 1984; Siegenthaler 1984; Fraser et al. 1986). The World Meteorological Organization (WMO 1983) panel concluded that these direct measurements point to an atmospheric CO_2 level in the last quarter of the 19th century in the range of 260 to 280 ppm, with a caveat that careful selection of these data is critical, because known systematic errors can result in CO_2 values either too high (urban pollution, diurnal bias) or too low (incomplete sample recovery, sample degradation during long storage times).

3.2.4 Historical CO_2 Levels from Absorption Spectra of the Smithsonian Solar Constant Program (1902-1962)

The analysis of spectroscopic data offers a method of estimating the concentration of CO_2 in the atmosphere for the period between the end of the ice-core record (~1850 A.D.) and the beginning of the Mauna Loa record in 1958 (Stokes 1982). The spectrum of the Earth's atmosphere has been recorded on a regular basis for nearly 100 years. As sunlight passes through the Earth's atmosphere, CO_2 molecules in the atmosphere absorb solar radiation at specific wavelengths, creating a series of characteristic absorption lines in the solar spectrum. The Smithsonian Solar Constant Program, which existed from 1902 to 1962, collected spectroscopic data for the purpose of measuring the transparency of the atmosphere. Spectra were collected almost daily at several locations on the Earth's surface. Most of the extant spectral plates were taken at Table Mountain Observatory in California (altitude 2286 m) from 1925 to 1962. The spectra are supported by considerable documentation, which will greatly facilitate the analyses (Stokes and Barnard 1986).

Although the technique for analyzing these spectra is still being refined, two relevant aspects must be considered: (1) the relationship between integrated column densities of atmospheric CO_2 and surface measurements, and (2) the potential for the old spectroscopic data base to be analyzed to the precision of $\pm 2\%$(± 6 ppm) required to make a useful contribution.

The attempt to establish a relationship between integrated column densities and surface measurements—calibrating the spectroscopic technique—is based on 4 years of spectra taken at the Kitt Peak National Observatory in Arizona on an almost monthly basis, weather permitting. A preliminary analysis of these data suggests a possible precision of ~1% and yields an average column-weighted CO_2 concentration of 340 ($\pm$ 10) ppm for the period from 1978 to 1981. Flask samples were also collected at Kitt Peak and analyzed at the NOAA CO_2 laboratory (Boulder, Colorado) to relate the seasonal variation of the integrated column CO_2 to the in situ ground-level values.

3.2.5 Ocean Chemistry

The history of changes in the atmospheric CO_2 level over the last several hundred years is also encoded in the carbonate chemistry of the deep waters of the world's oceans. To recover this signal, in particular to back calculate the initial atmospheric CO_2 level experienced by the surface waters in the early 19th century, very accurate chemical measurements (inorganic and organic carbon, nutrients, oxygen, alkalinity) are required for the deep waters and their surface waters (also see Chapter 5). Corrections must be made for the oxidation of organic matter and for the dissolution of carbonate shells. The procedure requires many assumptions regarding uncertain physical, chemical, and biological complexities; the approach is still quite controversial (Brewer 1978, 1983; Broecker et al. 1985; Chen and Miller 1979; Chen 1982a, 1982b; Chen et al. 1982). Because the history of atmospheric CO_2 is so difficult to extract from the carbonate chemistry of the ocean and so dependent on model assumptions, the ocean chemical approach must be regarded as a secondary rather than a primary method in the detailed reconstruction of the atmospheric CO_2 history since 1800 A.D. (WMO 1983). Nonetheless, the range of values ($CO_2 \simeq$ 260-290 ppm) derived from these oceanic chemical calculations is consistent with that obtained from other methods. As the deep-ocean mixing time is approximately 500 to 1000 years, the water column will contain little information about the atmospheric CO_2 concentrations in earlier millennia.

To document directly the fossil fuel CO_2 signal entering the surface waters in the 20th century will require the marine equivalent of the Mauna Loa CO_2 record; a well-designed network of ocean chemical measurements to extend over decades would be required (Scientific Committee on Ocean Research [SCOR] 1984; see Section 3.6).

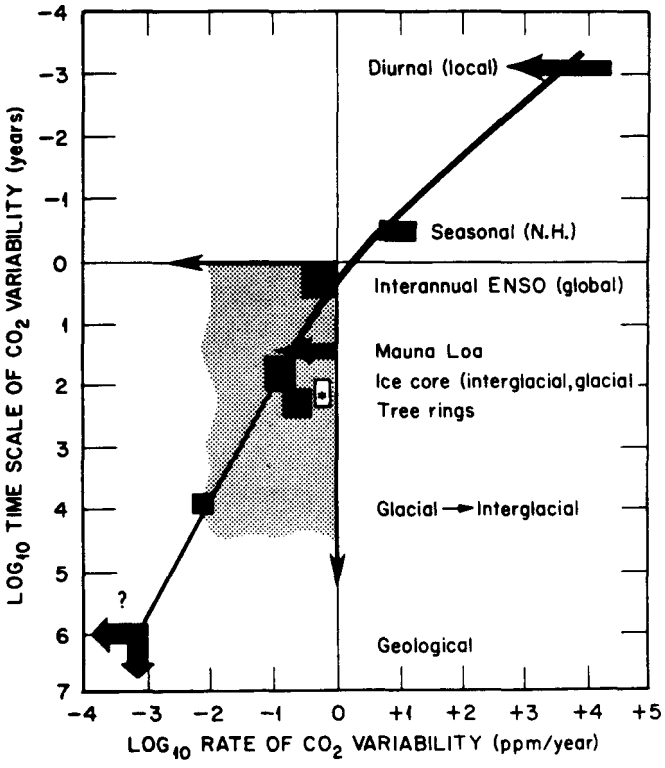


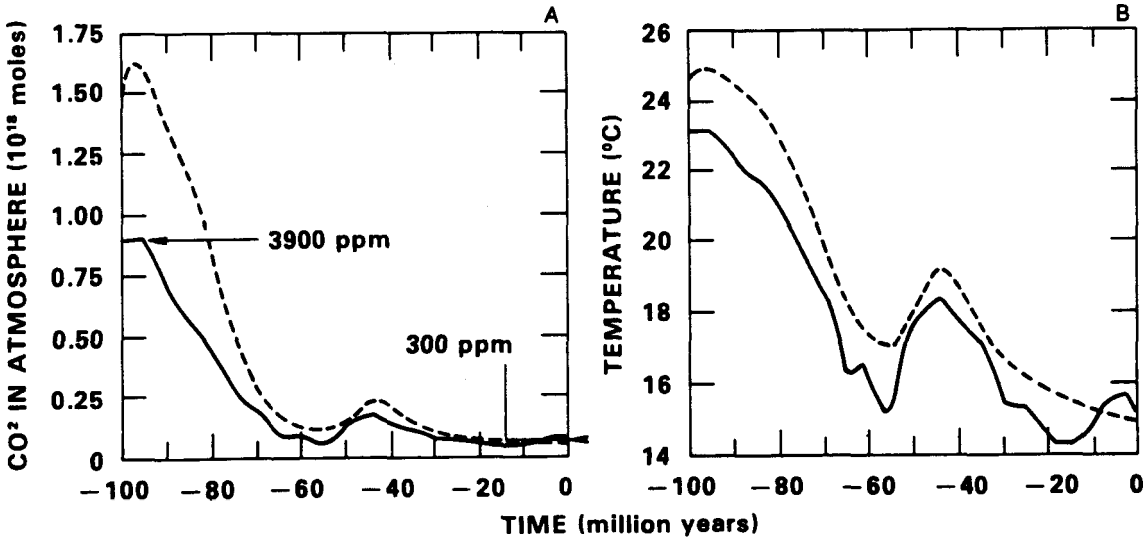
Figure 3.2. Summary of observed rates of natural CO_2 variability as a function of the time scale over which such changes occur. The most rapid changes occur during the diurnal cycle in forests (10^3 – 10^4 ppm/year over a 12-h period), the slowest is on the geological time scale of 10^6 years or longer (10^{-3} ppm/year or less). For all time scales longer than one year, the observed maximum rate of natural variability is ~1 ppm/year, for example, during strong ENSO events such as 1982/83. There is also ice-core evidence (Oeschger and Stauffer 1986) for rapid (0.5-1.0 ppm/year) CO_2 fluctuations over scales of a century during the last glacial maximum (* in figure).

3.3 NATURAL VARIATIONS OF ATMOSPHERIC CO_2 FROM THE MID-CRETACEOUS (10^8 YEARS B.P.) TO THE END OF THE LAST ICE AGE (10^4 YEARS B.P.)

3.3.1 Introduction

The history of the natural variability of atmospheric CO_2 has been estimated from four distinct approaches with overlapping time frames and differing assumptions (Table 3.1, Figures 3.1 and 3.2). On the longest time scales of interest here (10^5-10^8 years B.P.), geochemical carbon cycle models (constrained by sediment cores and by indices of sea-floor spreading, weathering, and volcanism) suggest that atmospheric CO_2 levels in the mid-Cretaceous were several times higher than the present value (Budyko and Ronov 1979; Berner et al. 1983; Lasaga et al. 1985; Figure 3.3). These model results may be uncertain by factors of 2 or more (Figure 3.3). A somewhat more direct approach is the proxy CO_2 record derived from analysis of measured carbon and oxygen isotopic changes in carbonate sediments in deep-ocean cores extending from 340,000 to 1000 B.P. (Shackleton et al. 1983; Shackleton and Pisias 1985; Figure 3.4). Although more direct and precise than the geological carbon cycle models, the sediment core method is still model-dependent and less precise than the direct measurement of the CO_2 content of trapped air bubbles in polar ice cores (Oeschger and Stauffer 1986).

Figure 3.3. Geochemical reconstruction of changing atmospheric CO_2 levels (A) and temperature (B) during the last 100 million years B.P. by Berner et al. (1983) (dashed line), and by Lasaga et al. (1985) (solid line) using an improved model (0.055×10^{18} mol = 300 ppm).



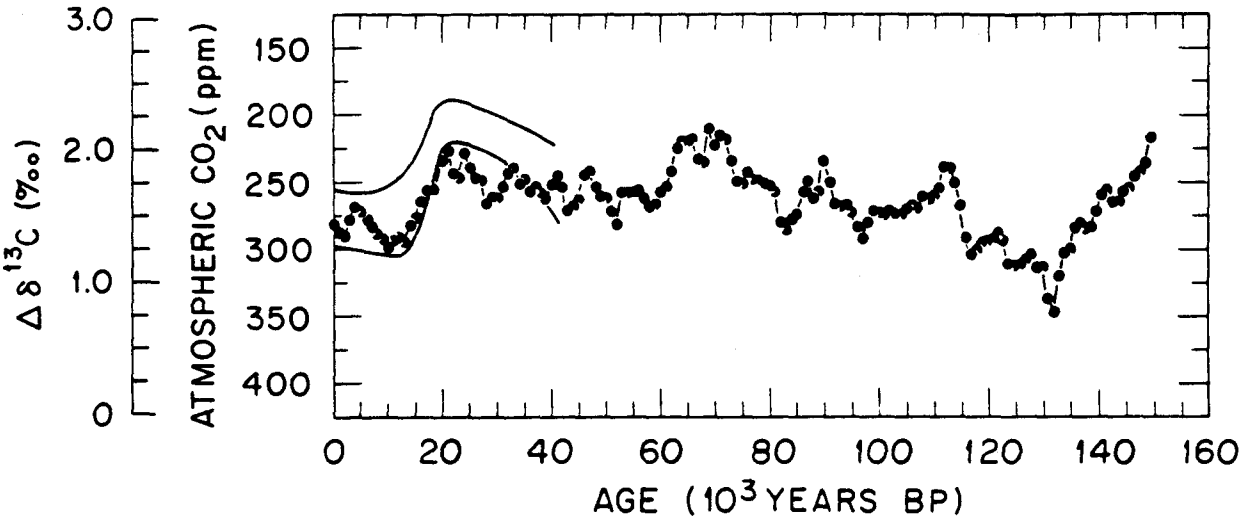


Figure 3.4. Atmospheric CO_2 as interpreted from isotopic records in sediment cores by Shackleton et al. (1983) for the last 150,000 years. Comparison with the ice-core results for the last 40,000 years is indicated by the heavy lines (Neftel et al. 1982). Notice the high CO_2 level for the last interglacial (~350 ppm about 130,000 years B.P.).

A summary of the importance of the recent ice- and sediment-core analyses to the CO_2-climate question was recently presented by Broecker (1984).

The ice-core measurements currently extend from ~50,000 years B.P. to ~1960 A.D. and seem to offer the best hope for the direct determination of the atmospheric CO_2 history during this period (WMO 1983). For this reason, polar ice was discussed in detail (Section 3.2). Finally, measurements of the $\delta^{13}C$ variations of annual tree rings can be modeled to yield a proxy atmospheric CO_2 record for the last several thousand years (Stuiver et al. 1984). The tree-ring record covers the transition period of postglacial (10^4 years B.P.) to preindustrial (1800 A.D.) times.

3.3.2 Natural Variations in Atmospheric CO_2 Between 10^4 and 10^6 B.P.

For information about atmospheric CO_2 before the 19th century, we must rely on indirect evidence such as the carbon-containing constituents of marine sediments and air trapped in continental ice sheets. From geological studies has emerged a general perception that the carbon content of the principal labile reservoirs—such as the biosphere, atmosphere, and near-surface oceans—has varied in the geological past, and that variable exchanges with the larger, more sluggish geochemical reservoirs have occurred. On the other hand, geologists have also compiled evidence for significant geochemical feedback mechanisms that act to limit the magnitude of many carbon cycle and CO_2 changes. Recent evidence strongly indicates that, despite these feedback mechanisms, atmospheric CO_2 levels varied significantly before the human influences of agriculture and fossil fuel combustion. The rates, magnitudes, and processes responsible for many of these natural CO_2 changes remain very uncertain. Moreover, understanding these changes requires consideration of a range of time scales that extends well beyond those generally associated with the anthropogenic CO_2 perturbation (Sundquist and Broecker 1985).

3.3.2.1 *Parallel Increases in Atmospheric CO_2 Level and Warming Climate at the End of the Last Ice Age*

The most conspicuous and direct evidence for preanthropogenic CO_2 variations is that obtained from measurements of air trapped in ice cores taken from the ice sheets of Greenland and Antarctica. For example, several ice-core studies (Berner et al. 1980; Delmas et al. 1980; Neftel et al. 1982) showed that atmospheric CO_2 concentrations rose rapidly from about 200 to about 280 ppm at the end of the last glacial period about 10,000 years B.P. (Figure 3.5). This correlation is tantalizing from the standpoint of climate model verification. Published models of the geochemical processes responsible for this correlation invoked climate change as the primary mechanism driving the postglacial CO_2 increase. Some studies (e.g., Shackleton 1977) hypothesized climatic influences

on CO_2 through changes in the terrestrial biosphere, whereas others (e.g., Thompson and Schneider 1981; Broecker 1982b) emphasized climatic influences on the exchange of CO_2 between the atmosphere and the oceans. Although it is clear that the direct chemical effects of sea surface warming could not account for the measured CO_2 increase (Broecker 1982a, 1982b), several studies hypothesized the transfer of CO_2 from the oceans to the atmosphere by means of climate-driven changes in the oceans' nutrient (Broecker 1982b; McElroy 1983; Broecker and Takahashi 1984; Knox and McElroy 1984; Toggweiler and Sarmiento 1985; Wenk and Siegenthaler 1985) and/or alkalinity budgets (Berger 1982). Although it would be very useful to know whether CO_2 was partially responsible for the climatic warming and contributed to the climate change at the end of the last glacial period, it is likely that the change in CO_2 was as much a result as it was a cause of the change in climate (also see Chapter 8).

3.3.2.2 *Natural Fluctuations in Atmospheric CO_2 Level During the Last Ice Age*

The interaction of CO_2 and climate is also apparent in ice core data spanning a period from about 30,000 to 40,000 years B.P. (Oeschger and Stauffer 1986). Although this period was entirely within the last ice age, a detailed ice-core record from Greenland shows rapid climatic fluctuations, indicated by changes in oxygen isotope ratios (Figure 3.6). These isotopic changes are accompanied by closely correlated CO_2 changes ranging from 180-200 ppm up to 250-270 ppm. Some of these changes appear to have occurred within 1 or 2 centuries. These data may have profound implications for anticipating the effects of anthropogenic CO_2. They demonstrate that some natural CO_2 changes have been comparable in both rate and magnitude to the changes measured during the last few decades. Moreover, these rapid CO_2 changes were likely associated with rapid changes in climate. These conclusions must be confirmed by comparable measurements in ice from Antarctica, and it should be emphasized that ice-core data covering the last 1000 years do not show any evidence for such large CO_2 variations (Oeschger and Stauffer 1986). Nevertheless, the Greenland ice-core CO_2 data appear to open the possibility of very significant feedbacks between the carbon cycle and the climate system on a time scale of less than a century.

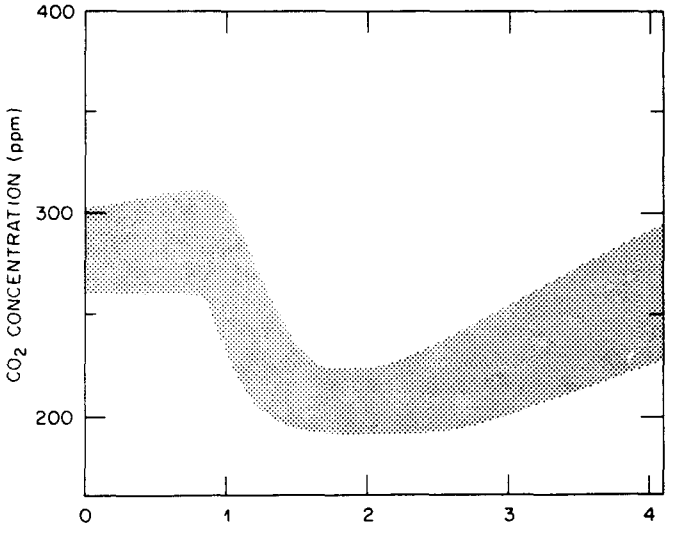


Figure 3.5. The variation in atmospheric CO_2 concentration as measured in polar ice cores by the Bern group (Neftel et al. 1982). Note the glacial low of ~200 ppm, rising abruptly at the end of the last ice age to ~280 ppm.

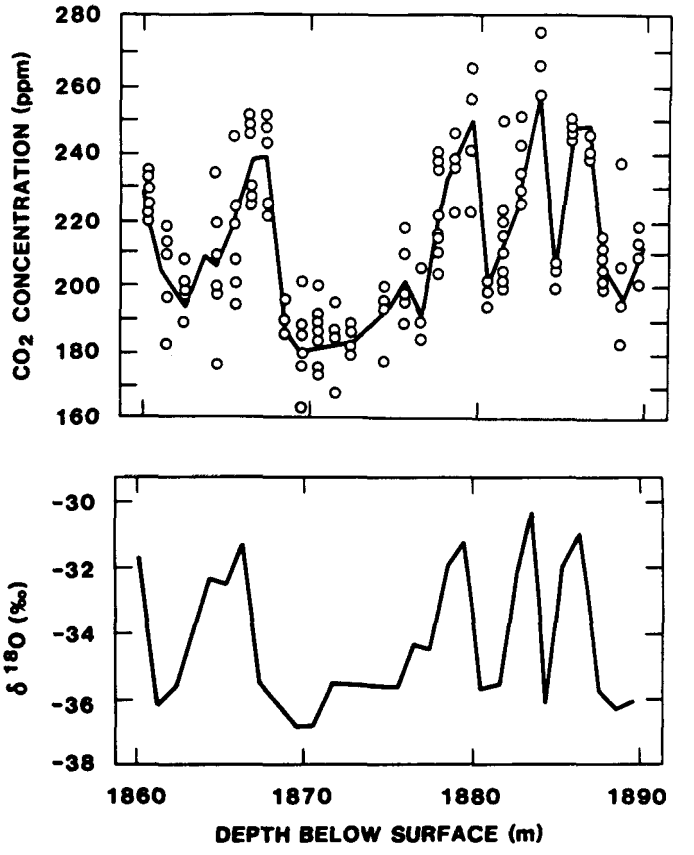


Figure 3.6. Rapid fluctuations (≥70 ppm in ~100 years) in the atmospheric CO_2 concentration during the middle of the last ice age (30,000-40,000) measured by the Bern group (Oeschger and Stauffer 1986). The ice-core sample was from Greenland. Note the excellent correlation between the CO_2 change (A) and the oxygen isotopic record (B), indicating that the climate changes as evidenced by sea-level precipitation rate and temperature occur concurrently. Open circles represent individual CO_2 measurements.

3.3.2.3 *Glacial-Interglacial Variations in Atmospheric CO_2 Level From Ocean Sediment Cores*

Although ice-core CO_2 data representing the last interglacial are not available, deep-ocean sediments appear to provide an independent record from which atmospheric CO_2 concentrations can be inferred for at least the last several hundred millennia (Shackleton et al. 1983; Shackleton and Pisias 1985). This approach rests on assumptions that can be tested with the ice-core and sediment-core data for the deglaciation of about 10,000 years B.P. These assumptions are: (1) the atmospheric CO_2 level during this time period responded primarily to the transfer of organic carbon to and from the oceans and to the redistributions of dissolved carbon within the oceans, and (2) the past abundance and distribution of carbon within the oceans can be calculated from the carbon isotope ratios of coexisting benthic and planktonic foraminifera extracted from sediments. Calculations proposed by Broecker (1982b) and Shackleton et al. (1983) showed that these assumptions are consistent with the ice-core CO_2 data and the sediment isotope data for the most recent deglaciation event (Figure 3.4). Shackleton et al. (1983) and Shackleton and Pisias (1985) therefore extrapolated backward to earlier times, using the same calculations applied to an extended carbon isotope record from a Pacific Ocean sediment core. Using a detailed benthic oxygen isotope record from the same core, these authors concluded that climatic fluctuations have been associated with regular CO_2 changes between limits of ~200 and 300 ppm throughout several glacial-interglacial cycles. On the basis of time series analysis they also argued that the CO_2 changes *preceded* the oxygen isotope changes, which reflect variations in global ice volume, suggesting that CO_2 has been a *cause* rather than merely a *consequence* of climate change. The connection between CO_2 and climate does not appear to require ice volume or sea-level changes. However, the precise mechanisms and causal relationships remain unknown.

3.3.2.4 *Geochemical Models of Changing Atmospheric CO_2 Levels and Geological Processes (10^6 to 10^8 Years B.P.)*

Whereas studies concerned with hundreds to thousands of years can teach us about CO_2 variations comparable to those of the last few decades, even longer time scales must be studied to find geologic analogs of CO_2 changes comparable to those anticipated from the burning of fossil fuels during the next century and beyond (Figure 3.1). Because significantly larger CO_2 fluctuations are plausible over longer time scales, geologists extended their studies of past atmospheric CO_2 levels to time scales of millions of years and more. Over these time scales, CO_2 can respond to changing rates of geological processes (e.g., volcanic emissions, weathering, ocean floor spreading) that are not as important to relatively short-term natural fluctuations (Figure 3.2). For example, geochemical models have been used to suggest that the relatively high activity of volcanoes about 100 million years B.P. was responsible for CO_2 concentrations about 13 times the present level (Lasaga et al. 1985). This hypothesis is corroborated in general by a climate model (Barron and Washington 1984) that suggests that conditions during the same period required CO_2 concentrations 2 to 10 times the present level. On the other hand, results from another recent study (Shackleton 1985) indicate that the carbon isotope record of ocean sediments deposited during the last 100 million years is not consistent with such high CO_2 levels.

3.3.3 Relevance to Fossil-Fuel-Induced Climate Change

Geological studies emphasized not only the importance of atmospheric CO_2 to climate, but also the ways in which climate can affect atmospheric CO_2. Current studies are concerned with the history of the atmospheric CO_2 level over time scales ranging from decades to millions of years and longer. Examinations of longer time scales, although often emphasizing processes that are unimportant over the time scales of most future CO_2 scenarios, appear to offer the prospect of providing information about geologic analogs of the more than twofold CO_2 increase that could ultimately be attained if fossil fuel consumption continues to increase. These long-term studies also lead to the conclusion that the geochemical response to fossil fuel CO_2 is likely to last for thousands of years and perhaps longer (Sundquist 1986). Most importantly, the "natural" history of CO_2 suggests very strongly that the climate system and the carbon cycle are intimately intertwined on all time scales.

3.4 HISTORY OF ATMOSPHERIC CO_2 CONCENTRATION FROM POSTGLACIAL TIMES (10^4 YEARS B.P.) TO THE INTERNATIONAL GEOPHYSICAL YEAR (1958)

3.4.1 Introduction

This section documents the period of nearly steady atmospheric CO_2 level since the end of the last ice age (10^4 years B.P.) until the beginning of the Mauna Loa record in 1958. In contrast to the glacial-interglacial

variations of the atmospheric CO_2 concentration (200-300 ppm), the atmospheric CO_2 level remained nearly constant (260-290 ppm) from 10,000 years B.P. until the beginning of the 19th century (Figure 3.1). This interval has been divided into two periods. The first period covers postglacial times to 1800 A.D., during which time human impact on the global carbon cycle is not observed in the proxy atmospheric records and apparent CO_2 fluctuations in tree-ring and ice-core data (± 10-40 ppm over 50- to 200-year intervals) are considered to be largely of natural origin. The second period, 1800-1950, is one in which the human perturbation of the natural global carbon cycle can first be well documented in the atmospheric record, and observed CO_2 concentration variations (Figures 3.1 and 3.7) are ascribed to both natural and anthropogenic causes. Analytical methods applicable to this period (10,000 B.P. to 1958 A.D.) were described in Section 3.2.

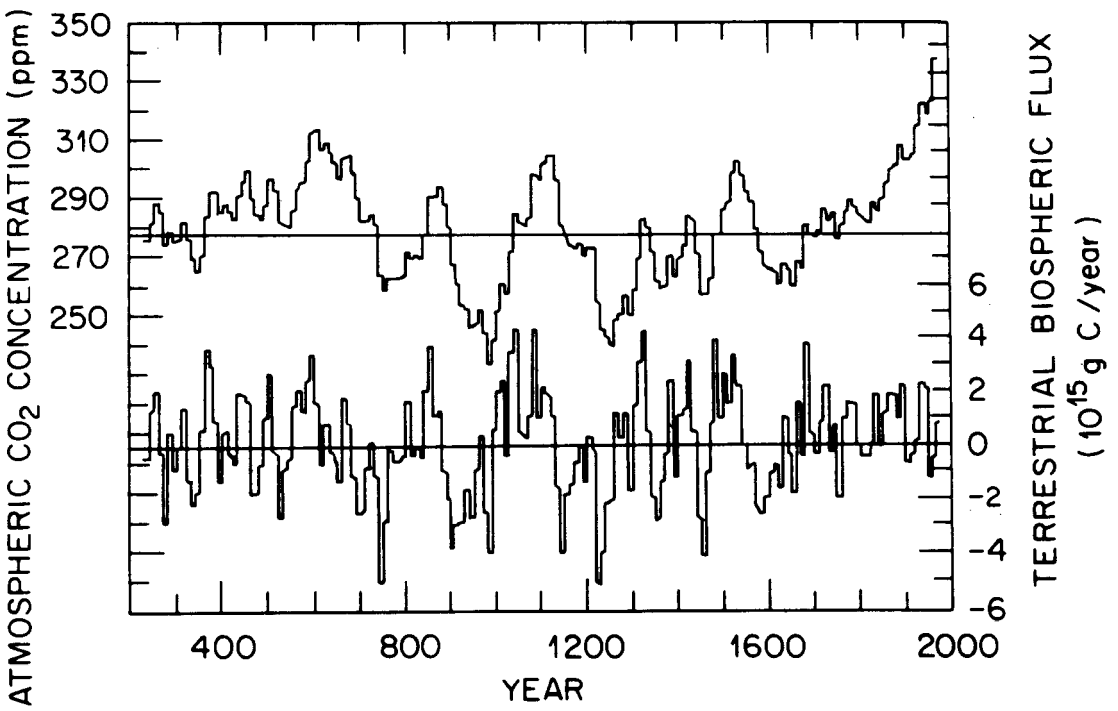
3.4.2 Natural Variation of the Atmospheric CO_2 Level from Postglacial Times to 1800 A.D.

Analyses of ice cores suggest that the atmospheric CO_2 level has been relatively steady since the end of the last ice age (Neftel et al. 1982). There is weak evidence from ice and sediment cores that during the warm period (altithermal) some 8000 years B.P. that CO_2 levels were 300-320 ppm, or approximately 10-20% above the postglacial mean (Figures 3.4 and 3.5).

To understand the variations in global atmospheric CO_2 observed in recent decades, it is important to assess the natural variability of the global carbon cycle during this earlier postglacial period when human impact was minimal (also see Chapter 8). The ice-core data from the Grenoble (Raynaud and Barnola, in press) and the Bern (Oeschger and Stauffer 1986) groups hint at natural CO_2 variations on the order of ± 10 ppm on century time scales (~0.1 ppm/year) during the last 1000 years. Recall that variations on much shorter time scales would be lost in the slow close-off of the trapped air bubbles in the ice.

The proxy CO_2 record from carbon cycle interpretations of the $\delta^{13}C$ isotopic time series suggests considerably larger fluctuations on the order of 20-50 ppm over century time scales (0.1-0.5 ppm/year) during the last 2000 years (Stuiver et al. 1984; Figure 3.7). Recall that tree-ring data have annual resolution and that the apparent variability of the derived CO_2 history undoubtedly still contains residual effects of plant physiology and local conditions, and hence must be considered as an upper limit to the climatic CO_2 variability. Averaging many individual $\delta^{13}C$ tree-ring

Figure 3.7. Proxy atmospheric CO_2 history for the period 200 A.D. to the present for carbon cycle model interpretation of $\delta^{13}C$ tree-ring measurements (Stuiver et al. 1984). The apparent variation (70 ppm peak to trough) over time scales of centuries probably reflects climatic signals mixed with residual effects of plant physiology.



records from globally dispersed sites may reduce this local plant signal and enhance the atmospheric CO_2 signal (Freyer 1986). Isotopic signals in banded corals provide yet another record of variations in the global carbon cycle (Druffel 1982).

It is significant that the CO_2 variability during postglacial times ($\leq$0.1-0.5 ppm/year) is much less than that recently found in the ice-core record for certain periods of the last glacial period (0.5-1.0 ppm/year increase over a century; Oeschger and Stauffer 1986). It will be important to evaluate the natural CO_2 variability during the previous interglacial period (120,000 years B.P.) when sufficiently deep ice cores became available.

The apparent rate of CO_2 increase recorded at Mauna Loa since 1958 ($\sim$1 ppm/year) lies at the extreme of all known natural CO_2 variability (but see later discussion on effects of ENSO on atmospheric CO_2), and is several times greater than the mean variability (0.1-0.5 ppm/year) inferred from tree-ring and ice-core records for the current interglacial period (Figure 3.2). A similar argument against a long-term, natural CO_2 fluctuation may be made for the observed 60- to 70-ppm increase during the past century. Natural variations of this magnitude and rapidity are simply not found in the postglacial records before 1800 A.D. Why atmospheric CO_2 fluctuations might be larger and faster during glacial than during interglacial times is an important question to be addressed in seeking a better understanding of the working of the global carbon cycle under different climatic regimes (Broecker 1984; Sarmiento and Toggweiler 1984; Siegenthaler and Wenk 1984).

3.4.3 Rising Atmospheric CO_2 Levels from 1800 A.D. to 1958

That a gradual increase in atmospheric CO_2 from a postglacial mean of $\sim$280 ppm began in the mid-19th century is documented in the tree-ring record (Stuiver et al. 1984), in direct chemical measurements (Wigley 1983; Fraser et al. 1986), by inferences from ocean chemistry (Brewer 1978; Chen 1982b; WMO 1983) in solar absorption spectra (Stokes 1982), and in ice-core air bubbles (Neftel et al. 1985).

The initial perturbation of the global carbon cycle was apparently the result of changing land-use practices and is measurable as falling isotopic ratios ($^{14}C/^{12}C$ and $^{13}C/^{12}C$) in tree rings as early as 1800 (Stuiver et al. 1984; Freyer 1986). Apparently, the concentration of atmospheric methane (CH_4), the second most abundant trace gas of the carbon cycle, had also begun to increase rapidly at about the same time or even earlier (Rasmussen and Khalil 1981).

Between the mid-1950s and mid-1960s, massive injections of ^{14}C to the atmosphere from nuclear testing began to compromise the $^{14}C/^{12}C$ record of human influences in tree rings. On the other hand, tropospheric measurements of ^{14}C from 1962 to 1980 proved to be very important for testing models of CO_2 exchange and more generally for studying the carbon cycle in nature (Chapters 5 and 7). The most comprehensive set of measurements was reported by Nydal and Lovseth (1983) and is represented by data from ground-level stations from latitude 78° (Spitsbergen) to 21° (Madagascar). The report by Nydal and Lovseth was designed to provide data for modeling research, and contains complete tables (and figures) of all atmospheric CO_2 data obtained through the year 1980 by the Radiological Dating Laboratory, Norwegian Institute of Technology, Trondheim.

Our knowledge of the atmospheric CO_2 increase is summarized in Figure 3.8 (based primarily on data from WMO 1983). The WMO panel concluded that a major nonfossil source of atmospheric CO_2 dominated the observed increase in the 19th and early 20th centuries, with the cumulative releases to the present time from fossil and nonfossil (i.e., land biosphere) sources comparable (150×10^{15} to 200×10^{15} g C; also see Chapter 6). Although the detailed reconstruction of the history of the biospheric release from tree-ring data is still quite controversial (Freyer 1986; Peng et al. 1983; Stuiver et al. 1984; Leavitt and Long 1983; Francey 1986), most studies indicate that both the rate and magnitude of the observed CO_2 increase during the last century are outside the range of the expected natural variability during postglacial times (Figure 3.2). The tree-ring and spectroscopic data constitute the best CO_2 records in the first half of this century, but it was not until Keeling began his infrared measurements at Mauna Loa, Hawaii, in 1958 that the era of precise ($\leq$0.1%) atmospheric CO_2 measurements began (Keeling 1983).

The initial results (Stokes et al. 1984) from analyses of spectrographic plates are encouraging, suggesting that annual mean CO_2 concentrations (total atmospheric column) are recoverable from the Smithsonian plates with an uncertainty of $\leq$2% (6 ppm). The mean CO_2 values derived to date [297.7($\pm$ 5.6) ppm for 1935, 298.4($\pm$ 6.3) ppm for 1941, and 305.3($\pm$ 4.6) ppm for 1948] appear to be reasonable (see Figure 3.8) as is the secular increase of 0.6($\pm$ 0.3) ppm/year (1935-1940).

Very recently, the detailed history of atmospheric CO_2 during the past 2 centuries was recovered with unprecedented time resolution from air bubbles trapped in an ice core from Siple Station, Antarctica, by the Bern group (Neftel et al. 1985). They found a time lag between the ice and its occluded air of $\sim$95 years and an effective close-off time of 22 years at this cold, high-accumulation site. These findings permit the top of the core to be matched to the early Mauna Loa

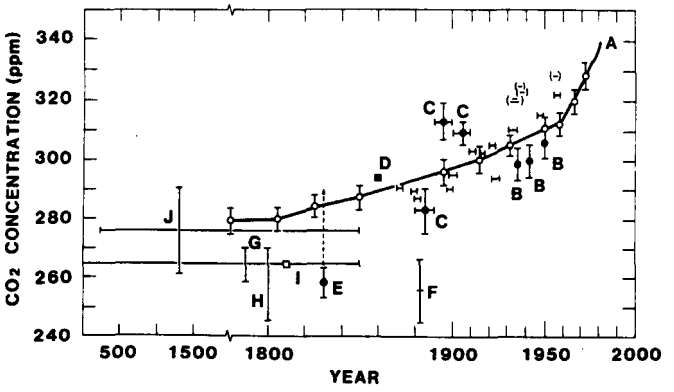


Figure 3.8. A summary figure of the history of atmospheric CO_2 from 100 A.D. to the present, combining results from all methods (WMO 1983). The different data sources are:

(A) Measurement from Mauna Loa Observatory (Keeling et al. 1982). These data are within 0.1% of the global tropospheric mean (solid line from 1958-1982).

(B) Based on analyses of archived solar spectra (Stokes et al. 1984; Stokes and Barnard 1986).

(C) Decadal average data from Montsouris Observatory, Paris, as recently presented by Stanhill (1982).

(D) Model calculation for comparison, extrapolating from (A), assuming fossil fuel combustion to be the only source of CO_2.

(E) Analysis of subsurface ocean waters giving pCO_2 at the time when the waters were in contact with the atmosphere, approximately 150 years ago (Brewer 1978). The mean and error bars represent random uncertainties and the dashed arrow indicates a possible underestimate of up to 25 ppm.

(F) Air sampled at Cape Horn by Muntz and Aubin in 1886. The standard deviation bar for 39 samples indicated that the variability is uncharacteristically high for Southern Hemisphere background air. Sample degradation on long storage may have been a problem.

(G) Typical model calculations assuming $\sim 150 \times 10^{15}$ g of nonfossil carbon released ca. 1900 (see, for example, Enting and Pearman 1982; Pearman et al. 1983b). Tree-ring data from Freyer and Belacy (1983).

(H) Estimate obtained using models to interpret $^{13}C/^{12}C$ content in tree rings (Peng et al. 1983).

(I) CO_2 concentrations in air bubbles trapped in glacial ice, based on results from French and Swiss laboratories (see Barnola et al. 1983), believed to be representative of atmospheric concentrations at the time of trapping. Average for years 1 to 1850 A.D.

(J) Same as (H) (235 to 1850 A.D.) with data from Pacific coastal tree rings normalized by annual ring growth (Stuiver et al. 1984).

– Early observations selected by Callendar (1958) as being representative of background air. Note Callendar chose to exclude the Montsouris (C) and Cape Horn (F) data from his analysis.

ϕ Very recent ice-core data from Siple Station, Antarctica, analyzed by the Bern group (Neftel et al. 1985).

record of Keeling, which began in 1958. The 1800 CO_2 concentration is found to be 280($\pm$ 5) ppm. This value is approximately 15 ppm higher than results of recent measurements from another Antarctic ice core reported by the Grenoble laboratory (Raynaud and Barnola 1985) and earlier results from both laboratories on the same Greenland ice core (Barnola et al. 1983). Figure 3.8 compares the Siple ice-core-CO_2 history with other proxy records. Although existing uncertainties about the absolute values of ice-core measurements may not yet permit unqualified adoption of the Siple record as the primary data source for atmospheric variations since 1800, when taken in combination with other data this very recent information appears to limit the year-1800 concentration to a range of 260 to 285 ppm.

3.5 ATMOSPHERIC CO_2 SINCE 1958

3.5.1 Overview

The modern period of precise atmospheric CO_2 measurements began with Keeling's pioneering determinations at Mauna Loa, Hawaii, and at the South Pole during the IGY. The Mauna Loa record (Figure 3.9) remains the single most valuable CO_2 time series in carbon cycle model verification. For very recent interpretations of atmospheric CO_2 measurements, the reader is referred to Rust et al. 1979; Cleveland et al. 1983; Fraser et al. 1983; Keeling 1983; Keeling et al. 1984; Komhyr et al. 1985; Machta 1983; Mook et al. 1983; Pearman et al. 1983b; Wong et al. 1984; Pearman and Hyson, in press).

In this section, the global atmospheric CO_2 records, particularly those of the NOAA/GMCC (Geophysical Monitoring for Climatic Change), are summarized for the period 1968 through 1983, with emphasis on the mean properties of the global carbon cycle as viewed from the atmosphere (i.e., global mean CO_2 concentration, latitude dependence of concentration and seasonal amplitude, atmospheric retention of anthropogenic CO_2, etc.). There is a brief discussion of the role of isotopic CO_2 measurements for separating oceanic from land biospheric exchanges with the atmosphere (Keeling 1983; Mook et al. 1983), with a contribution by Paul Fraser on the atmospheric $^{13}CO_2$ measurement program of the Commonwealth Scientific and Industrial Research Organisation (CSIRO), Australia.

To document the evolution of the global carbon cycle during the last decade, three examples are presented. First, an argument is made that the observed growth rate of atmospheric CO_2 at Mauna Loa has slackened during the past decade, particularly since 1979, in direct response to a slower global production of fossil fuels (Section 3.5.6). This is an important revision of the previous finding that "Until now we have not seen

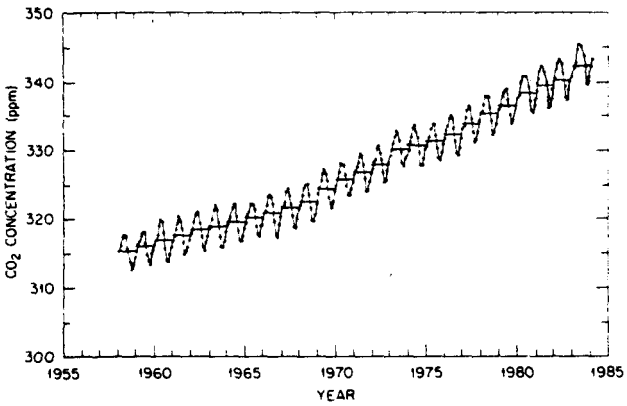


Figure 3.9. Concentration of atmospheric CO_2 at Mauna Loa Observatory, Hawaii, expressed as a mole fraction in parts per million of dry air. The dots depict monthly averages of visually selected data adjusted to the center of each month. The horizontal bars represent annual averages. Data were obtained by C. D. Keeling, Scripps Institution of Oceanography, University of California, La Jolla, California, and are from files in the Carbon Dioxide Information Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee.

any tendency for the rate of rise of CO_2 at either station (MLO, SPO) to fall off in response to a slowdown in fossil fuel production" (Keeling 1983, p. II-19).

A second example is the response of the global atmospheric CO_2 distribution to the extreme ENSO event of 1982/83 (Gammon and Komhyr 1983; Section 3.5.7); this perturbation may well be the last time that a natural fluctuation of the global carbon cycle is capable of temporarily halting the steady rise of atmospheric CO_2 due to fossil fuel combustion.

The third important indication that the global carbon cycle has begun to change is the well-documented growth in the CO_2 seasonal amplitude (1-2%/year) at Northern Hemisphere stations during the past decade (Pearman and Hyson 1980; Cleveland et al. 1983; Keeling et al. 1984; and Komhyr et al. 1985). Related changes in the growth rates and seasonalities of all the major trace gases of the carbon cycle [CO_2, CO, CH_4] suggest that the exchange of the land biosphere with the atmosphere began to change significantly in the past decade (Section 3.5.8).

3.5.2 Summary of Atmospheric CO_2 Measurements of NOAA/GMCC: Continuous Measurements at GMCC Baseline Observatories (1974-1983)

Background measurements of atmospheric CO_2 have been made continuously since the mid-1970s at four remote NOAA observatories (Pt. Barrow, Alaska [BRW]; Mauna Loa, Hawaii [MLO]; American Samoa [SMO]; and the South Pole [SPO]). The data acquisition and preliminary processing procedures have been described previously (Peterson 1978; Mendonca 1979; Peterson et al. 1982). The entire record of continuous CO_2 measurements at each station through 1983 was recently corrected, converted to the WMO X81 mole fraction scale (parts per million with respect to dry air), and edited to essentially final values. Monthly mean values of the background CO_2 concentration (Table 3.2) at the four observatories are plotted in Figure 3.10. The major episodes of line contamination and instrumental malfunction have been excluded from these records, and the previous problem of drifting calibration CO_2 standards (~0.1 ppm/year) with respect to the WMO standards maintained by Keeling at Scripps Institution of Oceanography (SIO) has now been resolved. The NOAA and SIO CO_2 records at Mauna Loa are in excellent agreement (Figure 3.11).

Each of the four NOAA stations shows a CO_2 increase of very nearly 9.0 ppm for the interval 1976-1982, yielding a mean global growth rate for atmospheric CO_2 of 1.5 ± 0.2 ppm/year. This value is only slightly affected by the anomalous growth rates of atmospheric CO_2 in 1982.

A second conclusion is that the latitude gradient of CO_2 concentration from BRW to SPO has remained near 3.5 ppm during the last decade. The observed interannual variations in this gradient (±0.5 ppm) are likely related to the ENSO events of 1976, 1979, and 1982 (Bacastow et al. 1980; Gammon and Komhyr 1983). The magnitude of the latitude gradient is directly related to the rate of release of fossil fuel CO_2 to the atmosphere of midnorthern latitudes. The fact that this gradient, which increased rapidly from ~1 ppm in 1960 to ~3 ppm in the mid-1970s (Keeling 1983),

Figure 3.10. Provisionally selected monthly mean CO_2 concentrations from continuous measurements at the NOAA/GMCC baseline observatories (Pt. Barrow, Alaska [BRW]; Mauna Loa, Hawaii [MLO]; American Samoa [SMO]; South Pole [SPO]). Values are in the WMO X81 mole fraction scale and have been provisionally selected for background conditions. The deseasonalized smooth curve is a 12-month running mean. (Refer to Table 3.2 for months with less certain means or for months in which flask means have been substituted for missing continuous means [Harris and Bodhaine 1983].)

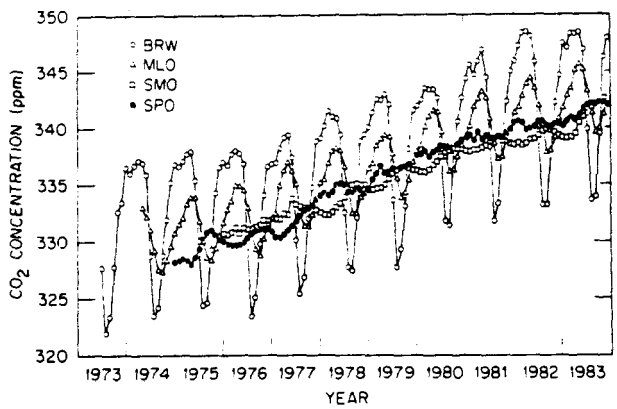


Table 3.2
Monthly Mean CO_2 Concentrations from Continuous Measurements at the Four NOAA/GMCC Baseline Observatories

Year	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
						Pt. Barrow Observatory						
1973	334.5[a]	335.5[a]	336.3[a]	337.0[a]	336.6[a]	334.0[a]	327.7[a]	322.0	323.4	327.8	332.7	333.5
1974	336.5	336.0	336.7	337.1	336.9	335.9	328.7	323.5	324.2	328.8	332.0	335.2
1975	336.8	336.6	337.1	337.7	337.9	335.4	330.1	324.4	324.6	329.3	334.3	336.5
1976	337.0	336.7	337.7	338.0	337.8	336.8	331.0	323.5	325.1	329.8	334.1	336.6
1977	336.8	336.9	338.2	339.2	339.4	337.4	330.1	325.4	326.8	332.0	334.7	338.8
1978	339.1	340.2	341.5	341.0	340.8	339.4	332.6	327.7	327.4	332.1	338.8	339.4
1979	340.1	341.3	342.5	342.4	342.9	342.0	333.6	327.7	329.3	333.1	338.1	341.5
1980	341.9	342.4	343.4	343.3	343.3	342.6	337.3	331.8	331.4	337.9	340.5	342.6
1981	344.3	345.5	344.6	345.8	346.8	344.4	338.2	331.8	333.3	338.8	342.2	344.9
1982	345.8	347.2	348.3	348.4	348.0	345.7	339.2	333.2	333.2	337.9	342.1	344.5
1983	347.5	347.0[b]	348.1	348.1	348.3[b]	347.0	340.4	333.8	335.5	340.6	346.5	346.8
						Mauna Loa Observatory						
1973	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1974	0.0	0.0	0.0	0.0	333.0	332.1	331.0	329.1	327.4	327.3	328.3	329.5
1975	330.7	331.5	332.1	333.3	333.8	333.9	331.8	330.0	328.5	328.3	329.3	330.5
1976	331.7	332.8	333.5	334.9	334.8	334.5	333.0	330.7	329.3	328.7	330.2	331.4
1977	332.0	332.3	334.9	336.2	336.9	336.1	335.0	332.6	331.3	331.3	332.3	333.6
1978	335.1	335.4	336.8	337.8	338.1	338.0	336.5	334.5	332.4	332.4	333.8	334.9
1979	336.3	336.8	338.3	338.9	339.3	339.1	337.6	335.5	333.8	334.2	335.3	336.8
1980	337.8	338.4	340.0	340.9	341.4	341.3	339.5	337.3	336.0	336.1	337.3	338.4
1981	339.2	340.0	341.7	342.4	343.1	342.5	340.7	338.5	337.1	337.3	338.7	339.9
1982	341.2	341.7	342.7	343.8	344.3	343.3	341.8	340.2	337.7	337.9	339.1	340.7
1983	341.4	342.7[b]	343.0	345.3	345.9	345.4	344.2	342.1	340.1	340.1	341.4	343.0
						Samoa Observatory						
1973	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1974	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1975	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1976	330.7[a]	330.6[a]	330.8[a]	331.2	330.7	331.2	331.1	331.1	331.3	331.5	331.9	331.8
1977	331.7	332.0	332.0	332.4	332.3	332.6[a]	332.6[a]	333.1	332.9	332.9[a]	333.1[a]	333.7[a]
1978	334.0[a]	334.2[b]	334.0[a]	334.3[a]	335.0[b]	335.0[b]	335.0[b]	334.3[a]	334.3[a]	334.5a	334.4[a]	334.2[a]
1979	334.6[a]	335.4[a]	335.9[a]	335.7	335.9	336.0	336.3	336.3	336.3	336.6	336.7	336.8
1980	337.5	337.6	337.9	337.3	337.8	338.0	338.4	338.3	338.2	338.1	338.5	338.9
1981	339.0	339.3	338.8	339.6	338.8	339.3	339.0	339.1	339.2	339.0	339.5	339.8
1982	340.4	340.5	340.4	339.8	339.9	340.1	340.5	340.4	340.1	340.0	340.2	340.2
1983	340.0[b]	340.5[b]	340.8[b]	340.6[b]	341.1[b]	341.6[b]	341.9[b]	341.9[b]	341.9[b]	342.1[b]	342.4[b]	342.6[b]

(Continued)

Table 3.2 (Continued)

Year	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
						South Pole Observatory						
1973	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1974	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1975	329.1[a]	328.8[a]	328.5[b]	328.3[b]	328.6[a]	328.6[b]	329.4[b]	329.8[a]	330.1[a]	330.3[a]	330.2[a]	330.3[b]
1976	330.1	329.8	329.6	329.6	329.7	329.9	330.3	330.7	331.0	331.1	331.1	331.1
1977	330.9	330.3	330.3	330.6	331.0	331.4	331.8	332.2	332.7	333.0	333.0	332.7
1978	332.6	332.4	332.3	332.5	333.0	333.3	333.8	333.4	334.9	335.0	334.9	334.7
1979	334.5[a]	334.5[a]	334.6[a]	334.6[a]	334.9[a]	335.4[a]	335.9[a]	336.2[a]	336.4[a]	336.5[a]	336.4[a]	336.2[a]
1980	336.1	336.0	336.0	336.2	336.6	336.9	337.4	337.7	337.9	338.1	338.0	338.0
1981	337.8	337.8	337.9	337.9	338.1	338.3	338.7	339.0	339.2	339.1	338.9	338.5
1982	338.5	338.6[a]	338.4	338.6	338.9	338.9	339.2	339.6	339.9	339.8	339.5	339.2
1983	339.7	339.5	339.5	340.0	340.4	340.5	341.0	341.6	341.7	341.8	341.8	341.7

Note: Values have been provisionally selected for background conditions and are expressed in the WMO X81 mole fraction scale (parts per million with respect to dry air).

[a]Substituted flask data.
[b]Less reliable or provisional data.

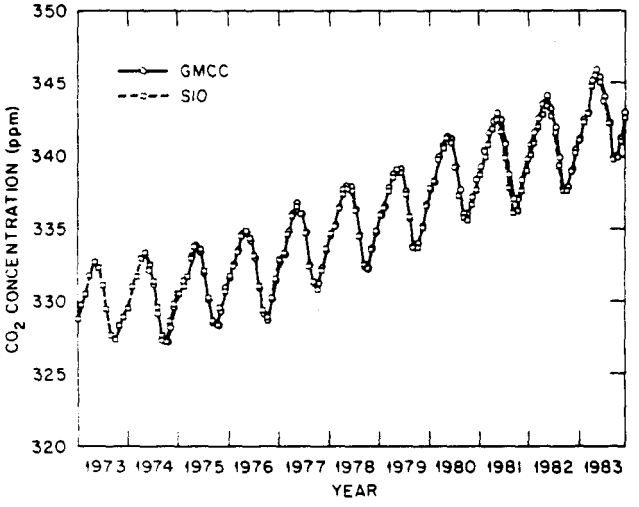


Figure 3.11. Comparison of the Keeling (SIO) and NOAA (GMCC) atmospheric CO_2 records at Mauna Loa Observatory for the common period since 1974. Values are edited monthly means from C. D. Keeling (SIO, personal communication, 1984) and from K. Thoning (NOAA/GMCC, personal communication, 1984); very similar to values given in Table 3.2).

is now nearly steady is a direct consequence of the abrupt decrease in the rate of global production and consumption of fossil fuels since 1973. (Fossil fuel growth rates have been less than 2%/year since 1974, with zero or negative growth rates from 1980 to 1982, as compared with the mean growth rate of 4.4%/year from 1950 to 1973, see Chapter 4.)

3.5.3 Global CO_2 Flask Sampling Network of NOAA/GMCC

In addition to the long time series of continuous CO_2 measurements made at the four GMCC observatories, flask samples of whole air are collected weekly at the approximately 20 cooperating sites of a global CO_2 network and returned to NOAA (Boulder, Colorado) for trace gas analysis (CO_2, CH_4, CO). Although the sampling began before 1970 at a few initial sites, the major expansion of the network to 15 stations took place in 1979 with support from the Department of Energy. This expanded flask network has allowed the synthesis of a detailed picture of the regional variation of atmospheric CO_2 concentration.

The distribution of cooperating flask sites around the globe (Figure 3.12; Table 3.3) clearly illustrates the interest of the GMCC program in selecting background marine locations and in avoiding the major land biomes. Previously reported details on each site in the flask network (Herbert 1980; DeLuisi 1981) have been updated by Komhyr et al. (1985). The most recent changes in the network have been the loss of the site in the Falkland Islands in February 1982 (replaced by Christchurch, New Zealand), the start-up of Cape Meares station on the Oregon coast, and the cooperative program with Battelle Laboratories at Kitt Peak, Arizona.

The entire flask data set (1968-1982) (corrected, converted to the WMO X81 mole fraction scale, and edited for background conditions) has been presented by Komhyr et al. (1985), and is available from the DOE Carbon Dioxide Information Center in Oak Ridge, Tennessee. For both the flask and continuous

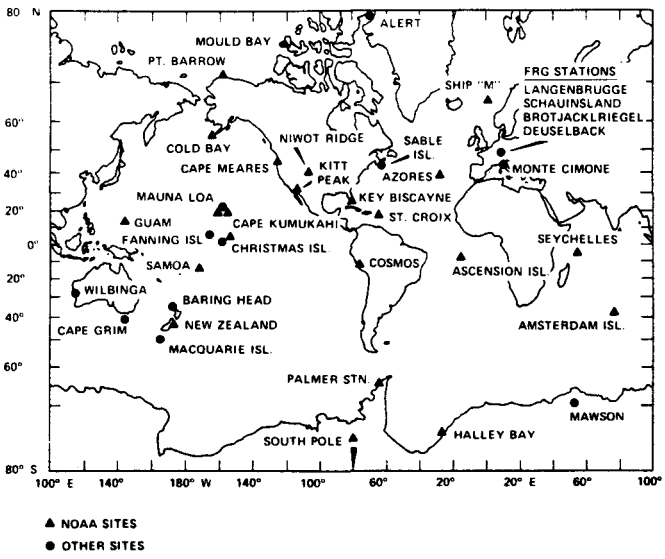


Figure 3.12. Geographic location of the cooperating sites of the NOAA/GMCC global CO_2 flask sampling network active in 1983. Refer to Table 3.3 for the coordinates and elevation of each sampling site (Harris and Bodhaine 1983; Komhyr et al. 1985).

CO_2 measurement programs, GMCC uses CO_2-in-air gas standards directly traceable to the WMO standards maintained by Keeling at Scripps Institution of Oceanography (SIO). These standards are now being converged with a set of reference materials developed by the U.S. National Bureau of Standards for future use.

As one check on the quality of the flask CO_2 measurements and on the reliability of the in situ CO_2 analyzers at the four GMCC baseline stations, the annual mean concentrations obtained at BRW, MLO, SMO, and SPO from the flask and continuous programs may be compared for the period of common record (1976-1983). The difference in the annual mean values is generally less than 0.2 ppm at each station (Table 3.4). As a second check on the flask results, there was initiated in mid-1981 a flask-flask comparison with Keeling's laboratory at Scripps Institution of Oceanography (SIO) for quasi-simultaneously collected flasks at BRW, MLO, Cape Kumukahi, Hawaii (KUM), and SMO. The agreement has improved steadily, and the flask-flask differences (NOAA data minus SIO data) now generally fall within the sampling and analysis error of ± 0.2 ppm. The annual mean CO_2 concentrations from the GMCC flask network are given in Table 3.5.

With the expansion of the flask network in 1979, the improved latitude coverage permitted the interpolation of the time series at the cooperating sites to create a CO_2 concentration surface representing zonally averaged global distribution (Figure 3.13). The grid points correspond to a latitude resolution of 10° and a time resolution of 20 d. Sub-grid-scale features of the global CO_2 distribution (e.g., local gradients associated with equatorial ocean upwelling or the intertropical convergence zone [ITCZ]) will not be well represented on this surface. Information about longitudinal gradients (perhaps important at the equator and at midnorthern latitudes) is suppressed in the zonal averaging by latitude bands. To create these CO_2 surfaces, the flask data were first edited for background conditions. The time series of background values at each station were then fitted to a cubic spline. The set of biweekly values of the spline function at

Table 3.3
The Cooperating Sampling Sites of the NOAA/GMCC Global CO_2 Flask Network in 1983

Code	Site	Latitude	Longitude	Elevation (m)
MBC	Mould Bay, Canada	76°N	119°W	15
BRW	Pt. Barrow, Alaska	71°N	157°W	11
STM	Ocean Station M	66°N	2°E	6
CBA	Cold Bay, Alaska	55°N	163°W	25
CMO	Cape Meares, Oregon	45°N	124°W	30
NWR	Niwot Ridge, Colorado	40°N	106°W	3749
AZR	Terceira Island, Azores	39°N	27°W	30
KPA	Kitt Peak, Arizona	32°N	112°W	2095
KEY	Key Biscayne, Florida	26°N	80°W	3
MLO	Mauna Loa, Hawaii	20°N	156°W	3397
KUM	Cape Kumukahi, Hawaii	20°N	155°W	3
AVI	St. Croix, Virgin Islands	18°N	65°W	3
GMI	Marianas Island, Guam	14°N	145°E	2
SEY	Mahe', Seychelles	5°S	55°E	3
ASC	Ascension Island	8°S	14°W	54
COS	Cosmos, Peru	12°S	75°W	4600
SMO	American Samoa	14°S	171°W	30
AMS	Amsterdam Island	38°S	78°E	150
NZL	Kiatorete Spit, New Zealand	44°S	173°E	3
PSA	Palmer Station, Antarctica	65°S	64°W	10
HBA	Halley Bay, Antarctica	75°S	27°W	10
SPO	South Pole, Antarctica	90°S	25°W	2810

Table 3.4
Differences in Annual Mean CO_2 Concentration Determined From Comparisons of Flask (f) and Continuous (c) Values From the NOAA and SIO Measurement Programs

	NOAA(f) - NOAA(c)				NOAA - SIO	
Year	BRW	MLO	SMO	SPO	(c-c) MLO	(f-f) SPO
1974	(+0.7)[a]	—[b]	—	—	−0.3	—
1975	(−0.6)[a]	—	—	0.0	0.0	+0.1
1976	+0.1	0.0	+0.3	−0.1	−0.1	−0.2
1977	−0.2	−0.1	0.0	−0.2	0.0	−0.3
1978	−0.2	−0.3	—	−0.1	+0.2	+0.1
1979	−0.3	0.0	−0.1	—	+0.2	+0.2
1980	+0.3	+0.3	+0.1	−0.4	+0.2	−0.1
1981	−0.3	+0.3	0.0	+0.1	+0.6	+0.4
1982	−0.2	−0.2	+0.2	+0.2	+0.4	+0.2
1983[a]	(+0.2)	(−0.3)	(0.0)	(+0.1)	(−0.1)	—
Mean Signed	−0.11	0.00	+0.08	−0.07	+0.13	+0.05
Absolute Value	0.23	0.23	0.15	0.20	0.27	0.23

[a]Provisional data, subject to change.
[b]— = missing data.

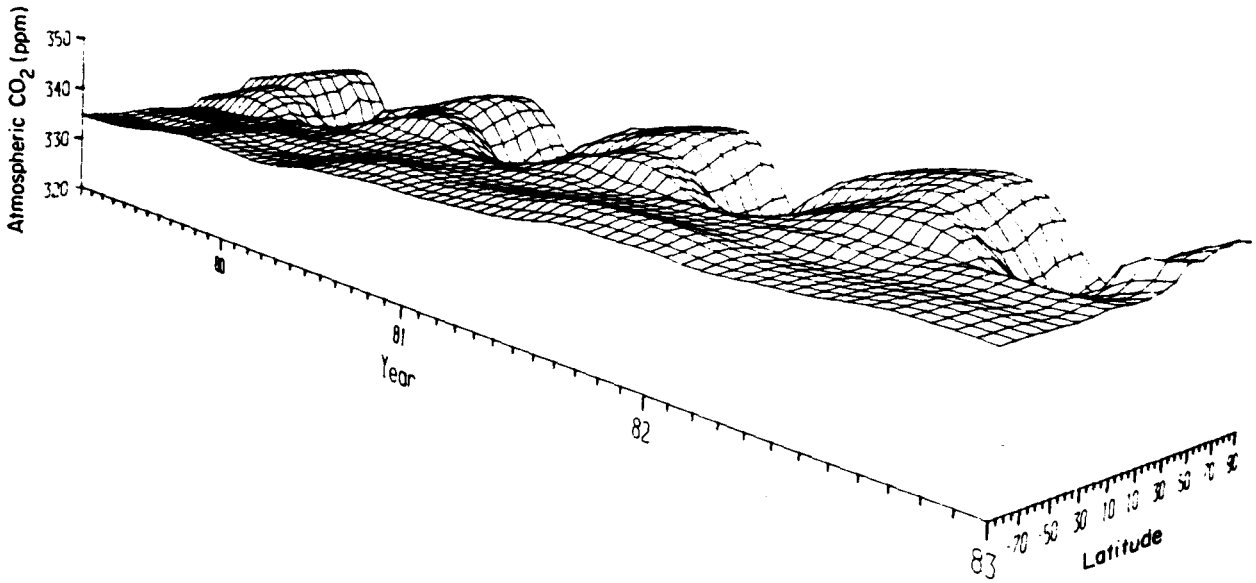


Figure 3.13. A three-dimensional perspective of the "pulse-of-the-planet," the variation of the global atmospheric CO_2 concentration in latitude and time based on flask measurements for 1979-1982. This zonally averaged surface has a resolution of 20 d and 10° in latitude and was synthesized from results of ~10,000 individual flask samples returned from the 15 remote sea-level sites shown in Figure 3.12 (Harris and Bodhaine 1983).

each station then served as the input to the computer graphics program which interpolated the concentration surface. Four years of flask data (1979-1982) were fit simultaneously, representing 10,000 separate flask measurements. Except for the South Pole, only island or coastal stations near sea level were included. This means that the seasonal amplitudes represent typical marine locations and as such are certainly minimum values. Much larger seasonal variations in CO_2 concentration are expected for midcontinental locations, particularly in Siberia and northern Canada (Fung et al. 1983; see also Figure 3.14).

The global distribution of the flask sites of the NOAA/GMCC network (Figure 3.12) permits a precise value for the global average atmospheric CO_2 concentration to be determined (Table 3.6). These annual global means are derived from latitude-weighted, sea-level sites, and represent the well-mixed troposphere (Komhyr et al. 1985). An approximation to a global mean, for the years prior to the network expansion in 1979, is provided by the simple average of the annual CO_2 means from the continuous measurements at the four GMCC stations (BRW, MLO, SMO, SPO).

Table 3.5
Annual Mean CO_2 Concentrations From Weekly Sampling at Cooperating Sites of the NOAA/GMCC Global CO_2 Network

STATION[a]	1968	1969	1970	1971	1972	1973	1974	1975
AMS	0	0	0	0	0	0	0	0
ASC	0	0	0	0	0	0	0	0
AVI	0	0	0	0	0	0	0	0
AZR	0	0	0	0	0	0	0	0
BRW	0	0	0	(327.3)	330.1	331.9	333.3	332.8
CBA	0	0	0	0	0	0	0	0
CMO	0	0	0	0	0	0	0	0
FLK	0	0	0	0	0	0	0	0
GMI	0	0	0	0	0	0	0	0
KEY	0	0	0	0	0	(330.7)	0	0
KUM	0	0	0	0	0	0	0	0
MBC	0	0	0	0	0	0	0	0
MKO	0	0	0	0	0	0	0	0
MLO	0		(325.2)	0	0	0	0	0
NWR	(323.1)	324.1	325.6	325.4	326.2	(330.2)	0	0
PSA	0	0	0	0	0	0	0	0
SEY	0	0	0	0	0	0	0	0
SMO	0	0	0	0	0	0	(330.9)	(330.7)
SPO	0	0	0	0	0	0	0	(329.4)
STC	0	324.0	326.4	326.8	(327.5)	0	0	0
STM	0	0	0	0	0	0	0	0

STATION	1976	1977	1978	1979	1980	1981	1982
AMS	0	0	0	(335.9)	337.8	339.3	339.4
ASC	0	0	0	0	(338.8)	339.8	340.7
AVI	0	0	0	(337.1)	339.6	340.4	340.9
AZR	0	0	0	0	(338.4)	339.6	341.1
BRW	333.8	334.8	336.5	337.6	340.2	341.4	342.6
CBA	0	0	0	337.7	339.7	341.2	341.8
CMO	0	0	0	0	0	0	341.2
COS	0	0	0	0	(338.7)	0	339.7
FLK	0	0	0	0	0	(339.7)	0
GMI	0	0	0	(337.7)	340.1	341.2	341.0
KEY	(332.7)	335.2	336.7	338.8	(340.1)	342.0	341.5
KUM	(332.3)	334.4	335.7	337.4	339.3	340.4	341.2
MBC	0	0	0	0	(340.3)	341.8	342.4
MKO	0	0	0	337.3	0	0	0
MLO	(332.1)	333.6	335.2	336.8	339.0	340.4	341.0
NWR	332.1	334.8	335.7	336.9	338.0	339.8	341.0
PSA	0	0	(333.9)	335.2	(337.5)	340.1	339.8
SEY	0	0	0	0	339.3	340.1	340.5
SMO	331.5	332.8	334.3	(336.0)	(338.1)	339.2	340.4
SPO	330.2	331.5	333.6	335.0	336.7	338.5	339.3
STC	0	0	0	0	0	0	0
STM	0	0	0	0	0	(341.8)	341.6

Note: Values in parts per million relative to dry air, expressed in the WMO X81 mole fraction scale. Values in parentheses represent less certain values and generally contain several months of missing data for which interpolated values have been used to estimate the annual mean (Komhyr et al. 1985).

[a]FLK = Falkland Islands; MKO = Mauna Kea, Hawaii; STC = Weathership Charlie (see Table 3.3 for definition of additional station acronyms) (See Komhyr et al. 1985.)

An important assumption often made by carbon cycle modelers is that the MLO record is a good approximation to the global CO_2 level for calculating atmospheric retention of anthropogenic CO_2, at least in the annual mean (Chapters 7 and 10). Data from Mauna Loa typically fall 0.2-0.3 ppm above the global annual CO_2 mean (Table 3.6), as determined from either the flask network or the four continuous NOAA stations. With steeper latitude gradients developing in the future from higher global fossil fuel usage, Mauna Loa will diverge from the global mean. For the present, modelers should use concentrations for MLO or MLO minus 0.2 ppm as being globally representative, and should *not* use the average of MLO and the South Pole, which lies well below the global mean CO_2 level (see analysis of airborne fraction in Section 10.2.7 of Chapter 10).

3.5.4 CSIRO (Australia) CO_2 Measurements From Aircraft

Since 1972, CSIRO has conducted a CO_2 observational program based on the analysis of discrete air samples

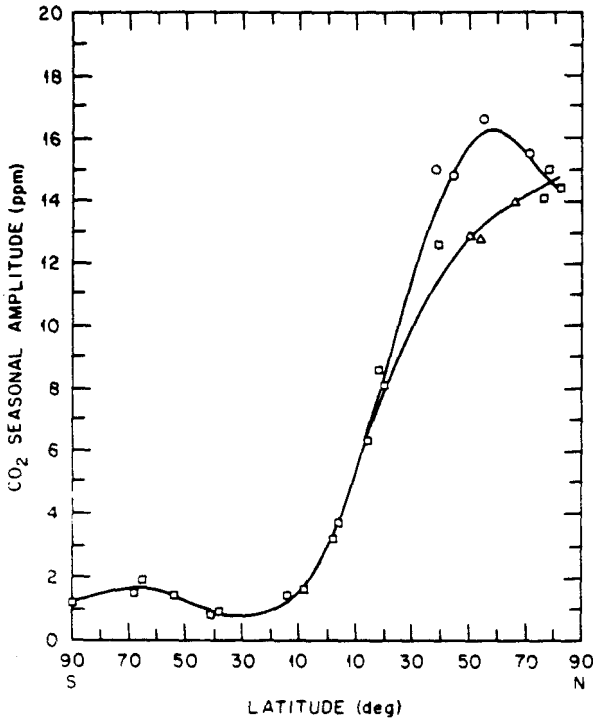


Figure 3.14. The latitude dependence of the seasonal amplitude of atmospheric CO_2 concentration observed at clean-air sites of NOAA GMCC (□) (Harris and Bodhaine 1983; Komhyr et al. 1985; see Figure 3.12), Ny Alesund (78°N) (W. Bischof, personal communication 1984), Canadian flask stations Alert (82°N), Weathership "P" (50°N), University of Stockholm, and Sable Island (44°N) (Wong et al. 1984); coastal Japan (~39°N) (Tanaka et al. 1983); Scripps Institution of Oceanography flask stations at Fanning Is. (4°N) and Christmas Island (2°N) (Keeling et al. 1984); and CSIRO (Australia) values from Cape Grim (4°S), Mawson (54°S), and Maguarie (68°S) (Fraser et al. 1983). Smooth polynomial fits to the data split at high northern latitude to indicate that midocean stations (Δ) (Weatherships, P, C, M) show lower CO_2 seasonal amplitudes than do continental and coastal sites (O) (Cape Meares, Oregon; Cold Bay, Alaska; Pt. Barrow, Alaska).

collected from aircraft throughout the troposphere and lower stratosphere over the Australian-New Zealand region and as far south as Antarctica. The CSIRO program complements the aircraft measurements begun in the 1960s in the Northern Hemisphere by Swedish researchers (Bolin and Bischof 1970; Bischof 1971). Comprehensive descriptions of the sampling network (Figure 3.15A) and the procedures for sampling and analysis have been published (Beardsmore et al. 1978; Pearman et al. 1983a; Pearman and Beardsmore 1984). The data are particularly important because, unlike those obtained from surface observations, they are unselected and, therefore, free of potential biases from selection procedures.

From data collected in the midtroposphere up to 1981 (Figure 3.15B), the CO_2 concentration increased, on average, by 1.3 ppm/year. The data exhibit an average annual cycle of 1.2 ppm, peak to peak, with a spring maximum and a fall minimum. The rate of increase does vary markedly from year to year between 0.4 and 1.9 ppm/year. There is evidence of a 2- to 4-year quasi-periodicity in the rate of CO_2 increase which agrees with observations from Mauna Loa and the South Pole. The midtropospheric data indicate an apparent fossil fuel airborne fraction (see definition in Section 10.2.7) of 0.57 during the period January 1973 to January 1981, compared with values of 0.59 and 0.52 from Mauna Loa and the South Pole, respectively,

Table 3.6
Global Annual Mean Tropospheric CO_2 Concentration Compared with Mauna Loa Annual Mean From NOAA/GMCC Continuous Records

		Global Mean		Difference MLO minus Global	
Year	MLO[a]	Flask[b]	Continuous[c]	Flask	Continuous
1976	332.1	—[d]	331.8	—[d]	+0.3
1977	333.7	—	333.2	—	+0.5
1978	335.5	—	335.1	—	+0.4
1979	336.8	336.6	336.6	+0.2	+0.2
1980	338.7	338.7	338.4	0.0	+0.3
1981	340.1	340.0	339.9	+0.1	+0.2
1982	341.2	340.6	340.8	+0.6	+0.4
1983	342.6	342.3	342.1	+0.3	+0.5
			Signed	+0.24	+0.34
1976-1983		Mean Difference			
			Absolute Value	0.23	0.12

[a]MLO = Mauna Loa annual means from NOAA/GMCC continuous data.
[b]Flask = global annual mean CO_2 concentration from sea-level sites of the NOAA flask network, latitude weighted (Komhyr et al. 1985).
[c]Continuous = global mean annual CO_2 from average of annual means from continuous data at BRW, MLO, SMO, and SPO (Harris and Bodhaine 1983).
[d]— = missing data.

during the 1960s and 1970s. An analysis of the monthly mean variability of midtropospheric CO_2 data shows that approximately one-half is due to synoptic features such as air mass trajectory coupled with the stability of the semipermanent subsidence inversion.

Upper tropospheric data exhibit an annual cycle larger in amplitude (2-3 ppm, peak to peak) than that observed in the midtroposphere, and the phase at higher altitude precedes that in the midtroposphere by approximately 1 month. This is probably due to the high-altitude meridional transport of Northern Hemisphere air into the Southern Hemisphere. The variability of upper tropospheric data is largely related to the location of the tropopause.

Vertical gradients through the mid and upper troposphere are generally of the order of 0.4 ppm/10 km, with lower concentrations at lower altitudes throughout most of the year. In the lower troposphere, below 2 km, gradients of up to 4 ppm/km can occur when air has traversed vegetated land.

The CSIRO aircraft-derived CO_2 data form the basis for ongoing studies that include (1) the determination of the relative importance of atmospheric transport processes and oceanic or biospheric exchange in the seasonal variation of Southern Hemispheric CO_2 data; (2) by comparison with other data, a description of the meridional concentration distribution which can be used to make qualitative estimates of meridional fluxes and surface exchanges; (3) the understanding of the variability of CO_2 on all time scales from which rational data selection criteria for surface monitoring stations can be developed.

3.5.5 CSIRO (Australia) Stable Carbon Isotope Measurement Program

Variation in the $^{13}C/^{12}C$ ratio can be used to distinguish between large-scale biospheric and oceanic CO_2 fluxes. For example, Pearman and Hyson (in press) quote a change in $\delta^{13}C$ of 0.052‰ for a 1-ppm change in CO_2 concentration for biospheric exchange, and about a 0.005‰ for a 1-ppm change for oceanic exchange. Keeling et al. (1979) reported a 0.03‰ per year change in $\delta^{13}C$ resulting from fossil fuel combustion, and more recently Mook et al. (1983), Keeling et al. (1984), and Francey (1986) demonstrated the influence of oceanic exchange on the isotope signal resulting from seasonal biospheric activity. At a site like Cape Grim, Tasmania (41°S, 145°E), the seasonal variation in CO_2 concentration is likely to contain contributions from the biosphere and the ocean (Pearman and Hyson, in press). A prime objective is the use of Cape Grim carbon isotope data to distinguish between these two components of the carbon cycle.

Since 1977, CSIRO has conducted a stable carbon isotope measurement program at Cape Grim, Tasmania involving in situ extraction of CO_2 from Cape Grim air. Descriptions of the sampling and measurement techniques employed have been given elsewhere (Goodman 1980; Baseline 1983, 1984; Francey 1986, in press). Measurements of mass ratios of carbon and oxygen isotopes are made on the samples and recorded as fractional differences ($\delta^{13}C$ and $\delta^{18}O$, respectively) relative to a CSIRO secondary CO_2 isotope standard which, in turn, has been calibrated against international standards.

Measurements of $\delta^{13}C$ at Cape Grim between October 1977 and November 1981 are characterized by a large scatter and a strong correlation between $\delta^{13}C$ and $\delta^{18}O$. This correlation has been used to correct $\delta^{13}C$ measurements to a constant $\delta^{18}O$ value for the pre-1981 data (Goodman 1980) under base-line conditions, and the resulting data show a $\delta^{13}C$ trend of $-0.03(\pm 0.01)$‰ per year, similar to that reported more recently by Keeling et al. (1980). The need for this correction disappeared in 1982, following several changes in technique and the maintenance of better quality control during sample collection (Francey 1986). The data since 1982 show a $\delta^{13}C$ trend of $-0.025(\pm 0.005)$‰ per year (Figure 3.16). A significant seasonal variation is present, despite considerable interannual variability possibly associated with the 1982/83 ENSO event. The average peak-to-peak amplitude is $0.055(\pm 0.014)$‰, with a maximum in late March, closely related to that in CO_2 concentration (mid-March). The amplitude of the seasonal variation of $\delta^{13}C$ at Cape Grim exceeds that at the South Pole by about a factor of 2. The phase is also different, with the South Pole lagging by approximately 3 months. Comparison of the annual cycles in $\delta^{13}C$ at Cape Grim and the South Pole with a recent two-dimensional carbon cycle model (Pearman and Hyson, in press) suggests that the model contains excessive biospheric exchange at high southern latitudes (Francey, in press).

Since late 1983, a new stable carbon isotope measurement program was initiated by CSIRO, using pressurized air samples collected in pairs of 5-L glass flasks. The air samples are collected at the NOAA/GMCC sites of Pt. Barrow, Alaska; Mauna Loa, Hawaii; and Cape Matatula, American Samoa; and at the South Pole, at Cape Grim, and in the midtroposphere over Southeast Australia using the same sampling procedure. The samples are returned to Aspendale (CSIRO) for subsequent $\delta^{13}C$ analysis. Preliminary results from the first 12 to 18 months of data collection are expected by late 1985.

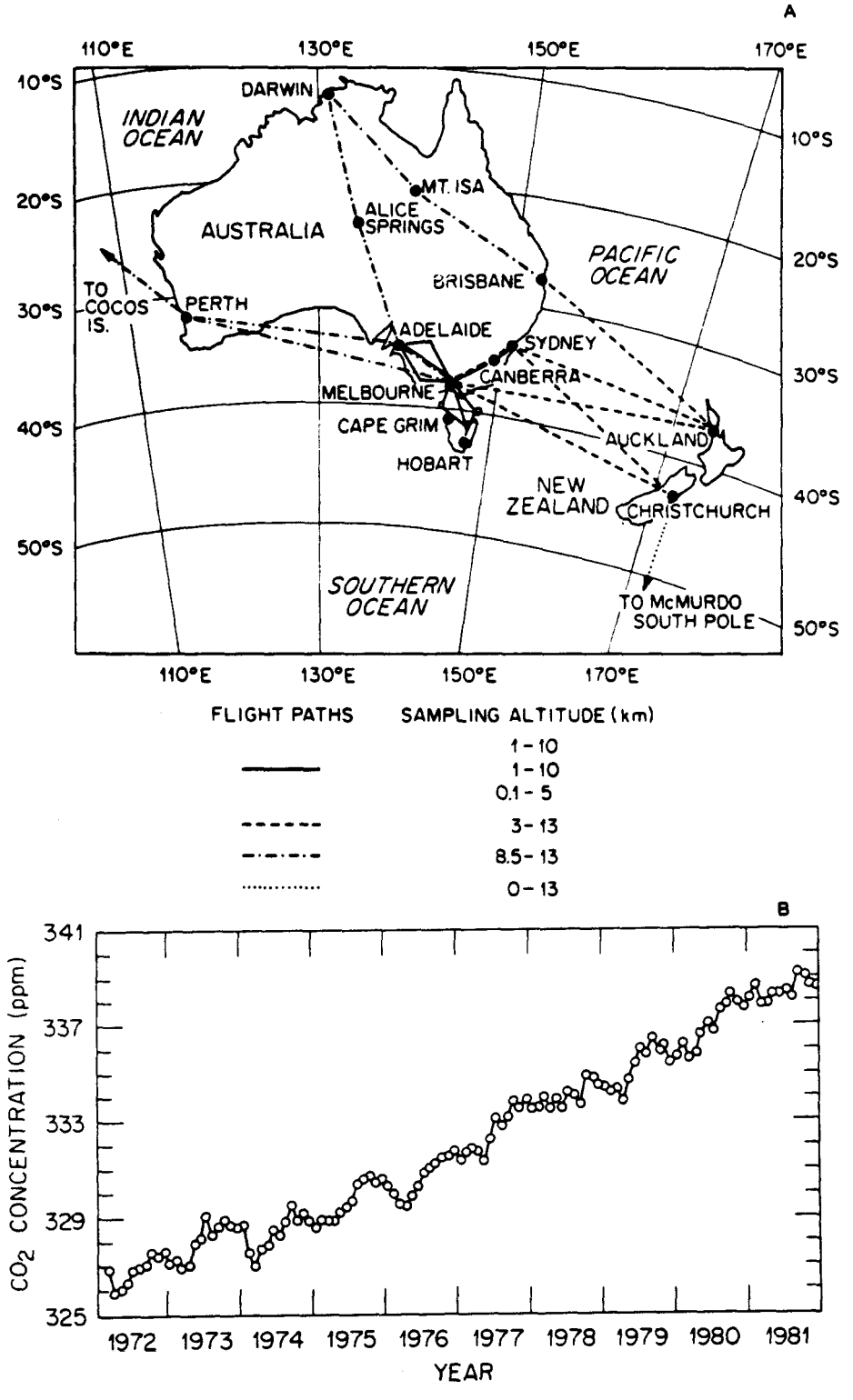


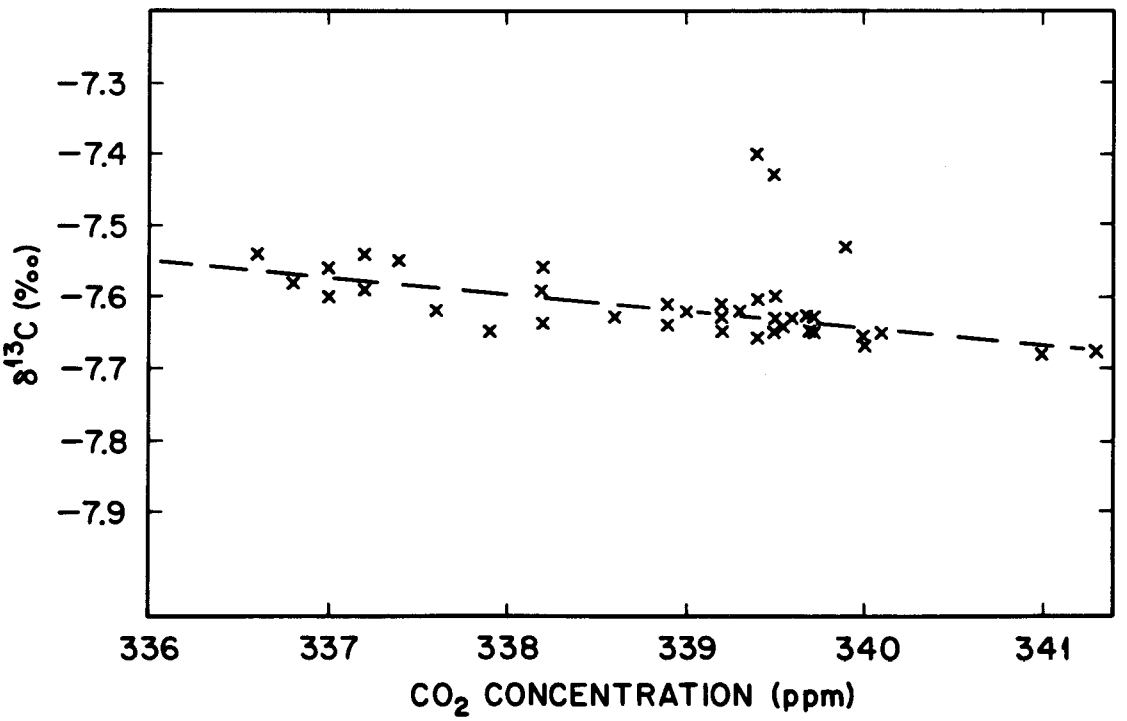
Figure 3.15. (A) CSIRO aircraft monitoring network showing details of flight paths, aircraft, and cooperating agencies. (B) Monthly mean CO_2 concentration variation in the midtroposphere over southeastern Australia as determined by the CSIRO air sampling program.

3.5.6 Potential Changes in Growth Rate of Atmospheric CO_2 at Mauna Loa Since 1973

The observed increase of atmospheric CO_2 at Mauna Loa, Hawaii, during the past quarter century may be compared with a theoretical atmospheric CO_2 increase scenario in which the 1973 global oil crisis does not occur and exponential growth of fossil fuel CO_2 production is instead allowed to continue at the 4.4%/year mean rate characteristic of the previous 25 years (see Chapter 4). The resulting difference (model minus observed) in atmospheric CO_2 reaches +4 ppm by 1983 (Figure 3.17). This difference lies well outside any possible uncertainties in the observed atmospheric levels, in the fossil fuel data, or in natural fluctuations in the source-sink relationships of the ocean and biosphere (see Section 10.2.7). The departure of the observed CO_2 atmospheric increase from the faster model increase has only become significant in recent years, with the most dramatic slowdown corresponding to the several years of global economic stagnation and declining fossil fuel usage beginning in 1979 (see Figure 4.2). The fact that the rate of the observed CO_2 increase closely followed the changing fossil fuel production rate during the last decade strongly suggests that this fossil fuel source dominates the atmospheric CO_2 rise, and that any changes in the source-sink properties of the biosphere and ocean in the global carbon cycle have been relatively minor during this period. The implication is that future atmospheric CO_2 increases will continue to reflect closely the actual fossil fuel production rates in the coming decades (also see uncertainty analysis in Section 10.2.7).

Small interannual departures from the quasi-exponential rate of increase are apparent in the atmospheric CO_2 record, and have been interpreted in terms of fluctuation in the natural carbon cycle associated with sea surface temperature (SST) anomalies and ENSO events (Bacastow 1976; Newell et al. 1978; Bacastow et al. 1980; Gammon and Komhyr 1983). Such ENSO-induced fluctuations represent fine structure, a natural noise level with fluctuations of the order of ±1 ppm persisting over 1 to 3 years, superimposed on the steadier manmade rise in atmospheric CO_2 due primarily to combustion of fossil carbon, which has averaged 1.4 ppm/year over the last decade. The major ENSO events of 1969/70, 1972/73, 1976/77, and especially 1982/83 are all apparent in the longer atmospheric CO_2 records. Together, these two source terms (i.e., fossil fuel and ENSO fluctuations) can account for all major (>0.5 ppm) features in the progression of annual mean CO_2 values at Mauna Loa and the South Pole over the full period of record since 1958.

Figure 3.16. Atmospheric $\delta^{13}C$ (uncorrected for N_2O) versus CO_2 concentration for baseline conditions (oceanic trajectories) at Cape Grim. Reprinted by permission of Springer-Verlag, R. J. Francey (1986).



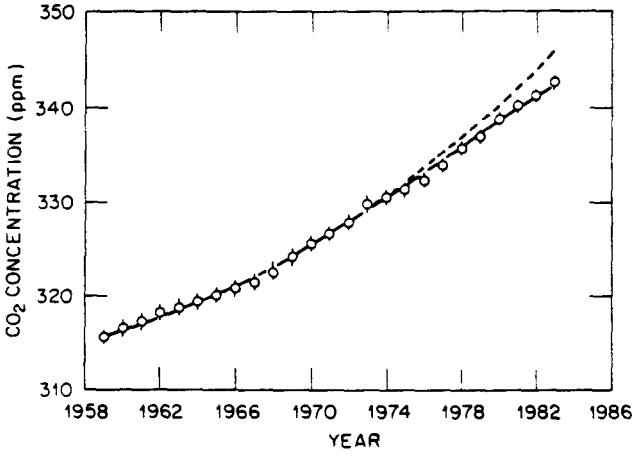


Figure 3.17. Global atmospheric CO_2 (solid line): Least squares fit of the Mauna Loa (MLO) annual CO_2 values (†) (from Keeling et al. 1982 until 1975; Harris and Bodhaine 1983) to a linear discrete-time perturbation model with fossil fuel production data (1860-1983) of Marland and Rotty (1984) and R. Rotty (Oak Ridge Associated Universities, Institute for Energy Analysis, personal communication, 1984). Fitted values: 1860 atmospheric CO_2 level, 296 ppm; effective tropospheric lifetime of fossil fuel CO_2, $\tau = 40$ years. The vertical bars (|) represent maximum possible errors in the atmospheric data which could affect the analysis; (dashed line) represents projection of simulated high-growth rate (4.4 %/year) of fossil fuel production since 1974, with other conditions held constant. The difference between the 1983 simulated and MLO values was 3.6 ppm.

3.5.7 ENSO Events as Recorded in the Global Atmospheric CO_2 Distribution

Background measurements of atmospheric CO_2 concentration represent a neglected resource in the study of the link between the tropical ocean and the global atmosphere. The CO_2 records extend over decades; are very precise ($\sim \pm 0.1\%$); and offer a spatially averaged history of the seasonal and interannual variability of the global atmospheric circulation, the regional air-sea exchange, and the integrated activity of the biosphere.

The equatorial ocean is a major source of CO_2 to the atmosphere. Under natural, steady-state conditions, this flux is returned from the atmosphere to the cooler surface waters of the world's oceans at higher latitudes in both hemispheres. This flux is not small, rivaling the current fossil fuel release of $\sim 5 \times 10^{15}$ g/year (Keeling 1983). Obviously, variations in the rate of either the equatorial upwelling source of CO_2 or the higher latitude ocean sink can significantly alter the observed atmospheric CO_2 level.

The ENSO perturbation of the global atmospheric CO_2 distribution is the strongest signal in the record after the local seasonal amplitude and fossil fuel trends are removed. These CO_2 fluctuations are well correlated with both the Southern Oscillation Index (SOI) and regional SST anomalies. The extraordinary 1982/83 ENSO event measurably perturbed the total global atmospheric burden of CO_2 as well as its detailed distribution by latitude, longitude, and altitude. The response of the global carbon cycle to this extreme circulation anomaly offers a unique opportunity to assess tropical-extratropical atmospheric linkages.

3.5.7.1 *Correlations Between CO_2 Variations and the Southern Oscillation Index*

After seasonal and fossil fuel trends are removed, the dominant residual signal in all time series of atmospheric CO_2 from background monitoring stations around the globe is an interannual fluctuation inversely correlated with the SOI. This was first shown by Bacastow (1976) in the CO_2 records for Mauna Loa, Hawaii (MLO), and the South Pole (SPO). The amplitude of the natural CO_2 fluctuation associated with an ENSO event is of the order of ± 0.5 ppm; for comparison, background air is currently $\sim$342 ppm CO_2 in the global mean in 1983. Bacastow found the best correlation to be between the SOI and the inverted CO_2 derivative of the MLO and SPO records after seasonal amplitudes and fossil fuel effects were removed from the time series; the CO_2 changes were found to follow the SOI by two (MLO) or three (SPO) seasons.

Natural atmospheric CO_2 fluctuations of the order of 1 ppm (or $\sim 2 \times 10^{15}$ g C global atmospheric equivalent) correspond to small perturbations (1-2%) in the annual exchanges between the atmosphere and the ocean (100×10^{15} g C/year) and between the atmosphere and the biosphere (60×10^{15} g C/year) (Keeling 1983). Bacastow realized that fluctuations in the source-sink balance of either the biosphere or the ocean could be responsible for the observed atmospheric CO_2 fluctuations; however, he concluded that variations in the strength of the ocean CO_2 sink at extratropical latitudes (40°-50° S) in response to variable westerly winds could well explain the observed correlated variability of SOI and CO_2 growth rate (Bacastow 1976, 1977).

Subsequent analyses by Bacastow and coworkers extended the original approach to other stations at both tropical (Fanning Island, MLO) and extratropical (Station P, SPO) latitudes, and confirmed the SOI-CO_2 signal in the major ENSO events of 1965, 1969, 1972, and 1976 (Bacastow 1977, 1979; Bacastow et al. 1980). The relative importance of anomalies in wind stress versus SST in perturbing atmospheric CO_2 levels remains an open question, as does the complication of volcanic cooling (the eruption of Agung in 1963 [as discussed by Bacastow 1979] or El Chichon in 1982).

3.5.7.2 *Correlations Between Atmospheric CO_2 Variations and SST Anomalies*

Although Bacastow established the correlation of CO_2 anomalies with SOI, others sought more direct links with sea surface properties in known source regions. Using empirical orthogonal function (EOF) analysis, Newell and colleagues demonstrated that the major contribution to natural variations in atmospheric CO_2 concentration is related in some way to changes in SST in the equatorial Pacific and in the upwelling region off the Peruvian coast (Weare et al. 1976; Newell and Weare 1977; Newell et al. 1978). The importance of the ENSO source region (0°-10°S) was independently confirmed in a two-dimensional, atmospheric tracer model of CO_2 variability by Machta et al. (1977).

Less is known about the possible relationship between SST and CO_2 fluctuations in extratropical latitudes. Hanson et al. (1981) established that anomalously low CO_2 levels at Weathership "P" (50°N, 145°W) were well correlated with large negative SST anomalies in the northwest Pacific (40°-45°N, 160°-180°E) during the ENSO years 1976/77. Schnell et al. (1981) found correlations between Pacific Ocean temperatures and CO_2 levels at Mauna Loa, Hawaii, and at Pt. Barrow, Alaska, and derived an empirical relationship between anomalies in SST and CO_2 of $\sim$0.6 ppm/°C.

After correcting for the fossil fuel CO_2 increase, there is a striking correlation between the year-to-year change in atmospheric CO_2 levels recorded at Mauna Loa and the year-to-year change in the mean air temperature in the northern subtropical zone (10°-30°N) (Figure 3.18). This is particularly apparent in the record since 1969, a period of stronger and more frequent ENSO events (e.g., 1969, 1972, 1976, 1979, 1982). The empirical relationship between interannual temperature change and observed CO_2 change is $\sim$0.8 ppm/°C, similar to that observed by Schnell et al. (1981) and deduced earlier by Bacastow (1979) on theoretical grounds. The observed relationship between zonal temperature change and CO_2 change does not permit the identification of the source (soil respiration, surface ocean?). However, this temperature dependence may be a local consequence of the reported correlation of east equatorial Pacific positive SST anomalies with subsequent atmospheric warming in higher latitudes (Angell and Korshover 1983). Even the month-to-month CO_2 growth anomalies observed at Mauna Loa appear to be correlated with changes in SST a few seasons earlier in the ENSO area (0°-10°S) of the eastern equatorial Pacific.[1]

[1]W. Elliott and J. K. Angell, NOAA, personal communication, 1984.

As a related but more oceanographic carbon cycle record of ENSO events over the last several decades, the weakened upwelling and warmer surface temperatures in the eastern equatorial Pacific during ENSO events are recorded in isotopic variations in banded corals (Druffel 1985). As discussed by Bacastow et al. (1980) and Druffel (1985), the detailed interpretation of atmospheric CO_2 anomalies in terms of changes in oceanic circulation and marine productivity remains problematical. For example, the reduced trade winds during ENSO imply reduced upwelling of CO_2 super-saturated deeper water, hence a weaker ocean source of atmospheric CO_2, while the simultaneous failure of the marine productivity in the upwelling regions implies a less effective ocean sink for carbon. Lower winds imply less efficient air-sea gas exchange in oceanic regions in which either sources or sinks are operating, and warmer surface waters can accommodate less dissolved CO_2 in saturation equilibrium than can colder waters. Models of air-sea exchange suggest that the role of changing wind stress and SST anomalies in extratropical latitudes may be more important in controlling global atmospheric CO_2 levels than previously considered (Komhyr et al. 1983).

3.5.7.3 *The 1982/83 ENSO Event*

From the established correlation of the SOI with anomalies in the CO_2 growth rate (Bacastow 1976; Bacastow et al. 1980), it was expected that the extreme ENSO event which began in mid-1982 would measurably perturb the global atmospheric CO_2 distribution. The expected CO_2 anomaly was first reported in analyses of the NOAA/GMCC global CO_2 records by Gammon and Komhyr (1983). They documented an anomalously low observed rate of CO_2 increase in mid-1982 and predicted a post-ENSO maximum in the CO_2 year-to-year increase for late 1983 or early 1984 of 2-3 ppm/year.

The response of the global atmospheric CO_2 distribution. This prediction was confirmed in the latest results from the NOAA/GMCC network (Figure 3.19). The globally averaged year-to-year change in atmospheric CO_2 plotted by season (Figure 3.19) clearly shows the effect of the 1982/83 ENSO event on the global atmospheric burden of CO_2 over at least a 2-year period, from mid-1982 to mid-1984. Comparison of the atmospheric CO_2 concentrations in the fall of 1982 and the fall of the previous year showed no increase (0.0 ± 0.3 ppm/year) in global CO_2, which means that the equivalent of approximately 5×10^{15} g of fossil carbon burned during that interval was removed from the atmospheric reservoir for storage (temporary) in the near surface ocean, and (probably) in the land biosphere as well. Now, more than a year after the minimum in the SOI in January of 1983, this missing 5×10^{15}

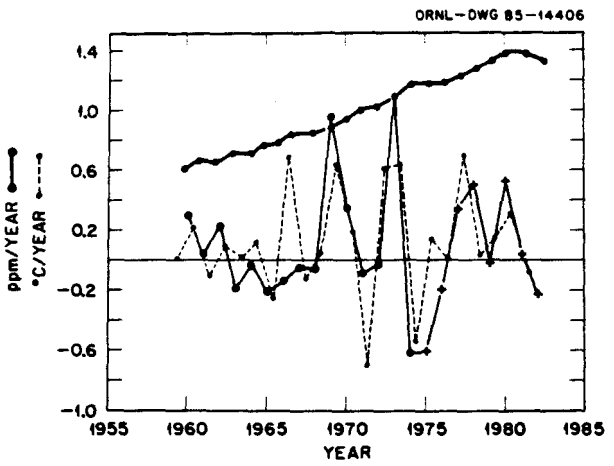


Figure 3.18. Correlation of year-to-year variation in annual mean CO_2 at Mauna Loa, Hawaii (solid line), after correction for fossil fuel increase as indicated (heavy solid line, top of figure), with year-to-year change in mean surface temperature in northern tropical zone 10-30°N (dashed line) (data from Angell and Korshover 1983). CO_2 annual means are from C. D. Keeling (SIO, personal communication, 1984) until the beginning of the NOAA/GMCC program in 1975 (+; Harris and Bodhaine 1983). Major ENSO events occurred in 1969/70, 1972/73, 1976/78, 1979/80, and 1982/83.

g C seems to have been returned to the atmosphere; the year-to-year CO_2 increase peaked in early 1984 at 2.2(±0.3) ppm/year, well above the mean growth rate of 1.4 ppm/year over the past decade. Recall that the past 5-year period has been one of nearly constant global production of fossil fuel ($\sim 5 \times 10^{15}$ g C/year), so that the observed large interannual variations in atmospheric CO_2 concentration are clearly not the result of changing rates of fossil fuel usage.

Previous investigators focused interpretations on the high CO_2 growth phase, which lags the SOI minimum by a few seasons, depending on the latitude of the CO_2 monitoring station (Bacastow et al. 1980); however, the precursor phase of low CO_2 growth is equally prominent in the 1982/83 event and may offer predictive possibilities for future ENSO events. There is also the distinct possibility that the El Chichon eruption (April 1983) is somehow linked to the subsequent atmospheric circulation anomalies in the spring and summer of 1982, which marked the unusually phased beginning of this ENSO event. If, in addition to the well-established stratospheric heating, tropospheric cooling could be clearly attributed to the El Chichon dust veil, such a cooling might have deepened the mid-1982 CO_2 slow-growth dip. One hope in unraveling the intertwined effects of El Chichon and ENSO on global circulation and on the global carbon cycle rests in the inherently longer time scale of the climatic signal of the volcanic stratospheric dust. As the circulation and SST anomalies attributable to the 1982/83 ENSO event fade, the residual and more persistent volcanic signal, and its global climatic and carbon cycle signature, may emerge.

Probably more important than the complicating effects of El Chichon is the very large uncertainty in assessing the global response of the land biosphere to the altered temperature and precipitation patterns that accompany a major ENSO event. Although regional crop failures, floods, and droughts are well documented, the globally integrated response and its effect on the atmospheric CO_2 level are more difficult to estimate. There is some recent evidence from satellite monitoring of the "vegetation index" of land plant activity that 1982 was a better growing season globally than was 1983 (Tucker et al., in press). This suggests that some fraction of the 5×10^{15} g C missing from the global atmosphere in the fall of 1982 may have been temporarily stored in the annually recycled biosphere.

An added complication is the role of perturbations in the marine biosphere during ENSO events, particularly in the productive upwelling regions in the eastern equatorial Pacific and off the South American coast. Marine biologists documented the drastically reduced productivity of this region during the October 1982 to July 1983 period of the most recent ENSO event, and estimated that the marine biospheric carbon sink may have fixed $\sim 1 \times 10^{15}$ g C less during this period than normal (Barber and Chavez 1983). Such a failure of the marine biospheric pump might account for a minor fraction of the observed atmospheric rise during this period (Figure 3.19; also see Chapter 5). As a research priority for the future, the evaluation of the relative importance of the marine biosphere versus variations in upwelling and surface warming of CO_2-rich deeper water will require detailed physical, chemical, and biological measurements in the ENSO source regions of the eastern equatorial Pacific throughout the full evolution of a major event. On a broader scale, separating the relative carbon cycle response of the land biosphere/soil from that of the ocean to the perturbed atmospheric/oceanic conditions of a major ENSO event remains a central priority for improving our understanding of the variability of the global carbon cycle.

The latitude dependence of atmospheric CO_2 response. In the previous section, the globally averaged response of atmospheric CO_2 to the 1982/83 ENSO event was summarized and the possible biospheric and volcanic complications briefly reviewed. A more detailed reconstruction of the latitude-dependent CO_2 response to this extreme circulation perturbation is possible from an analysis of the 24 monitoring sites of the NOAA/GMCC global atmospheric CO_2 network (Komhyr et al. 1985; Harris and Bodhaine 1983; Gammon et al. 1986). The weekly sampling at these primarily marine sites permits a zonally averaged

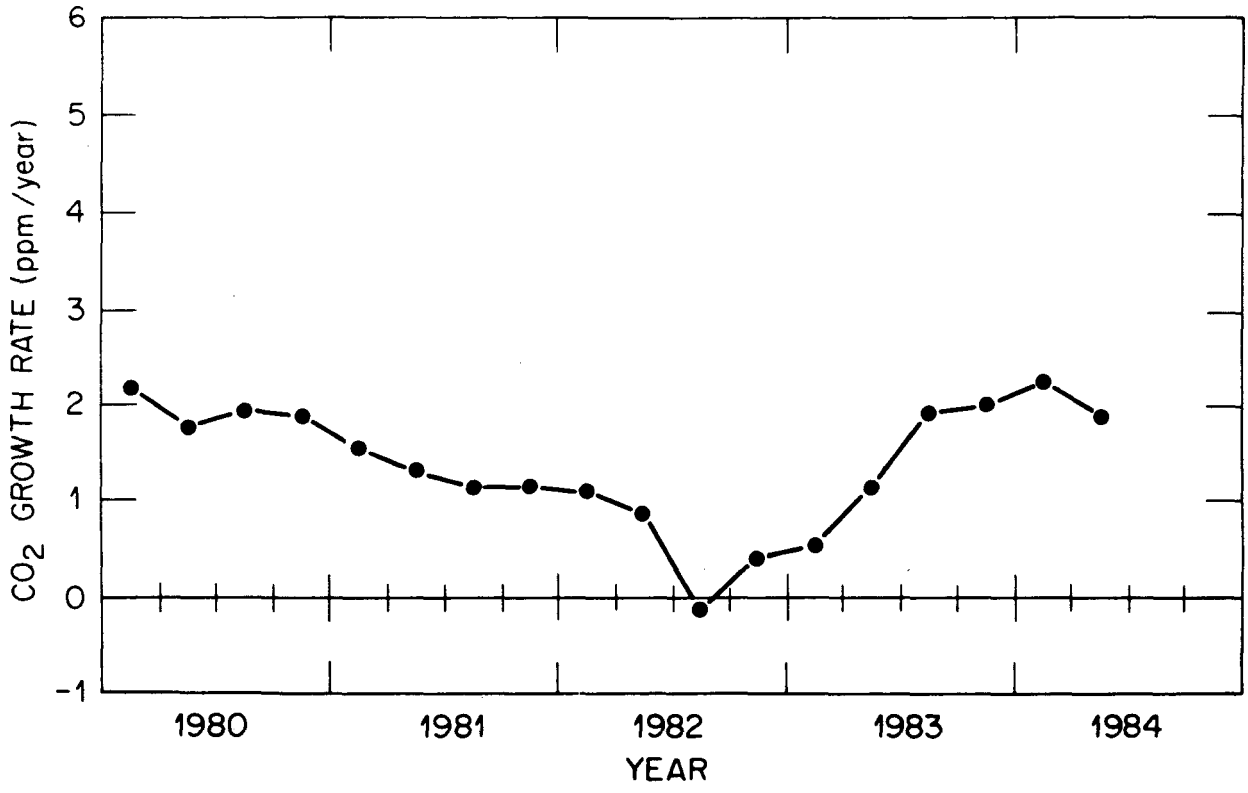


Figure 3.19. Year-to-year global atmospheric CO_2 change by season, plotted in the forward year. Data are from the 24-station global flask network of NOAA/GMCC (Komhyr et al. 1985; R. H. Gammon, NOAA, personal communication, 1984). Seasonal global means are area weighted. The decade-averaged CO_2 growth rate is ~1.4 ppm/year primarily due to fossil fuels.

CO_2 concentration surface to be constructed (Figure 3.13), the overall representativeness of which has been verified by direct atmospheric CO_2 sampling aboard ships of opportunity during the past 2 years (Harris and Nickerson 1984).

The regular breathing of the global biosphere is dominated by the activity of Northern Hemisphere land plants and soil (Figure 3.13). The Southern Hemisphere seasonal amplitude is much smaller. Even in this zonally averaged representation, several clear anomalies in the CO_2 distribution anticipate or accompany the 1982/83 ENSO event. First, the normal seasonal migration of the intertropical convergence zone (ITCZ), here defined as a region of steep CO_2 latitudinal gradient in the equatorial zone, appears perturbed in 1982, shifting far south of its normal position and crossing the equator in the fall of 1982. This southward dislocation of the ITCZ is characteristic of major ENSO events and was also reported in the meteorological data for 1982/83 (Philander 1983).

The second anomaly in the zonally averaged CO_2 distribution is the failure of the high southern latitudes to register the expected 1.5 ppm/year fossil fuel increase from 1981 to 1982 (e.g., in Figure 3.20A, the closed contours at 340 ppm, 30°–90°S, mid-1981 to the end of 1982). The year-to-year CO_2 changes by latitude and season are better visualized in Figure 3.20B, which is simply the difference between the concentration surfaces for two consecutive years plotted in the forward year.

The most obvious feature shown in Figure 3.20B is the strength and location of the low to negative CO_2 year-to-year change in 1982, which clearly suggests that extratropical CO_2 sinks in both hemispheres intensified during this period. At least in the high southern latitudes, neither changes in the response of the land biosphere in 1982 nor the direct radiative effects of stratospheric dust from El Chichon could have been responsible for the observed CO_2 decrease. This leaves as the only possible explanation a more effective southern ocean sink for CO_2 during this period in combination with a less efficient meridional transport of fossil fuel CO_2 from the north.

The interpretation of the negative CO_2 growth anomaly in the Northern Hemisphere is complicated, both by the presence of the El Chichon dust veil as well as by the unusual terrestrial biospheric response in 1982 (Tucker et al., in press). As shown in Figures 3.20 and 3.21, the CO_2 growth rate anomalies occurred in phase at both northern subtropical and southern subantarctic latitudes, a possible teleconnection not consistent with

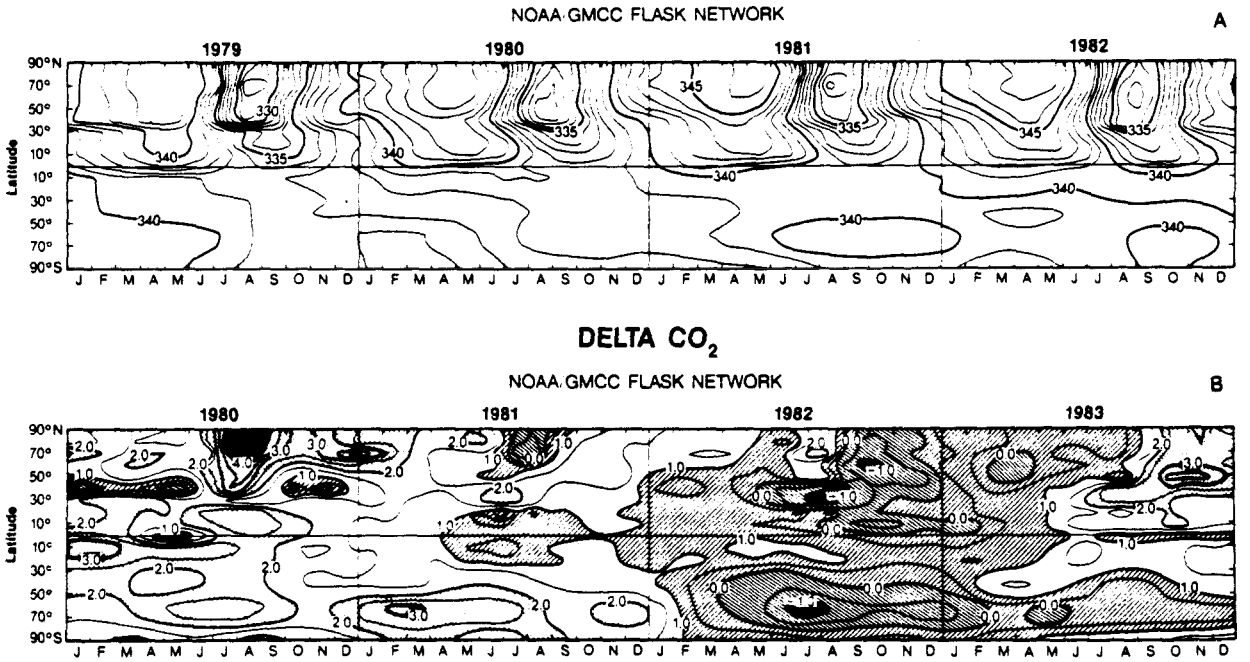
the concept of equatorially generated CO_2 perturbations migrating poleward. The Northern Hemisphere ocean sinks for CO_2 were probably more effective during this period, given the extensive and persistent negative SST anomalies (1°–3°C) in the far North Atlantic and Northwest Pacific, known source regions of deep and intermediate water formation (Climate Analysis Center [CAC] 1983).

An important conclusion from a review of the latitude-dependent CO_2 response to the 1982/83 ENSO perturbation is that the effect signal seems not to be generated at the equator, but rather appears nearly simultaneously at midlatitudes to high latitudes in both hemispheres, and displays features that both precede (low CO_2 growth) and follow (high CO_2 growth) the minimum of the SOI by several seasons.

Drawing on the pattern of the latitude-dependent CO_2 response to the weak 1979/80 ENSO event, it was expected and observed that the recovery to a more rapid than normal CO_2 increase would begin at southern tropical latitudes soon after the minimum in the SOI and peak in high northern latitudes in the late summer/fall of the same year, with the peak at high southern latitudes perhaps a season later. Although the Southern Hemisphere CO_2 results are incomplete for 1984, the atmospheric CO_2 response to the 1982/83 ENSO event by latitude and season seems to have followed the general pattern of the previous, weaker 1979/80 event. By mid-1984, the year-to-year CO_2 changes at most stations of the NOAA/GMCC flask network had returned to normal values (~1.5 ± 0.3 ppm/year), signaling the end of the CO_2 response to this extraordinary ENSO event.

CO_2 response in the equatorial zone. The latitude gradient of annual mean CO_2 concentrations (Figure 3.22) has been typically 3 to 4 ppm in recent years, the result of current fossil fuel releases of 5 × 10^{15} g C annually from sources heavily concentrated in the Northern Hemisphere (Komhyr et al. 1985). The second most prominent feature of the latitude distribution (Figure 3.22) is a 1-ppm equatorial enhancement in atmospheric CO_2, usually ascribed to the ocean upwelling source (Keeling 1983) or to tropical deforestation. An important consequence of the 1982/83 ENSO event is the obvious suppression of this equatorial feature in 1982 and its weak recovery in 1983 well south of its normal location (5°N) near the ITCZ. This is the first well-documented case of an ENSO modulation of this equatorial atmospheric CO_2 feature. This result, together with analyses of the isotopic record (Keeling et al. 1984), essentially rules out a major land source (i.e., tropical deforestation) for this feature, leaving only upwelling in the equatorial ocean, particularly the eastern Pacific, as an explanation.

Figure 3.20. Global atmospheric and delta CO_2. (A) Zonally averaged CO_2 concentration for 1979-1982 from the NOAA/GMCC global flask network, representing the sea-level, atmospheric mean CO_2 level (Komhyr et al. 1985). Approximate resolution of the original data used to construct the surface is 10° in latitude and biweekly in time. (B) Year-to-year difference of the concentration surfaces of Figure 3.20(A), plotted in the forward year showing the latitude and time dependence of the zonally averaged CO_2 growth rate anomalies (<1.0 ppm/year or >2.0 ppm/year) associated with the 1982/83 ENSO event.



Much less is known about the evolution of CO_2 perturbation in the upwelled surface waters of the east equatorial Pacific Ocean during the 1982/83 ENSO event. While some atmospheric CO_2 samples were taken aboard the NOAA R/V DISCOVERER in the equatorial zone in the spring of 1983 (Harris and Nickerson 1984), continuous measurements of the partial pressure of CO_2 (pCO_2) in the equatorial surface waters were not made until the fall of 1983, during the U.S./U.S.S.R. joint expedition of the A. KOROLEV between Hawaii and Tahiti. At that time (November-December 1983) the surface waters registered peak pCO_2 values of ~440(±10) ppm (130°–160°W, 5°-10°S). The degree of CO_2 supersaturation (~100 ppm above the atmospheric level) is not unusual relative to values observed during the First Global Geophysical Experiment expedition (1979/80) in the same location (Keeling 1983); however, the latitude of the maximum pCO_2 in the surface waters during the KOROLEV expedition is anomalous, well south of the expected position, much closer to the equator for that season and longitude. This marine CO_2 perturbation parallels the observed excursions of the equatorial atmospheric CO_2 feature and the meteorological ITCZ well south of their normal positions during the 1982/83 ENSO event. The DISCOVERER sampled the pCO_2 in these same equatorial surface waters in March 1984 and found that the latitude of peak supersaturation had relaxed northward toward its normal position on the equator, although broad regions of anomalously warm surface water were still encountered in south equatorial latitudes.[2] These fragmentary pCO_2 measurements can but hint at major fluctuations in the strength and location of the upwelling in the equatorial Pacific during the 1982/83 event. Only a detailed time series of chemical, physical, and biological measurements in these surface waters throughout the full evolution of the next major ENSO event can begin to separate the biological, oceanographic, and meteorological components.

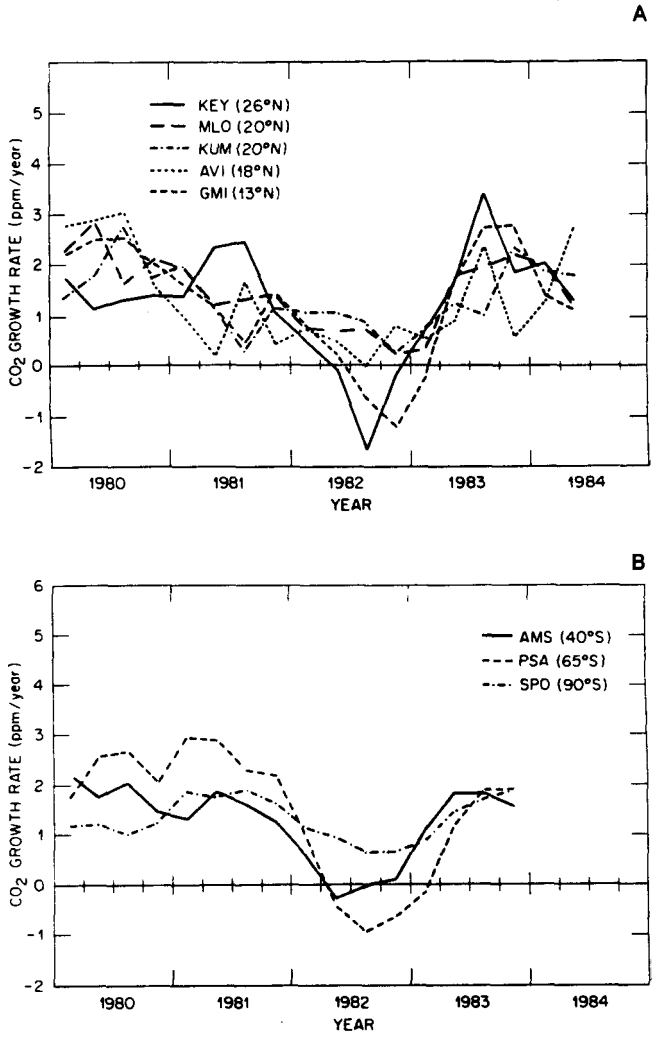


Figure 3.21. GMCC flush network CO_2 growth rates. (A) Year-to-year change in seasonal mean atmospheric CO_2 concentration at five northern tropical sites of the NOAA/GMCC flask network (Komhyr et al. 1985; R. H. Gammon, NOAA, personal communication, 1984). KEY = Key Biscayne, Fla.; MLO = Mauna Loa, Hawaii; KUM = Pt. Kumukahi (Hilo), Hawaii; AVI = American Virgin Islands; GMI = Guam, Marshall Islands. (B) As in (A), but for mid-to-high southern latitudes (Komyr et al. 1985; R. H. Gammon, NOAA, personal communication, 1984); AMS = Amsterdam Island (Indian Ocean); PSA = Palmer Station, Antarctica; SPO = South Pole. The southern data for 1984 are incomplete.

3.5.8 Increasing Amplitude of the CO_2 Seasonal Cycle

The regular variation of atmospheric CO_2 levels by season and latitude in the Northern Hemisphere is dominated by the seasonal exchange between the atmosphere and the land biosphere-soil carbon reservoir (Figure 3.14). Any major change in either photosynthetic uptake or respiratory release of carbon will alter the observed atmospheric CO_2 seasonality, which represents the local imbalance of these two opposing processes. Although the first attempts to find such long-term changes in the CO_2 seasonal amplitude at Mauna Loa (Hall et al. 1975) were not successful, more recent investigations found large and significant growth rates in the seasonal amplitude at several Northern Hemisphere locations (MLO, Weathership "P", BRW), especially during the past decade (Pearman and Hyson 1980; Cleveland et al. 1983; Komhyr et al. 1985; Keeling 1983). At Mauna Loa, the mean rate of increase was 0.66%/year from 1958 to 1981, resulting in a nearly 1 ppm (17%) increase (from 5.5 to 6.4 ppm) in the

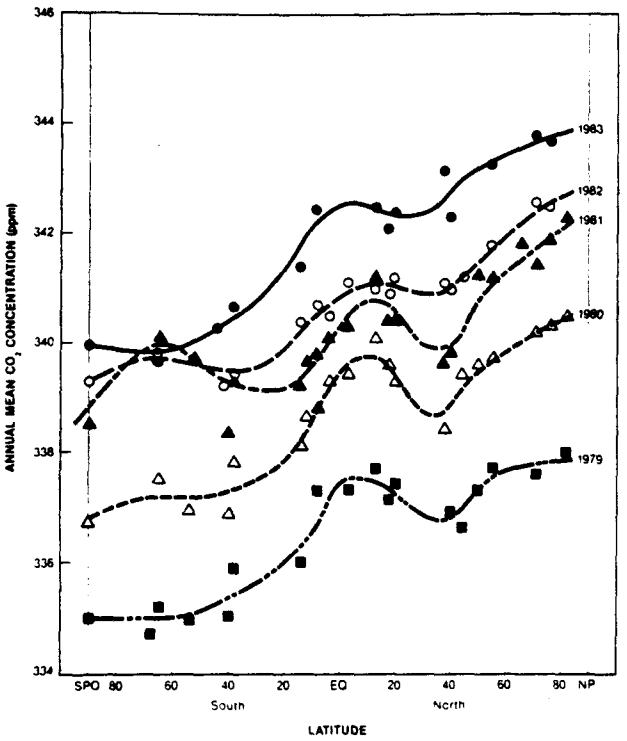


Figure 3.22. Latitude dependence of the annual mean CO_2 concentration for sea-level sites; data are from NOAA/GMCC, Canada (Wong et al. 1984), C. D. Keeling, SIO, personal communication, 1984), and Australia (Pearman and Beardsmore 1984). Figure is adapted from Kombyr et al. (1985). The smooth curves have been drawn by objective spline fits to selected station data.

CO_2 seasonality over this interval. For comparison, the corresponding CO_2 percentage increase was only 8%. The increase before 1975 was slight (~0.2%/year); since 1975, the growth has been very rapid (~2%/year or ~0.1 ppm/year). Similar growth rates, ~0.7%/year with respect to the local CO_2 seasonality, were recently reported for the long CO_2 records at more northerly stations (e.g., Weathership "P"; Keeling et al. 1984), again with even more rapid rates of increase since 1975. Although the NOAA/GMCC flask and continuous records at MLO and BRW extend only over the last decade, these also exhibit CO_2 seasonality growth rates of 1-2%/year over this period. The South Pole record is complex, showing a long-term increase (~0.5%/year, 1958-1978, Cleveland et al. 1983), but a rapid decrease during the past decade.[3,4] The Southern Hemisphere seasonality is small (1-2 ppm) and is of a complex origin; isotopic data suggest a predominant oceanic influence (Keeling et al. 1984). A growing Northern Hemisphere seasonality could perhaps be transported to high southern latitudes with a lag of ~1 year, arriving out of phase and causing the SPO seasonality to decrease at the same time that northern seasonalities are increasing. The changes at all latitudes are comparable in absolute terms (0.1-0.3 ppm/year), with high northern latitudes (BRW) showing the largest absolute rates of increase (0.3 ppm/year).

As discussed by Keeling (1983), an increasing CO_2 seasonal amplitude indicates "increased plant activity," but not necessarily increased net carbon storage (also see Chapter 6). The observed percentage increase in Northern Hemisphere CO_2 seasonality is more than twice that of the atmospheric CO_2 level, which suggests that any "fertilization effect" of enhanced photosynthesis by the land biosphere in a higher CO_2 environment could explain only a small fraction of the observed change in seasonal amplitude.

The seasonality of the growing fossil fuel release is also insufficient to explain the increase (Pearman and Hyson 1980). A more likely explanation is to be found in the strong temperature dependence of respiration of CO_2 from plants and soils and in the observed warming trend over the last decade in the hemisphere. A harmonic analysis of the secular changes in CO_2 amplitude at each site may reveal whether enhanced photosynthesis (annual) or enhanced respiration (semiannual) is more important. A first look at the NOAA/GMCC continuous CO_2 records since 1975 suggests that the changing seasonality of MLO is largely due to enhanced summer photosynthetic fixation of CO_2, while at BRW, spring and fall respiration pulses are as important as the enhanced summer drawdown in causing the CO_2 amplitude to grow. If, as it seems, the change is proportional to the local seasonality, then the largest absolute changes are predicted to be occurring in the latitude band 50° to 60° N where the CO_2 seasonal amplitude itself is maximum (Figure 3.14). This idea can be tested when the NOAA/GMCC flask records become long enough (> 10 years) for the secular trend of seasonal amplitude as a function of latitude to be separated with confidence from competing signals (fossil fuel, ENSO).

An interesting observation is that CH_4, second most important trace gas of the carbon cycle, is now known to be increasing in the atmosphere at ~1%/year, and that the percentage rate of the CH_4 increase has become more rapid during the last decade. If the latitude band of the source region driving the atmospheric CH_4 concentration increase is found to be the same as that for the CO_2 seasonal amplitude increase, one could then argue that both phenomena represent a response of the Northern Hemisphere biosphere and soil carbon pool to the incipient climate change predicted for the coming decades.

[3]C. D. Keeling, SIO, personal communication, 1984.
[4]D. A. Gillette, NOAA, personal communication, 1984.

3.6 SUMMARY AND RESEARCH RECOMMENDATIONS

3.6.1 Summary: Atmospheric CO_2 and Current Questions about the Carbon Cycle

A synthesis of the atmospheric CO_2 measurements from clean-air sites around the world represents the most precise and integrated record available of the workings of the global carbon cycle, of the fluctuating CO_2 exchanges of the atmosphere with both the ocean and the biosphere, and of the response of this complex system to strong perturbations of both natural (ENSO) and human (fossil fuel combustion) origins. The global climate change predicted for the coming generations will undoubtedly impress its signature on this CO_2 atmospheric record. This signature may offer one of the most sensitive possibilities for early detection of climate change. The research recommendations given in detail below are directed toward improving the understanding of the natural carbon cycle as it operated in the past and the perturbed carbon cycle today and in the future.

The history of atmospheric CO_2 before major human impact must be recovered with greater certainty from more refined and comparative analyses of cores from both deep sea sediments and polar ice, and from a more reliable separation of the $\delta^{13}C$ tree-ring record into its components of variable plant physiology and global CO_2 change.

The ice-core time series of atmospheric CO_2 variations should be extended to the full glacial cycle and compared in detail with corresponding records derived from $\delta^{13}C$ measurements of carbonates in deep-ocean sediment cores (Shackleton et al. 1983) for the period 125,000 to 1000 years B.P. Similarly, the atmospheric CO_2 record derived from $\delta^{13}C$ studies of annual tree rings for the period 10^4 to 10^2 years B.P. can be smoothed to the lower resolution (30-300 years) of the ice-core data for detailed comparison of CO_2 concentration variability. Of particular interest is the range and rate of natural CO_2 variability on decade-to-century time scales. Such fluctuations may rival the anthropogenic perturbation of recent decades (Δ ~70 ppm at ~0.5 ppm/year). The extant CO_2 absorption spectra in the Smithsonian archives offer the promise of a CO_2 history in the important time interval extending from the end of the last century to the era of modern measurements starting in the late 1950s.

The current period of precise atmospheric CO_2 records began with Keeling's measurements at Mauna Loa and the South Pole in 1958. These longer records are complemented by the more recently begun, but more geographically complete, CO_2 time series of several national and individual monitoring programs coordinated by WMO. A coordinated interpretation of all such data must be applied to the outstanding questions of the global carbon cycle (e.g., Is the biosphere currently a net source or sink? What will be the future source/sink roles of the biosphere and ocean when the carbon cycle is more strongly perturbed by greater rates of fossil fuel input?). Important clues are certainly imbedded in the complex response of the global carbon cycle to extreme natural fluctuations like the 1982/83 ENSO event. Global atmospheric CO_2 monitoring networks must be extended to the major land biomes in both tropical and higher latitudes in close coordination with intensified research on the use of satellite remote sensing to quantify the photosynthetic activity of these biomes.

Atmospheric CO_2 records in the future may well record changes in the apparent airborne fraction of the annual fossil fuel CO_2 input as the dynamic response of the global ocean and biosphere adjusts to changes in the fossil fuel growth rate. Detecting a change in the airborne fraction will, in turn, require a much-improved understanding of the competing signal, the natural interannual variability of the global carbon cycle responding to large-scale perturbations (ENSO, volcanoes).

The CO_2-induced climate change predicted to occur in the coming decades will measurably alter the "breathing of the world" as monitored by the international network of CO_2 sites. The details of the CO_2 response to changing climate will depend on many factors (rate of fossil fuel use, ocean thermal lag, role of other greenhouse gases, etc.); however, it seems certain that, under altered regional temperature and precipitation regimes, the regional and global seasonality of atmospheric CO_2 will be different than is currently observed. Such CO_2 records now give the most precise and spatially integrated signal of the regional imbalance of respiration and photosynthesis. An evolving biospheric-atmospheric exchange, as now appears to be occurring in the growing CO_2 seasonal amplitude in the Northern Hemisphere, may be the early signature of the predicted climate change. For the global biosphere of the future, the relative importance of enhanced photosynthetic uptake (CO_2 fertilization, β-factor response) versus enhanced respiration in the warmer, higher-CO_2 world of the next century cannot yet be predicted with confidence and will probably only be determined from the future CO_2 record itself. Specific detailed recommendations for future research are given below.

3.6.2 Specific Recommendations

3.6.2.1 *Pre-1958 History of Atmospheric CO_2*

Geological carbon cycle models and ocean sediment core analyses

1. Refine geological models to better define probable CO_2 levels from 10^8 to 10^6 years B.P.
2. Extend carbon isotopic studies in ocean sediment cores to all major ocean basins.
3. Quantify phase relationships of implied CO_2 variations with temperature variations at glacial-interglacial frequencies.
4. Extend the comparison of proxy CO_2 records for sediment cores with contemporaneous direct CO_2 records from ice cores since the last interglacial period.

Atmospheric trace gas measurements in trapped air bubbles in polar ice cores

1. Take cores to greater depth to extend the atmospheric CO_2 record back to the previous interglacial period.
2. Perform a detailed comparison to confirm unusual features (e.g., CO_2 fluctuations during the last glacial period) in Greenland versus Antarctic cores.
3. Begin isotopic ($^{13}CO_2$) and other trace gas (CH_4, CO, N_2O) measurements in ice cores.
4. Improve ice dating by developing trace ^{14}C accelerator techniques.
5. Quantify ice close-off times by calibrating top of core with fluorocarbon dating.
6. Relate the measured CO_2 history from ice cores to that derived from both sediment cores and $\delta^{13}C$ tree-ring records.
7. Resolve the methodological and sample-size effects currently extant in reported ice-core data.

Ocean chemistry

1. Reduce model uncertainties in back-calculating preindustrial atmospheric CO_2 levels from ocean chemical measurements (Redfield ratio, respiration, carbonate dissolution, seasonality of preformed values, mixing).
2. Improve the data base of relevant surface chemical properties (carbon cycle properties, nutrients, oxygen) in all major ocean basins in all seasons. Build on the results of GEOSECS and TTO to assess chemical tracer fields in all oceans at least once each decade.
3. Develop improved analytical techniques to directly measure the penetration of fossil fuel CO_2 into the upper ocean (e.g., total CO_2 manometrically to $\pm$ 1 μmol/kg).
4. Undertake a coordinated international program of ocean CO_2 monitoring as outlined by SCOR-Working Group 75 (1984).

Carbon isotopic measurements in tree rings

1. Improve models of plant physiology and variable ^{13}C fractionation to derive cleaner atmospheric CO_2 records from $\delta^{13}C$ tree-ring time series.
2. Investigate CO_2 fertilization and regional pollution (e.g., acid rain, metals) affecting the $\delta^{13}C$ tree-ring record.

Spectroscopic plates (1902-1962)

1. Extend the time coverage of these spectral analyses as far as possible, both back in time to the original Langley spectrum of the turn of the century and forward in time to the end of the Table Mountain record in the late 1950s, to achieve the overlap with Keeling's early flask measurements in southern California. There is no alternative approach or data set that offers comparable temporal resolution on the growth of atmospheric CO_2 in the first half of this century when biospheric releases were probably peaking.
2. Analyze the remaining 90% of the plates from Table Mountain and other sites in order of priority (with respect to date, location, and plate quality) to verify the global representativeness of the Table Mountain record.
3. Substantiate the results obtained to date by analyses of other CO_2 spectral features in the archived plates, and by careful studies of ratio spectra (CO_2/O_2, CO_2/N_2O) as a means to normalize instrumental effects and drift.
4. Continue the observational program of modern spectra at Kitt Peak National Observatory (Arizona), both for use in interpreting the historical CO_2 plates and as an important historical record of the evolution of the infrared opacity and of the abundance of CO_2 and other greenhouse gases in the Earth's atmosphere at the close of this century.

3.6.2.2 *Modern Atmospheric Carbon Dioxide Measurement Programs*

Current measurement networks

1. Continue all existing CO_2 measurement programs at remote clean air locations.
2. Converge NOAA and SIO CO_2 records at Mauna Loa and critically assess their level of agreement over the common interval since 1974.
3. Streamline the prompt archive of provisional CO_2 values from WMO/CO_2 programs for use by the carbon cycle modeling community within 1 year after observation.
4. Initiate a global CO_2 intercalibration experiment to determine the level of absolute agreement of participating laboratories.

New measurement programs

1. Expand the existing flask sampling networks into major land biomes if field tests confirm that regionally representative values can be obtained.
2. Through WMO, expand CO_2 measurement programs cooperatively into developing countries of South America, Africa, and Asia (initially as flask sampling sites).
3. Establish the "Mauna Loas of the ocean," a global network of ocean CO_2 monitoring stations in the major ocean basins for direct recording of the penetration of fossil fuel CO_2 into the sea (see proposal of SCOR [1984]), in collaboration with ongoing GEOSECS/TTO chemical surveys of the world's oceans.
4. Assess the utility of isotopic measurements in banded corals in equatorial waters as a proxy record for ENSO and atmosphere-ocean CO_2 anomalies over the last 100 years.
5. Undertake systematic, ongoing determination of $^{13}CO_2$ at representative subsets of the clean-air sites of the global CO_2 programs, sufficient to define the $^{13}CO_2$ latitude dependence and regional seasonal variations. Such isotopic time series will be invaluable for separating the relative importance of land data from oceans as important sources and sinks of CO_2.
6. Undertake the ongoing measurement of CH_4 and CO (and eventually $^{13}CH_4$ and ^{13}CO) in global CO_2 flask sampling networks. Analyses of the full suite of all three principal gases of the carbon cycle is needed to clarify oceanic versus biospheric exchanges with the atmosphere and to specify regionally important biosphere and soil processes (e.g., CO_2 fertilization versus tropical slash-and-burn versus soil respiration). Of particular interest is the latitude dependence of the secular growth rate of CH_4 and CO (tropical termites and deforestation or retreat of the boreal permafrost).
7. Establish ongoing programs for collection at selected sites (including South Pole) of large volumes of whole air for permanent archive and for ^{14}C analysis ($^{14}CO_2$, $^{14}CH_4$, ^{14}CO) to distinguish fossil fuel from more recent carbon sources.

3.6.2.3 *Interpretation of Atmospheric Records*

Airborne fraction. Critically assess all CO_2 records for evidence of variation in the apparent airborne fraction in response to abrupt changes in the rate of use of fossil fuels (e.g., since 1973). This will require a cleaner separation of the secular trend from the interannual variation associated with ENSO events.

ENSO. Investigate the response of the global carbon cycle to the extreme, natural perturbations of the atmospheric and oceanic circulation (e.g., 1982/83).

Remote sensing. Refine the use of satellite remote sensing methods ("greenness index") to quantify the photosynthetic activity of land plants in the major biomes with respect to the direct measurement of atmospheric CO_2 uptake registered by global CO_2 monitoring stations. Improve the net primary productivity (NPP) and net ecosystem productivity (NEP) values used in current models to be more consistent with the larger observed CO_2 seasonal amplitudes at high northern latitudes.

General circulation models (GCM). Extend the use of GCMs to interpret the global tracer fields of carbon cycle trace gases (CO_2, CH_4, CO) from combined monitoring networks by more realistic parameterization of the exchanges of the atmosphere with the biosphere and ocean, constrained by $^{13}CO_2$ observations as they become available. For models of CH_4 and CO, realistic tropospheric photochemistry will also be required.

Changing CO_2 seasonal amplitude. Analyze all available CO_2 records of more than 5 years length for evidence of changing seasonal amplitude. Seek to establish the latitude dependence for the observed amplitude increase in the Northern Hemisphere, relating it to long-term changes in other carbon cycle trace gases (CH_4, CO).

Atmospheric CO_2 and the high latitude oceans. Investigate the seasonal and interannual variability of atmospheric CO_2 levels at high southern latitudes in relation to regional anomalies in SST, wind stress, ice cover, and marine productivity.

3.6.2.4 *New Technologies*

Critically assess new measurement approaches for their usefulness to the CO_2/carbon cycle question, including gas chromatography and diode laser spectroscopy (as alternatives to the standard method of in situ CO_2 measurement, namely nondispersive infrared analysis); eddy correlation systems employing fast CO_2 sensors applied to the local flux measurement of air-sea CO_2 exchange; satellite remote sensing of the photosynthetic activity of the land biomes and of trace gas concentrations; and Fourier transform infrared spectroscopy for simultaneous ground-based measurements of the total atmospheric column abundance of many climate-related trace gases.

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4. CARBON DIOXIDE FROM FOSSIL FUEL COMBUSTION: TRENDS, RESOURCES, AND TECHNOLOGICAL IMPLICATIONS

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4.1 SOURCES OF CARBON DIOXIDE

Human population growth as well as increasing industrialization has resulted in the release of carbon from long-term storage in some of the natural reservoirs of the Earth. Conversion of large forests and other natural ecosystems to agricultural land to feed the growing population has been a source of carbon dioxide (CO_2) during both the 19th and 20th centuries. During the 20th century, as a result of industrial growth, the annual rate of CO_2 release from fossil fuel combustion has increased such that it now exceeds the amount released from deforestation (Chapter 6 this volume; henceforth all chapters cited are from this volume). There are strong indications that fossil fuels will continue to be the dominant source of CO_2 to the atmosphere (Chapters 9 and 10).

In this chapter we look at both the history and the potential of fossil fuel use as a source of CO_2. The analysis is divided into two parts: estimates of past fossil fuel CO_2 emissions and estimates of remaining resources and potential CO_2 emissions.

4.2 PAST FOSSIL FUEL DATA AND ESTIMATED CO_2 EMISSIONS

Estimates of CO_2 emissions from fossil fuels depend on three factors: production rate of the fuel, the amount of carbon in the fuel, and the fraction of the fuel that is oxidized. The reliability of the estimates depends on the quality of the available quantitative data concerning these factors. Fuels are categorized as solids, liquids, and gases and the three factors are analyzed for each category (Marland and Rotty 1983, 1984). For each fuel type the CO_2 emissions are given by:

$$CO_{2_i} = (P_i)(FO_i)(C_i),$$

where CO_2 represents the carbon emissions for a particular fuel type *i*, *P* is the annual production of that fuel, *FO* is the fraction of that year's fuel production that is oxidized, and C is the average carbon content of that fuel.

The Marland and Rotty (1983, 1984) calculations rely heavily on United Nations (U.N.) data for world fuel production. The U.N. Statistical Office uses a variety of data sources and makes the necessary adjustments to resolve inconsistencies (Marland and Rotty 1984). The U.N. master data file is continually adjusted and updated as better data become available, even for past years. The U.N. data are in common energy units (joules, tons of coal equivalent, etc.) as well as mass or volumetric units. This is important because there is a wide range of resource quality and chemical composition. For each fuel type, chemical composition can be correlated with heating value; for example, with solid fuels, a metric ton of coal equivalent (mtce) (29.31×10^9 J) defines the carbon content much more precisely than does a "ton of coal."

Fuel production itself must be carefully defined. The quantity of interest for natural gas production is, for example, what the U.S. Department of Energy would define as marketed production minus extraction loss. Reinjected gas is not counted, flared gas is treated separately, and gas liquids are grouped with liquid fuels. Hence, the extraction loss in gas processing is essentially a transfer to the liquid accounts. Liquid fuel production is defined as the sum of crude oil and natural gas liquids because carbon contents are approximately equal (Marland and Rotty 1984).

Carbon content of the fuel is determined in the many analyses of chemical composition made by both fuel producers and fuel users. Examination of some of these data enabled Marland and Rotty (1983, 1984) to estimate a mean value for the carbon content of each fuel type.

Limited information exists on the fraction of fuel produced that is actually consumed in a given year. Global estimates are obtained by extrapolating data for particular industries or regions. Some data are available for the amounts of petroleum and natural gas used as feedstocks, but the division between the portion likely to be oxidized and that remaining unoxidized is usually not well documented. Data on combustion efficiency are sufficient to estimate the fraction of unburned carbon, and, although this fraction has changed with time, little error is introduced by using an average value for all years of interest. Diversion to nonfuel uses is harder to estimate, and rapid changes are evident in recent decades; large uncertainties are currently assigned to these values (Marland and Rotty 1983, 1984).

The largest uncertainty in estimating CO_2 emissions from fossil fuels is associated with fuel production data.

4.2.1 Pre-1950 Data and Emissions

The energy data file of the U.N. Statistical Office contains extensive information on fossil fuel production and consumption and is published annually as the Energy Statistics Yearbook (formerly Yearbook of World Energy Statistics). The U.N. energy data file (U.N., each year) includes consistent annual fuel information from 1950 to present, but reliable worldwide information about fossil fuel production before 1950 is far less well documented. Keeling (1973) calculated the CO_2 production from fossil fuels for the years 1860-1953 from estimates of the production of commercial sources of energy compiled by the

Table 4.1
CO_2 Emission Factors for Fuel Data Before 1950

Fuel	Carbon Content	Oxidized to CO_2 (percent)	CO_2 Factor (tons C/ton fuel)
Coal	70%	99	0.693
Lignite	28%	99	0.277
(Coal plus Lignite as coal equivalent)	(69%)	(99)	(0.683)
Crude Petroleum	84%	91.5	0.769
Natural Gas	540 g/m^3	97	0.524 × 10^{-3}[a]

[a]Tons of carbon per m^3.
Source: Keeling (1973).

Department of Economic and Social Affairs, United Nations (1956). Little effort was made to ensure uniform fuel quality in updating fuel quantities in the 1956 U.N. report. The resultant variation in heating value and thus in carbon content made it necessary for Keeling to estimate a global average for the carbon content of each fuel. Analyses of more extensive recent fuel data suggest that Keeling's estimates for carbon content in pre-1950 fuels cannot be improved substantially.

After estimating the carbon fraction and the fraction oxidized to CO_2 (Table 4.1), Keeling (1973) used the fossil fuel data compiled by the United Nations for the years 1860-1953 (U.N. 1956) to produce a time series of global CO_2 emissions for that period. The fuel data for the pre-1950 period contain so many assumptions that the resultant estimates for CO_2 emissions have much greater uncertainty than those based on later U.N. fuel data, although the results for 1860–1949 are still accepted. The annually revised fuel data of the U.N. Statistical Office are now used for estimating CO_2 emissions after 1950 (see following sections).

The global CO_2 emissions estimates for 1860–1949 are shown in the semilogarithmic plot in Figure 4.1. The straight line segments indicate periods of relatively steady exponential growth of CO_2 emissions, and the breaks represent interrupted growth during the two world wars and the economic depression of the early 1930s. Except for the interruptions, the growth rate appears to be nearly steady at about 4.2%/year. Past exponential growth should not be taken as indicative of the future because such growth cannot continue indefinitely. It is unlikely that the world could have sustained the 4+% growth rate into the 1970s without the discontinuities of World War I, the Depression, and World War II (Elliott 1983).

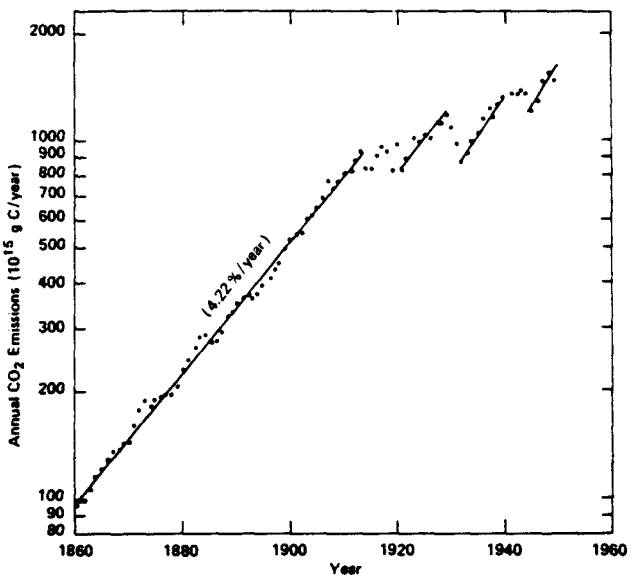


Figure 4.1. Annual CO_2 emissions resulting from fossil fuel combustion, 1860–1949.

4.2.2 1950–1973 Data and Emissions

The Energy Statistics Yearbook (U.N., each year) is based on a master data set maintained, revised, and updated by the U.N. Statistical Office. This data set includes complete fuel information for each year beginning in 1950. Many published sources of energy data are considered in maintaining and updating this information, and the master data set is promptly and appropriately revised when necessary. This may include changes in values for any or all years in the record. As a consequence, the energy data set of the U.N. Statistical Office has become the standard for estimating global CO_2 emissions.

Marland and Rotty (1983, 1984) reviewed available fuel information and updated the conversion factors from Keeling (1973) for application to the current U.N. data (Table 4.2). With current global fuel data, error limits cannot be assigned with scientific precision; the plus or minus values indicated in the table represent subjective judgments for an approximate 90% confidence interval. This is distorted somewhat when we recognize that some factors are based primarily on data

Table 4.2
Factors and Units for Calculating Annual CO_2 Emissions from Global Fuel Production Data

$$CO_{2_i} = (P_i)(FO_i)(C_i)$$

From Natural Gas Production

CO_{2_g} = CO_2 emissions in 10^6 t C
P_g = Annual production in thousands of 10^{12} joules ($\pm$~10%)
FO_g = Effective fraction oxidized in year of production = 0.98 $\pm$ 1%
C_g = Carbon content in 10^6 t per thousands of 10^{12} joules = 0.0137 $\pm$ 2%

From Crude Oil and Natural Gas Liquids Production

CO_{2_l} = CO_2 emissions in 10^6 t C
P_l = Annual production in 10^6 t ($\pm$~8%)
FO_l = Effective fraction oxidized in year of production = 0.918 $\pm$ 3%
C_l = Carbon content in tons C per ton crude oil = 0.85 $\pm$ 1%

From Solid Fuel Production

CO_{2_s} = CO_2 emissions in 10^6 t C
P_s = Annual production in 10^6 tce ($\pm$ ~11.2%)
FO_s = Effective fraction oxidized in year of production = 0.982 $\pm$ 2%
C_s = Carbon content in tons C per tce. = 0.746[a] $\pm$ 2%

From Natural Gas Flaring

CO_{2_f} = CO_2 emissions in 10^6 t C
P_f = Annual gas flaring in 10^9 m^3 ($\pm$ ~20%)
FO_f = Effective fraction oxidized in year of flaring = 1.00 $\pm$ 1%
C_f = Carbon content in tons per 10^3 m^3 = 0.525 $\pm$ 3%

Note: Units are consistent with fuel production data compiled in U.N. Energy Statistical Yearbook in 1983. All masses are in metric tons (10^6 g).

[a]This value includes a heating value adjustment to recognize that the carbon content, developed on a higher heating value basis, must be increased when used with U.N. production data (U.N. 1982) based on "net" or lower heating values.

from the most recent decade and that time-dependent biases have been introduced by employing these factors for the entire period after 1949.

The resultant emissions calculations (Figure 4.2) show a steady exponential growth of 4.44%/year from 1950 through 1973 and then a sharp change in the growth rate. During this same period the global emissions from the burning of oil and gas grew at rates above 7 and 8%/year, respectively (Figure 4.2). In 1950, CO_2 emissions from coal combustion were more than twice those from the combined combustion of oil and gas. However, the rapid growth in the use of oil and gas was sufficient to overtake coal as the primary emissions source by 1964.

In 1968, CO_2 emissions from oil alone exceeded coal—and this situation is still evident in most recent data (Figure 4.2). However, it appears that emissions from coal may soon surpass those from oil.

4.2.3 1973-Present Fuel Data and Emissions

The U.N. energy data sets also were used with the procedures documented in Marland and Rotty (1983, 1984) to obtain the CO_2 emissions for this most recent period. Calculations show a much-reduced rate of increase in emissions (Figure 4.2). The exponential growth rate for the 1973–1982 emissions is 1.46%/year, but the obvious scatter in the points makes it impossible to attribute significance to this value. Coal and gas emissions appear to exhibit continuing exponential growth, albeit at significantly different rates than before 1973, but oil emissions follow a much different pattern.

In 1973, when the Organization of Petroleum Exporting Countries (OPEC) began increasing the price of crude petroleum, the world economy was affected, and there was a resultant change in the growth rate of annual CO_2 emissions. In 1979 oil prices were raised sharply and the resultant changes in fuel use again were observed in the CO_2 emissions. The rate of growth in the use of petroleum products dropped sharply, particularly in areas most dependent on OPEC oil (Western Europe and Japan). Coal consumption, particularly in the United States, has increased to compensate partially for reduced oil use.

Total CO_2 emissions declined from 1979 to 1982 (Figure 4.2). Much of the variance in emissions data can be attributed to fluctuations in the world economy, and the data for the oil contribution in particular reflect this. The price and availability of oil have a major influence on the economy; and the economy, in turn, influences the demand for fuel. There is a lag between

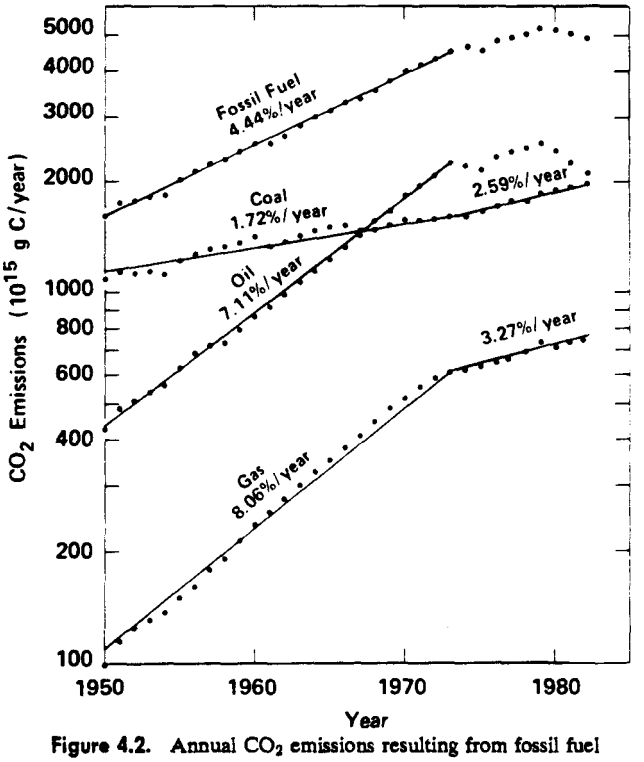


Figure 4.2. Annual CO_2 emissions resulting from fossil fuel combustion, 1950–1982.

the price increases and the production or consumption of oil because it takes time for industries and nations to develop alternatives. CO_2 emissions from oil have not grown exponentially during 1973 to 1982. Coal and gas CO_2 emissions, however, continue to exhibit exponential growth patterns, even though growth rates for natural gas have fallen dramatically since 1973.

4.2.4 Summary of Past Emissions of CO_2 From Fossil Fuel Combustion

For the period before 1950, the procedures of Keeling (1973) have been used to estimate the CO_2 emissions for each fuel type (Table 4.3). For the period after 1950, the procedures described by Marland and Rotty (1983, 1984) were used, and data for CO_2 emissions generated by cement production have been added to the tabulation (Table 4.4).

Although the data for total CO_2 emissions in Tables 4.3 and 4.4 are given to four significant figures, the accuracy of results is much less than that implied in the tabulation. It is evident from the results in Table 4.2, and as stated earlier, that the greatest uncertainty is associated with the global fuel data. Marland and Rotty (1983, 1984) estimated the post-1950 estimates o

Table 4.3
Global CO_2 Emissions (in 10^{12} g C) From Fossil Fuels, 1860–1949

Year	Coal	Lignite	Crude Petroleum	Natural Gas	Total CO_2
1860	91.5	1.7	0.1	0.0	93.3
1861	96.7	1.8	0.2	0.0	98.7
1862	96.1	2.0	0.3	0.0	98.4
1863	103.5	2.1	0.3	0.0	106.0
1864	112.5	2.4	0.2	0.0	115.1
1865	119.0	2.6	0.3	0.0	121.9
1866	125.7	2.6	0.4	0.0	128.7
1867	134.7	2.8	0.4	0.0	137.9
1868	133.3	3.0	0.4	0.0	136.7
1869	138.1	3.2	0.5	0.0	141.8
1870	141.0	3.4	0.6	0.0	145.0
1871	157.4	3.9	0.6	0.0	161.9
1872	170.9	4.3	0.7	0.0	175.9
1873	182.5	4.8	1.2	0.0	188.4
1874	177.3	5.3	1.2	0.0	183.8
1875	182.8	5.3	1.1	0.0	189.2
1876	184.9	5.5	1.2	0.0	191.6
1877	188.8	5.5	1.7	0.0	196.0
1878	188.9	5.7	1.9	0.0	196.5
1879	199.1	6.0	2.5	0.0	207.6
1880	217.5	6.4	3.2	0.0	227.1
1881	234.9	6.8	3.4	0.2	244.4
1882	251.3	7.1	3.8	0.3	262.5
1883	268.8	7.6	3.2	0.4	280.0
1884	269.9	7.8	3.8	0.6	282.1
1885	263.8	8.0	3.8	0.8	276.4

Table 4.3 (Continued)

Year	Coal	Lignite	Crude Petroleum	Natural Gas	Total CO_2
1886	264.4	8.3	5.0	1.0	278.7
1887	282.2	8.6	5.0	1.3	297.7
1888	305.8	9.2	5.5	1.5	321.9
1889	310.5	9.8	6.5	1.7	328.5
1890	328.9	10.8	8.1	2.0	349.7
1891	342.2	11.5	9.6	2.0	365.4
1892	345.5	11.7	9.4	2.1	368.7
1893	337.6	12.2	9.7	2.2	361.6
1894	352.9	12.5	9.4	2.3	377.1
1895	371.7	13.6	10.9	2.5	398.6
1896	382.4	14.5	12.1	2.7	411.6
1897	400.0	15.7	12.8	3.0	431.5
1898	421.5	16.7	13.2	3.2	454.6
1899	462.2	17.7	13.8	3.5	497.3
1900	485.6	19.8	15.8	3.7	524.9
1901	497.2	21.3	17.7	4.1	540.3
1902	508.5	20.9	19.1	4.4	552.9
1903	559.5	21.7	20.5	4.7	606.5
1904	563.0	22.6	23.0	4.9	613.4
1905	594.2	24.1	22.7	5.6	646.6
1906	641.6	25.8	22.5	6.1	696.1
1907	708.7	28.3	27.8	6.4	771.2
1908	670.9	29.3	30.1	6.3	736.6
1909	700.2	29.8	31.5	7.5	769.0
1910	732.2	30.0	34.5	8.0	804.8
1911	746.2	31.2	36.3	8.1	821.8
1912	785.7	34.5	37.1	8.9	866.2
1913	842.9	35.7	41.3	9.2	929.0
1914	752.7	33.5	43.0	9.3	38.4
1915	741.2	34.3	45.4	9.9	830.8
1916	798.4	36.5	48.1	11.8	894.8
1917	842.2	37.4	53.2	12.5	945.3
1918	828.1	39.2	53.4	11.3	932.0
1919	720.7	36.9	59.6	11.7	828.9
1920	826.4	43.6	76.3	12.6	958.9
1921	688.3	46.6	82.6	10.4	828.0
1922	735.8	50.6	92.2	12.0	890.6
1923	835.4	45.3	108.8	15.8	1005.3
1924	824.0	47.9	108.7	17.9	998.5
1925	821.3	51.8	114.4	18.9	1006.4
1926	815.7	52.0	117.5	20.8	1006.0
1927	883.7	56.2	134.7	22.8	1097.5
1928	863.5	60.8	141.6	24.9	1090.9
1929	918.3	64.4	158.9	30.3	1171.9
1930	843.3	54.6	151.2	28.4	1077.5
1931	744.6	50.3	145.7	27.6	968.2
1932	662.0	47.1	139.0	25.7	873.8
1933	693.0	48.3	151.6	25.9	918.8
1934	754.0	53.0	160.1	29.4	996.5
1935	770.3	57.0	174.3	30.1	1031.7
1936	866.3	56.8	188.4	34.9	1146.4
1937	902.3	69.8	215.3	38.8	1226.4
1938	841.3	73.1	209.9	37.0	1161.4
1939	892.6	80.9	219.6	39.8	1232.9
1940	944.6	88.4	244.5	42.9	1300.4
1941	985.4	91.7	214.6	45.4	1337.1
1942	990.3	92.5	202.2	49.4	1334.4
1943	991.0	96.4	221.5	55.1	1364.0
1944	955.6	87.5	249.2	59.8	1352.2
1945	823.3	51.9	263.0	65.5	1203.6
1946	842.7	70.9	289.1	67.8	1270.5
1947	952.2	75.9	318.4	75.0	1421.5
1948	990.3	82.9	359.9	84.5	1517.5
1949	930.0	92.0	358.4	89.2	1469.6

Source: Adapted from Keeling (1973).

Table 4.4
World CO_2 Emissions (in 10^{12} g C) from Fossil Fuels and Cement, 1950–1982

Year	Solid Fuel	Liquid Fuel	Gas Fuel	Flared Gas	Cement	Total CO_2 From Fuels and Cement
1950	1077	423	97	23	18	1638
1951	1137	479	115	24	21	1776
1952	1127	504	124	26	22	1803
1953	1132	533	131	27	24	1847
1954	1123	557	138	27	27	1872
1955	1215	625	150	30	30	2050
1956	1281	679	161	32	32	2185
1957	1317	714	178	35	34	2278
1958	1344	732	192	35	36	2339
1959	1380	790	214	36	40	2460
1960	1430	850	235	39	43	2597
1961	1341	906	253	41	46	2587
1962	1373	981	276	44	49	2723
1963	1431	1054	300	47	52	2884
1964	1481	1138	328	51	57	3055
1965	1503	1220	351	55	59	3188
1966	1524	1323	380	60	63	3350
1967	1466	1421	409	66	66	3428
1968	1497	1551	445	73	70	3636
1969	1524	1669	487	80	74	3834
1970	1577	1833	516	88	78	4092
1971	1572	1944	553	90	84	4243
1972	1584	2054	583	95	90	4406
1973	1604	2238	608	112	96	4658
1974	1613	2245	616	107	96	4677
1975	1683	2131	621	96	96	4627
1976	1727	2311	645	110	102	4895
1977	1779	2391	668	108	109	5055
1978	1795	2423	695	106	117	5136
1979	1892	2527	730	105	119	5373
1980	1935	2412	718	102	121	5288
1981	1947	2270	727	93	122	5159
1982	1999	2162	727	92	122	5102

Source: Adapted from Marland and Rotty (1984).

global CO_2 emissions to have an uncertainty of between ±6 and ±10%. Keeling (1973) estimated a relative error of approximately ±13% after 1960, and suggested that errors in pre-1949 data could well be more than ±20%. Retroactive improvement in the U.N. post-1949 fuel data has reduced the estimate of errors in annual emissions below the ±13% figure cited by Keeling.

A perspective on the magnitude of the values in Tables 4.3 and 4.4, and the influence of fossil fuels on the increase of CO_2 in the atmosphere can be gained by comparing the emissions with the increased CO_2 in the atmosphere. From January 1958 through December 1982, CO_2 emissions from fossil fuels (including a small contribution from cement manufacture) totaled 98.79 × 10^{15} g C. Between March 1958 and March 1983,[1] the seasonally adjusted atmospheric CO_2 concentration at Mauna Loa Observatory increased 26.8 ppm[2] (57.0 × 10^{15} g C). Thus, the observations represent an amount equivalent to 58% of the releases from fossil fuel combustion.

4.2.5 Implications of Recent Trends for the Future

Immediately after World War II, the world was dominated by a few nations. Not only was that true from a political and military standpoint, but this was also the case for industrial activity and energy use and the associated CO_2 emissions. In the more than 35

[1]March 1958 to March 1983 was used rather than January 1958 to January 1983 because the measured record does not begin until March in 1958.

[2]C. D. Keeling, Scripps Institution of Oceanography, personal communication, 1984.

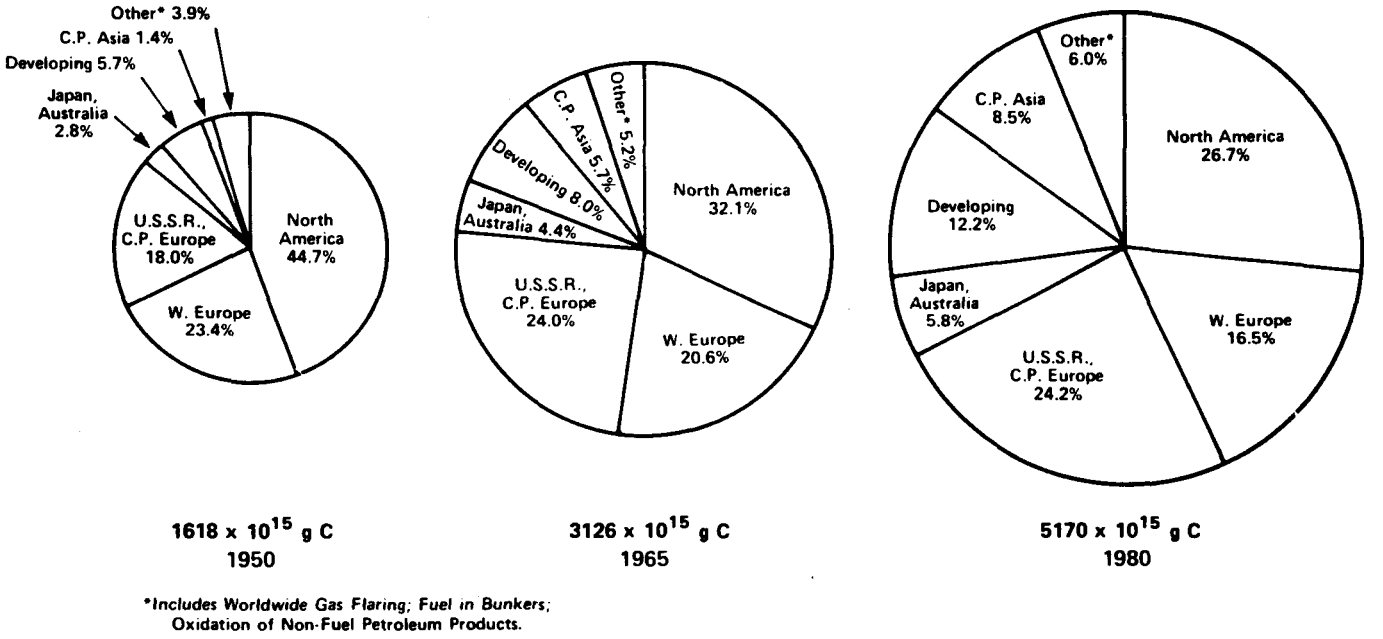


Figure 4.3. The changing patterns of global CO_2 emissions. C. P. = Centrally Planned Economies (e.g., Soviet Bloc).

years since that time, the world has undergone highly significant changes, and attempts to construct scenarios for the future must be based on an understanding of underlying causes for the most recent trends.

Since 1950, the global pattern of CO_2 emissions has been changing in the same way as the pattern of industrialization (Rotty and Marland 1986). In 1950, North America and Western Europe produced over 68% of all the CO_2, but by 1980 these areas produced only 43% of the expanded total (Figure 4.3). In contrast, the fraction of global CO_2 emissions attributable to the People's Republic of China and the developing countries in South Asia, Africa, and Latin America grew from 7% to over 20% of the increasing world total.

Even in the face of slowed development caused by current economic conditions, the developing world should continue to assume a greater share of global CO_2 emissions because of enhanced population and economic growth (Chapter 9). Although the percentage of CO_2 emissions from the developing world (not including the People's Republic of China) may not double (to 24 or 25%) in the next 30 years as it did in the past 30 years, the evidence is strong that the patterns of growth will be different than in the Organization for Economic Cooperation and Development (OECD) countries. Thus, the changing economic and energy needs of different world regions must be seriously considered when developing reasonable emissions scenarios for the future (see Chapter 9).

We are confident that an increasing percentage of the world's energy supply will be converted to and used as electricity. This is supported by the increasing trend toward more electrification in virtually all parts of the world (U.N., each year). In developed regions, the increased use of electrical energy is occurring in more and more types of industrial operations (Burwell 1983). There is also a strong possibility that use of electricity in transportation will increase, especially in the developing world. An increase in the percentage of the total energy used as electricity has implications for the CO_2 issue that can be addressed at least qualitatively (Chapter 9).

A shift toward greater use of electricity changes the fuel mixes that are possible in meeting future energy demand. A substantial number of new technologies (or old technologies, revised or modernized) that can provide significant quantities of electrical energy are emerging, and in the context of the CO_2 issue, some of these take on special importance. Many produce electricity without using fossil fuels, and their development will result in a trend toward reduced CO_2 emissions. Technological developments in energy conservation, nuclear energy, photovoltaics, and wind energy conversion will result in reduced CO_2 emissions as the world experiences increased emphasis on electricity. However, some of the developing tech-

nologies for enhanced coal extraction and utilization, and deep drilling to increase the supply of natural gas, could cause increases in CO_2 emissions.

4.3 FOSSIL FUEL RESOURCES AND POTENTIAL CO_2 EMISSIONS

The magnitude of fossil fuel resources is sufficiently large that refining current estimates is not likely to appreciably affect uncertainties in atmospheric CO_2 levels. The total quantity of carbon in recoverable fossil fuels is many times greater than that in any other active reservoir. The quantities of carbon extracted during the next several centuries will depend far more on socioeconomic and technological factors (Chapter 9) than on the size of the ultimate resource.

It is difficult to reconcile the shortages experienced during the 1970s with available resources. Even the shortage of petroleum, seemingly so acute during the last decade, appears inconsistent with the estimates of readily recoverable resources remaining in conventional petroleum deposits. Shortages resulted from inadequate production and thus relate to reserves rather than to resources. Resources may be very large, but if they are presently undiscovered, reserves may be limited and shortages can occur.

Annual CO_2 emissions from fossil fuels are currently about 5×10^{15} g C. The potential for further releases or increases in emissions rates is determined by the amount of fossil fuels remaining to be recovered and the rate at which they are exploited. The terms reserves and resources are easily confused, and they are used in different ways by analysts from different disciplines, often leading to confusion or errors of interpretation.

In addressing the question of potential CO_2 releases to the atmosphere, the amount of fuel that can be recovered and used is the quantity of interest. Three resource terms are commonly used by oil and gas analysts and we will also apply them to coal (Kalisch and Wingenroth 1983):

1. *Demonstrated reserves*—resources already discovered that can be produced with current technologies and under current economic conditions;
2. *Undiscovered recoverable resources*—resources that are thought to exist and that can be produced, given technological and economic conditions likely in the foreseeable future;
3. *Remaining total resources*—all remaining deposits such that some portion could be economically produced.

The potential CO_2 emissions should be estimated from the sum of the first and the second of these resource categories—those classed as demonstrated reserves and undiscovered recoverable resources.

The very nature of the definition makes undiscovered recoverable resources an uncertain quantity even in well-defined and well-explored regions, and especially in areas of the world where reliable data are sparse. Each of the fossil fuel categories will be examined in this chapter and an attempt will be made to determine a best estimate of potential CO_2 emissions (based on "consensus" of demonstrated reserves and undiscovered recoverable resources) and an upper limit speculation that includes estimates of remaining total resources. Most of the discussion will be limited to fossil fuels from conventional sources because rates of production from unconventional sources are not likely to be of such dimension as to sustain them as major contributors to global CO_2 emissions during the foreseeable future.

4.3.1 Solids

Solid fossil fuels include a variety of coals and coal-like materials. Coals are usually classified as anthracite, bituminous, subbituminous, and lignite, and within each class they are grouped by heating value. "Hard coal" includes both anthracite plus bituminous, and "soft coal" includes subbituminous and lignite. Table 4.5 has been adapted from the American Society for Testing Materials (ASTM) classification of U.S. coals. In some areas of the world, the term "brown coal" is used for poor grade subbituminous coal or lignite. Peat may be included among the solid fossil fuels in data from some countries, but on a global scale neither the present rate of use nor the resources of peat are large in comparison to those of the total solid fossil fuels. Therefore, data for peat are included here only as a small addition to the world totals.

The World Energy Conference (WEC) has periodically provided estimates of proved reserves and energy resources. The data assembled by the WEC were compiled by an international advisory panel and published as the *Survey of Energy Resources* (WEC 1980). Although resource estimates for coal have been less variable than those for oil or gas, the WEC and other estimates generally have increased with time because world population and economic pressures have resulted in more exploration for new energy sources. The uncertainty in these estimates is still very large despite the lack of variability from one estimate to another. Table 4.6 was constructed from WEC data, and estimates of recoverable resources were made by Parent (1983). His estimates were generally made by taking 50% of the total resources reported in WEC (1980), but they included some adjustments for the

Table 4.5
Classification of Coals

Class	Group	Fixed Carbon Content (percent)	Heating Value (MJ/kg)
Anthracite	1. Meta-anthracite	98	
	2. Anthracite	92–98	
	3. Semianthracite	86–92	
Bituminous	1. Low volatile	78–86	
	2. Medium volatile	69–78	
	3. High volatile A	<69	>32.5
	4. High volatile B		30.2–32.5
	5. High volatile C		26.8–30.2
			24.4–26.8
Subbituminous	1. Subbituminous A coal		24.4–26.8
	2. Subbituminous B coal		22.1–24.4
	3. Subbituminous C coal		19.3–22.1
Lignite	1. Lignite A		14.6–19.3
	2. Lignite B		<14.6

Source: Adapted from ASTM (1974).

United States, the United Kingdom, and India. For these countries Parent provided a range of values. This range reflected disagreements among data sources. Most of the resource numbers in Table 4.6 could be substantially higher.

Estimates of both proved reserves and total resources have changed with time, largely because of new discoveries. However, the estimates of recoverable resources have been affected as much or more by changes in economic conditions and improvements in the efficiency of extraction. Even though the uncertainties in the coal resources are large, the revisions made in the numbers during the past decade have been small—partially because the numbers are so large that there appears to be little to be gained in refining the estimates.

Table 4.6
World Coal Resources
[in tons and tons of coal equivalent (tce)]

	Demonstrated Reserves[a]		Resources		
			Total in Place[b]		Recoverable[c]
	(10^9 tons)	(10^9 tce)	(10^9 tons)	(10^9 tce)	(10^9 tce)
Africa	32.7	32.6	220.2	217.8	109
Asia[c]	119.0	116.1	1649.8	1605.6	803-813
Europe[c]	203.2	136.9	767.7	664.8	331-350
Oceania	59.5	36.4	784.1	570.2	335
U.S.S.R.	233.0	165.5	5926.0	4631.8	2316
W. Hemisphere	234.4	199.9	4219.8	3270.0	960-1501
Total Coal	881.8	687.4	13,477.6	11,060.2	4854-5421
Peat	15.8	5.8	318.3	122.0	61
Total Solid Fuels	897.6	693.2	13,795.9	11,182.2	4914-5485
Potential CO_2 Emissions @ 700 kg C/tce					3440-3840

[a]Data from WEC (1980).
[b]Estimates from Parent (1983) (taken as half the total resources, based largely on WEC data, but with some minor revision).
[c]Not including U.S.S.R.
Source: Adapted from WEC (1980) and Parent (1983).

CUMULATIVE PRODUCTION	IDENTIFIED RESOURCES	UNDISCOVERED RESOURCES PROBABILITY RANGE		
64		95%	MODE	5%
ECONOMIC	RESERVES			
	103	46	79	202
MARGINALLY ECONOMIC	MARGINAL RESERVES (Each 1% world average recovery increase yields 5×10^9 tons to reserves)			
SUBECONOMIC	SUBECONOMIC RESOURCES (REMAINDER IN PLACE)			
	324	89	153	393

Figure 4.4. Classification of world crude oil resources × 10^9 tons oil. Subeconomic quantity based on world average conventional recovery of 34%; marginally economic undiscovered resources, no data available. Terminology has been simplified but does not result in significant variance with classification of resources as reported in World Petroleum Conference. (Identified resources and production as of January 1, 1981, undiscovered resources as of March 1983. Adapted from Masters et al. 1983.)

4.3.2 Liquids

The classification of liquid fuel resources is based on a combination of economic assumptions and knowledge of petroleum geology. Because there are occasional misunderstandings in definitions of terms, especially among professionals not working in the petroleum field, the apparent uncertainties in the values reported for each classification frequently are confusing. To assist in the discussion here, the format of the U.S. Geological Survey resource classification diagram has been adopted and is shown in Figure 4.4.

Figure 4.4 shows the estimates for both reserves and resources. In the figure, the degree of economic potential decreases from top to bottom and geologic confidence in the data decreases from left to right. Values are provided for economically recoverable resources of conventional crude oil as well as for the subeconomic remainder. The estimates for identified and undiscovered resources include crude oil that has greater than 10° API (American Petroleum Institute) gravity that is economically recoverable by conventional primary, water-flood, and pressure maintenance techniques. Natural gas liquids are not included, nor are the unconventional resources such as extra-heavy oils (less than 10° API), for example, in the Orinoco Petroleum Belt of Venezuela or in Athabascan tar sands of Canada.

Reserves represent the total amount expected to be recovered from known fields by conventional techniques including water-flood whether or not actually installed. Prior to 1970, industry custom was to report as measured or proved reserves (a fraction of reserves as defined here) only that portion that had actually been developed enough to permit reliable flow testing. Since 1970, reporting has been based on an estimate of the maximum expected conventional recovery. The remainder of the identified resources in Figure 4.4 are marginally economic and subeconomic resources. Marginally economic reserves will eventually include additional quantities from enhanced oil recovery (EOR); however, EOR is still under development.

The undiscovered resources column in Figure 4.4 is also divided into economic, marginally economic, and subeconomic resources. The quantities are classified by the estimated probability of being exceeded—95%, 5%, and a measure of central tendency, the mode. The sum of undiscovered resources and identified resources is the quantity of interest for global carbon cycle analyses.

The undiscovered economic resources of the world have been assessed by techniques used earlier by the U.S. Geologic Survey in domestic petroleum assessment. The process is subjective: results of the geological study of each area and the corresponding resource calculations, which are used as scaling factors in the assessment, are given to a team of 10 to

Table 4.7
Estimate of Recoverable Resources (10^9 tons) of World Crude Petroleum

Region	Cumulative Production	Demonstrated Reserves	Undiscovered Recoverable Resources			Estimated Total Remaining Recoverable (mode)
			Probability Range			
			95%	Mode	5%	
North America	20.3	9.0	14.9	23.3	46.0	32.3
South America	6.7	4.9	2.9	4.7	9.9	9.6
Europe (except U.S.S.R.)	1.6	3.8	1.7	2.9	7.0	6.7
U.S.S.R.	9.7	10.0	8.4	15.3	49.0	25.3
Africa	4.6	7.4	4.0	6.6	15.0	14.0
Middle East	17.7	63.4	10.3	17.9	48.1	81.3
Asia/Oceania (except U.S.S.R.)	3.0	4.9	4.7	8.3	25.1	13.2
Antarctica	0	0	0	0	2.7	0
WORLD TOTAL (in 10^{15}g C)	63.6 (54)	103.4 (87)	45.9 (39)	78.6 (66)	202.4 (170)	182.0 (153)

Source: Adapted from Masters et al. (1986).

12 specialists, and each specialist makes an estimate. After the estimates are posted, the results are debated from the perspective of personal experiences. When necessary, the estimation procedure can be repeated until consensus is established. The final assessment is an average of individual estimates, and probabilistic methods are used to determine aggregates of the ranges of values determined for different areas.

Approximately 90% of the world's undiscovered resources of crude petroleum have been assessed in this manner during the past 3 years, and the process has been estimated to have utilized about 18 man years. The world total of 153 × 10^{15} g C in Table 4.7 is believed to represent the most probable amount of remaining recoverable crude petroleum. This total is less than other estimates (e.g., Heede 1983; Rotty and Marland 1980; WEC 1980; Edmonds and Reilly 1985), but falls within the range of uncertainty.

Estimates of the amount of recoverable natural gas liquids are subject to more assumptions; data bases are poorer than those for either crude petroleum or natural gas. The *Survey of Energy Resources,* Part A (WEC 1980) states:

> Figures for reserves of gas condensates with an accuracy of better than ±50 percent can only be made under very favorable conditions, . . . estimates for reserves and resources of gas condensates . . . based on one overall average value for the condensate/gas relationship, which is the method used by Meyerhoff (1979), can only lead to very general and unreliable results. (p. 98)

The estimates for natural gas liquids made by WEC (1980) have been accepted here (Table 4.8). The addition of the total remaining recoverable carbon in natural gas liquids to that in crude petroleum gives a probable carbon pool of 168 × 10^{15} g C for conventional liquid fuels yet to be recovered.

Although unconventional sources of petroleum exist in large quantities as extra-heavy crude oil and bitumen in various parts of the world (Meyer et al. 1984), most of it is very expensive to extract and use. For the CO_2 issue, we are interested in how much will be produced rather than how much is in the ground. For unconventional petroleum, reserves are defined as the amount produced over a 25-year expected service life of production facilities. Otherwise recoverable resources are classified as marginal reserves, and the remainder of the resources is classified as subeconomic. Approximately 10% of the total resource is economically recoverable (Table 4.9). At present, world production from tar sands and extra-heavy oil deposits is only 11 × 10^6 tons/year because extra-heavy oil and bitumen cannot compete economically with conventional crude oil. Thus the use of these deposits in quantities of concern to the CO_2 issue will occur in the distant future; although the resource is very large, its probable exploitation is highly speculative.

4.3.3 Gases

Much of the previous discussion of liquid fuel resources and potential CO_2 emissions applies to natural gases. One important difference finds its origin in the fact that part of the gas that is associated with oil production is reinjected for pressure maintenance in oil fields, and thus classifying all natural gas as recoverable would result in an overestimate of the amount burned and

Table 4.8
Estimates of Recoverable Resources (in 10^9) of Natural Gas Liquids

Region	Proved Recoverable Reserves	Additional Reserves Resources	Estimated Total Remaining Recoverable
North America	0.69	1.10	1.79
Latin America	0.39	0.25	0.64
Western Europe	0.36	0.55	0.91
U.S.S.R. and E. Europe (also includes People's Republic of China)	2.47	6.33	8.47
Africa	0.47	0.45	0.92
Middle East	1.89	2.10	3.99
Asia/Oceania (except U.S.S.R. and People's Republic of China)	0.31	0.85	1.16
WORLD TOTAL	6.58	11.60	18.18
(in 10^{15} g C)			(15.27)

Source: Adapted from WEC (1980).

Table 4.9
World Resources in 10^9 barrels (1 bbl = 0.159 m^3) of Extra-Heavy Oil and Bitumen

Cumulative Production	Extra-heavy oil	0.227
	Bitumen	0.427
Reserves	Extra-heavy oil	0.2
	Bitumen	1.5
Marginal Reserves	Extra-heavy oil	59
	Bitumen	172
Subeconomic Resources	Extra-heavy oil	500
	Bitumen	1500

Source: Masters et al. (1986); data from Meyer et al. (1984).

oxidized to CO_2. Some of this gas is vented or flared, resulting in CO_2 emissions. However, flared gas does not appear in the marketed gas statistics (Section 4.2.2). This complicates matters and necessitates the separate consideration of flared gas (e.g., in Table 4.4.)

Recoverable gas resources include gas that might be flared or reinjected. One path (flaring) will directly produce CO_2 emissions, but the other will not. Although flared gas must be considered in current gas production data and CO_2 emissions (Section 4.2), this distinction is not particularly important in resource analysis because other factors in resource estimates introduce much greater error.

The data in Table 4.10 were based largely on the data of Parent (1983) and WEC (1980). These works include sets of data that are not totally independent, but the values were analyzed by independent experts using different methodologies.

4.3.4 Resource Constraints on Future Emissions

Because of the very large coal deposits in the United States, the People's Republic of China, and the U.S.S.R., the fossil fuel resources of the world that are classed as recoverable are so vast that their exploitation could cause a severalfold increase in the atmospheric CO_2 concentration. The rate of use as well as the total amount of fossil fuel used will play a major role in determining the increase in atmospheric CO_2 concentration.

Although fossil fuel resources are vast, their concentration in a few localities can, and likely will, result in political constraints on some fossil fuel use, at least in the short term. With free global exchange there would be no practical physical resource constraints on the emissions of CO_2 from combustion of solid fossil fuels. Recoverable resources of conventional oil and gas are sufficiently limited that it is almost certain that their use as a fuel will decline through most of the 21st century; however, the global coal resources are sufficient to ensure an almost unlimited supply, especially if some improvements in technologies for extraction and utilization are assumed. Some technologies now exist for conversion of the solid fossil fuels to liquids and gases and, even without much more technological refinement, their application could ensure that fossil fuel utilization will not be limited by the supply.

However, as the market supply of conventional oil and gas decreases and if extraction and production costs

Table 4.10
Estimate of Recoverable Resources (in 10^{12} m^3) of World Conventional Natural Gas

Region	Cumulative Production	Proved Reserves	Undiscovered Resources	Estimated Total Recoverable
North America	20.0	8.4	28-36	36-44
South America	3.4	4.9-5.2	7.7-10.8	12.6-16
Western Europe	2.2	5.4-5.6	4.2-6.0	9.6-12
U.S.S.R. and E. Europe	5.9	15.5	13-25	28-40
Africa	1.0	5.7-6.3	7.5-28	13-34
Middle East	2.5	17-24	30-37	47-61
Asia/Oceania	0.5	4.5-6.0	29-36	33-42
Antarctica	0	0	2.3-4.0	2.3-4
WORLD TOTAL	35.5	61-71	122-183	182-253
(in 10^{15}g C)	(19.2)	(33-38)	(66-99)	(99-137)

Source: Adapted from WEC (1980) and Parent (1983).

of large amounts of unconventional oil and gas remain relatively high, nonfossil fuels are likely to assume a larger portion of the total energy supply picture. Thus, the main constraints on the future use of fossil fuels, at least during most of the 21st century, will be economic and technological (Chapter 9). The cost of obtaining suitable energy supplies using new technologies will be a major factor in determining how much of the total fossil fuel resource will be exploited.

If fossil fuel costs through the next century are high, then the global economy will have difficulty sustaining rapid growth, and less fuel will be used. If environmentally acceptable alternative fuel systems can be developed to produce energy at lower costs than those using conventional fossil fuels, then the alternatives will displace fossil fuels, at least partially, and less fossil fuel will be used. Energy-economic models address the issue of comparative fuel costs (see Chapter 9), but making effective estimates of the costs of fuel from new technologies is even more difficult than relating the energy supply to the economy.

The full costs of alternative energy systems are just as important as the rate of global economic growth in determining the constraints on future fossil fuel use. Unless other ways of accomplishing the tasks needed by society are found (i.e., using less energy), the problem of CO_2 emissions from fossil fuels will have a circular element because the global economy determines fuel use, but the economy is constrained to a large degree by the cost of available energy. The cost of fossil fuels relative to the cost of alternative systems is central in obtaining a full picture of future CO_2 emissions (Chapter 9).

4.4 SUMMARY AND RESEARCH RECOMMENDATIONS

4.4.1 Need to Account, Anticipate, and React

Even though fossil fuel combustion may not be the only source of CO_2 contributing to the observed increase in atmospheric CO_2 concentration, it is certainly the major one at present (Chapter 6). Because of the close link between burning fossil fuels and increasing atmospheric CO_2 (Chapters 3 and 10), we must be continually sensitive to this cause-and-effect relationship and to the possible feedbacks involved. Information about fossil fuel use is important to the CO_2 problem not only because fossil fuel combustion is (and will be) the primary cause of the CO_2 increase in the atmosphere, but also because information about fossil fuel use can serve as basic input in studies of the carbon cycle and the interactions of humans and their total environment.

Just as the CO_2 observational records (Chapter 3) have documented the increases in atmospheric CO_2 content, studies of the fossil fuel sources are essential to account for the carbon being exchanged among the reservoirs of the global carbon cycle. The accounting process requires accurate information for a long period of time for both CO_2 in the atmosphere and CO_2 sources disturbing the natural cycle.

The second way that a greater understanding of fossil fuel use can contribute to the overall assessment of the CO_2 issue is in the need to anticipate what is likely to happen (or can happen) under a given set of circumstances. Information about the fossil fuel resources and the rates of extraction and potential use is required to estimate future atmospheric CO_2 concentrations. The extent to which improved efficiency

in the use of energy can effectively reduce the demand for energy, and for fossil fuels in particular, must be evaluated. Trade-offs among fossil fuels, other energy sources, and ways to reduce energy use will occur as both technologies and economics change in unpredictable ways. In assessing future atmospheric CO_2 concentrations, some of these changes can be anticipated even though the details may be clouded by uncertainty (Chapter 9).

The third element is in the area of information about how society can or will react through modifications of the global energy system. Will the reactions be productive, spontaneous, ineffective, counterproductive? What is possible, or probable? As a result of economic growth and the introduction of new technologies, society is offered opportunities to react in ways that can either increase or decrease dependence on fossil fuels. However, if society determines that a move away from dependence on fossil fuels is desirable or necessary, the opportunities to react will be restricted by both economics and the availability of suitable alternative technologies.

4.4.2 Specific Needs and Recommendations

1. Procedures to estimate the CO_2 emissions from world fuel data are well established, and efforts to substantially improve the accuracy of the estimates have such a low probability of success that the cost would be difficult to justify. Improving the reliability of the fuel data in those parts of the world for which the U.N. Statistical Office receives limited information is an exception. Because the U.S.S.R. and the People's Republic of China are second and third, respectively, in the world in fuel use, narrowing the uncertainty in data on both the fuel quantity and fuel quality from these nations would increase confidence in the estimates of total CO_2 emissions.

The current time series of estimates of global CO_2 emissions from fossil fuels (having an estimated accuracy of $\pm 10\%$) should be maintained and updated each year. Not only should the most recent year's estimate be added to the series, but the entire U.N. fuel data sequence should be examined for revisions and new CO_2 estimates should be made when required.

2. Fuel consumption data should be disaggregated both geographically and seasonally. This would allow atmospheric CO_2 observations at different times and places to be correlated with CO_2 releases; analyses of these relationships could lead to better understanding of the carbon cycle (Chapters 3, 7, 10, and 11). Current efforts in this direction should be expanded.

3. Grasses, tree branches, and a variety of other biotic materials have been used as fuels for centuries. Much of this material is natural waste, subject to rapid decay, and thus its use as a fuel does not affect the carbon cycle. However, such biotic materials constitute an important energy source, and, because the rate of energy extraction from this source is limited, additional energy demands from people now using this source will be supplied by fossil or other commercial fuel sources. Efforts should be made to obtain better estimates of the amount of energy derived annually from biospheric wastes and the percentage of the human population dependent on this source of energy.

4. During the past half century, the use of wood as a commercial fuel has declined in favor of the readily available fossil fuels. Most recently, increases in fossil fuel prices have caused a resurgence in the use of wood as a fuel, including the development of fuelwood farms. A time series of fuelwood use should be developed, not so much as a CO_2 source, but rather as a guide for its potential as a substitute for fossil fuels. An investigation to determine the amount of wood that could be harvested each year on a sustainable basis would be helpful in determining the potential for wood as an energy source in the long term.

5. The availability, geographic distribution, and estimated costs of a wide variety of nonfossil energy resources must be taken into account in analyzing future energy systems. The geographic partitioning of some nonfossil energy resources among world regions is much more even than that of fossil fuels. How are these differences in distribution of energy resources likely to affect future energy systems? How much biomass energy production can be sustained? How much of the vast amount of energy in the wind can be used effectively? How are the relative costs of nonfossil versus fossil energy production likely to vary among different world regions? Improved understanding of the future energy system will depend, in part, on obtaining better answers to such questions.

6. Among the tools for estimating the quantities of fossil fuel used and the rates of change of use in the future are energy-economic models. A review of energy-economic models and related projections of CO_2 emissions from fossil fuel combustion for the next century obtained from these models is the subject of Chapter 9. Also covered in Chapter 9 are (a) a quantitative assessment of uncertainties associated with these projections, including identification of principal contributors to uncertainty, and (b) recommendations for research to improve our capability for understanding how we can deal with these uncertainties.

7. Research is needed to improve understanding of the systems on which the models are built. Research into the dependence of economic growth and development on energy demand is needed, especially into how the dependence changes as the process of industrial development proceeds. Energy is required in meeting the basic human needs of food, water, and shelter, but individuals make trade-offs among sanitation, transportation, and a variety of conveniences and comforts. With continuing population growth in the developing areas of the world, meeting basic human needs will be the highest energy priority. It is especially important to study the influence of energy prices on economies in rigidly controlled political environments with rapidly growing populations, that is, different in characteristics and responses than OECD nations. Thus, research should consider a number of world regions analogous to, but not necessarily identical to, the nine regions of the Edmonds and Reilly (1985) energy-economic model. This effort could establish further limits on acceptable future energy systems, especially with respect to energy conservation.

8. An assessment of future energy technologies is of utmost importance. The increase in the percentage of world energy used as electricity requires that realistic scenarios consider the physical factors in energy transfers and substitutions, with distinctions between primary energy supply systems, energy carriers, and types and purposes of energy consumption. Technological developments are occurring at an unprecedented rate and many of the developments will have profound impacts on the global energy system of the next 50 to 100 years. Not only new supply technologies, but also new energy transformation and energy-use technologies will affect the total quantities of energy used and CO_2 released. Energy storage, fuel extraction, solar energy, and advanced, safer nuclear reactors are but a few examples of rapidly changing technologies that must be examined in order to develop realistic scenarios for future energy demand.

9. With the exception of energy conservation, energy systems that can be effective in reducing CO_2 releases on a global scale are capital intensive. Nuclear reactor plants are expensive, and solar cells are now and are likely to continue to be expensive, although neither of these is likely to be so expensive as to prohibit its use. An attempt to reduce CO_2 emissions by rapidly constructing nonfossil, capital-intensive energy systems could actually increase the use of fossil fuels for a few decades (e.g., Perry et al. 1982). Careful study is needed to determine the rates at which shifts to other energy supply systems would be ineffective or even counterproductive.

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5. UPTAKE OF CARBON DIOXIDE BY THE OCEANS

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5.1 THE OCEANS IN THE CARBON CYCLE

The oceans of the world contain about 50 times the amount of carbon currently in the atmosphere in the form of carbon dioxide (CO_2). All but a small fraction of this carbon is dissolved in the form of the bicarbonate ion, an accumulated product of the reaction of CO_2 with basic substances of the Earth's crust. This carbon pool dwarfs the other rapidly exchanging pools in the terrestrial carbon cycle as well as the reserves of fossil carbon that can be recovered economically. The capacity of the oceans to take up additional CO_2 is so great that without doubt it is the ultimate sink for the majority of the fossil carbon that will be released by mankind. Therefore, the principal question to be addressed in this chapter is not whether but how rapidly this uptake can occur.

The need to answer this crucial question has inspired a growing number of investigations over the past 2 decades. Given the intricacies of ocean topography, circulation, chemistry, and biology, credible estimates of the response of the world's oceans to increasing atmospheric CO_2 can come only from models. Just how detailed such models must be before their predictions are reliable remains to be determined, however. This chapter will consider first the state of the knowledge of various features and processes in the ocean, which ultimately may be incorporated into such models, then, briefly, the state of the development of the models themselves, and finally the additional observational data and model development that may be needed before the future uptake of CO_2 by the oceans can be predicted reliably. A more detailed treatment of carbon cycle modeling is provided in Chapter 7 (all chapter references are to this volume), and the prehistoric role of the oceans in the carbon cycle is treated in Chapter 8.

5.1.1 Ocean Circulation

Approximately three-quarters of the Earth's surface is covered by the oceans to an average depth of about 4000 m, equal to only 1/1600 of the Earth's radius. This thin layer of water is highly structured, however, because of uneven heating at its surface. From the surface downward, the waters are well mixed to an average depth of only about 75 m. This mixed layer is about 30 m deep in the tropics but deepens with increasing latitude. The surface temperature decreases from 30°C near the equator to -1.5°C at the highest latitudes. Because colder water has a higher density, it tends to circulate downward and toward the Equator. This movement establishes a stratified structure in which successively deeper layers have decreasing temperatures, increasing densities, and reach the surface (outcrop) at higher and higher latitudes (Figure 5.1). This plausible representation of ocean structure is greatly oversimplified, however.

The actual pattern of these so-called isopycnal surfaces in the ocean is much more complex for several reasons. (1) The density of water also depends on salinity, which is affected by evaporation and precipitation at the surface and by freezing and thawing in polar regions. (2) The circulation of water is strongly affected by the ocean's boundaries and by the Earth's rotation. (3) The conditions at the surface that drive ocean circulation—wind, temperature change, evaporation, precipitation, thawing, and freezing—can vary greatly with the seasons and location.

The gross features of deep-ocean circulation have long been known. The densest water is produced near the surface at high latitudes in winter, mostly in the North Atlantic and around Antarctica. While the Atlantic Ocean is thus supplied from both north and south with new bottom water, the Indian and Pacific oceans are supplied only from the south by deep water from the easterly Antarctic Circumpolar Current. There is a resultant net flow of deep water, usually below 3000 m, from the Atlantic to the Indian and to the Pacific Oceans. In each ocean basin, overlying water masses that left the surface at lower latitudes can be traced over long distances as they mix with adjacent water masses. These waters are in turn overlaid by the so-called thermocline, through which the temperature rises from about 5°C toward the temperature of the surface waters. The lower boundary of the thermocline is about 1000 m deep in the tropics and rises with latitude.

The traditional view has been that circulation of the ocean below the thermocline is largely "thermohaline," that is, driven by density differences arising from changes in temperature and salinity and involving

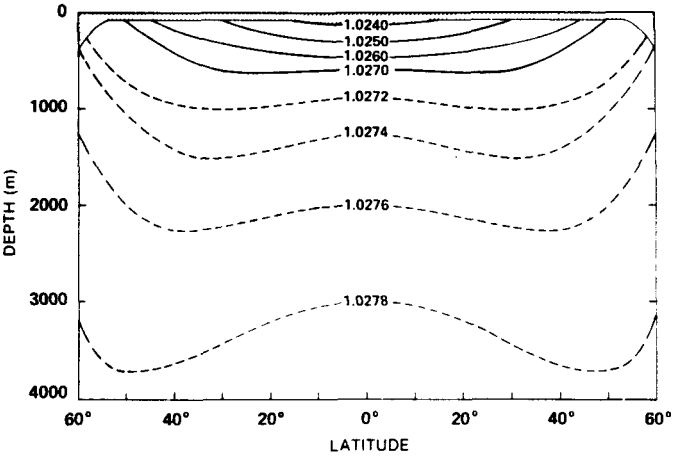


Figure 5.1. An idealized representation of isopycnal (constant-density) surfaces in a north-south section of the ocean. The vertical scale is exaggerated by a factor of about 2200.

the movement (advection) of water along isopycnal surfaces. Circulation in the upper ocean is thought to be largely driven by the wind and by uneven heating. More recent studies suggest, however, that circulation in the upper and lower ocean is more closely coupled than previously thought. In each ocean basin there are great anticyclonic motions (gyres) thousands of kilometers in diameter, which, while weakening with depth, may extend from the surface to great depths (Worthington 1976). Until recently, these and other ocean currents were thought to be fairly uniform, predictable, and modulated only by seasonal and longer-term climatic changes. Within the last 2 decades, however, so-called mesoscale eddies were discovered and studied. These eddies are rotating water masses on the order of 100 km in diameter that separate from the edges of the ocean currents associated with gyres and move slowly about for months or years before they dissipate. So many other smaller-scale, persistent features of circulation have since been identified that, in a recent review, Rossby (1982) has referred to them as "omnipresent, often very energetic fluctuations of ocean currents that occur over a wide spectrum of length and time scales. The variety of oceanic motion that falls under the general rubric of eddy motion is extraordinary. Indeed, we now wonder whether the presumption of a steady-state circulation is valid or merely reflects our inability to understand the very long time-scale secular changes of the oceans" (p. 137).

In the same publication (Brewer 1982), several other investigators (Munk 1982; Niiler 1982; Turner 1982; Whitehead 1982) reviewed the present state of observational knowledge of ocean circulation. With the many techniques of measurement now available (including free-floating buoys; submerged, neutrally buoyant, acoustic-ranging floats; current meters; and satellite remote sensing), this knowledge is growing rapidly; however, a coherent representation of ocean circulation based on direct observation has not yet emerged.

The gross features of ocean circulation can be simulated with three-dimensional hydrodynamic models that incorporate topography and functions for wind, heat, evaporation, and precipitation to drive circulation (Sarkisyan 1977; Bryan and Lewis 1979). Promising simulations have also been done by the diagnostic method, that is, directly from the observed gradients of temperature, salinity, and density in the oceans (Wunsch 1978; Sarmiento and Bryan 1982). However, three-dimensional models have not yet been applied with success to the uptake of excess CO_2.

Surface waters exchange about one-seventh of their carbon with the atmosphere each year (Section 5.3.3), but the capacity of this relatively small, well-mixed volume to take up additional CO_2 is quite limited. It is the transfer of excess CO_2 to the deep oceans by water flow (advection) and mixing along isopycnal surfaces, and by diffusive mixing across isopycnals that appears to be most important in determining the short-term response of the oceans to increasing atmospheric CO_2 concentrations. The details of ocean circulation that will determine this transfer rate, the overturning of water to great depths at high latitudes in winter and the advection and mixing of water in the upper ocean at lower latitudes, remain to be described adequately. Sarmiento (1986) reviewed the state of development of three-dimensional models of ocean circulation with respect to the problem of the uptake of CO_2 (Section 5.7.2).

5.1.2 Ocean Composition

Despite their stratification, the oceans of the world are surprisingly uniform in composition. Although variations in the salinity of the water provide an important means of identifying water masses and determining the nature of circulation, these variations are generally small. The average salinity is about 34.7 g/kg, and in the open ocean deviations rarely exceed 1% except near the surface. The mole ratios to sodium of the more important minor constituents—magnesium (0.114), sulfate (0.0600), calcium (0.0218), potassium (0.0211), bromide (0.00177), and boric acid (0.00093)—vary even less. The concentrations of inorganic carbon, oxygen, and the nutrient elements (nitrogen, phosphorus, and silicon) are more variable because of the biological processes.

5.1.3 Processes Affecting the Distribution of Carbon

As noted above, relatively rapid exchange of CO_2 occurs between the surface waters and the atmosphere. Although the pressure of CO_2 generated by waters can differ from the local atmospheric partial pressure by more than 100 μatm[1] in either direction (Takahashi et al. 1983), the area-weighted average of the absolute difference is much smaller, about 16 μatm. Generally, the seawater pressure is lower in polar regions, where cooling and downwelling occur, and higher in the equatorial regions, where upwelling and warming occur (Figure 5.2). This partial pressure is determined to a great extent by the abundance of the various forms of carbon in these waters. This in turn is

[1] Humidity, temperature, and pressure correction are needed to convert these values to atmospheric CO_2 ppm-equivalents; typically, values change by a few percentage points.

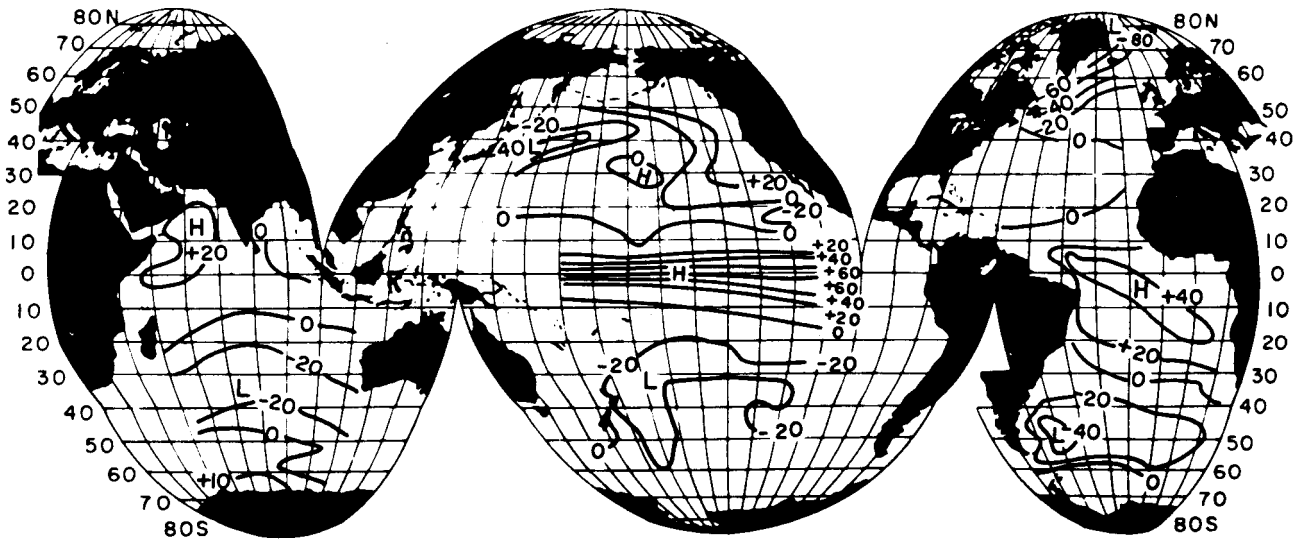


Figure 5.2. Global distribution of the difference [pCO_2 (surface waters)-pCO_2 (atmosphere)] in microatmospheres observed during the GEOSECS Program (from Takahashi and Azevedo 1982). The data represent summer conditions in both hemispheres.

determined by the biological processes of photosynthesis and decay, the biologically mediated precipitation of calcium carbonate, and water circulation. All but the last process (and sometimes it as well) act to produce a nonuniform distribution of carbon in the oceans.

Because there is much more carbon in the ocean than in the atmosphere, it is apparent that the amount of CO_2 in the atmosphere may be controlled to a considerable extent by the oceans. Equally apparent is the need to understand and then model the processes in the ocean that affect the partial pressure of CO_2 (pCO_2) in the surface waters.

5.2 THE DISTRIBUTION OF CO_2 BETWEEN OCEAN AND ATMOSPHERE

The state of knowledge of the processes that affect the distribution of carbon within the ocean and hence between the ocean and atmosphere will be considered in this section.

5.2.1 The Inorganic Chemistry of Carbon in the Oceans

Carbon dioxide is not an inert gas. Its reaction with water is usually represented by

$$CO_2(g) + H_2O \rightleftharpoons H_2CO_3(aq), \qquad (5.1)$$

where g denotes the gaseous state and aq the aqueous and H_2CO_3 denotes the total of dissolved CO_2 and molecular H_2CO_3 (referred to collectively as carbonic acid). Carbonic acid can dissociate in two steps to form first the bicarbonate ion (HCO_3^-),

$$H_2CO_3\,(aq) \rightleftharpoons H^+ + HCO_3^-, \qquad (5.2)$$

and then the carbonate ion (CO_3^{2-}),

$$HCO_3^- \rightleftharpoons H^+ + CO_3^{2-}. \qquad (5.3)$$

In pure water, these ionization reactions occur to only a limited extent. In natural waters over geologic time, however, basic minerals have consumed the hydrogen ion produced by these reactions, and bicarbonate and carbonate have accumulated in the oceans such that at present approximately 90% of the carbon in the oceans is in the form of bicarbonate ion and most of the remainder is in the form of carbonate ion.

These reactions of CO_2 are rapid and local equilibrium within a small volume of seawater can be assumed. This condition can be represented quantitatively by the so-called equilibrium constants in Table 5.1, in which one quotient of concentrations is defined by each of the three reactions given above. In addition, there is a mass-balance equation that defines the total dissolved inorganic carbon (TC in Table 5.1) and a charge-balance equation that defines the alkalinity (TA in Table 5.1), both of which are accurately measurable quantities. Appropriate substitution of these equilibrium constants (along with others for minor reactions affecting TA) yields expressions for TC and TA that involve only two other measurable quantities: the pCO_2 in the water sample and the hydrogen ion concentration [H^+]. It is possible, therefore, to calculate any two of the four quantities TC, TA, pCO_2, and [H^+] from the remaining two by these two equations. The equilibrium constant parameters needed for this calculation are in fact constant only if the salinity, pressure, and temperature are constant. Accordingly,

Table 5.1
The Inorganic Chemistry of Carbon in the Oceans

Equilibrium Constants	Value[a]
$K_1 = [H_2CO_3]/pCO_2$	3.746×10^{-2} mol kg^{-1} atm^{-1}
$K_2 = [HCO_3^-]\,[H^+]/[H_2CO_3]$	1.106×10^{-6} mol/kg
$K_3 = [CO_3^{2-}]\,[H^+]/[HCO_3^-]$	7.816×10^{-10} mol/kg
Total Dissolved Inorganic Carbon and Alkalinity	
$TC = [H_2CO_3] + [HCO_3^-] + [CO_3^{2-}]$	
$TA = [HCO_3^-] + 2[CO_3^{2-}] + \Sigma\,[J]$	
Relationships of TC, TA, pCO_2, and $[H^+]$	
$TC = K_1\,pCO_2\,(1 + K_2/[H^+] + K_2K_3[H^+]^2)$	
$TA = K_1K_2\,pCO_2/[H^+]\,(1 + 2\,K_3/[H^+]) + F(M_j, K_j, [H^+])$	

Note: Brackets denote the concentration of a species in the ocean, including minor ones (j) that can react with acid. pCO_2 denotes the partial pressure of CO_2. K denotes an equilibrium constant (K_1, K_2, and K_3 for reactions 5.1, 5.2, and 5.3, respectively, in the text, and K_j for reactions of the minor species). With all the K values known as a function of temperature, pressure, and salinity, and the concentrations (M_j) of the minor substances known, then only two of the four measurable quantities TC, TA, pCO_2, and $[H^+]$ are independently variable (Figure 5.2).

[a]At 15°C, 35 g/kg salinity, 1 atm. From Almgren et al. (1975) and Weiss (1974).

their values have been measured carefully as a function of these variables (Hansson 1973; Almgren et al. 1975; Weiss 1974). It is customary to measure TC and TA in weight units (μmol of carbon/kg and μmol of alkalinity/kg, respectively) and to calculate pCO_2 and pH (-log $[H^+]$) for the temperature, pressure, and salinity of the water sample.

The accuracy with which pCO_2, TC, and TA can now be measured is quite impressive. Uncertainties have been reduced to about ± 3 μatm (in surface waters), ± 5 μmol C/kg, and ± 5 μmol alkalinity/kg, respectively. The uncertainties in TC and TA produce an uncertainty in the calculated value of pCO_2 of about 10 μatm. Brewer et al. (1986) recently described the state of the art of the titrimetric procedures involved in the determination of TA and TC. The principal uncertainty that remains seems to be related to the problem of preserving the samples before they are analyzed. Although different sets of equilibrium constants are in use, they differ mainly in the way that the pH scale is defined. When used consistently, they yield pCO_2 values that agree well with observed values.

It should be clear from the above that the tendency of a volume of seawater to lose or take up CO_2 is ascertainable from measured values of TC and TA and from its temperature, pressure, and salinity. The relationships involved are illustrated in Figure 5.3, in which curves of constant pH and pCO_2 are shown on a plot of TA versus TC for surface water at 15°C and a salinity of 35 g/kg. The pCO_2 increases approximately exponentially as TC increases and decreases in a similar fashion as TA increases (Figure 5.4). A given fractional change in pCO_2 produces a much smaller fractional change in TC. For average surface water, the ratio of small fractional changes, the so-called buffer (or Revelle) factor,

$$\beta = \frac{\delta \ln pCO_2}{\delta \ln TC}, \tag{5.4}$$

is about 10. The value of this derivative increases as the carbon content of seawater (TC) increases, and thus the capacity of seawater to take up additional CO_2 decreases. This effect is evident from the curves shown in Figure 5.4.

The buffer factor (β) has been used widely in models to represent the chemical processes that determine the equilibrium distribution of CO_2 between surface ocean waters and the atmosphere. Because it varies considerably with TA and temperature as well as with TC (Takahashi 1979), its usefulness for this purpose is limited. Fortunately, the direct calculation of pCO_2 from TC, TA, and temperature is relatively trivial compared with the other calculations necessary in ocean models.

Lowering the temperature of seawater lowers pCO_2. At 15°C

$$\frac{\delta \ln pCO_2}{\delta t} = 0.042/°C, \tag{5.5}$$

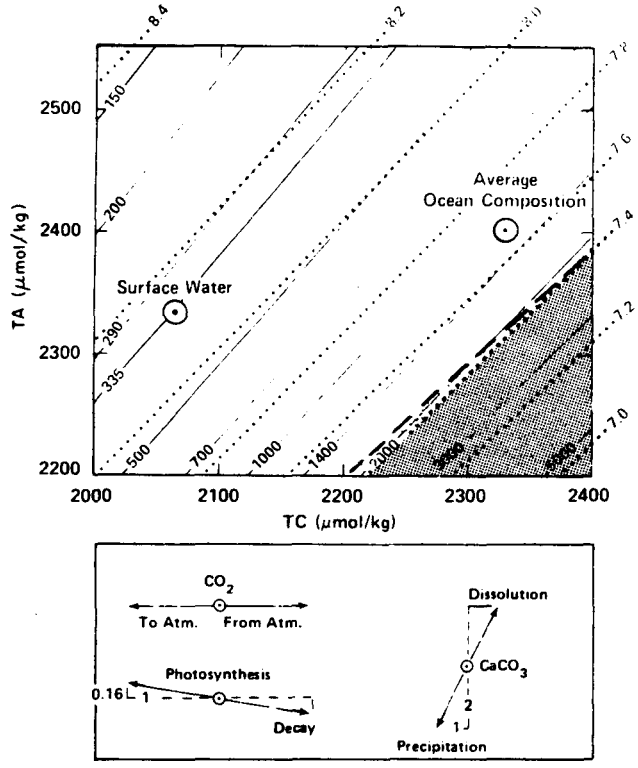


Figure 5.3. Composition diagram of ocean surface waters at 15°C and 35 g/kg salinity. The solid curves are compositions with the same value of pCO_2 (μatm). The dotted curves are compositions with the same pH. The shaded region is undersaturated and the unshaded region is oversaturated with calcium carbonate (calcite). Surface water and average ocean compositions are indicated by the large circled-dot symbols on the upper graph. The vector diagrams denote the direction of composition changes accompanying the indicated processes (from Baes 1981).

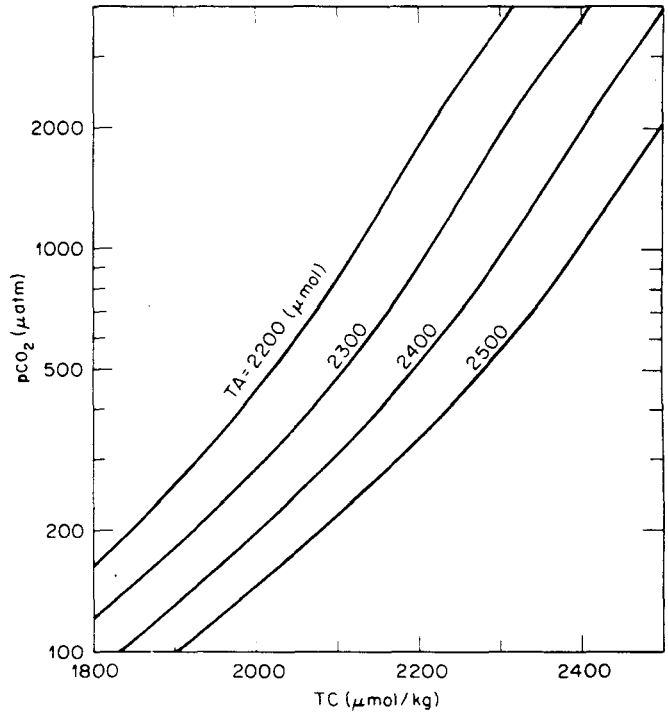


Figure 5.4. The dependence of pCO_2 on TC at specified values of TA in seawater (35 g/kg salinity, 15°C, and atmospheric pressure).

that is, pCO_2 in surface water changes about 4%/°C provided that TC and TA are held constant. MacIntyre (1978) pointed out, however, that because the amount of carbon in the well-mixed surface waters is comparable to that in the atmosphere, the transfer of CO_2 between ocean and atmosphere can cause an appreciable change in TC, and this reduces the temperature dependence. If isolated, well-mixed volumes of seawater and atmosphere could be equilibrated, this transfer of CO_2 would reduce the temperature dependence of pCO_2 by the factor $(1 + \beta/R)$, where R is the ratio of the amount of carbon in the water to that in the atmosphere (Bacastow 1979; Baes 1981). For a column with a constant cross section extending from the bottom of the mixed layer (about 75 m deep) to the top of the atmosphere, R is about unity. Because β is about 10, the temperature dependence of pCO_2 for this isolated system would be about 0.4%/°C.

The actual fluctuation of pCO_2 in surface waters has in fact been found by Takahashi et al. (1983) to be as high as 4%/°C for the colder waters of the North Atlantic and about 1%/°C for the warmer waters (Figure 5.5). There is considerable scatter in the data, doubtless the result of processes other than the transfer of CO_2 between ocean and atmosphere that can change TC and TA in surface waters. The high temperature dependence in cold water might be caused by temperature changes that are rapid compared with the rate of all other processes that can alter TC and TA.

As noted earlier in this chapter, the carbon composition of the oceans is largely the result of the reaction of the acidic CO_2 liberated by volcanic activity with basic substances of the Earth's crust. One product of this reaction is calcium carbonate ($CaCO_3$), an important constituent of sediments in much of the ocean. It is precipitated from surface waters in two forms, calcite and the more soluble aragonite, almost entirely by organisms that fabricate their shells and skeletons from these materials. The saturation equilibrium for either form may be written:

$$CaCO_3(c) \rightleftharpoons Ca^{2+} + CO_3^{2-}, \qquad (5.6)$$

where c denotes the crystalline state and equilibrium constants are defined as the product of ion concentrations, $K = [Ca^{2+}][CO_3^{2-}]$, for calcite and aragonite.

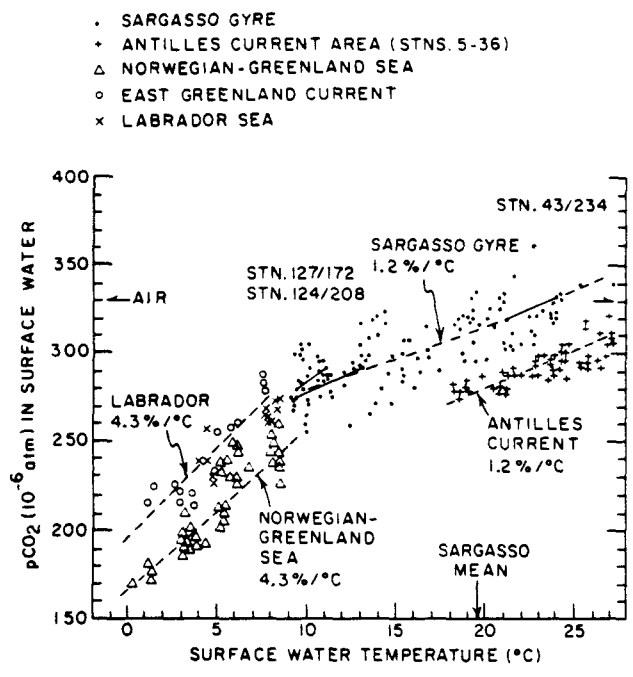


Figure 5.5. pCO_2 in various surface waters in the North Atlantic versus temperature (from Takahashi et al. 1983).

These values, like those in Table 5.1, are known quite accurately as a function of salinity, temperature, and pressure (Ingle 1975). They do not vary greatly with temperature but increase fairly rapidly with pressure. Because of this, both forms become increasingly soluble with depth and are largely redissolved in the deep ocean. The circulation of the oceans recycles the dissolved material, and, as a consequence of these processes, the middle depths of the ocean are near saturation with respect to calcite, the upper ocean is supersaturated (Figure 5.3), and the deeper ocean is undersaturated.

5.2.2 The Biological Chemistry of Carbon in the Oceans

One of the most striking features of TC and TA values in the oceans is that they are generally lower near the surface than in deep water. Although the decreases are small, the effect on pCO_2 is substantial. If deep water of the typical composition shown in Figure 5.3 is brought to the surface and warmed to 15°C, the pCO_2 generated (about 1200 μatm) would be several times that in the atmosphere. If the surface water deficiency in TC and TA were absent, the pCO_2 in the atmosphere would be substantially increased. (Although the increase in TA would tend to lower pCO_2, this effect is more than overcome by the increase in TC.)

The biological processes are the most responsible for these deficiencies. This can be understood from an idealized representation of marine photosynthesis (Redfield et al. 1963):

$$106CO_2 + 16NO_3^- + H_2PO_4^- + 17H^+ + 122H_2O \rightarrow (CH_2O)^{106}(NH_3)^{16}(H_3PO_4) + 138O_2(g), \quad (5.7)$$

and the other biologically mediated process, the precipitation of calcium carbonate (the reverse of the reaction in Equation [5.6]). In the case of photo synthesis, CO_2 is withdrawn from the water, which lowers TC. Although there is a rise in alkalinity (TA) owing to the simultaneous consumption of H^+ ions, this is more than overcome by the loss of dissolved carbonate to the formation of calcium carbonate, which lowers both TA and TC. The manner in which these reactions can alter the TA, TC, pCO_2, and pH values of surface waters is indicated by the vector diagrams shown at the bottom of Figure 5.3.

The carbon and alkalinity thus taken out of the inorganic carbon pool falls from the surface water as detritus. Most of the organic detritus is decomposed by the reverse of the reaction shown in Equation (5.7) in the first 1000 m (Suess 1980), returning inorganic carbon (increasing TC) and also the nutrient elements nitrogen and phosphorus (as nitrate and phosphate ions) to the waters in this region. The inorganic (carbonate) detritus is redissolved by the reaction shown in Equation (5.6) at considerably greater depths, first the aragonite and then the calcite forms, thus delivering carbon and alkalinity to the deeper ocean.

The rate at which organic carbon and calcium carbonate are exported from the surface waters is probably determined in most locations by the rate at which the nutrient elements nitrogen and phosphorus are delivered to the surface by upward circulation of deep water. The fact that these elements are often rate limiting is clearly indicated by the low concentrations of nitrate and phosphate found in temperate and tropical surface waters, except in regions of intense upwelling. At higher latitudes, their concentration increases, suggesting that some other condition, probably light supply (Raymont 1963), limits biological productivity.

Peterson (1981) estimated the global net rate of export of organic carbon from the surface waters conservatively at about 2×10^{15} to 3×10^{15} g C/year from a simple two-box model representing ocean circulation and the rate at which nutrients are supplied to the surface waters. A similar result of 4×10^{15} g C/year was obtained from a model described by Broecker and Peng (1982, p. 531). Bolin et al. (1983), using a multibox model to represent the world ocean, obtained 5×10^{15} to 6×10^{15} g C/year. These estimates are 10-20% of the net primary production (NPP) of

organic carbon: estimated at 16×10^{15} g/year by Ryther (1969); 23×10^{15} g/year by Koblentz-Mishke et al. (1970); and as much as 43×10^{15} g/year by De Vooys (1979). This percentage range is consistent with a correlation suggested by Suess (1980) to represent the downward flux of detritus as a function of depth. Thus, organic carbon is recycled many times in the surface water before it falls out as organic detritus.

5.2.3 Gross Features of Carbon Distribution

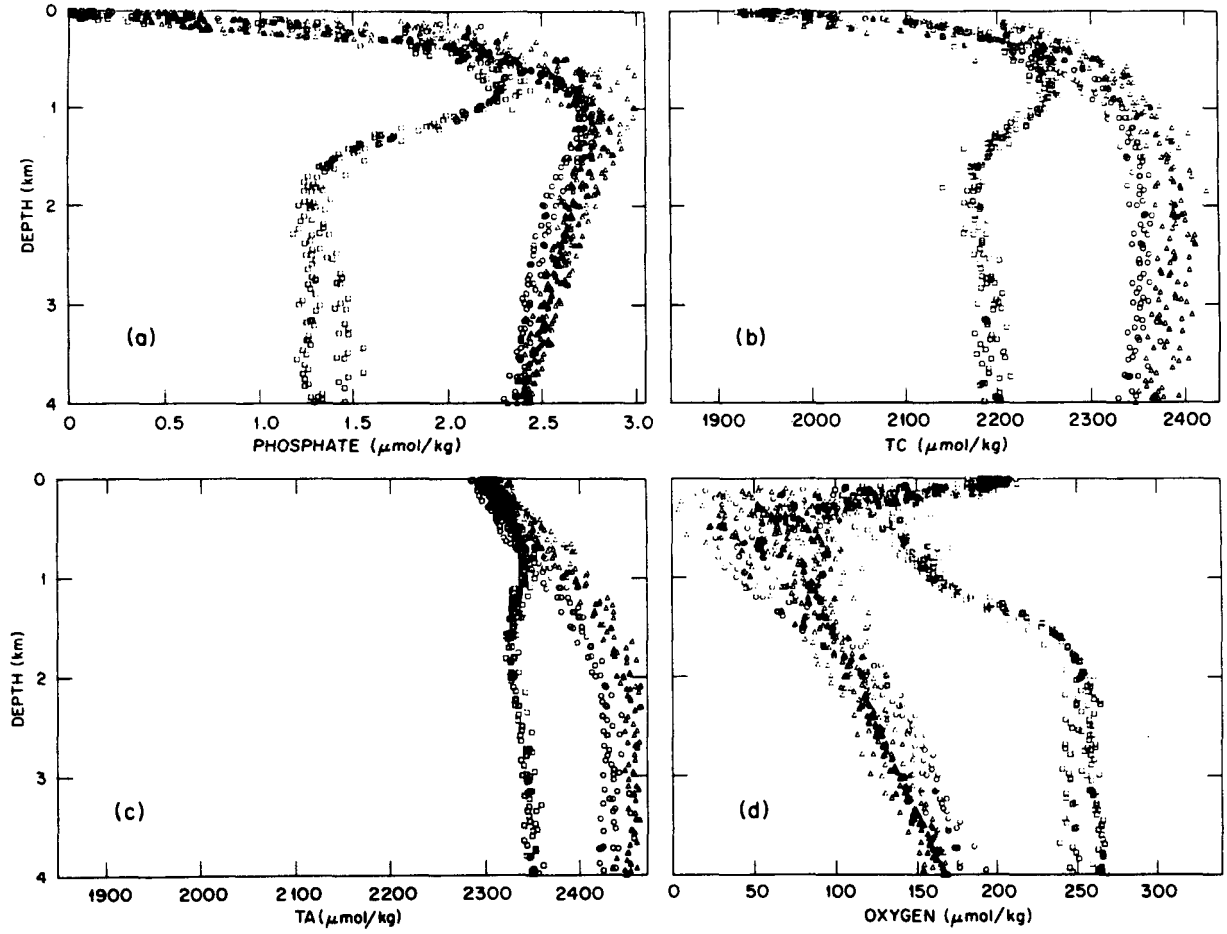
From the foregoing discussion, it is understandable why the surface waters are lower in TC and TA than are the deeper regions of the ocean. The profiles of these quantities, as well as dissolved nutrients and oxygen, show other interesting features as well (Figure 5.6). Most striking is the general increase of TC, TA, and nutrient values, and the decrease of dissolved oxygen, in the deep waters of the Indian and Pacific Oceans compared with those in the Atlantic. These differences have long been recognized to be the result of the general movement of deep water from the Atlantic. As deep water proceeds to the other oceans, it delivers nutrients to the surface by upward circulation and collects carbon, alkalinity, and nutrients from the fallout of detritus along the way.

The distribution of carbon, as well as nutrients and oxygen, in the oceans is thus the result of biological processes and water circulation. Because this distribution has a strong influence on the amount of CO_2 in the atmosphere, these processes must be modeled well enough to at least approximate the observed profiles of these quantities before one can presume to predict reliably the rate at which additional CO_2 may be taken up by the oceans.

5.3 TRACING OCEAN CIRCULATION AND CARBON TRANSPORT

Much of what is known of large-scale ocean circulation and the movement of carbon was deduced from

Figure 5.6. Observed depth profiles of (a) phosphate, (b) dissolved inorganic carbon (TC), (c) alkalinity (TA), and (d) oxygen. The points (normalized to 35 g/kg salinity) are from GEOSECS stations within 5° of the Equator in the Atlantic (□), the Indian (○), and the Pacific (△) oceans.



observations of the distribution of physical, chemical, and radiochemical tracers. This knowledge is growing rapidly as results become available from ocean surveys conducted within the last decade. In this section, this rapidly growing data base is outlined and its application to modeling the uptake of CO_2 is indicated.

5.3.1 Tracers in the Ocean

5.3.1.1 *Nonradioactive Tracers*

The traditional tracers of the ocean are salinity and temperature. These properties and internal pressure are routinely measured with a high degree of accuracy (temperature to 0.001°C and salinity to 0.001 g/kg). The in situ density of seawater depends almost entirely on these three quantities, and they are used to establish isopycnal surfaces in the ocean. Salinity and temperature changes (after correcting for the effects of pressure) are used to trace the movement and mixing of water masses as well as the effects of evaporation, precipitation, thawing, and freezing. Contours of temperature, salinity, and density provide a large body of observational data that can be used to test three-dimensional models of ocean circulation.

The chlorofluoromethanes, CCl_3F and CCl_2F_2, are virtually inert gases that have been widely used as refrigerants and aerosol propellants. They were released to the atmosphere at an exponentially increasing rate until the mid-1970s (Gammon et al. 1982). Direct measurements of their atmospheric concentration, which began at about that time, show that they have been increasing at a linear rate since then. With no natural sources and the only other sink thought to be their destruction by free-radical reactions in the stratosphere, these gases appear to be good transient tracers as they penetrate the oceans. Accurate oceanographic measurements were recently published by Gammon et al. (1982), Gammon and Bullister (1982), and Bullister and Weiss (1983).

Other nonradioactive tracers are the constituents of the ocean affected by the biological processes, such as total dissolved inorganic carbon (TC), alkalinity (TA), nitrate, phosphate, silicate, and oxygen. These tracers reflect the effects of circulation on biological productivity and the disposition of the products.

5.3.1.2 *Excess Carbon*

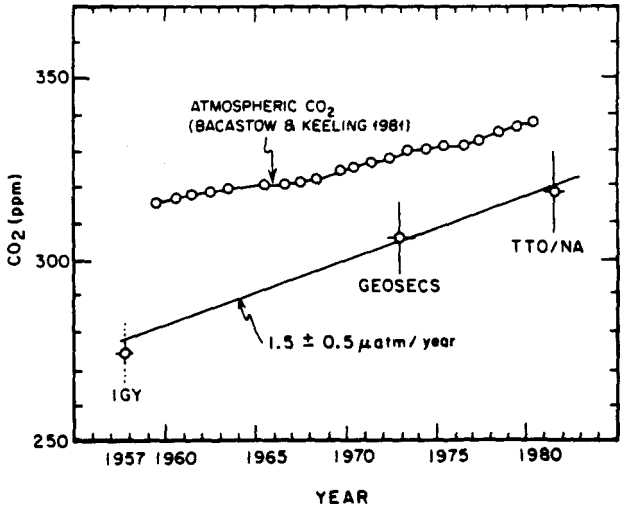
Of great interest is the appearence of excess carbon in the upper ocean resulting from increasing atmospheric CO_2 concentration. Takahashi et al. (1983) observed a long-term increase in pCO_2 in Sargasso Sea surface waters, based on direct measurements during the successive International Geophysical Year (IGY, 1957), Geochemical Ocean Sections (GEOSECS in 1972), and Transient Tracers in the Oceans (TTO in 1981) surveys. Although the best straight-line fit to the data does not exactly parallel the atmospheric increase during the same period (Figure 5.7), the agreement is satisfactory, considering the uncertainties in the measurements and possible changes other than the uptake of CO_2 that could alter pCO_2 in these waters during the 24-year period.

Considerable effort has been expended in an attempt to detect the signature of excess CO_2 in the ocean by correcting TC and TA for the effects of the biological processes that occur in the water between the time of exposure to the atmosphere and the time of sampling at the ocean depths (Brewer 1978; Chen and Millero 1979; Chen 1982; Broecker et al. 1985a). Perhaps the largest uncertainty is the estimation of TC and TA at the time of exposure to the atmosphere (the so-called preformed values), because surface water conditions can change with time as well as location (see Section 5.3.3). The desirability of using excess carbon in the ocean as a tracer has spurred efforts to improve the accuracy of measurement of TC and TA and to refine the methods of isolating this signal from the other processes that alter these quantities (Chen and Poisson 1984). Once the history of atmospheric CO_2 levels is firmly established (e.g., by ice-core studies), the TC distribution in the ocean can be used to constrain ocean models for the uptake of fossil fuel CO_2.

5.3.1.3 *Radioactive Tracers*

Of the radioactive tracers, ^{14}C was the first used to study ocean circulation. It is produced in the upper atmosphere by cosmic ray bombardment and enters the

Figure 5.7. Comparison of mean pCO_2 values in surface waters of the Sargasso Gyre with atmospheric values versus time (from Takahashi et al. 1983).



ocean as $^{14}CO_2$. The profiles of ^{14}C at various locations reveal the increasing age of seawater with depth and distance from the principal sources of deep water in the Atlantic Ocean. Although this tracer is the most useful thus far for the calibration of ocean models for uptake of CO_2 from the atmosphere, the record in the ocean region of greatest interest, the first 1000 m of depth, is complicated by the ^{14}C produced by atmospheric nuclear testing, mostly during the period from 1955 to 1964. Moreover, its radioactive half-life (5730 years) is long compared with the time scale (10 to 100 years) of circulation in the upper ocean. Another complication, discussed by Oeschger et al. (1980), is the changing rate of production of natural ^{14}C caused by the variation of solar activity with time.

Atmospheric nuclear testing also produced tritium (^3H, half-life 12 years), which has entered the oceans as ^3H_2O. Current ocean models are being tested against the distribution of tritium, but because its distribution reflects the strongly advective nature of ocean circulation, multidimensional models are required for a proper representation of the invasion of this tracer into the oceans (Sarmiento 1983). The same is true of other transient tracers, but in the case of ^3H such models are especially needed because the injection occurred for only a few years in the early 1960s, a period too short to permit complete atmospheric mixing. The distribution of stable ^3He from the decay of ^3H in the upper ocean provided additional information about the circulation of water in the thermocline (Jenkins 1980).

The tracer ^{39}Ar offers promise for the study of ocean circulation. Like ^{14}C, it is produced in the atmosphere by cosmic ray bombardment but was not produced in detectable amounts during nuclear testing (Oeschger et al. 1974). Also, its shorter half-life (269 years) is more suitable for providing information on the thermohaline circulation. The use of ^{39}Ar as a tracer has been limited thus far because its determination is difficult, requiring extraction from large volumes (e.g., 2000 L) of seawater to obtain adequate counting rates. Similar problems have limited the use of anthropogenic ^{85}Kr as an oceanic tracer.

The nuclide ^{222}Rn (half-life 3.9 d) is produced by the decay of ^{226}Ra in ocean sediments and in seawater. Radon-222 enters seawater from the sediments, passes through the water column to the surface, and from there to the atmosphere. The resultant distribution of this nuclide was used to give information about both the circulation of water near the ocean-sediment interface and the rate of gas exchange at the atmosphere-ocean interface (Broecker and Peng 1982, 1984).

5.3.2. Ocean Surveys

5.3.2.1 *Geochemical Ocean Sections Study (GEOSECS)*

The most extensive chemical and radiochemical survey of the oceans carried out thus far has been that of the GEOSECS program. The Atlantic Ocean was surveyed first (1972 to 1973), and then the Pacific Ocean (1973 to 1974), and finally the Indian Ocean (1977 to 1978). The number of locations (stations) occupied in each ocean was 121, 147, and 141, respectively, and at each station, samples were taken at intervals (typically 50) throughout the water column. For nearly all the samples taken, pressure, temperature, and salinity were determined, which together gave the depth, potential temperature, and density. Oxygen, silica, phosphate, nitrate, inorganic carbon (TC), and alkalinity (TA) were usually determined for the smaller volume (30-L) samples. Larger volume (270-L) samples were taken for determination of radiochemical tracers, including ^{14}C, ^3H, and ^{222}Rn. The pCO_2 in surface waters was also measured directly (Takahashi and Azevedo 1982).

The quantities of TC and TA were determined by titration, with an overall precision of about $\pm 0.5\%$ (± 10.5 μmol/kg) and $\pm 0.4\%$ (± 9 μmol/kg), respectively. Broecker and Takahashi (1978) noted, however, that the results from the Pacific and the Indian Oceans led to calculated CO_2 partial pressures that were systematically about 15% greater than the values measured directly. Bradshaw et al. (1981) subsequently found systematic errors in the reduction of the titration data that may be corrected by subtracting from TC the sum of 12 μmol/kg plus the dissolved phosphate concentration (usually <3 μmol/kg).

The extensive GEOSECS data set is currently available on computer-readable magnetic tapes and in tabular form. The results were extensively reviewed and discussed by Broecker and Peng (1982).

5.3.2.2 *Transient Tracers in the Ocean (TTO)*

The TTO program began with a survey conducted in the North Atlantic during 1981. A total of 250 stations were located at latitudes ranging from 15° to nearly 80°N. In addition to the quantities measured during the GEOSECS survey, the radioactive tracers ^3He, ^{85}Kr, and ^{39}Ar were determined as well. About 9000 samples were analyzed for salinity, oxygen, and nutrients; 6000 for TC and TA; 4000 for tritium; and 1200 for ^{14}C. In addition, pCO_2, CH_4, and N_2O were measured continuously in surface waters.

Special attention was given to improving the titrimetric determination of TC and TA (Brewer et al. 1986). The reduction of the titrimetric data included explicitly the effects of CO_3^{2-} and OH^- ions, and phosphate and silica. In addition to the shipboard determinations, a separate set of samples was stored for later analysis on shore. The determinations of TA at sea and on shore in duplicate samples showed an average deviation of ±9.5 μmol/kg (±0.4%), with the mean difference being only 0.8 μmol/kg (0.03%). The shore-based determinations of TC, however, averaged 21.3 ± 8.6 μmol/kg (1%) higher than the shipboard values. The shipboard determinations are preferred because they agree well with determinations of TC by an independent method and they yield calculated CO_2 partial pressures that agree well with measured values. Although the reason for the higher values of TC found in samples stored for several months is not yet known, there seems little reason to doubt the accuracy of the shipboard determinations.

Analyses of the TTO North Atlantic survey data are just now appearing in the literature. A preliminary review was given by Brewer (1983). The program was continued with a survey of the tropical Atlantic in 1982-1983, but the planned extension to the South Atlantic and, eventually, the other oceans has not been carried out.

5.3.2.3 *Other Surveys*

The First Global Geophysical Experiment (FGGE) and the North Pacific Experiment (NORPAX) were intended to supplement the GEOSECS results. Many of the same chemical and radiochemical tracers were determined in selected locations and as a function of time. In addition, surface winds and drifting buoys were observed, and measurements were taken by geostationary satellites. A principal activity included in NORPAX was the Hawaii to Tahiti Shuttle Experiment, consisting of repeated round trip surveys by ships and planes between these two locations over a period of 17 months beginning in early 1979 (Wyrtki et al. 1981). The principal objective of this experiment was to develop the scientific basis for a monitoring system for the tropical Pacific.

Another effort was concerned with the ^{14}C, ^{90}Sr, and $^{18}O/^{16}O$ ratios in the annual growth rings of corals. Data were collected by Druffel (at Woods Hole Oceanographic Institution) and Broecker and Toggweiler (at Lamont-Doherty Geophysical Observatory) from coral (aragonite) growth rings in the Pacific, Atlantic, and Indian Oceans to provide records over various intervals between 1640 A.D. and the present. The results give information on the tracer content of surface waters during this period (Section 5.3.3).

As already noted, measurements of the distribution of the chlorofluoromethanes in the ocean are now appearing in the literature. They include data from two locations in the eastern North Pacific (Gammon et al. 1982), two locations in the North Atlantic off Greenland and Norway (Bullister and Weiss 1983), and several locations in the western North Atlantic as part of the TTO survey (Gammon and Bullister 1982; Brewer 1983).

Finally, Levitus (1982) recently completed a critical compilation of measurements, accumulated over the last 80 years, of the temperature, salinity, and oxygen content of ocean waters. Average salinity, potential temperature, and potential density on a 1° grid are available on magnetic tapes. Levitus presents tables listing zonal averages of these quantities at 5° intervals of longitude and at selected depths for the world ocean and for each ocean basin.

5.3.3 Application of the Data

The useful information extracted from the large body of oceanographic tracer data is growing rapidly and is already much too extensive (Broecker and Peng 1982) to be described properly here. The following discussion will review briefly the relatively direct applications of these data to the study of processes that affect the uptake of CO_2 by the oceans. The use of tracers in carbon cycle modeling is considered further in Section 5.5 and a more detailed treatment is available in Chapter 7.

5.3.3.1 *Ocean-Atmosphere Exchange*

The rate at which CO_2 is exchanged between the ocean and atmosphere was derived from the natural ^{14}C content of surface waters. Except in locations where strong upwelling or deep overturning occur, this concentration is determined primarily by a balance between the decay rate of ^{14}C in surface waters and the rate at which it can be supplied from the atmosphere. The direct observations prior to the beginning of nuclear testing in the mid-1950s (summarized by Broecker and Peng 1982, p. 415) show that in temperate and tropical regions this concentration was 50–60 parts per thousand (‰) less than that in the atmosphere. The results from the annual growth rings of corals formed prior to 1957 are in the same range (Druffel and Suess 1983; Broecker and Peng 1982, p. 418). By adopting a value of −50‰ for the surface water deficiency of ^{14}C, Broecker and Peng (1982, p. 531) obtained a rate constant of 0.063 mol m^{-2} μatm^{-1} $year^{-1}$. Similar results were obtained by others, and another estimate based on bomb-^{14}C buildup in surface waters yields a CO_2 exchange rate only 10%

higher than that based on natural ^{14}C (Broecker and Peng 1984). If the rate-controlling step in the transfer is assumed to be diffusion of CO_2 through a stagnant liquid film at the surface, then the observed transfer rate for natural ^{14}C suggests the average thickness of this film is about 30 μm.

The concentration of ^{222}Rn in surface waters is similarly determined by a balance involving the rate of transfer between the ocean and the atmosphere. In this case the balance includes the rate of production of ^{222}Rn from the decay of ^{226}Ra dissolved in the water, the decay of ^{222}Rn, and its loss to the atmosphere. The observed deficiency of ^{222}Rn adjacent to the surface is used to calculate the thickness of the stagnant surface film. The results (summarized by Broecker and Peng 1982, p. 122–128) indicate that the film thickness varies from about 15 to 60 μm, decreasing with increasing wind velocity in the range of 7 to 12 m/s. Film thicknesses determined as a function of wind velocity in wind tunnels show a similar dependence but are generally lower by a factor of about 2. The averaqe film thickness given by the radiocarbon results is consistent with the ^{222}Rn results. Other methods for estimating the gas exchange rate either failed to yield consistent results or are not applicable to global estimates (Broecker and Peng 1984).

Given a transfer rate for CO_2 (0.063 mol m^{-2} μatm^{-1} $year^{-1}$), the area of the oceans (3.6×10^{14} m^2), and the present atmospheric pCO_2 in 1983 (342 μatm; Chapter 3), the calculated exchange rate of CO_2 is 93 $\times 10^{15}$ g C, about one-seventh the carbon content of the well-mixed surface waters, per year. The principal uncertainty in this result (about ±25%) arises from uncertainty about the depletion of natural radiocarbon in surface waters relative to the atmosphere (about ±15‰). Although individual determinations of the surface water ^{14}C depletion are more accurate, especially those for corals, there is an appreciable uncertainty about the overall effect of mixing with deeper waters, which are even more depleted in radiocarbon. Indeed, the results from coral growth rings formed at various times and in different locations (Druffel 1981, 1982; Druffel and Suess 1983; Toggweiler 1983) constitute a valuable record of upper-ocean circulation.

The global average exchange rate for CO_2 is therefore deduced principally from ocean models that describe the distribution of natural radiocarbon, and this rate is thus affected by the choice of the parameter(s) describing circulation (cf. Broecker et al. 1980, and Figure 4 therein). Because ocean model CO_2 uptake appears to be relatively insensitive to the value assigned to this exchange rate, its uncertainty does not currently appear to be a limiting factor (Chapter 7). However, if the global inventories of bomb radiocarbon are to be applied usefully to constrain global ocean circulation models, better models of regional CO_2 exchange, and wind stress dependence in particular, will have to be developed (Broecker et al. 1985b).

Using a value for the exchange rate constant of 0.067 mol m^{-2} μatm^{-1} $year^{-1}$, Takahashi and Azevedo (1982) estimated net flows of CO_2 to and from various portions of the ocean induced by the differences between surface water and atmospheric pCO_2 (Figure 5.2). The gyres of the temperate zones are the largest CO_2 sink ($\sim 2.2 \times 10^{15}$ g C/year) because of the reduced pCO_2 in waters over these large areas. The equatorial regions are a strong source ($\sim 0.9 \times 10^{15}$ g C/year). Although the northern high-latitude seas are an intense sink, at least during the summer months, their small area produces an uptake only about one-fourth that of the gyres. The area-weighted average of the differences in partial pressure (retaining the sign) is -8(±8) μatm, and the net transfer is from the atmosphere to the ocean at 2.1(± 2) $\times 10^{15}$ g C/year (about 40[±40]% the present rate of release of fossil carbon to the atmosphere). For this direct method of estimating the total uptake of CO_2 to achieve the desired accuracy, much more extensive monitoring of pCO_2, coupled with a more accurate knowledge of how the rate constant for CO_2 exchange varies with seawater conditions, will be required.

5.3.3.2 *Ocean Circulation*

The GEOSECS data for natural ^{14}C were used extensively to establish the gross rate of deep-ocean circulation. Stuiver et al. (1983) summarized these data and conclude that replacement times of the waters below 1500 m in the Atlantic, Indian, and Pacific Oceans are, respectively, about 275, 250, and 510 years. The data for ^{14}C in the deep oceans helped establish the diffusivity parameter used widely in one-dimensional (depth) ocean models to represent the rate of mixing in the deep ocean (Section 5.5). These data also reflect the general trend of deep circulation from the Atlantic to the other oceans and are of value in establishing the circulation parameters in multidimensional models.

Transient tracers produced by human activities provide perhaps the most useful information about circulation in the upper ocean. Most dramatic is the rapid penetration of tritium into the deep North Atlantic Ocean at high latitudes (Figure 5.8) and the advance southward evidenced by a comparison of the TTO (1981) results and the GEOSECS (1972) results (Östlund 1983). This penetration reflects the deep overturning at high latitudes. In the temperate oceans, mean penetration depths of 300 to 700 m were seen in the GEOSECS data (Broecker and Peng 1982, p. 403), indicating strong ventilation of the thermocline. Penetration was least (<200 m) in the equatorial oceans.

Penetration of excess ^{14}C from atmospheric nuclear testing into the upper ocean was similarly seen in

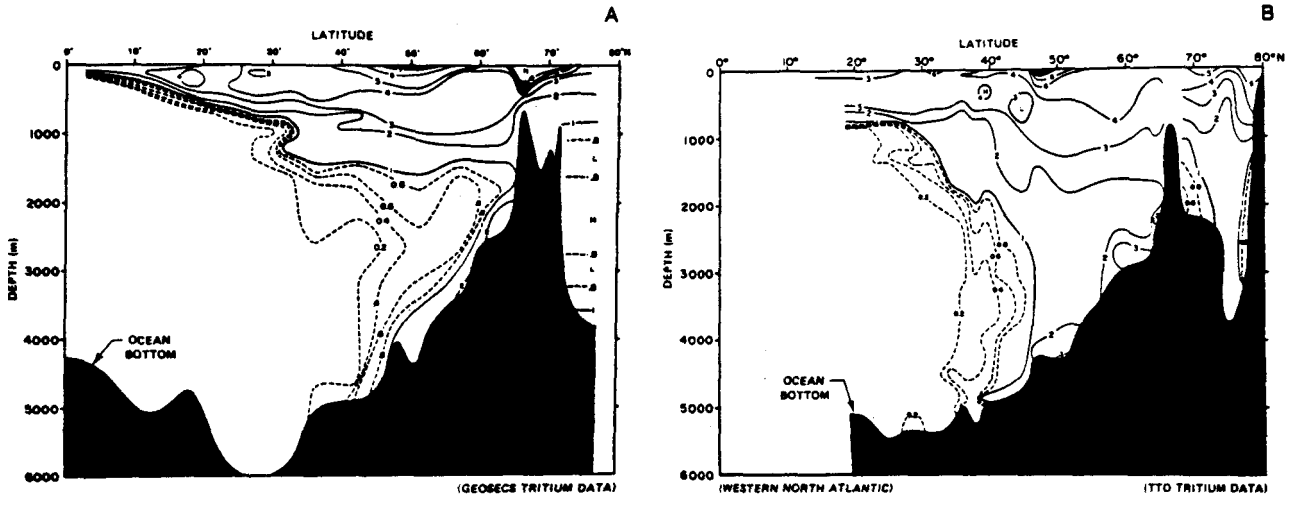


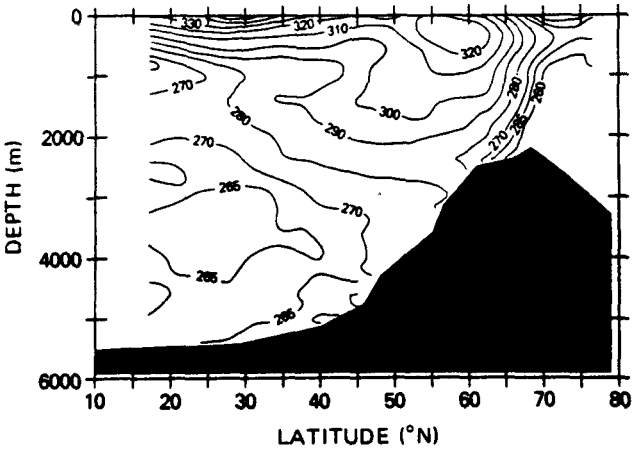
Figure 5.8. Penetration of tritium into the western basin of the North Atlantic Ocean: (A) GEOSECS data (1972), approximately 9 years after the release of tritium by atmospheric nuclear testing; (B) TTO data (1981), approximately 18 years after its release. The tritium concentrations are expressed in the T.U. unit (1 T.U. = the atomic ratio $^1H/^3H \times 10^{18}$) and are corrected to January 1, 1981 (from Östlund 1983).

the GEOSECS and the TTO results (Östlund 1983), although the measurement of this signal is more difficult because it must be separated from the background of natural radiocarbon (Broecker et al. 1985b). Because the release of ^{14}C to the atmosphere by nuclear testing was more prolonged (10 years) than was the release of tritium (2 years) and its rate of exchange with the surface waters is slower, the time scale over which the effects of circulation are revealed by ^{14}C is greater.

The release of the chlorofluorocarbons to the atmosphere occurred over an even longer (40-year) period. If their release rate can be established with suitable accuracy (Gammon et al. 1982), they should provide yet another time scale with which to view ocean circulation. From measurements in the Greenland and Norwegian Seas, Bullister and Weiss (1983) found mixing time in the deep Greenland Sea to be about 40 years and an exchange time with the deep Norwegian Sea of 20 to 30 years.

As the excess of CO_2 in the oceans continues to increase, it too should become a valuable transient tracer, revealing directly the effects of circulation and the biological processes on the uptake of CO_2. Unfortunately, excess carbon cannot as yet be accurately measured. In his review of the preliminary TTO results, Brewer (1983) examined the TC and TA data, the most accurate obtained thus far, for the signature of excess CO_2 in the western North Atlantic Ocean. After a brief but thorough discussion of the methods used and the uncertainties involved in the reduction of the data, he presented contours of pCO_2 in a plot of latitude versus depth (Figure 5.9). These are estimated partial pressures that prevailed in various water masses when they were last exposed to the atmosphere. Although the results suggest that deeper, older water was exposed to an atmosphere containing less CO_2 than at present, Brewer (1983) points out that without a similar plot from the preindustrial era, it is not possible at present to know how much of this variation is caused by atmospheric changes and how much is caused by normal changes in the surface water pCO_2 value with location and season. This point is well illustrated by the fact that for a line of stations (Nos. 235 to 250) extending from the Sargasso Sea north to Nova Scotia, the preformed pCO_2 values in the core of the so-called western boundary undercurrent (at 3.5 km depth) were not elevated significantly, even though chlorofluorocarbon and tritium data showed that this water had been exposed to the atmosphere quite

Figure 5.9. Back-calculations of pCO_2 of deep water of the western Atlantic. These are estimated partial pressures (in μatm) when water was last exposed to the atmosphere (from Brewer 1983).



recently (within the last few years). The signal of excess CO_2 in the oceans will not be seen clearly until it rises further above the present background fluctuations in pCO_2 or until these fluctuations are better known and understood (also see Broecker et al. 1985a).

The translation of the results from transient tracers into a coherent picture of ocean circulation requires the use of models. The simple models often employed for this purpose (Section 5.5 and Chapter 7) have, of necessity, simulated the spread of transient tracers in the oceans by averaging their distribution in one or two dimensions. As the data base for these tracers grows, more elaborate models are being developed to interpret their behavior. An excellent example is the two-dimensional (depth × latitude) model of the upper waters of the equatorial Pacific employed by Quay et al. (1983) to interpret tracer data (tritium, ^{14}C, TC, dissolved oxygen, and salinity) collected mostly during the Hawaii to Tahiti Shuttle Experiment. The results indicated that equatorial upwelling in the Pacific Ocean is 47 × 10^6 m^3/s from a maximum depth of 225 m. Similar applications of many different tracer data sets (e.g., Wunsch 1984; Broecker et al. 1985a), including the large ones for temperature and salinity, to other regions of the oceans can be expected to yield similar quantitative information as more sophisticated models are developed (Sarmiento 1986).

An important result from the TTO survey was the finding that, north of 50° in the North Atlantic, the deep waters and especially the the surface waters were less saline than 20 years previously (Brewer et al. 1983). The change occurred over such a wide expanse of ocean that climate change is thought to be responsible. These results suggest that the temporal as well as the spatial coverage of ocean surveys, especially at high latitudes in winter, will have to be improved substantially.

5.3.3.3 *Biological Processes*

The extensive data available on the distribution of TC, TA, oxygen, and nutrients in the oceans of the world (only a small portion of which is shown in Figure 5.6) are being used widely to calibrate or validate various simple models that include simulations of the biological processes and the exchange of CO_2 with the atmosphere. The nonuniform distribution of these quantities is of course what renders them valuable for this purpose, but the observed variabilities also show the need for more complete spatial and especially temporal coverage of the oceans than thus far has been achieved. As will be discussed in Section 5.5 and in Chapter 7, more realistic models are being developed that can employ this more extensive data.

The annual growth rings of corals, in addition to giving information on the history of the natural radiocarbon content of surface waters, recorded in $^{18}O/^{16}O$ ratios the long-term variability in sea-surface temperature on a time scale of centuries (Druffel 1982). Recent changes such as the El Niño phenomena were also clearly detected from bomb radiocarbon distributions and $^{18}O/^{16}O$ ratios in Pacific corals (Druffel 1981, 1985).

5.4 OTHER SOURCES AND SINKS OF CARBON IN THE OCEANS

In addition to direct transfer of CO_2 between the atmosphere and the oceans, a number of other pathways and processes may significantly affect the net uptake of CO_2 by the oceans. These include the transport of organic carbon by rivers and the deposition of organic carbon in estuarine, coastal-ocean, and deep-ocean sediments. In this section is presented current estimates of these fluxes and, with less certainty, estimates of changes in them over the past 100 years that are needed to balance the carbon cycle. A summary is provided in Table 5.2.

5.4.1 Influx from Rivers

The current flux of organic carbon to the oceans from rivers was estimated to be about 0.4 × 10^{15} g/year (Schlesinger and Melack 1981; Meybeck 1982). However, this value could be low for a variety of reasons: (1) lack of measurements of transport of large objects (e.g., logs, floating debris); (2) inadequate measurements of particulate organic carbon flux in rivers with high suspended sediment loads, particularly in Asia; and (3) relatively few measurements in large tropical rivers and large Asian rivers in which carbon flux is likely to be high. Moreover, recent data reported for some tundra and taiga rivers of the eastern U.S.S.R. are 25–50% higher than values previously reported for these types of ecosystems (Degens et al., in press). Some argued that the global flux of organic carbon from rivers could be as high as 1 × 10^{15} g/year (Richey et al. 1980; U.S. Department of Energy 1981), but this is based on the assumption of relatively high organic content of transported sediments and probably represents an upper bound on organic carbon flux from rivers. In addition to organic carbon, about 0.2 × 10^{15} g/year of inorganic carbon (about one-half the global flux of inorganic carbon from rivers) is produced from the oxidation of organic carbon in transit to the sea (Meybeck 1979, 1982). A major effort is currently under way by the SCOPE/UNEP International Carbon Unit to improve global estimates of the present river discharge of plant nutrients and carbon to the oceans (Degens et al. 1983, in press).

Table 5.2
Estimates of Current Carbon Flux and Recent (last 100 years) Increases in Those Fluxes for Potentially Important Oceanic Carbon Sinks

Process	Current Carbon Flux Best Estimate	Current Carbon Flux Range	Recent Changes in Carbon Flux
River Flux Accumulated in Oceans			
Organic	0.5	0.2–1.0	0–0.2
Inorganic[a]	0.2	0.1–0.3	0–0.1
Organic Flux from Surface to Deep Oceans[b]	0.4	0.02–0.6	0
Accumulation of Coastal Net Primary Productivity			
Phytoplankton	0.8	0.2–1.5	0.1–0.3
Macrophytes	0.8	0.4–1.5	
Carbonate Mineral Dissolution	?	?	0.1–0.2
Total			0.2–0.8

Note: All fluxes are in 10^{15} g C/year.

[a]Derived from the oxidation of organic carbon.
[b]Approximately one-half buried in sediments and one-half added to refractory dissolved organic carbon pool.

Estuarine systems at river mouths are highly efficient traps for organic carbon transported in rivers (Meade 1972; Sholkovitz and Price 1980; Wangersky 1981; Gibbs 1981). Accumulation of carbon in deltas formed at the mouths of the larger rivers could account for most of the estimated annual carbon flux from rivers (Mulholland 1981), although accumulation of marsh plant debris also contributes to delta formation (see Section 5.4.3). Some of the organic carbon flux from rivers is oxidized to CO_2 in coastal areas; however, this is probably a small fraction of the total flux (Wollast and Billen 1981). Finally, the more refractory portion of the riverine dissolved organic carbon (DOC) is transported out to sea, where it adds to the refractory DOC pool of the oceans or is eventually oxidized.

Very little data exist on recent changes in the carbon flux from rivers to the oceans, although relatively high values in heavily populated regions of the world indicate that it may have increased within the last 100 years (Meybeck 1982; U.S. Department of Energy 1981). Increased rates of soil erosion, resulting from forest cutting and agriculture, and increased inputs of sewage probably increased organic carbon flux from rivers. Organic carbon concentrations in some European rivers are two to four times greater than those in less populated regions (Meybeck 1982). One study estimated a recent river carbon flux increase of up to 0.15×10^{15} g/year (U.S. Department of Energy 1981). Increased flux of inorganic carbon from rivers to the oceans because of the oxidation of increased inputs of organic carbon is estimated to be from 0.02×10^{15} to 0.04×10^{15} g/year (Meybeck 1982; Kempe 1984).

5.4.2 Accumulation of Organic Carbon in the Deep Ocean

The flux of organic carbon from oceanic primary production to the deep-ocean sediments and to refractory DOC constitutes an additional ocean carbon sink. Estimates range from 0.02×10^{15} g/year, based on a few measurements of organic carbon content and sedimentation rate in deep ocean sediments (Berner 1982), to 0.6×10^{15} g/year, based on ocean net primary production (NPP) estimates and an average oxidation efficiency of 97% computed from laboratory studies (Menzel 1974; Eppley and Peterson 1979). Field measurements of organic carbon accumulation in sediments in the open ocean indicate that only about 1% of NPP escapes oxidation and is buried in sediments (Muller and Suess 1979), giving an organic carbon accumulation rate of about 0.2×10^{15} g/year. An additional flux of DOC to the deep ocean of 0.2×10^{15} g/year was estimated (Williams et al. 1969; Williams and Linick 1975), although some of this may actually be a contribution from rivers.

There is little reason to believe that the flux of organic carbon from the surface to the deep ocean significantly increased in recent decades (e.g., Peng and Broecker 1984). Although nutrient inputs to the oceans by rivers increased, the consequent fertilization effect on ocean NPP was probably confined to the ocean's coastal regions (see following section).

5.4.3 Processes on the Continental Shelves

Various chemical and biological processes occurring on the Earth's continental shelves constitute sinks for atmospheric CO_2. These include the accumulation in sediments of carbon fixed by phytoplankton, macrophytes, and bacterial autotrophs, and the dissolution of carbonate minerals.

Although NPP by phytoplankton in coastal waters is only about one-third as great as that in the open ocean (Whittaker and Likens 1973), it is concentrated in a much smaller area, and a much greater fraction escapes oxidation and accumulates in coastal sediments. During late winter and spring blooms of phytoplankton in temperate coastal regions, zooplankton populations are small and unable to efficiently utilize the algal production. Shallow water and high sediment deposition in coastal regions reduce the oxidation of sinking and sedimented algal remains (Walsh 1976; Malone and Chervin 1979; Walsh et al. 1981). Assuming that (1) primary production is nitrogen limited in coastal waters (Perry and Eppley 1981; McElroy 1983), (2) the ocean nitrogen concentration is at steady state, and (3) the carbon-nitrogen ratio in sediments is 10:1 (Krom and Berner 1981; Walsh et al. 1981), organic carbon accumulation from the stimulation of coastal NPP by nitrogen inputs from rivers can be calculated. Recent estimates of fixed nitrogen input to the oceans by rivers range from 0.020×10^{15} g/year (Soderlund and Svensson 1976) to 0.075×10^{15} g/year (Van Bennekom and Solomons 1981), which would result in organic carbon accumulation of 0.2×10^{15} to 0.8×10^{15} g/year in coastal sediments. Results of recent studies (Broecker 1982; McElroy 1983) suggested that during periods of rising sea level, as at present, nutrient concentrations in the ocean are not at steady state but are being depleted by increased biological activity and organic sediment accumulation on the continental shelves. If this is so, current organic carbon storage in sediments could be as much as 0.5×10^{15} to 1.0×10^{15} g/year higher than that resulting from nutrient input from rivers alone (McElroy 1983; Walsh 1984).

Macrophyte (larger marine plants such as eel grass, kelp, and intertidal marsh grasses) communities in coastal areas occupy only about 10% of the continental shelf (Smith 1981), but their biomass is greater than that of ocean plankton and their NPP is on the order of 1×10^{15} to 2×10^{15} g/year (Woodwell et al. 1973; DeVooys 1979; Smith 1981). Macrophyte tissue is somewhat more refractory than that of phytoplankton, and thus a higher proportion of macrophyte NPP escapes oxidation and is buried in marsh and other coastal sediments. Assuming that (1) coastal marshes cover 0.4×10^{12} m^2 globally (Woodwell et al. 1973); (2) river inputs, coastal phytoplankton, and macrophytes contribute equally to marsh sediments (Haines 1977; Meade 1982); (3) average accretion rate of sediment is 1 cm/year (Delaune et al. 1978); and (4) average marsh organic sediment density is 0.3 g/cm^3 (Delaune et al. 1978), then about 0.4×10^{15} g/year of macrophyte production accumulates in marsh sediments. Accumulation of macrophyte detritus in all other coastal sediments could be of the same order of magnitude, giving a total sediment sink of about 0.8×10^{15} g/year for marine macrophytes.

Coastal NPP and organic sediment accumulation probably increased over the last century as a result of global eutrophication from increases in population, fertilizer use, and deforestation. However, there is large uncertainty concerning the magnitude of this increase. Walsh (1984) estimates that organic carbon burial, mostly in slope sediments, rose by about 0.5×10^{15} g/year as a result of a 10-fold increase in the flux of fixed nitrogen from rivers over the preindustrial flux of about 0.005×10^{15} g N/year, based on comparisons of data from a few rivers in unpopulated and populated regions. In a more detailed accounting of the river flux to the oceans, Meybeck (1982) estimated that the preindustrial flux of dissolved fixed nitrogen was 0.035×10^{15} g N/year, and in 1970 was only 0.007×10^{15} g/year greater than the preindustrial value. From Meybeck's estimate for the dissolved nitrogen flux, Peterson and Melillo (1984) concluded that the increased flux of organic carbon to sediments in the coastal ocean from recent eutrophication was 0.04×10^{15} to 0.08×10^{15} g/year. Considering the additional effects of increased particulate nitrogen inputs, the nitrogen inputs directly to the ocean from coastal cities, and the increases in the river fluxes of nitrogen since 1970 (Kempe 1984), the resulting increase in organic carbon fixation and burial was probably in the range of 0.1×10^{15} to 0.3×10^{15} g/year over historical levels.

The increased invasion of CO_2 from the atmosphere caused by higher atmospheric CO_2 concentrations increased the dissolution of magnesian calcites in coastal regions and further increased the uptake of atmospheric CO_2 by an estimated 0.07×10^{15} g C/year (Garrels and Mackenzie 1981). Increased coastal primary production and subsequent reoxidation of organic carbon may also have increased the magnesian calcite dissolution, perhaps by as much as 0.1×10^{15} to 0.2×10^{15} g C/year.

5.5 MODELING THE OCEANS

To construct a quantitative picture of how rapidly an excess of CO_2 will spread from the atmosphere into the oceans, one must translate the physical, chemical, and biological processes described earlier in this chapter into mathematical language. The product of this translation is usually a set of differential equations expressing the time rate of change of the amounts of carbon in the

atmosphere and in various regions of a model ocean in terms of these same quantities and related ones, along with the water flows and various boundary conditions of the model (see Chapter 7). The goal in developing such CO_2-ocean models generally has been to formulate the simplest representation that is consistent with the relevant data. As will be seen, insufficiencies in simpler models have led to more sophisticated ones. This process presumably will continue until it becomes clear that the predicted uptake of excess atmospheric CO_2 no longer is significantly affected by further refinements in the models. In this section this development is briefly described up to the present; in Section 5.6.2 some predictions based on different models will be summarized; and in Section 5.7.2 recommendations concerning the future development of such models will be made. The use of tracer data for parameterizing ocean models is considered further in Chapter 7.

5.5.1 One-Dimensional Models

The primary requirement of even the simplest ocean model is to represent the slow vertical circulation of water that is usually the rate-determining condition in the uptake of CO_2. The simplest way to account for the slowness of vertical mixing has been to visualize the oceans as a system of two boxes, one above the other (Figure 5.10, Model B-B). The uppermost box receives CO_2 from the atmosphere, but because it is only 75 m deep, it cannot store large amounts of carbon. The lower box is 4 km deep, but it exchanges material with the surface box only slowly. The exchange can be estimated by noting that the abundance of natural ^{14}C in the lower and upper boxes, respectively, should be about 84 and 95% of that in the atmosphere. Maintenance of these pools against radioactive decay requires that about 1 part in 1100 of the water in the lower box be exchanged each year with that in the upper box, and that, as previously noted, about one-seventh of the carbon in the upper box be exchanged each year with the atmosphere.

Keeling (1973) analyzed the distribution of CO_2 from fossil fuel combustion using this model. Similar computations were made by others. Bacastow and Björkström (1981) showed that if fossil fuel usage is simulated by an exponentially growing function, more than 80% of the CO_2 released will remain in the atmospheric box (Table 5.3). The observed increase in atmospheric CO_2 from 1958 to 1980, however, is equivalent to only about 58% of the fossil fuel emissions (assuming that no other sources exist; see Chapters 6, 7, and 10). Thus, unless the land biota has been a net sink for CO_2, the oceans seem to have been more than twice as efficient as a sink than this simple model suggests.

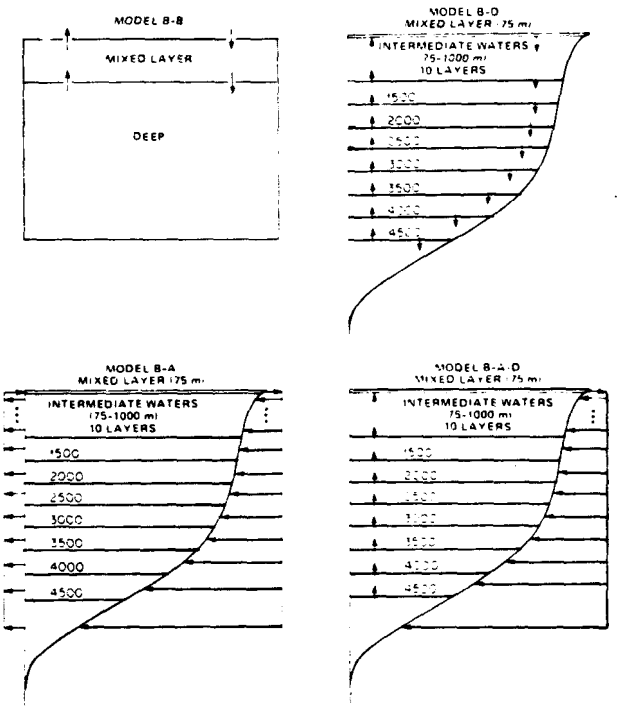


Figure 5.10. One-dimensional box models of the ocean (from Killough and Emanuel 1981). The numbers in deep boxes denote the depth (m) of the lower boundary.

(The role of the land biota is a matter of some debate; see Chapters 6, 7, and 10.)

The agreement can be greatly improved by increasing the depth of the upper box to about 500 m. Then both the ^{14}C distribution and the observed increase in atmospheric CO_2 would be reasonably well accounted for. However, the well-mixed surface waters average about 75 m in depth worldwide, not 500 m. The conclusion to be drawn, therefore, is that the two-box model contains oversimplifications, the most conspicuous of which is the implicit assumption that the entire ocean below the mixed layer is a homogeneous region. It is evident that the internal structure of the ocean must be taken into account.

Oeschger et al. (1975) described an alternative model (similar to that shown in Figure 5.10, Model B-D) for the ocean below the mixed layer in which the downward penetration of CO_2 was assumed to obey the equation normally used to describe the diffusion of molecules in one dimension. Vertical mixing in the ocean is not brought about by molecular-scale motion; however, it is a reasonable hypothesis that the many large- and small-circulation cells that perform mixing, taken together, exert an effect statistically similar to molecular diffusion, only on a much larger scale.

Actually, the use of diffusion models is not new in oceanography. Similar models were used by Wyrtki

Table 5.3
Uptake of CO_2 by Various Model Oceans

Model[a]		Time Period	Airborne Fraction[b] $\frac{100 \times \Delta C_a}{\Delta C_a + \Delta C_o}$		Source
		One-Dimensional Models			
B-B	75-m Mixed Layer	1860–1980	81		Bacastow and Björkström (1981)
	75-m Mixed Layer	1860–1975	77.7	(80.3)	Killough and Emanuel (1981)
	260 m Mixed Layer	1860–1975	65.7	(68.5)	Killough and Emanuel (1981)
B-D	Constant Diffusivity	1860–1973	66.6		Oeschger et al. (1975)
	Constant Diffusivity	1860–1973	62.5		Broecker et al. (1980)
	Variable Diffusivity	1860–1975	58.3	(61.3)	Killough and Emanuel (1981)
B-A		1860–1975	67.0	(70.3)	Killough and Emanuel (1981)
B-A-D		1860–1975	56.0	(58.7)	Killough and Emanuel (1981)
		Multidimensional Models			
B-D	Polar, Temporate, Equatorial Regions	1860–1973	57.4		Broecker et al. (1980)
B-D	Atlantic, Indian-Pacific, Polar Regions:				
	0% Polar Outcrop	1959–1978	67.7		Crane (1982)
	10% Polar Outcrop	1959–1978	64.3		Crane (1982)
	20% Polar Outcrop	1959–1978	60.5		Crane (1982)
B-D	No Isopycnal Outcrop	1959–1978	66.7[c]		Siegenthaler (1983)
	10% Isopycnal Outcrop	1959–1978	60.5[c]		Siegenthaler (1983)
	20% Isopycnal Outcrop	1959–1978	58.5[c]		Siegenthaler (1983)
B-D	Deep-Water Circulation to Surface			(61)	Peng et al. (1983)
12-Box	World Ocean	1860–1980	72–73	(74-75)	Bolin et al. (1983)
91-Box	Idealized Half Ocean	1860-1980	63-68	(59-69)	Baes and Killough (1985)

[a]The model configurations denoted by the abbreviations B-B (two-box), B-D (box-diffusion), B-A (box-advection), and B-A-D (box-advection-diffusion) are shown in Figure 5.10.
[b]ΔC_a and ΔC_o denote increases of carbon content in the atmosphere and in model oceans, respectively. The values in parentheses are for the period 1958–1980 covered by the Mauna Loa record. The apparent percentage of carbon from fossil fuel use and cement production remaining airborne during this period was 58% (Chapters 4 and 10).
[c]Models were calibrated with natural ^{14}C. Calibration with bomb-^{14}C yielded airborne fractions of 61.1 and 52.8% for 0 and 10% polar outcrops, respectively.

(1962), among others, to explain the distribution of oxygen, by Munk (1966) to explain the distribution of temperature, salinity, and ^{14}C, and by Craig (1969) to explain the distribution of TC and ^{14}C. Oeschger et al. (1975), however, were the first to apply this technique to the analysis of the uptake of excess atmospheric CO_2. Like the two-box model, their so-called box-diffusion model was homogeneous in all horizontal directions, with conditions being variable only in the vertical direction.

The diffusive lower ocean was represented by a stack of many boxes with a uniform cross section, and the flows between adjacent boxes were determined by a single diffusivity parameter. When this parameter was adjusted to account for the average distribution of natural ^{14}C in the oceans, it produced an airborne fraction (AF*) of about 67%, lower than that given by the two-box model, but still substantially higher than the observed AF of 58% based on the fossil carbon release. (The airborne fraction, "AF*", referred to here is defined in Table 5.3 as 100 times the ratio of the increased amount of carbon in the atmosphere to the sum of the increases in the ocean plus the atmosphere. A number of alternative definitions of the airborne fraction have been used and these are analyzed in Chapter 10.)

Broecker et al. (1980) tested the simple box-diffusion model more exhaustively, using it to simulate the response of the oceans to ^{14}C and ^{3}H introduced by atmospheric nuclear testing, and the decrease of ^{14}C in the atmosphere resulting from the introduction of fossil carbon free of ^{14}C (the Suess effect). They found a somewhat higher diffusivity but obtained an AF* of only about 63%.

Another noteworthy study of the simple box-diffusion and other one-dimensional models is that of Killough and Emanuel (1981), who tested the effect of the pattern of water flow. To simulate diffusion, it is assumed that water flow occurs only between adjacent boxes (Figure 5.10, Model B-D). In another model with the same arrangement of boxes, they simulated advective flow of water directly between each box and the surface (Model B-A). In a third model, water was supplied to each box directly (by advection) from the surface, but was returned (diffusively) upward through the overlying boxes (Model B-A-D). The flows in all three models were adjusted to produce a standard profile of ^{14}C based on the GEOSECS data. The B-D

model differs from the original box-diffusion model in that the effective diffusivity varies with depth. The AF* given by these models (Table 5.3) was found to depend substantially on the flow pattern assumed. The B-A-D model gave 56% (the lowest); the B-D model, with a variable diffusivity, gave 59%; and the B-A model gave 67%.

5.5.2 Multidimensional Models

The variable results obtained with one-dimensional models suggest that it is an oversimplification to treat the ocean as being horizontally homogeneous (see Chapter 7). Downward mixing cannot be simulated by diffusion alone. An important part of the vertical transport in the ocean is by circulation cells that cannot be properly described unless one considers how conditions in the oceans depend on latitude and longitude. In particular, it is natural to suspect that the neglect of the role of deeply penetrating cold surface waters formed at high latitudes causes the one-dimensional models to take up less excess atmospheric CO_2 than would otherwise be the case (Sarmiento 1983, 1986).

Broecker et al. (1980) considered a model in which the polar, temperate, and equatorial regions of the oceans were represented by separate stacks of boxes with different diffusivities. This model gave an AF* of 57% (Table 5.3).

Viecelli et al. (1981) constructed a somewhat similar model and determined the diffusivities for each region from the GEOSECS data on tritium penetration. The area-weighted average value obtained (5.1 cm^2/s) was considerably higher than the previous estimate of Oeschger et al. (1975), based on natural radiocarbon (1.26 ± 0.52] cm^2/s), and gave the model enough capacity to account for the fossil carbon release plus a substantial release of CO_2 from the land biota. Li et al. (1984) used all the available tritium data and the box-diffusion model of Oeschger et al., adapted to each of the ocean regions, to determine their diffusivities. The area-weighted average was $1.7(\pm 0.2)$ cm^2/s. The large difference between these two efforts to represent the tritium penetration data in terms of vertical diffusivities may arise partly from model differences and partly because Viecelli et al. (1981) employed data only from the Northern Hemisphere, including deep-ocean data at high latitudes excluded by Li et al. (1984).

Crane (1982) constructed a model to simulate the effects of longitude rather than latitude with two separate diffusive reservoirs, one representing the Atlantic Ocean and the other the Indian and Pacific Oceans. The diffusivity of each was adjusted on the basis of the ^{14}C content of each of these oceans. Because, as already noted, deep water in the Atlantic is younger than that in the other two oceans, the diffusivity in this case turned out to be four times larger. Yet the AF* given by this model was only a few percentage points lower than that given by the simple, one-region box-diffusion model with a single diffusivity.

Craig (1963), Broecker et al. (1971), Siegenthaler (1983), and others explored different consequences of a direct CO_2 exchange between the atmosphere and the deep ocean. Siegenthaler conducted a particularly interesting series of experiments with a model that combines the deep-ocean outcropping with the idea of vertical diffusion. For a hypothetical case in which 10% of the deep ocean is in direct contact with the atmosphere, Siegenthaler found that the AF* decreased from 67 to 61% for a model calibrated with natural ^{14}C, and from 61% to 52–53% for a model calibrated with bomb-^{14}C (see Chapter 7).

When considering these results, one should realize that the outcrop model is a "grossly oversimplified representation of nature, mainly regarding the infinitely fast, quasihorizontal mixing by which fossil CO_2 is immediately mixed into the whole deep sea," and that the B-D models with and without outcrops are "extreme cases with respect to the role of direct exchange atmosphere-deep sea, the real ocean being situated somewhere in between" (Siegenthaler 1983, p. 3607).

Crane (1982) modeled outcropping by means of a separate box representing the oceans at high latitudes. Airborne fractions with this model ranged from 60 to 65%, depending on how large the surface of the high-latitude reservoir was assumed to be. Bacastow and Björkström (1981) described a modification of the simple box-diffusion model in which transport of water occurs directly between the mixed layer and the bottom of the diffusive reservoir. This transport, however, changes the computed results very little (Chapter 7).

5.5.3 Incorporation of Chemistry and Biology

In the box models described thus far, the goal has been to find the rate of exchange of carbon and water among the boxes. That which goes on inside the boxes is dealt with only slightly. Customarily, the implicit assumption is made that each box is undergoing internal mixing at a rate that is rapid on the time scale being modeled. Current opinion about the future carbon cycle in the oceans stems, in large part, from studies based on models of this kind. Because it is unsatisfactory to conceive of large volumes of the ocean being mixed instantaneously, it is natural to try to use as large a number of boxes as possible. The disadvantage, however, is that the number of unknown fluxes may then become so large that the available information does not suffice to determine them with any certainty.

It has already been illustrated how the abundance of ^{14}C and other isotopic tracers provide useful information for the determination of fluxes between boxes. The

distribution of quantities such as those of TC and TA (defined in Section 5.2.1), nutrients, and oxygen is additional information that can help fix the many fluxes in a multibox model. Because these distributions are determined by the inorganic and biological processes of carbon, it becomes necessary to represent these processes in such models. This is at the cost of a greatly increased number of fluxes between boxes, including those of inorganic and organic detritus, as well as rate constants for photosynthesis, decay, and calcium carbonate dissolution. However, because the same water motions transport all substances, most of the new fluxes are interrelated. Even if the distribution of each substance can be explained in a number of ways, the uncertainty is greatly reduced if a single pattern of ocean circulation must explain the distribution of all substances simultaneously (also see Chapter 7).

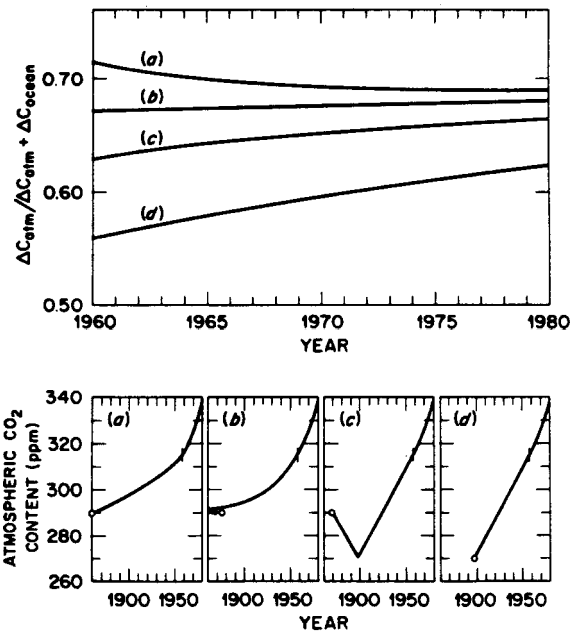
Keeling and Bolin (1968) exploited this idea in a study of the circulation of the Pacific Ocean using a three-box model. A more elaborate application of the method was made by Bolin et al. (1983). The Atlantic and Pacific Oceans were divided into four boxes each, which, together with the polar oceans, gave 12 boxes in all. Based on the data of the GEOSECS survey, the particular solution was derived that best explains the distribution of ^{14}C, alkalinity, phosphorus, and oxygen, simultaneously. The circulation pattern thus computed is oceanographically sound and the method is promising for future model development. However, when this model was used to compute the present distribution of excess atmospheric CO_2, it was found that well over 70% of the total CO_2 released remained in the atmosphere. In this important respect, the model is probably further away from reality than some of the simpler ones. This may be caused by the poor resolution in the model of the upper ocean, in which occur most of the processes important in the short-term response to increasing atmospheric CO_2 concentrations. In a later version of this model, Bolin and coworkers are using a larger number of boxes to represent just one ocean basin (see Chapter 7).

Baes and Killough (1985, 1986) modeled one-half of an idealized world ocean with a total of 91 boxes arranged by latitude and depth into isothermal regions that simulated the appearance of isotherms and isopycnals. In this way, the upper ocean was represented in greater detail. The many water flows were assigned by the use of two diffusivity parameters, one for flows within an isothermal region and the other for flows between adjacent isothermal regions. The chemical and biological processes were included and the parameters of the model adjusted to simulate the average distribution of TC, TA, nutrient, oxygen, and natural ^{14}C. Time integrations of the model to 1980 give AF*s in the range of 59 to 68%. As Siegenthaler et al. (1978) found for a one-dimensional, box-diffusion model, the AF* obtained depended on the assumptions made concerning the preindustrial level of atmospheric CO_2 and how it increased in the period prior to the beginning of the Mauna Loa measurements (Figure 5.11).

5.5.4 Time Scales

Few if any of the models described above and listed in Table 5.3 appear to be able to explain satisfactorily the apparent uptake of carbon by the oceans, especially if it is true that the biosphere has been a substantial source of CO_2 over the last 100 years (Chapter 7). This may be due in large part to the wide range of time scales on which the various oceanic processes operate. For example, chemical equilibrium between the atmosphere and the surface waters is established in less than 1 year, whereas ventilation of the most remote parts of the sea requires more than 1000 years. The processes that govern the ocean's response to slowly varying or constant external conditions, such as the production of ^{14}C in the atmosphere, are not the same as those that govern its response to a quasi-exponentially growing

Figure 5.11. Variation of the airborne fraction during the period of the Mauna Loa data as a function of past atmospheric CO_2 history (model of Baes and Killough 1985). The lower curves show the variation of atmospheric CO_2 assumed in the four cases: the point indicates the steady-state value at the beginning of each time integration; the vertical line marks the beginning of the portion of the curve that represents the Mauna Loa data.



input with a historical doubling time of 15 years. Simple box models or one-dimensional diffusion models are not flexible enough to represent the many response patterns that are involved (Chapter 7).

Not surprisingly, when models are compared against the accurate Mauna Loa record (since 1958), the AF* obtained can depend strongly on the prior variation of atmospheric CO_2 that was assumed for the calculations. Bacastow and Keeling (1978) showed that in a linear model with exponentially growing CO_2 emissions, the AF* will tend to approach a limiting value asymptotically. The time constant for the approach to this value is given by the time constant for the emissions. Baes and Killough (1985) found a similar result for their large two-dimensional box model (Figure 5.11, curve b). As already noted, they and Siegenthaler et al. (1978) obtained different AF*s when alternative scenarios were used to describe changes in atmospheric CO_2 prior to 1958. Restricting comparisons of AF*s given by various models to the period 1958-1980 (Table 5.3) removes some of the variability caused by the assumption of different model input functions; however, a substantial amount remains.

The lumping together of large parts of the ocean in many of the models may have biased the computed uptake in the conservative direction. Among the most interesting steps to be taken in the future will be the construction of ocean models with greater resolution. Yet it is noteworthy that the more detailed ocean models are in modest agreement, namely, that the oceans should have taken up 30 to 40% of the excess fossil carbon released to the present (Table 5.3). If more sophisticated models do not substantially increase this capacity, the role of other sources and sinks of carbon will remain an important question (see Chapter 10).

5.6 FUTURE UPTAKE OF CO_2 BY THE OCEANS

5.6.1 Detection of the Effects of Excess CO_2

As previously noted (Section 5.3.3), the signature of excess CO_2 in the oceans has not yet been clearly seen because it is difficult to separate from the temporal and spatial variations of TC and TA that normally occur in the upper ocean, especially at high latitudes. As more is learned of these natural variations through monitoring, the detection of excess CO_2 in the oceans should provide a valuable tracer to add to those already used for the validation of predictive models.

Another effect to be sought as the atmospheric CO_2 concentration increases is an increase in TA resulting from the dissolution of calcium carbonate in the upper ocean (Figure 5.3). This is currently thought to occur largely in confined terrestrial and coastal regions (Section 5.4.3), in which local pCO_2 generated by biological oxidation (decay) greatly exceeds that in the atmosphere. At high latitudes in the open ocean, the extent of oversaturation of surface waters with calcium carbonate is lower than elsewhere and dissolution here may also become significant. To the extent that dissolution occurs, there is a net uptake of atmospheric CO_2. This can be seen in Figure 5.3 by noting that as the water composition moves upward along a vector with a slope of 2, the pCO_2 decreases.

The monitoring program necessary to detect these effects will require a substantial increase over the present levels of effort, especially in view of recent evidence from the TTO program that temporal variation of oceanographic properties may be larger than previously thought (Brewer et al. 1983).

5.6.2 Predicted Future Uptake

Questions about the effects of increased CO_2 on the Earth's future climate address widely differing time scales. On one hand are the very long-term prospects for the global climate, for example, whether the next glaciation will be delayed or avoided, or whether there will be substantial melting of the polar ice caps. To answer these questions, a detailed knowledge of ocean circulation may be less important than other processes that control the steady-state distribution of carbon between the ocean and the atmosphere. On the other hand, there are important questions about the changes in atmospheric CO_2 and climate over the next 100 or 200 years that will require more sophisticated ocean models, ones that will respond differently to various kinds of driving conditions.

The box-diffusion model was used by Bacastow and Björkström (1981) to investigate the longer-term problem. The most important assumed condition was the total amount of CO_2 to be released. Assuming that all the economically recoverable fossil carbon is burned in the next 200 years, the CO_2 content of the atmosphere will rise to a value of at least five times the preindustrial value. After the peak, the concentration of CO_2 will decline only slowly and will be expected to level off at a higher steady-state value after several thousand years. The decline to the new steady state is slow because of the slow rate of deep-ocean circulation. A higher steady-state value of atmospheric CO_2 will result because the amount of carbon added to the ocean has significantly increased TC (10%) and because this increase causes a substantial rise in pCO_2 in the surface waters (Figure 5.4). Baes and Killough (1986) used their two-dimensional model involving the chemical and biological processes to calculate that 900 μatm of CO_2 would remain in the atmosphere when steady state had been re-established after the release of 4.1×10^{18} g of fossil carbon. The excess fossil carbon added to the atmosphere-ocean system ultimately may be expected

to react with the carbonate minerals in rocks and ocean sediments, thus shifting the oceans back toward their original composition, but this can be expected to take much longer than a few thousand years (Broecker and Peng 1982, p. 552; Sundquist 1986).

As predictive tools for the next century, the present carbon cycle models are less reliable, mainly because of uncertainties about the nature of circulation in the upper ocean generally and about episodic deep overturning at high latitudes (Sarmiento 1986). The importance of circulation is illustrated by the calculations of Killough and Emanuel (1981) using one-dimensional models with different flow patterns (Figure 5.10, Table 5.3). Another uncertainty, of course, is the rate at which fossil carbon will be burned during this period. In general, the higher the release rate, the higher the airborne fraction is predicted to be.

The farther into the next century that predictions are attempted with models, the better the models must be. The pace of model improvements may be determined by the rate at which observational data can be gathered and assimilated. For example, the enhanced dissolution of calcium carbonate in the upper ocean caused by increasing atmospheric CO_2 would be extremely difficult to predict with a model, but it might readily be detected as a secular increase of TA in certain areas and then parameterized in models. Similarly, feedback effects from climate change are so difficult to predict with present models (see Section 5.6.3) that their onset may be observed before they can be predicted. Finally, the expansion of data on all the transient and natural tracers is likely to permit more rapid improvements in the models than may be gained from increased theoretical understanding of ocean processes.

5.6.3 Feedback Effects and Other Uncertainties

Broecker (1982) was perhaps the first to point out that if for some reason the nutrient (phosphate) content of the ocean had changed significantly over geologic time, then the CO_2 content of the atmosphere would be affected (see Chapter 8). This comes about through changes in biological productivity and the accompanying effect on the surface water deficiencies of TC and TA. Baes and Killough (1986) found with their model that if the nutrient content of the ocean were doubled, then at steady state the atmospheric content of CO_2 would be halved. Significant increases in nutrient content presumably could occur over the very long time periods required for naturally occurring changes in sea level or perhaps more quickly from anthropogenic inputs (see Section 5.4). In any case this feedback would take many centuries to become significant.

Sarmiento and Toggweiler (1984) and Siegenthaler and Wenk (1984) used simple models of the ocean to explore possibly more rapid changes in atmospheric CO_2 that might result from changes in circulation and biological productivity at high latitudes. Their goal was to explain the rapid changes (in a few hundred years or less) in atmospheric CO_2 that have been associated with past climate variations. The models consisted of a high-latitude surface box, one other surface box covering the rest of the ocean, and a single box for the deep ocean. Sarmiento and Toggweiler (1984) explored the dependence of atmospheric CO_2 at steady state on three variables: (1) a circulation flow of water from the deep box to the warm surface box, then to the cold surface box, and back to the deep box; (2) an exchange flow (representing deep overturning) between the cold surface box and the deep box; and (3) a flux of detrital carbon from the cold surface box to the deep box. They found that the CO_2 content of the atmosphere at steady state is generally lowered if the circulating flow is increased, if the overturning flow is decreased, or if the biological productivity of the cold box is increased. In the model of Siegenthaler and Wenk (1984), which is very similar, increased circulating flow or reduced overturning similarly lowers atmospheric CO_2 concentration.

Using their more complex model, Baes and Killough (1986) found that reducing the water flows by half caused about a 10% decrease in atmospheric CO_2 concentration at steady state. The effect is in the same direction as that found for decreased overturning at high latitudes in the two models described above. Although the decreased circulation lowered the biological productivity, it also permitted a larger deficiency of TC and TA to develop in surface waters, and this produced a net uptake of CO_2. When the average water temperature was increased (Baes and Killough 1985), the accompanying increase in atmospheric CO_2 concentration was close to the dependence of 4%/°C for isochemical conditions. The effects of increasing temperature and decreasing circulation that might result from increasing atmospheric CO_2 concentration thus may be opposite to one another.

The models of Sarmiento and Toggweiler (1984) and Siegenthaler and Wenk (1984) suggest that changes in water circulation and biological productivity at high latitudes may have a strong influence on the concentration of CO_2 in the atmosphere. This could be an artifact of the simplicity of their models, yet the more detailed model of Baes and Killough (1985) gives similar responses to changes in circulation and biological productivity insofar as the results may be compared. It is safe to conclude, however, that accurate predictions of the effects of a warming climate on ocean circulation, biological productivity, and atmospheric CO_2 concentrations over the next 100 years will require greater knowledge than is currently available of the

relevant oceanographic processes, coupled with more realistic ocean models to simulate them (Sarmiento 1985).

5.7 NEEDED RESEARCH

5.7.1 Monitoring

From the information presented above, it can be seen that as more detailed models are developed to estimate CO_2 uptake, larger data bases will be required for their validation and calibration (Sarmiento 1986). Although success was achieved in constructing simple models that account for the average behavior of a large number of tracers (Broecker and Peng 1982; Peng et al. 1983), it appears likely that the trend toward more complex models must continue. The consistency of increasingly complex models with the predictions of simpler ones will be the ultimate test of whether the body of tracer data has achieved enough spatial and temporal resolution.

Perhaps the most important deficiency in the present knowledge of the oceans concerns circulation, both the short-term, episodic events that may be important in determining the rate at which CO_2 is sequestered in deep water and the long-term average water movements in the upper ocean that may determine the overall rate of CO_2 uptake elsewhere. Recent ocean studies such as TTO produced an explosion of knowledge about ocean circulation, and such efforts should be strongly supported and continued. This includes not only direct observation of water movements but especially the monitoring of chemical and radiochemical tracers, which are the best indicators of time-averaged effects.

Measurements of the difference in pCO_2 between surface waters and the atmosphere reflect the effect of circulation and the biological processes on the transfer of CO_2 between the ocean and the atmosphere. The spatial and temporal coverage of such measurements should be increased, not only as a means of monitoring this transfer more closely, but also to gain a better understanding of all processes that control it.

More intense monitoring of tracers in the upper ocean seems to be especially desirable. The variations in TC, TA, and nutrients may permit the tracking of excess amounts of carbon entering the oceans. These data will also provide more information about biological productivity, the export of biocarbon from the surface, and the variability of the Redfield ratios (e.g., Takahashi et al. 1985).

Finally, efforts to determine any changes that may be occurring in the rate of accumulation of carbon in a variety of coastal environments should continue. Unfortunately, the present methods of monitoring have large uncertainties, and improved methods are badly needed (see Chapter 11).

5.7.2 Improved Models

As noted in Section 5.6.2, no simple CO_2-ocean model can encompass all of the time scales of the various processes that occur in the oceanic carbon cycle. Model improvement should therefore be guided by the time scale that is of interest. With regard to human-induced climatic warming, the most important question is probably how much CO_2 can the oceans take up in the next century for a given fossil fuel release.

In 100 or 200 years, the oceans do not undergo a full mixing cycle. Excess carbon that sinks into the abyssal ocean can probably be regarded as permanently removed. The deep ocean can, therefore, be treated as a moderate number of well-mixed boxes. The mixed layer should be modeled in enough detail to represent the variation of temperature, TC, and TA with latitude. There remains the main oceanic thermocline. The redistribution of excess CO_2 from the atmosphere to a location in this region occurs on a time scale of 10 to 100 years. It becomes important, therefore, to explore the pattern of penetration in some detail. Until now, global carbon cycle models have represented the main oceanic thermocline either with five or fewer boxes or with a one-dimensional diffusive medium. Neither alternative may be an adequate approximation of reality. It is important in future modeling to improve the resolution with which this region is represented (see Chapter 7).

In the study of dynamic oceanography, models of the circulation in one or both of the major oceans were constructed with a resolution down to a few degrees of latitude and longitude and with several vertical layers (Bryan and Lewis 1979; Sarmiento and Bryan 1982). These models permit greatly detailed descriptions of oceanic flow patterns. It is, in principle, possible to equip such models to simulate the movement of carbon, radiocarbon, and the other radiochemical and chemical tracers as they are transported in the water and affected by the biological processes (Sarmiento 1986). Preliminary results with a grid point model of the oceanic carbon cycle have been presented by Maier-Reimer (1984), and he and coworkers are continuing development of such a model. Its validation will depend on the availability of large amounts of data on water circulation and the temporal and spatial distribution of tracers.

Despite their crudeness, box models have certain advantages over large grid-point models. The pattern of tracer distributions on which the box models are based reflect the mixing and circulation in the ocean as averaged over time. For the CO_2-climate issue, this is what is needed provided the averaging time is properly chosen. Moreover, it is possible to use temperature and salinity data and the density fields that they imply in such models to deduce time-averaged water and heat flows, much as is done in the large grid-point models. The

introduction of such hydrodynamic and thermodynamic constraints into one- and two-dimensional box models is now under way (Volk and Hoffert 1985) and should, in the near future, produce more credible box models of the ocean.

Three-dimensional CO_2-ocean models will probably be needed to test the predictions of ocean uptake made by simpler models. Sarmiento (1986) reviewed the current state of development of such models and concludes that many years will be needed to produce a useful predictor of the uptake of CO_2. In the meantime, one must "retain a healthy scepticism about the predictions of simpler models."

5.8 SUMMARY AND RESEARCH RECOMMENDATIONS

The oceans are the ultimate sink for most of the excess CO_2 released to the atmosphere by humans. The rate at which this excess can be taken up will depend on ocean circulation, chemistry, and biology, all of which affect the distribution of carbon. Prediction of the uptake will require a sound knowledge of these processes and the development of models that can represent them adequately.

5.8.1 Ocean Circulation and Chemistry

Ocean circulation is the most important single factor that determines the rate of uptake of excess atmospheric CO_2. Its gross features are suggested by the idealized representation of constant-density (isopycnal) surfaces in Figure 5.1. There is a well-mixed surface layer that is about 30 m deep at the Equator and increases in depth toward the poles. Circulation occurs by water movement along isopycnals and by cross-isopycnal surface mixing. Deep overturning and deep-water formation occur only in winter and primarily in the north and south Atlantic Ocean. Although knowledge of ocean circulation in all of its complexity is growing rapidly through the acquisition of data and the development of ever more sophisticated models, a coherent picture sufficient for the present need has not yet emerged.

Because the amount of carbon in the oceans greatly exceeds that in the atmosphere, the oceans virtually control the atmospheric content of CO_2 at steady state. About one-seventh of the carbon content of the well-mixed surface waters is exchanged with the atmosphere each year, a process that is reflected by the global distribution of the differences between pCO_2 in surface waters and that in the atmosphere (Figure 5.2).

The value of pCO_2 in surface waters is determined by the local concentrations of the three forms of dissolved inorganic carbon (carbonic acid, bicarbonate ion, and carbonate ion), along with the temperature and salinity. Because the chemical reactions of these forms of carbon (Equations [5.1]-[5.3]) are rapid, the four quantities pCO_2, pH, total dissolved inorganic carbon (TC), and total alkalinity (TA) are interrelated in a precisely known way. Only two of the four can be treated as independent variables (usually TC and TA) at a given temperature, pressure, and salinity (Figure 5.3). At constant TA, pCO_2 increases rapidly as TC increases (Figure 5.4).

The biological processes in surface waters lower TC by photosynthesis (Equation [5.7]) and both TC and TA by the precipitation of calcium carbonate (the reverse of Equation [5.6]). The net effect is to substantially lower pCO_2 in surface waters, and therefore in the atmosphere, below what it would otherwise be (Figure 5.3). In most locations the biological productivity is controlled by the supply of nutrients in upwelling water. It is important to understand and to simulate with models the consequent interrelationships among biological productivity, circulation, and pCO_2.

The rate at which the organic carbon formed in the surface waters is exported to the deep ocean, currently estimated to be in the range of 2×10^{15} to 6×10^{15} g/year, is determined primarily by the rate at which limiting nutrients (phosphate and nitrate) are delivered to the surface waters by circulation. Most of the descending organic detritus is decomposed in the first 1000 m of depth, whereas most of the calcium carbonate detritus is dissolved at greater depths. The profiles of nutrients, TC, TA, and dissolved oxygen (Figure 5.6) reflect these processes.

5.8.2 Ocean Tracers

The rapidly growing knowledge of ocean circulation and the movement of carbon therein has been gained through the use of many tracers. The important naturally occurring tracers include salinity and temperature, variations of which permit the tracking of water masses; the quantities (Figure 5.6) affected by chemical and biological processes; and radioactive materials such as ^{14}C and ^{39}Ar produced in the atmosphere by cosmic ray bombardment. Important transient tracers include the chlorofluorocarbons released in substantial quantities in the past several decades, the excess CO_2 released (Figure 5.7), and radioactive tritium (Figure 5.8) and ^{14}C produced by atmospheric nuclear testing.

Knowledge of the distribution of all these and other tracers in the oceans came principally from the extensive GEOSECS survey carried out in the early 1970s. Two recently completed surveys, TTO in the North Atlantic and the Hawaii to Tahiti Shuttle Experiment, provided much additional information and

revealed secular changes that occurred in the decade since the GEOSECS survey was carried out.

The distribution of natural ^{14}C in the oceans gives a direct measurement of the rate of exchange of CO_2 between the surface waters and the atmosphere ($\sim$100 $\times$ 10^{15} g/year) and the replacement times of waters in the ocean basins (275-510 years). The transient tracer tritium proved to be the most revealing indicator of the deep overturning of water at high latitudes and the rapid advective movement of deep water masses (Figure 5.8).

Although the secular increase of pCO_2 at selected surface water locations (Figure 5.7) gives a clear indication of the entry of excess carbon into the oceans, efforts to detect excess carbon at greater depths have been hampered thus far by the smallness of the signal compared with the effects arising from spatial and temporal fluctuations of TC and TA that occur naturally in surface waters.

The data on ^{14}C and tritium were widely applied to the calibration and validation of CO_2-ocean models. Data on the distribution of nutrients, TC, TA, and oxygen (Figure 5.6) are being used increasingly in models that simulate the biological processes. As more sophisticated three-dimensional models are developed, the large amount of data on salinity and temperature can also be included in more detailed simulations of ocean circulation.

5.8.3 Continental Shelves

In addition to the excess CO_2 taken up directly from the atmosphere by the oceans, excess carbon may be sequestered on the continental shelves because of anthropogenic increases in the carbon and nutrient content of river outflow. The enhanced carbon uptakes are estimated (Table 5.2) to be 0 $\times$ 10^{15} to 0.3 $\times$ 10^{15} g/year from rivers, 0.1 $\times$ 10^{15} to 0.3 $\times$ 10^{15} g/year from NPP in coastal regions, and 0.1 $\times$ 10^{15} to 0.2 $\times$ 10^{15} g/year from atmospheric CO_2 consumed by reaction with carbonate minerals in these regions. The total uptake, 0.2$\times$ 10^{15} to 0.8 $\times$ 10^{15} g/year, may be appreciable but remains quite uncertain (also see Chapters 8, 10, and 11).

5.8.4 Model Development

Given the intricacies of ocean topography, circulation, chemistry, and biology, credible estimates of the response of the world's oceans to increasing atmospheric CO_2 can come only from models. Until recently such models were usually one-dimensional arrangements of boxes in which water flows and the exchange of CO_2 with the atmosphere were the principal adjustable parameters. The most popular were the so-called box-diffusion models (Figure 5.10, Model B-D) in which water flows are adjusted in such a way as to simulate a diffusion of excess carbon into the deep ocean. More realistic simulations of the world oceans were achieved by linking several such arrays of boxes, each representing different ocean regions (e.g., tropical, temperate, polar), or by adding direct connections between polar surface waters and the deep-ocean boxes. Although the arrangement of water flows and other differences have a significant effect on the ability of such ocean models to take up CO_2 (Figure 5.10, Table 5.3), the rate of uptake is generally substantially less than that necessary to account for the estimated anthropogenic carbon releases during the last 100 years.

The more sophisticated CO_2-ocean models developed recently included the chemical and biological processes that affect the distribution of carbon. Although this introduces additional adjustable parameters and fluxes between boxes, many of the fluxes are related by water movement. These models evolved from the one-dimensional box-diffusion configuration, first to two-dimensional (depth $\times$ latitude) arrangements containing as many as 91 boxes, and then to three-dimensional representations of an ocean basin. They were found to take up excess CO_2 at no greater, and sometimes a lesser, rate than do the simpler models.

Because the response times of the various processes in the ocean affecting the uptake of CO_2 differ widely, simple ocean models cannot be expected to describe ocean behavior reliably over time periods ranging from decades to millenia. For the period of greatest interest, the next 100 or 200 years, more sophisticated models will be needed than have been developed thus far.

5.8.5 Future Uptake of CO_2

Direct evidence of the uptake of CO_2 by the oceans, such as the appearance of excess carbon and perhaps increased alkalinity from the dissolution of carbonate minerals, should become distinguishable above background fluctuations with the passage of time and increased monitoring activities. In particular, it is important that monitoring be increased in both spatial and temporal coverage.

The present CO_2-ocean models are not yet reliable predictive tools for the next century, mainly because of uncertainties about the nature of circulation in the upper ocean generally and episodic deep overturning at high latitudes. Feedback effects of climate warming on ocean response are even more difficult to predict than is the primary uptake of CO_2. These will require more knowledge and better modeling of the biological processes as well. For the longer-term future, the level of atmospheric CO_2 reached will depend primarily on the amount of fossil carbon released to the atmosphere.

5.8.6 Conclusions

To paraphrase one of the reviewers of this chapter (H. G. Östlund): There is a tendency, especially within that part of the community not directly involved in the technical and scientific details, to believe that modeling alone will forecast the CO_2-climate relation. This is definitely not the case. Nor is it true that measurements alone will suffice. Rather, it will be the hand-in-hand cooperation between experimentalists and modelers that will advance our capability to make such a forecast.

It is clear from the present state of development in both observational oceanography and ocean modeling that a stronger cooperative effort must be developed. The detailed information gained from oceanographic surveys cannot be interpreted without a model, and, at the same time, there is little hope of developing the needed CO_2-ocean models without much more tracer data.

Just how realistic the model or how extensive the data that will be needed cannot be foreseen. When formulated, an adequate model will be recognized because it yields the same results for CO_2 uptake as a more detailed model. The levels of temporal and hydrographic resolution in the data needed to support the models should become apparent as the development proceeds.

All that can be said with certainty at present is that the state of the art of oceanography and that of modeling are not yet sufficient for the task of predicting the future response of the oceans to increasing atmospheric CO_2 concentration.

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6. CARBON DIOXIDE EXCHANGE BETWEEN THE ATMOSPHERE AND TERRESTRIAL ECOSYSTEMS

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6.1 INTRODUCTION

This chapter addresses the role of the terrestrial biota and soils in the global carbon cycle. Its emphasis reflects that of research on this topic—it deals primarily with estimates of the amount of carbon released to the atmosphere in recent years as a result of changes in land use. The compilation of data from ecosystems throughout the world and the improvement of information on tropical deforestation have narrowed the range of estimates of a biotic release of carbon to between 0.6×10^{15} and 2.6×10^{15} g/year. The lower end of the range is now consistent, within limits of error, with the other terms in the global carbon balance.

The agreement is not altogether satisfying, however, because the flux of carbon from land-use change may only be part of the net biotic flux. The response of terrestrial systems to changes in climate could dwarf the fluxes from changes in land use, yet ecologists are not in agreement about the direction of the climate response. Current understanding of the biota may be sufficient to predict future concentrations of carbon dioxide (CO_2) only in the absence of climate change.

This chapter reviews estimates of the carbon stored in the terrestrial biota and in soils, estimates of the annual flux of carbon between the atmosphere and undisturbed terrestrial ecosystems, and estimates of the net flux of carbon between the atmosphere and disturbed systems. The chapter also reviews three potential methods for measuring a net flux that includes not only the flux due to changes in land use but also the fluxes due to other changes (i.e., changes in climate, atmospheric CO_2 concentrations, or other environmental factors). Although resolution of the net flux of carbon from changes in land use is coarse and should be refined with further research, the need to address the biotic response to climate change is critical.

6.2 DEFINITIONS

The organic carbon of terrestrial ecosystems is stored in the biota and soils. Because the vegetation contains more than 99.9% of the carbon in terrestrial biota, other components are not considered. In this report all living material and standing dead material are considered to be a part of the biota; fallen and belowground dead material are included in soils.

Literature sources used to estimate the carbon content of vegetation commonly report results in units of dry mass per unit area. The carbon content of plant dry mass is variable. For example, the carbon content of leaf material (42%) is lower than that of woody or root material (47-52%, Ajtay et al. 1979). Global estimates of the carbon content of forests used both 45% (Whittaker and Likens 1973, Ajtay et al. 1979) and 50% (Brown and Lugo 1982) to convert dry mass to carbon. Data compiled by Ajtay et al. (1979, their Table 5.4) suggest that the carbon content of forest ecosystems (aboveground mass) is approximately 50% of dry mass. However, this conversion factor could cause an error in the carbon content of forests by as much as 10%, and some refinement may be desirable.

Soil organic matter exists in many forms, including fresh, incompletely decomposed, plant matter on the soil surface (litter or detritus) and humus dispersed through the mineral soil horizons. The turnover of these fractions varies from less than 1 year to greater than 1000 years. The pool of soil organic matter includes peat, which in many areas accumulates to great depths, forming a continuum with the geologic pools of lignite and coal. In arid and semiarid regions, carbon also is stored as carbonate minerals that precipitate in the lower soil horizons in deposits known generally as caliche or calcrete (also see Chapter 8—all chapter references are to this volume).

6.3 AMOUNT OF CARBON IN THE BIOTA AND SOILS OF THE EARTH

6.3.1 Biota

Estimates of the amount of carbon in terrestrial biota, worldwide, range from 421×10^{15} to 827×10^{15} g C (Table 6.1). The differences among these estimates are due almost entirely to estimates for tropical forests, which have received less study than ecosystems in the temperate zone. The range of estimates for carbon held in tropical forests is 102×10^{15} to 460×10^{15} g C, whereas the range for all other systems is 319×10^{15} to 368×10^{15} g C (Table 6.1).

The methodology used by different investigators to estimate the carbon content of tropical forests was (1) classification of the forests into different types, (2) determination of the areas in each type, (3) determination of the carbon stocks (carbon content of vegetation per unit area) in each type from literature values, and (4) calculation of carbon content as the sum of the products of areas and their carbon stocks. Because each of the six studies listed in Table 6.1 used a different scheme of classification, comparisons among studies are almost impossible to make for any category other than the tropics as a whole. Even if each of the six studies had used the same literature values for carbon stocks, the division of these values among different categories and the assignment of areas to each would be unique for each study.

Table 6.1
Estimates of the Amount of Carbon Held in the World's Terrestrial Biota

Tropical Forests			Ecosystems Other Than Tropical Forests	All Terrestrial Ecosystems
Area (10^8 ha)	Average Carbon Stocks (10^6 g/ha)	Carbon Total (10^{15} g)	Carbon Total (10^{15} g)	Carbon Total (10^{15} g)
24.5[a]	188[a]	460[a]	364[a]	827[a]
14.5[b]	165[b]	240[b]	319[b]	560[b]
15.1[c]	125[c]	189[c]	368[c]	558[c]
18.4[d]	124[d]	228[d]	-	-
	114[e]	-	-	-
19.3[f]	53[f]	102[f]	-	(421-470)[g]

[a]Whittaker and Likens (1973).
[b]Ajtay et al. (1979).
[c]Olson et al. (1983).
[d]Brown and Lugo (1982).
[e]Olson et al. (1978).
[f]Brown and Lugo (1984).
[g]Values of 319 × 10^{15} to 368 × 10^{15} g C assumed for systems other than tropical forests.

Part of the differences among the estimates in Table 6.1 can be attributed to the availability of a more extensive data base for the more recent studies. For example, 22 of 37 of the biomass studies summarized in Brown and Lugo (1982) were published after the review by Whittaker and Likens (1973).

Most of the studies were based on major units of vegetation (Whittaker and Likens 1973; Olson et al. 1978; Ajtay et al. 1979), but Brown and Lugo (1982) used environmental factors to categorize types of forests. They used the life zone approach of Holdridge (1947) and showed that the carbon stocks of tropical vegetation were highly correlated with life zones (Figure 6.1). The highest carbon contents were associated with tropical moist forests, with carbon content decreasing slightly in wetter life zones [temperature/precipitation (T/P) ratio < 0.01] and sharply in drier ones (T/P ratio between 0.012 and 0.035).

The data used to estimate the carbon content of tropical forests are insufficient. Field studies are not numerous (less than 40), typically cover small areas (less than 30 ha), and tend to be concentrated in a few life zones. Much more information, from a broader geographical area and from more and larger plots, is available on standing timber volumes of tropical forests. This type of information served as the basis for an alternative estimate of the carbon content of tropical forests (Brown and Lugo 1984). In a recent assessment of tropical forests the Food and Agriculture

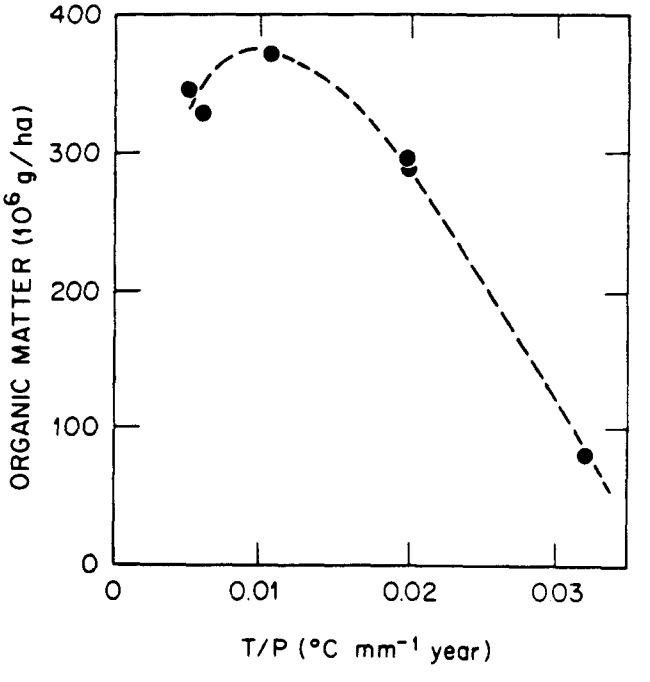


Figure 6.1. The relationship between the organic matter content of tropical forest vegetation and temperature to precipitation ratio (T/P). The relationship is statistically significant. The T/P ratio is an index of the life zone. Low ratios correspond to humid climates (T/P < 0.02), and humidity decreases as the ratios increase (left to right on the X-axis) (adapted from Brown and Lugo 1982).

Organization/United Nations Environment Programme (FAO/UNEP) (FAO 1981) gave detailed information on forest areas and on their corresponding stand volumes (volume of the free bole including bark to a top diameter of 7 cm for all living trees with a diameter ≥ 10 cm) for the three tropical regions of the world. Stand volumes were converted to total biomass by multiplying volume by an average wood density and by an expansion factor: 1.6 for closed forests and 3.0 for open forests (Brown and Lugo 1984). A summary of the forest areas, volumes, and resulting biomass values for the different forest types is given in Brown and Lugo (1984). Total volume-derived biomass for all tropical forests was about 205×10^{15} g dry weight or 102×10^{15} g C according to these conversions. The average carbon stocks were 53×10^6 g/ha, a value considerably lower than the 124×10^6 to 188×10^6 g/ha obtained from estimates based on biomass (Table 6.1).

The major discrepancy between the estimate based on volume and the estimate based on mass occurs in the broadleaf closed forests. The carbon stocks of these forests based on biomass studies was 155×10^6 to 164×10^6 g/ha compared with the volume-based estimate of 88×10^6 g/ha. The discrepancy between the two estimates for the open forests was not as large: 40×10^6 to 46×10^6 g/ha (based on biomass studies) compared with 31×10^6 g/ha (Brown and Lugo 1984).

The differences in estimated carbon stocks are one of the largest contributors to the uncertainties in estimating the current net flux between biota and the atmosphere. The reasons for the discrepancies are as yet unclear. Potential sources of error exist in the original volume estimates, the wood densities, and the expansion factors used in the volume-based estimates. There is considerable speculation but little documentation of the biases introduced in the selection of stands by foresters versus ecologists. Are the volume-derived estimates low or are the mass-derived estimates high? The uncertainty because of the wood densities is minimal (less than 5%; Brown and Lugo 1984). On the other hand, the expansion factor used to convert merchantable timber to total biomass can vary significantly with the size class of the stand (Johnson and Sharpe 1983). The expansion factor determined by Brown and Lugo (1984) from ecological studies in the tropics was 1.6. Johnson and Sharpe (1983) calculated an average expansion factor of 2.3 (excluding litter) for forests in the eastern and western United States. This latter value (2.3) was higher than that (~1.7) used previously for midlatitude forests, primarily because it included carbon in trees of less than 12.7 cm diameter at breast height (dbh), in nonmerchantable trees, and in nonmerchantable portions of merchantable trees (e.g., branches and leaves). If the expansion factors for the closed forests of the tropics were as large as 2.3, the weighted carbon stocks for undisturbed closed forests would be 126×10^6 g/ha, closer to the range of 155×10^6 to 164×10^6 g/ha based on biomass studies. Recent extensive measurements in tropical moist forests of Colombia and Venezuela (Saldarriaga 1985) suggest the higher estimates for undisturbed forests.

In summary, the carbon stored in tropical forest vegetation is probably within the range of 100×10^{15} to 200×10^{15} g. The average carbon stocks of tropical closed forests are between 88×10^6 and 160×10^6 g/ha, and the carbon stocks of open forests vary from 30×10^6 to 45×10^6 g/ha. To narrow this range, field research is needed to determine the expansion factors required to convert volumes to total biomass.

Despite the uncertainty associated with the total amount of carbon in tropical forest vegetation, estimates of the change in that amount can be made with considerably more precision (Section 6.5.1). Changes in the area of forests are moderately well known and the changes in carbon storage associated with deforestation are large.

6.3.2 Soils

There have been several attempts to estimate the storage of soil organic matter and litter in world ecosystems. Most of these estimates fall within the range of 1200×10^{15} to 2200×10^{15} g C, based on compilations by soil groups (Bohn 1976, 1982; Buringh 1984), by vegetation types (Schlesinger 1977, 1984; Ajtay et al. 1979; Bolin et al. 1979), and by life zones (Post et al. 1982; Zinke et al. 1984), or based on modeling of plant production and decomposition (Meentemeyer et al. 1981). The most extensive data set is that used by Zinke et al. (1984) (also see Post et al. 1982), which included values from 3583 measured soil profiles and allowed patterns of soil organic content to be mapped for world life zones (Figure 6.2). The best estimate of the pool of soil organic carbon is probably 1500×10^{15} g, with a ± 20% uncertainty, much of which lies in the treatment of Histosols, the peatland soils of the world (Armentano 1979; also see Chapter 8, this volume). In vegetation-based compilations, peatlands are included in categories such as tundra, boreal forests, and wetlands. These categories contain 489×10^{15} g C, or 34% of Schlesinger's (1977) estimate of the global soil carbon pool. The content of soil organic matter in tropical soils is fairly well known; differences among estimates of the global pool in tropical soils are mainly related to differences in the area assigned to tropical regions (Schlesinger 1986; Brown and Lugo 1982; Post et al. 1982).

Although not organic, caliche develops in the soil

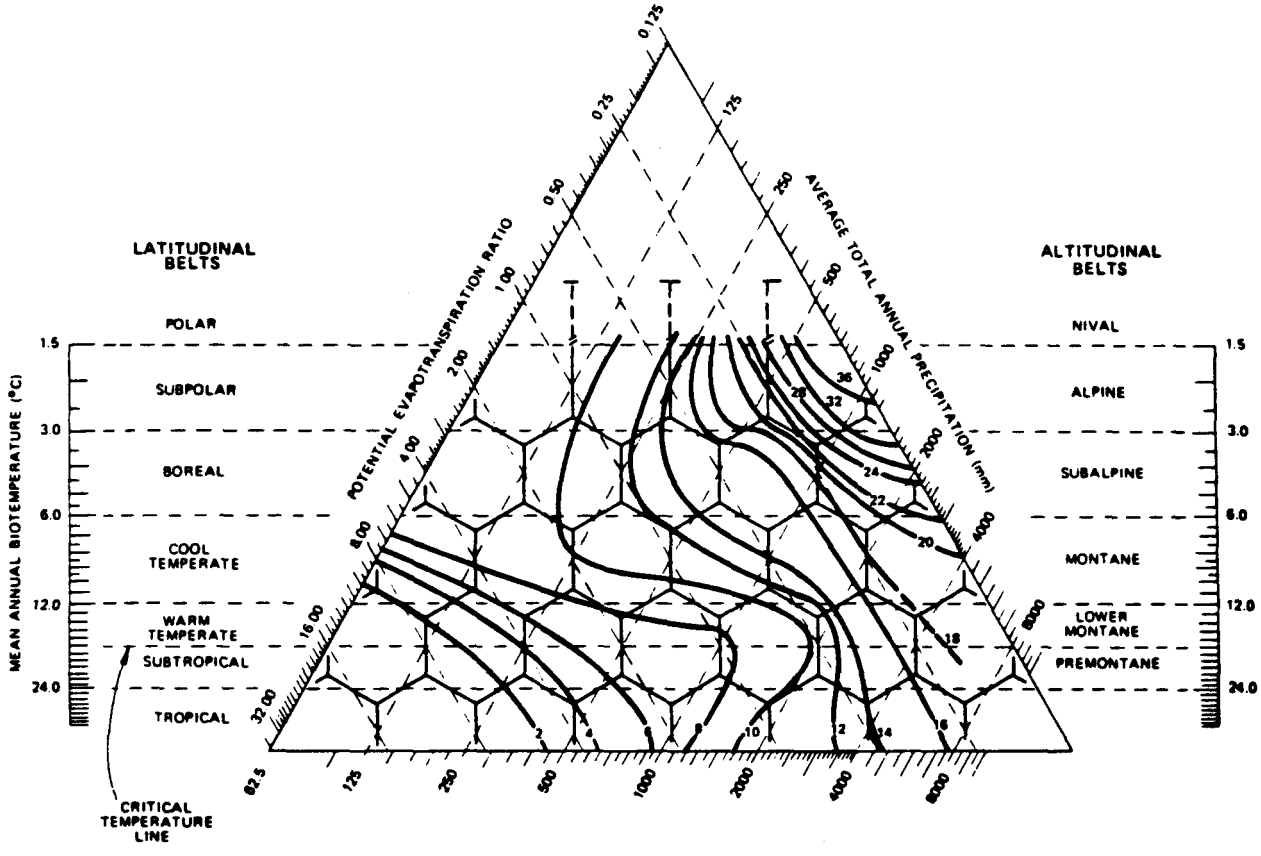


Figure 6.2. Contours of soil carbon (kg/m^2) plotted on a Holdridge (1967) life-zone chart (source: Zinke et al. 1984).

horizons affected by biotic processes and must be included in estimates of the soil carbon pool in the biosphere. Caliche is composed primarily of calcium carbonate that precipitates in a variety of forms in arid soils (Lattman 1973). Schlesinger (1982) found that the soils of Arizona contained 4-5.7 times more carbon stored as carbonate than as soil organic matter. Assuming that the Arizona data can be extrapolated to arid soils in other areas of the world, the amount of carbon stored in caliche of the world's deserts is from 800×10^{15} to 930×10^{15} g. A range in uncertainty in this estimate may extend from 550×10^{15} to 1800×10^{15} g C (Yaalon 1981; Schlesinger 1985). Some of these deposits are very old; as in the case of peat, caliche spans the realm between the geologic and biospheric carbon pools (see Chapter 8).

6.4 FLUX OF CARBON BETWEEN THE ATMOSPHERE AND UNDISTURBED TERRESTRIAL ECOSYSTEMS

Gross primary production (GPP) is the rate at which green plants transform inorganic CO_2 to organic carbon. It corresponds to the rate of carboxylation of ribulose biphosphate. The GPP is not the rate of exchange of carbon between plants and the atmosphere, because some of the CO_2 molecules fixed in photosynthesis are generated inside plant tissue from the oxidation of organic carbon. Photorespiration of leaf tissue in the light is one source of CO_2 within the plant. Mitochondrial respiration is another potential source of CO_2 within plant tissue.

The measurement of GPP is difficult because photosynthesis and respiration occur simultaneously in leaves. However, photorespiration is generally small enough that GPP can be estimated from field measurements of the rate of CO_2 uptake in the light plus the rate of CO_2 release in the dark. This calculation assumes that the rate of mitochondrial respiration at night applies in the light as well. It underestimates GPP by the rate of photorespiration.

Global estimates of GPP for the world's terrestrial ecosystems range from 90×10^{15} to 120×10^{15} g C/year, approximately twice net primary production

Table 6.2
Net Primary Productivity of the Earth's Terrestrial Vegetation

Net Primary Productivity (10^{15} g C/year)	Source
48.3-52.9	Whittaker and Likens (1973, 1975)
45.1	Lieth (1975)
53.4-56.0	Lieth and Box (1972); Lieth (1973)
59.9	Ajtay et al. (1979)
60.2	Olson et al. (1983)
56.7-62.1	Esser et al. (1982)

(NPP) (Table 6.2). The NPP is the rate at which carbon is fixed less than that respired by plants. The average turnover time of carbon in the vegetation component of terrestrial systems is somewhere between 5 and 20 years, depending on whether GPP or NPP is used as the annual input.

Most estimates of NPP are not based on measurements of CO_2 exchange but on measurements of growth or of change in biomass. Although such measurements require less sophisticated technology than direct measurements of CO_2 exchange, they are complicated by the mortality and disappearance of plant tissues during the course of a growing season or year (also see Chapter 8). Annual GPP and NPP include the turnover of plants and their parts, as well as the exudation of soluble organics from live tissues. The GPP includes, in addition, the turnover of organic molecules within plants (i.e., photorespiration and nonphotorespiration).

Although estimates of GPP are low by the amount of photorespiration, they still represent overestimates of the exchange of carbon between terrestrial ecosystems and the atmosphere, because some carbon fixed in photosynthesis is derived from respiration within the plant canopy or soil. The fraction of carbon in NPP that comes from sources other than the atmosphere is commonly 30-40% (Monteith et al. 1964; Biscoe et al. 1975; Monteith 1976; Houghton and Woodwell 1980), but was as high as 80% in a tundra ecosystem (Coyne and Kelley 1975). There is isotopic evidence for this cycling within plant canopies (e.g., Vogel 1978; Medina and Minchin 1980), although Francey and Farquhar (1982) argue that vertical differences in the isotopic ratios of leaves reflect, in part, differences in the exposure of leaves to sunlight and not necessarily differences caused by respiratory versus atmospheric sources of CO_2.

If 30-40% of the carbon fixed in GPP is cycled within the plant canopy, including cycling within leaf tissues, the exchange between terrestrial ecosystems and the atmosphere may be 60×10^{15} to 80×10^{15} g C/year rather than 90×10^{15} to 120×10^{15} g C/year. This exchange is limited to a small volume of the atmosphere.

Estimates of the exchange of carbon between the biota and the atmosphere, based on seasonal, latitudinal, and altitudinal variation in concentrations of CO_2, indicate a seasonal exchange of only 4×10^{15} to 6×10^{15} g C for the Northern Hemisphere (Bolin and Keeling 1963; Pearman and Hyson 1980). GPP for the terrestrial biota in the same area is 26×10^{15} to 32×10^{15} g C/year (Whittaker and Likens 1973; Ajtay et al. 1979). The difference between the net seasonal exchange and GPP is related to the recycling of carbon within the lower atmosphere (Section 6.5.3).

The carbon from NPP either accumulates in vegetation or soil or is released to the atmosphere through heterotrophic respiration or fires. If most undisturbed systems neither accumulate nor lose carbon in the long term, an amount of organic carbon equivalent to that generated from NPP is decomposed each year. Exceptions to this are described below, and the more general question of steady state is addressed in Section 6.5.2. Under the assumption of steady state, the loss of CO_2 to the atmosphere by decomposition should equal the inputs of plant debris to soil, because losses to streams and groundwaters are relatively small (Schlesinger and Melack 1981; Meybeck 1982; see also Chapters 5 and 8). Estimates of litter production are, therefore, independent estimates of NPP. Reiners (1973) estimated that the global production of plant litter above- and belowground was 37.5×10^{15} g C/year, based on a compilation of data from field studies in natural ecosystems. Meentemeyer et al. (1982) calculated a regression between aboveground litter production and actual evapotranspiration in world ecosystems. Coupled to a map of the global distribution of actual evapotranspiration, their regression indicated a global flux of 54.8×10^{15} g C/year from aboveground plant debris. However, root and mycorrhizal respiration along with decomposition of large amounts of root detritus (e.g., Edwards and Harris 1977), poorly measured in most field studies, produce an additional CO_2 flux from soils that must be considered. Thus, the release of CO_2 from world soils may be as much as 75×10^{15} g C/year (Schlesinger 1977). In steady-state conditions, the flux of carbon through the pool of soil organic matter should be close to the annual NPP of land ecosystems (45×10^{15} to 62×10^{15} g C/year) (Table 6.2), but uncertainty may actually be greater (37.5×10^{15} to 75×10^{15} g C/year).

Not all soils are currently in steady state for soil organic matter. However, before widespread human occupation, significant net carbon accumulations ($> 0.1 \times 10^{15}$ g C/year) were probably confined to peatland and wetland soils, where current rates of accumulation range from 5 to 150 g C m^{-2} $year^{-1}$ (Armentano 1979; Armentano et al. 1984; Miller et al. 1984). Estimates of

Table 6.3
Estimates of the Current Net Annual Flux of Carbon Between the Atmosphere and Terrestrial Ecosystems (10^{15} g C)

World Total	Tropical Zones	Temperate and Boreal Zones	Source
0.4-4			Adams et al. (1977)
0.4-1.6	0.4-1.6	0	Bolin (1977)
	6		Brunig (1977)
2.5-20			Woodwell and Houghton (1977)
1.9	1.8[a]	0.1[a]	Wong (1978)
7.8 4 to 8 probable 2 to 18 possible	5.6	2.2	Woodwell et al. (1978)
	2.6		Hampicke (1979)
2.4			Pankrath (1979)
		−0.19 to −1.2	Armentano and Ralston (1980)
		−1.6 to −1.9	Johnson and Sharpe (1983)
0.9-1.6			Chan et al. (1980)
−2 to 2	0.8-1.6		Seiler and Crutzen (1980)
	0.4-4.1		Hall et al. (1985)
2.2-4.7			Moore et al. (1981)
0.8 (0.5-2.0)	2.0[b]	−1.2[b]	Olson (1982)
1.8-4.7	1.3-4.2	0.5	Houghton et al. (1983) Woodwell et al. (1983a)
	0.6-1.1		Molofsky et al. (1984)
	1.2-1.5		Detwiler et al. (1985)
	0.5-1.8		Detwiler et al. (1984)
	0.9-2.5		Houghton et al. (1985)
		0.15	Melillo et al. (in press)

Note: Positive values indicate a net release to the atmosphere.
[a]Release of carbon from soils (0.4×10^{15} g) assigned to tropical and temperate zones in the proportion 3:1.
[b]Release of carbon from agricultural soils (0.3×10^{15} g) assigned to tropical and temperate zones in proportion 2:1.

the areas of these ecosystems suggest that the average prehistoric accumulation of carbon in organic soils of the world was 0.14×10^{15} g C/year (Armentano 1980) to 0.3×10^{15} g C/year (Miller 1981; also see Chapter 8).

The turnover time of soil organic carbon is 20 to 40 years (pool divided by the annual input of plant debris). However, ^{14}C dating shows the vast majority of the soil organic matter to be older, because most material decomposes within a few years, leaving only small increments, to long-term storage. Before widespread human disturbance of soils, the net increase in soil carbon was estimated at 0.01-0.02%/year, principally confined to peatlands. Similarly, Schlesinger (1982, 1985) estimated that the rate of carbon storage in caliche is about 0.01×10^{15} to 0.02×10^{15} g C/year, with a turnover time of 85,000 years (also see Chapter 8).

6.5 FLUX OF CARBON BETWEEN THE ATMOSPHERE AND DISTURBED ECOSYSTEMS

When forests are cleared for agricultural crops, the carbon stored in trees is oxidized and released int

he atmosphere, either rapidly if the trees are burned, or slowly if they are left on the ground to decay or are converted to paper, lumber, and other wood products. Similarly, some of the organic matter of soils is oxidized after cultivation. Reductions in terrestrial carbon stocks occur after forest harvest and forest clearing for cropland, pasture, or other uses. Forest regrowth following harvest, abandonment of agriculture, and establishment of plantations increases carbon storage in vegetation and soils. The balance between the clearing and regrowth of forests is the major factor in determining changes in the net storage of carbon in terrestrial systems. Most carbon lost from terrestrial systems is released as CO_2 into the atmosphere; increased concentrations of carbon in runoff account for only a small portion of the carbon lost from disturbed systems (see Section 6.5.1).

The net flux of carbon between terrestrial ecosystems and the atmosphere has been impossible to measure directly, but many methods have been used to estimate the flux (or a portion of it) indirectly. Early global models based on atmospheric and oceanographic data required land biota to accumulate a fraction of the carbon released from combustion of fossil fuels. More recent studies incorporating data on $^{13}C/^{12}C$ ratios in tree rings have generally shown that the biota has released carbon to the atmosphere, although the pattern of release through time varies considerably among analyses (see Chapter 10).

An independent approach toward defining the role of the biota was through an analysis of land-use change together with an accounting of the resultant changes in carbon in vegetation and soil. Estimates of the current net annual flux varied from a high of 20×10^{15} g C to a low of -2.0×10^{15} g C (Table 6.3). Various methods and flux estimates will be reviewed and analyzed in the following section.

6.5.1 Analyses Based on Land-Use Change

Estimates of the net biotic flux of carbon are shown in Table 6.3. Although many analyses based on land-use change were made, passage of time has made some of them obsolete. Others are not documented in enough detail to show how the authors obtained their results or, therefore, to allow identification of similarities and differences among studies. In the review that follows, the major comparisons will be of the analyses by Armentano and Ralston (1980), Houghton et al. (1985), Richards et al. (1983), Detwiler et al. (1984), Molofsky et al. (1984), and Melillo et al. (in press). The major reasons for differences among these estimates seem to be differences in the data used for carbon stocks in vegetation, for changes in the carbon content of soils following disturbance, for rates of deforestation, and for rates of decay of organic matter. Different methodologies and assumptions of steady state also explain some of the differences.

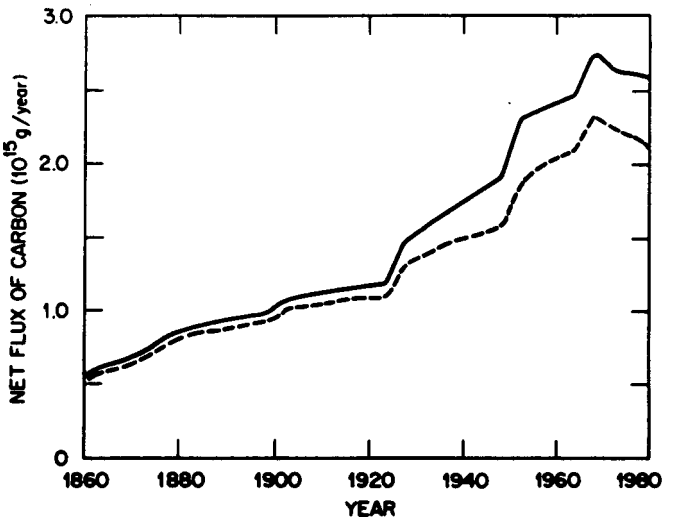


Figure 6.3. The annual net release of carbon from the terrestrial biota and soils of the Earth according to different assumptions (from Houghton 1985): (solid line) estimates of biomass from Whittaker and Likens (1973); (dashed line) estimates of tropical biomass derived from timber volumes reported by FAO/UNEP (FAO 1981).

6.5.1.1 *Carbon Stocks in Vegetation*

Recent reviews of the carbon stocks of vegetation (Ajtay et al. 1979, Brown and Lugo 1982; Olson et al. 1983) indicate that stocks reported by Whittaker and Likens (1973) are too high and that estimates of the flux of carbon based on these values (e.g., Seiler and Crutzen 1980; Moore et al. 1981; Houghton et al. 1983; Woodwell et al. 1983a) are, therefore, overestimates. When the values from Brown and Lugo (1982) are used for tropical forests, the calculated flux is about 30% lower (Detwiler et al. 1985; Houghton et al. 1983).

An analysis based on even lower estimates of carbon stocks for tropical forests (Brown and Lugo 1984) produces a further reduction in the estimated net carbon flux: 0.5×10^{15} g C lower in 1980 and 22×10^{15} g C lower between 1860 and 1980 (Figure 6.3; Houghton 1986). The calculated flux of carbon based on the lowest estimates is not 50% less for the world, because soils and land uses other than permanent clearing, such as logging, contribute to the overall net flux. Further, the lower carbon stocks reduce the release of carbon from cleared forests, but they also reduce the total accumulation of carbon in regrown forests.

If the FAO/UNEP (FAO 1981) assessment of stand volumes on which the lowest estimates of biomass are

based included small or disturbed stands that ecologists have avoided, what is the origin of these stands? Are these stands small because of natural disturbance, or have they been reduced in stature from human use? What has the history of human disturbance been in forests? The clearing of forests for agriculture can be documented throughout history. In contrast, it is difficult to determine a gradual reduction of carbon stocks within existing forests (see Section 6.5.1.7, The History of the Net Biotic Flux).

In the end, the carbon stocks in areas actually cleared or harvested are important in the analyses of land-use change. If the locations of disturbed forests and their carbon stocks can be determined accurately, a large fraction of the uncertainty in current flux estimates will be reduced (Hall et al. 1985; Houghton 1986).

6.5.1.2 *Change in the Carbon Content of Soils*

Logging and cultivation reduce the annual production of plant residues and increase decomposition because of increased soil temperature, aeration, and moisture (Lutz and Chandler 1946; Moller 1947; Hart 1961; Ovington 1968; Witkamp 1971; Marks and Bormann 1972). Erosion losses are also likely to increase, but most of the soil carbon loss is believed to be to the atmosphere as CO_2. Losses were documented in boreal (Tamm and Pettersson 1969), temperate (Morey 1942; Sartz and Huttinger 1950; Trimble and Lull 1956; Covington 1977, 1981), and tropical forests (Popenoe 1957; Cunningham 1963; Nye and Greenland 1964; Zinke et al. 1978; Aweto 1981; Glubczynski 1983) and in grassland soils (Haas et al. 1957). Increases in the organic matter of some agricultural soils under intensive agricultural management (Lemon 1977) are not believed to occur in most of the world's agricultural lands.

The total loss of carbon from soil because of man's activities is calculated by knowing the areas of natural ecosystems that were converted to agriculture and the percentage of carbon loss from undisturbed soil profiles in these ecosystems. The data and assumptions used for percentage loss have a major effect on calculations of net release of carbon from soils. Various modeling efforts assumed a decline of 15% (Revelle and Munk 1977) to 50% (Bohn 1978) in soil organic content upon cultivation.

Using a reduction of 50% in the organic content of soils converted to agriculture and a 25% reduction in the soils of forests converted to pasture, Houghton et al. (1983) estimated that soils contributed 0.87×10^{15} g C to the flux in 1980. The clearing rates used in this analysis were intermediate, but the losses of soil carbon were probably too high. Schlesinger's (1986) review indicates that the loss of soil organic matter is about 30% during a 20-year period of agricultural conversion, averaging 21% in tropical forest soils and about 34% in temperate forest soils. The tropical estimate may be low because few areas were studied longer than a few years. Temperate grasslands appear to lose about 29% of their original carbon content in response to long-term cultivation.

Drainage and cultivation of the organically rich peatlands of the world are apparently causing a carbon release of about 0.03×10^{15} g C/year. Because these soils are thought to have accumulated carbon at 0.14×10^{15} g C/year prior to agricultural disturbances, the net change represents a shift of about 0.17×10^{15} g C annually (Armentano et al. 1984). We know little about changes in soil organic matter in tropical savannas and in arid soils, in which changes in carbonate content are also seen during irrigation (Magaritz and Amiel 1981; Schlesinger 1986).

Forest harvest followed by immediate reforestation results in little net loss of soil carbon (e.g., Gholz and Fisher 1982); net accumulation of soil carbon is also possible (Pastor and Post, in press). The difference in the loss of organic carbon between the clearing of forests for cultivation and for pasture is also important.[1] In the analysis by Houghton et al. (1983), conversions of forest to pasture were assumed to result in a 25% loss of soil carbon (Krebs 1975; Buringh 1984), one-half the amount used for conversions to agriculture. A recent analysis of the pasture soils in Puerto Rico and Costa Rica showed no consistent reduction in carbon relative to mature forest soils (Glubczynski 1983); in some instances, soil carbon may also increase after conversion of forests to pasture.

A 20% rather than a 50% loss of soil carbon associated with clearing of forests and subsequent cultivation reduced the estimate of the net annual release in 1980 by 0.25×10^{15} g C worldwide (Houghton et al. 1983). The release from soils accounted for about 25% of the total net flux after this modification. Estimates of biotic flux based on models of land-use change that do not account for changes in the carbon of soils (e.g., Hall et al. 1985) may, therefore, be somewhat low.

The net release of carbon from soils of the world because of agricultural clearing probably ranges from 0.1×10^{15} to 0.7×10^{15} g/year when the FAO/UNEP (FAO 1981) clearing rates of tropical forests are used (Houghton et al. 1985; Detwiler et al. 1984; Schlesinger 1984). The upper value is about 10% of the annual release from fossil fuels, and represents about a 1% perturbation of the steady-state flux of soil carbon resulting from inputs and decomposition.

[1] R. P. Detwiler, personal communication, 1984.

Table 6.4
Estimates of the Area Worldwide of Forests and Woodlands

Area (10^9 ha)	Date	Remarks	Source
6.15	Preagricultural		Matthews (1983)
6.656	Preagricultural		Olson et al. (1978)
5.64	1860		Olson et al. (1978)
3	1923	Woody growth of economic importance	Zon and Sparhawk (1923)
2.684	1947	U.S.S.R. and Mongolia not included	FAO (1948)
5.65	1950	Includes shrublands	Whittaker and Likens (1973)
5.70	1950		Lieth (1975)
3.796	1953		FAO (1955)
4.401	1958		FAO (1960)
4.281	1963		FAO (1966)
4.72	1970		Olson et al. (1978)
4.030	1974		Persson (1974)
5.83	1979	Includes savanna and chaparral	Ajtay et al. (1979)
3.763	1980		Barney (1980)
5.237	1980		Matthews (1983)
5.817	1980	3.079 if "interrupted woods" are excluded	Olson et al. (1983)

6.5.1.3 *Rates of Deforestation*

The forest areas have decreased since preagricultural times, but both the timing and the extent of the decrease are difficult to determine from the range of estimates available (Table 6.4). In the past few years, estimates of deforestation in the tropics ranged over more than an order of magnitude, from 3×10^6 to 30×10^6 ha/year (Table 6.5). Although the current range of values is perhaps less than twice the mean, it is probably not as small as 12% of the mean, as suggested by a comparison of Myers (1980) and the FAO/UNEP (FAO 1981) study (Melillo et al. 1985). A country-by-country comparison of Myers and FAO/UNEP reveals discrepancies of more than 50% (Myers 1984). For open- and closed-canopy forests combined, the estimates of deforestation by FAO/UNEP (FAO 1981) and FAO Production Yearbooks (FAO 1983) agree within 8% (Houghton et al. 1985). The agreement within regions is not as good as that for the tropics as a whole but, at worst, is within 5%.

Different estimates of the rate of deforestation in the tropics produced a wide range in estimates of the flux of CO_2 to the atmosphere. Houghton et al. (1983) derived estimates of deforestation for the tropics that varied from 3.22×10^6 to 17.1×10^6 ha/year (Table 6.5). These rates yielded a range of flux estimates for the tropics of 1.3×10^{15} to 4.2×10^{15} g C/year. A subsequent study (Houghton et al. 1985) showed that both flux estimates were based on extreme deforestation estimates. When the rate of deforestation is based directly on changes in forest area rather than derived from changes in agricultural area, the low rate of 3.22×10^6 ha/year used by Houghton et al. (1983) becomes 9.41×10^6 ha/year (Houghton et al. 1985) (Table 6.5). The high rate of 15.15×10^6 ha/year for deforestation of closed forests is based on Myers' (1980) study. This becomes 7.5×10^6 ha/year when the distinction is made between permanent clearing of mature forests and repeated clearing of secondary forests (fallow) (Melillo et al. 1985).

The most recent analyses of the carbon flux from tropical regions show a reduced range of estimates.

Table 6.5
Rates of Tropical Deforestation (10^6 ha/year)

Tropics	Tropical America	Tropical Africa	Tropical Asia	Source
24	7.5	8	8.5	FAO (1966) (from Wong 1978)
30				Brunig (1977)
12	6	2	4	Bolin (1977)
7.5	1.3	1.2	5.0	Wong (1978)
9.3-16.6 (5.0-9.1)[a]				Seiler and Crutzen (1980)
11.0-18.5 (6.7-11.0)				Detwiler et al. (1985) (from Seiler and Crutzen 1980)
3.22 (2.84)	2.21	0.07	0.795 (0.465)	Houghton et al. (1983) (from FAO Production Yearbooks)
17.1 (15.15)	5.40	2.85	8.7 (6.8)	(from Myers 1980)
8.65 (5.76)	2.36 (2.36)	2.86 (1.91)	3.286 (1.39)	(based on population)
(19.95)	(10.05)	(1.73)	(8.18)	Barney (1980)
11.07 (7.27)	5.39 (4.12)	3.68 (1.33)	2.00 (1.82)	FAO (1981)
9.41	5.43	3.09	0.90	FAO Production Yearbook (FAO 1983)
(7.5)[b]	(2.5)[b]	(1.4)[b]	(3.1)[b]	Melillo et al. (1985) (from Myers 1980)

[a]Values in parentheses refer to closed forests only; open forests or woodlands have been omitted from these values.
[b]Values are for closed, broad-leaved forests only.

Detwiler et al. (1984) estimated the net release in 1980 to be between 0.5×10^{15} and 1.8×10^{15} g C. The range of deforestation rates used in the analysis varied from 11.0×10^6 to 18.5×10^6 ha/year. The range of flux values included analyses with both high (Brown and Lugo 1982) and low (Brown and Lugo 1984) estimates of carbon stocks. Using the lower of the two estimates of clearing (FAO/UNEP [FAO 1981]) and medium to high biomass values, Molofsky et al. (1984) calculated a net release of 0.6×10^{15} to 1.1×10^{15} g C/year for 1980. In both studies, variations in deforestation rates and in carbon stocks contributed equally to the range of estimates of net carbon flux.

Recent analyses by Houghton et al. (1985) suggest a net flux of carbon from the tropics of 0.9×10^{15} to 2.5×10^{15} g in 1980. The narrower range and lower values than the earlier study (Houghton et al. 1983) reflect improved estimates of deforestation, lower estimates of carbon stocks (Brown and Lugo 1984), and several other model improvements (Houghton et al. 1985). If only the net flux from deforestation is considered, the range is almost exactly the same as the range obtained by Detwiler et al. (1984) of 0.5×10^{15} to 1.8×10^{15} g C/year. The difference between the two studies, 0.4×10^{15} to 0.8×10^{15} g C/year, is the annual net release of carbon to the atmosphere from the permanent clearing of secondary or fallow forests.

The FAO/UNEP assessment (FAO 1981) did not consider deforestation of forests previously cleared for shifting cultivation. Such forests are traditionally in a cycle of clearing, cropping, and fallow (regrowth) and not a source of timber. On the other hand, fallow forests are a source of carbon to the atmosphere if they are cleared permanently. In recent decades the traditional form of shifting agriculture has been replaced in many areas with a permanent subsistence agriculture, referred to as "forest farming" by Myers (1980, 1984) or as "sedentary shifting cultivation" by

the National Research Council (NRC 1982). The ratio of permanently cleared to rotationally cleared fallow forests varies among regions but was estimated by Myers (1980) to be roughly 2:1 for the tropics as a whole. The permanent clearing of fallow forests is estimated to be 10×10^6 ha/year (Houghton et al. 1985), larger than the 7×10^6 ha/year cleared from mature forests.

The loss of fallow forests depends on the rate of deforestation attributed to shifting versus sedentary cultivators. This distinction, as well as that between mature forest and fallow forest, may be resolved with extant and future data from satellites. Monitoring tropical forests with LANDSAT may allow investigators to determine rates of deforestation (Woodwell et al. 1983b, 1984), whether the deforestation was permanent or temporary, and whether the cleared forest had been disturbed previously. LANDSAT also can be used with reference data from field studies to help establish indirectly the carbon stocks of fallow forests that have entered the cycle of shifting cultivation since 1972.

Given the technical and economic difficulties of measuring forest areas on the ground, the use of satellites appears essential for measurement of changes in forest area. LANDSAT data are available for much of the world, starting in 1972, and the successful use of LANDSAT (multispectral scanner data of 80-m resolution) to measure changes in the area in temperate and tropical forests was demonstrated by Woodwell et al. (1983b, 1984). The use of Advanced Very High Resolution Radiometry (AVHRR) on NOAA6 and NOAA7 satellites offers another, less expensive, method for measurement of change, but the resolution of 1 km is not fine enough to show small areas of change. A scheme using AVHRR to identify locations of intense activity and LANDSAT for accurate measurement of change in those locations offers the most promise (Tucker et al. 1986; Woodwell et al. 1984).

6.5.1.4 *Northern Temperate and Boreal Zones*

Recent estimates of the net flux of carbon from northern temperate and boreal systems differ as much as estimates for the tropics (Table 6.3). Johnson and Sharpe (1983) estimated that the northern regions were accumulating 1.6×10^{15} to 1.9×10^{15} g C/year; Houghton et al. (1983) calculated a net release of 0.5×10^{15} g C/year. Given negligible recent clearing of northern forests for agriculture and given the general agreement on carbon stocks in these forests (Table 6.1), the wide range of estimates is surprising.

The high estimates (Johnson and Sharpe 1983) were based, in turn, on the higher of two estimates from Armentano and Ralston (1980), who included two different approaches, one yielding a net storage of 0.19×10^{15} g C/year, the other yielding a net storage of 1.0×10^{15} to 1.2×10^{15} g C/year. The higher value was calculated from the difference between estimates of merchantable timber made in 1953 and 1975 by different investigators. The errors inherent in such estimates make the difference between two of them an unreliable assessment of change, even when the estimates are made by the same agency (Persson 1974; Houghton 1986). The 23-year difference in this case was 6% of the standing stock of carbon in temperate and boreal forests (Ajtay et al. 1979; Olson et al. 1983).

The second method used by Armentano and Ralston (1980) was based on data from the Continuous Forest Inventory (CFI) of the U.S. Forest Service and from analogous records in other countries. The CFI reports the area of forests, the annual net increment of wood, and the amount of wood harvested. The difference between harvest and increment is the change in merchantable timber and is generally regarded as an accurate value for the forests inventoried. The CFI statistics yielded a net annual accumulation of carbon for the northern temperate and boreal forests of about 0.19×10^{15} g, a factor of 5 to 6 less than the estimate based on differences between independent estimates of timber resources. However, the lower estimate applied only to commercial forest stands. Neither noncommercial forests of North America and the U.S.S.R. nor any forests of the Peoples Republic of China, Korea, Eastern Europe, and several other countries were considered in this analysis. The conversion from timber to carbon included 0.6 g/cm^3 for the density of wood, 0.5 for the carbon content of wood, and an expansion factor of 1.75 to account for nonmerchantable biomass (i.e., roots, stumps, branches, and leaves).

Johnson and Sharpe (1983) increased the expansion factor to 2.7, based on a comprehensive analysis of forests in Virginia; the higher figure represents additional carbon in small-sized, dead, and other nonmerchantable trees and in litter.[2] The application of this conversion factor to the two estimates by Armentano and Ralston increases the calculated net accumulation in midlatitudinal forests to 0.3×10^{15} g C/year in one case and 1.6×10^{15} to 1.9×10^{15} g C/year in the other.

The net amounts of carbon estimated by Houghton et al. (1983) and Melillo et al. (in press) to be stored annually in trees of the northern latitudes following harvest (1.2×10^{15} and 0.9×10^{15} g C, respectively) fall between the estimates of Armentano and Ralston (1980) and Johnson and Sharpe (1983). When forest soils, harvested products, and agricultural lands are included in the analysis, however, a net release of carbon to the

[2]Estimates for eastern and western U. S. forests averaged 2.6 with litter included and 2.3 without litter.

atmosphere resulted—0.5 × 10^{15} g C/year (Houghton et al. 1983) and 0.15 × 10^{15} g C/year (Melillo et al., in press), respectively, in 1980. The analysis by Melillo et al. (in press) is thought to provide a better estimate of flux than the analysis of Houghton et al. (1983) or Woodwell et al. (1983a), because it includes a more detailed treatment of harvest in the Soviet Union and an improved estimate of the change in soil organic matter following disturbance (a temporary reduction of 20% rather than 50%). Additional details are provided below.

Forest soils may release carbon for 10 to 25 years following harvest and then act as carbon sinks (Section 6.5.1). The two fluxes should balance over centuries, but the increasing rates of harvest in northern latitudes currently put a larger proportion of soils in the loss phase. Hence, the net flux of carbon from soils is the reverse of the flux into vegetation. In the analysis by Melillo et al. (in press), the fluxes associated with forest harvest and regrowth were 0.9 × 10^{15} g C/year into vegetation and 0.75 × 10^{15} g C/year out of soils, yielding a net storage of 0.15 × 10^{15} g C/year.

If the wood removed from forests were stored permanently, the CFI data would provide a good estimate of the change in carbon stocks. The nonmerchantable parts of trees, fuelwood, paper, and lumber do not accumulate indefinitely, however, and the approach used by Armentano and Ralston (1980) does not account for oxidation of these harvested products. In the analysis by Melillo et al. (in press), the net release of carbon in 1980 from the products accumulated through 1980 was 0.20 × 10^{15} g C for northern ecosystems.

Including the conversion of forests to cultivated lands added another 0.10 × 10^{15} g C/year release, for a total net release of (−0.15 +0.2 +0.1 = 0.15) × 10^{15} g C from temperate and boreal zones in 1980. Because the analysis by Melillo et al. (in press) included soils, wood products, and agricultural changes as well as logging practices, the best current estimate of flux from the midlatitudinal zones is thought to be a release of about 0.15 × 10^{15} g C/year.

6.5.1.5 *Charcoal and the Redistribution of Carbon in Soil*

Seiler and Crutzen (1980) first drew attention to the reduction of organic matter to elemental carbon as a potential mechanism for long-term storage of carbon. They calculated that as much as 0.3 × 10^{15} to 1.3 × 10^{15} g C might be sequestered annually as a result of fires caused by humans. To the extent that the Brown and Lugo (1984) estimates of carbon stocks are more accurate than those of Whittaker and Likens (1973), the estimate of charcoal formation would be 0.2 × 10^{15} to 0.7 × 10^{15} g C/year. This range of values was found also by Detwiler et al. (1985), using different assumptions, data, and a model that included shifting cultivation Houghton et al. (1983) calculated that the formation of charcoal, according to the data of Seiler and Crutzen (1980), decreased the net annual flux by only 0.1 × 10^{15} g C/year from a reference analysis. Their calculation did not include shifting cultivation, which is a major source of the charcoal produced, and which may annually affect an area twice as large as the area of net deforestation (Melillo et al. 1985).

If Seiler and Crutzen are correct in assuming that 75% of the aboveground vegetation does not burn when a tropical moist forest is cleared for shifting cultivation and that the charcoal formed is on average 25% of this unburned material, the amount of charcoal produced from a secondary forest of 80 × 10^6 g C/ha (with 25% of its biomass belowground) would be about 11 × 10^6 g C/ha. After 10 rotations, the amount of charcoal produced would equal the amount of organic carbon originally in these soils (117 × 10^6 g/ha to a depth of 1 m; Schlesinger 1977). Charcoal has been found in soils but not at such high concentrations (in fact, about an order of magnitude lower; Saldarriaga 1985; Sanford et al. 1985). If the rates of formation proposed by Seiler and Crutzen are correct, the charcoal produced during burning must be oxidized later or eroded from its site of formation. The observation of charcoal in lake and ocean sediments confirms the theory that particulates are transported in air and water, but quantification of the annual accumulation of this material has not been attempted.

The possibility that losses of soil organic matter during cultivation may not be released to the atmosphere but are redistributed downslope within the same landscape or downriver to alluvial sediments poses a similar question. The small export of organic carbon by the world's rivers (Schlesinger and Melack 1981; Meybeck 1982; Chapter 5) indicates that the material is not reaching the oceans, but the possibility of small accumulations of organic carbon over large areas cannot be discounted.

6.5.1.6 *Summary of the Current Estimates of Flux Based on Land-Use Change*

Based on the most recent analyses (Detwiler et al. 1984; Molofsky et al. 1984; Houghton et al. 1985; Melillo et al., in press), the net flux of carbon from the terrestrial ecosystems of the world to the atmosphere as a result of changes in land use is estimated to fall within the range of 0.6 × 10^{15} to 2.6 × 10^{15} g in 1980. Almost all of the flux is from the tropics. The range is about 1 × 10^{15} g C narrower than estimated by Houghton et al. (1983) and Woodwell et al. (1983a), largely from re-evaluations of the rate of tropical deforestation (Section

6.5.1.3). The range could be reduced further through the use of satellite imagery to distinguish the temporary from the permanent clearing of forests and through field studies to determine the carbon stocks of forests being cleared. The estimated flux is also lower than earlier estimates, because subsequent studies suggested smaller losses of carbon from disturbed soils (Palm et al., in press; Schlesinger 1986) and lower carbon stocks in tropical forests (Brown and Lugo 1982; Brown and Lugo 1984).

The data used in analyses of land-use change improved markedly in recent years, but the emerging consensus exists largely because "independent" analyses are now using the same data in models that are virtually identical. For example, the analyses of both Detwiler et al. (1984) and Houghton et al. (1985) included estimates of biomass and deforestation derived from the FAO/UNEP study (FAO 1981; Lanly 1982; Brown and Lugo 1984). Although it seems satisfying to have more agreement now than in the past (Table 6.3), agreement should not be confused with accuracy.

Two processes that have not been adequately addressed in current modeling efforts, in addition to the several others noted earlier, are fire suppression and land degradation. The suppression of fires in North America and other regions of the Northern Hemisphere probably increased the carbon stocks within extant forests and led to the expansion of forests into woodlands or grasslands. The increased storage of carbon on land as a result of this phenomenon is not known, but the effect may be small in comparison to the net release of carbon associated with deforestation in the tropics (see Chapter 8).

The degradation of tropical open forests, primarily as a result of logging, harvest of fuelwood, deliberate burning, and grazing, was observed in many countries (FAO 1981; Lanly 1982) but not quantified. The process of degradation has been gradual and the associated reductions in the carbon stocks of vegetation and soil are variable, so the process is not as well documented as deforestation. The areas affected are large, however, and include lands outside the tropics. The flux of carbon to the atmosphere from this type of land degradation is unknown but, hypothetically, might be of the same order as the flux from the clearing of fallow lands, 0.4×10^{15} to 0.8×10^{15} g C/year.

The overall range suggested by the most recent studies, 0.6×10^{15} to 2.6×10^{15} g C/year, may be slightly low, but is believed to represent the best current estimate of uncertainties in the biotic flux resulting from changes in land use worldwide. Real net fluxes outside of this range are less likely to result from surprises in the data used in the analyses than from processes not directly related to land-use change (see Section 6.5.2).

6.5.1.7 *The History of the Net Biotic Flux*

Recently, it was argued that the rates of increase in fossil fuel combustion and in atmospheric CO_2 concentrations have been parallel since 1957, and hence that the net biotic flux either must be insignificant or must be increasing at the same rate as fossil fuels (Elliott et al. 1985). This argument is not valid if there were significant releases of carbon from the biota prior to 1957. Recent changes in the CO_2 content of the atmosphere reflect not only the sources since 1957, but the response of the global carbon system to pre-1957 concentrations (Oeschger and Heimann 1983; Peng et al. 1983). Documentation of the history of land use is necessary to define whether the biota contributed significant amounts of carbon to the atmosphere. Estimates based on historical records of land use are needed, independent of those derived from proxy data such as isotopic ratios of carbon in tree rings. The latter reflect the atmospheric summation of all sources and sinks and thus may not produce an accurate record of biospheric fluxes (also see Chapter 10).

Only four studies attempted to estimate the net biotic flux based on land-use data since 1860 (Revelle and Munk 1977; Moore et al. 1981; Houghton et al. 1983; Richards et al. 1983), and of these only two analyses (Moore et al. and Houghton et al.) estimated the pattern of net flux for the years 1860 through 1980. The estimates of Revelle and Munk and Richards et al. considered only the expansion of agricultural area; Houghton et al. also considered the harvesting of wood and, to a limited extent, the conversion of forest to pasture. The three studies generally agree on the area in agriculture worldwide (Table 6.6; Figure 6.4). There is less agreement on the calculated net fluxes, even when the flux from agricultural expansion is considered alone (Table 6.6).

The largest difference between the calculated fluxes from agriculture results from different assumptions about the average change in carbon stocks that occurs when land is converted to agriculture. Differences in the estimates of carbon stocks are addressed in Sections 6.3.1 and 6.5.1.1. But only part of the difference between the estimates of Richards et al. (1983) and Houghton et al. (1983) can be explained by their use of carbon stocks estimated by Olson et al. (1983) versus Whittaker and Likens (1973), respectively. Different assumptions were made for the types of vegetation cleared for agriculture and the history of the reductions in carbon stocks; that is, whether carbon stocks were reduced before or after 1860. Information about whether grassland, woodland, or forest was cleared is not always available in statistical records. Nevertheless, the division of the world into countries rather than broad regions permitted Richards et al. (1983) to make a more

Table 6.6
Comparison of Estimates of the Net Release of Carbon from Terrestrial Systems as a Result of Changes in Agricultural Area Since 1860

Source	Period	Release of Carbon from Changes in Agricultural Area (10^{15} g)	Change in the Area of Cultivated Land (10^6 ha)
Revelle and Munk 1977	1860-1970	72	853
Richards et al. 1983	1860-1978	62	852
Houghton et al. 1983	1860-1980	116	911
	1860-1970	100	813

detailed assessment of the vegetation types that have been cleared for agriculture.

Estimates of the carbon stocks of these lands, especially after various intensities of wood gathering, grazing, or burning, are difficult to obtain. Changes in successive estimates of the carbon held in the world's biota (Whittaker and Likens 1973; Brown and Lugo 1982; Olson et al. 1983; Brown and Lugo 1984) (Section 6.3.1) could represent a real reduction of the Earth's biota. On the other hand, the different estimates may be a succession of refinements in accuracy, with no real change in terrestrial carbon implied. In this case the most recent estimates of carbon stocks should be used for calculations of both the 1860 and 1980 fluxes. Olson et al. (1978) estimated that the carbon content of the world's terrestrial vegetation decreased from 1050 × 10^{15} g in preagricultural times to 696 × 10^{15} g in 1860 and to 557 × 10^{15} g in 1970. These changes amount to an average annual flux from vegetation of 1.26 × 10^{15} g C between 1860 and 1970. If the change between the 1950 estimate (827 × 10^{15} g C) of Whittaker and Likens (1973) and the 1970 estimate of Olson et al. (1978) is real, it amounts to an average annual release of 13.5 × 10^{15} g C from 1950 to 1970.

No one believes this last example to be correct, but it serves to help illustrate two important points. First, it is important to distinguish changes in accuracy from real changes in carbon stocks. Second, reductions in the world's terrestrial carbon occur both from changes in the area of vegetation types and from changes in the carbon stocks within vegetation types. The two contributed equally to the reductions estimated by Olson et al. (1978), but estimates of the biotic flux based on land-use analysis generally neglect the latter of these fluxes. Such analyses included, for the most part, only the changes associated with the transformation of forests to cultivated land and pasture.

The analysis by Houghton et al. (1983) included some reduction in carbon stocks, because harvested primary forests regrew to only 75-90% of their original carbon stocks before they were harvested again. Changing this restriction to permit complete recovery decreased the net loss of carbon between 1860 and 1980 by about 10%, or 16 × 10^{15} g C. In all other

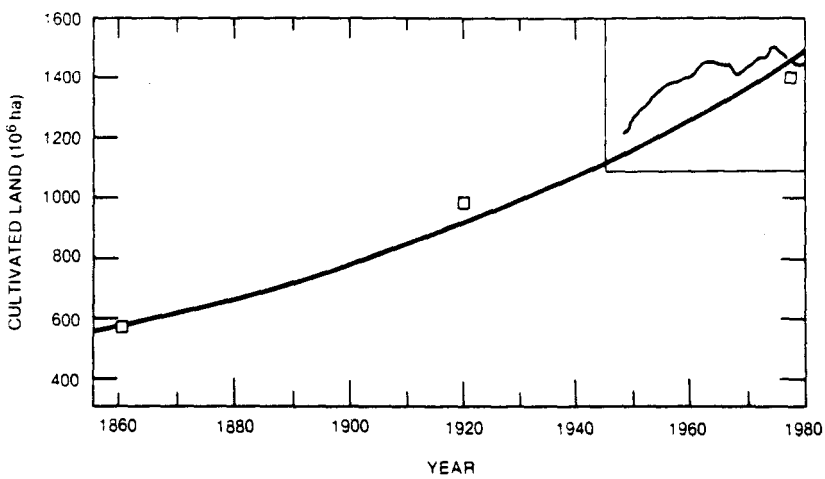


Figure 6.4. Estimates of the area in cultivated land worldwide according to analysis by Houghton et al. (1983) (solid line); Richards et al (1983) (□); and Production Yearbooks of the FAO (1949-1981) (inset).

respects the analysis assumed that no reductions in carbon stocks occurred before 1860. Richards et al. (1983) assumed that reduction of some carbon stocks occurred both prior to 1860 and during the 1860-1978 period. However, the associated release of carbon to the atmosphere (e.g., from shifting cultivation, forest harvest, or other degradation) was not included in their analysis. Only the releases associated with the conversion to cropland during the periods 1860-1920 and 1920-1978 were calculated. Hence, the carbon losses associated with land-use change were larger in the analysis by Houghton et al. than in that by Richards and associates. Estimates of the release of carbon should eventually take into account changes in carbon stocks within ecosystems that have not changed in area. Identification of the systems disturbed and documentation of the extent of degradation would seem to require the close collaboration of historians and ecologists.

Standard methods of historical and historical-geographical research can appreciably reduce the uncertainties discussed above. Documentary sources extending back into the early- to mid-19th century are extant for nearly all world regions (sub-Saharan African sources begin somewhat later). Beginning in the late 19th century, national agricultural and forestry statistics become increasingly reliable and comprehensive. Official administrative reports, travelers' accounts, pioneer journals, and surveyors' logs, among other sources, contain detailed descriptions of vegetation and land cover. The same sources also reveal the type and scale of various kinds of human intervention—whether it be land clearing, wetland drainage, irrigation, or various modes of timbering.

A detailed historical analysis of land-use change in South and Southeast Asia is currently underway at Duke University.[3] The unit of analysis reaches below the nation to subprovincial units or administrative districts (2000-10,000 km^2). After adjustments for shifts in boundaries over time, the land-use and vegetation history of each district is charted from 1820 to 1980. A standard vegetation classification is used (Olson et al. 1983). A full land-use accounting of all area within the district boundaries is attempted for each reference date. Decreasing the size of each study unit, using more detailed sources, and moving to a decennial time series are all efforts to increase the reliability of the estimates. Carbon release can be estimated for individual districts for each land-use and vegetation-cover category as the areas change over time. District figures can be aggregated to develop provincial and national totals. The end result will be carbon emission figures for tropical South and Southeast Asia that invoke a greater degree of confidence than the disparate estimates provided to date.

To the extent that detailed historical research can critically use a variety of sources to refine our understanding of land-use and vegetation changes in South and Southeast Asia over the last two centuries, similar studies may be appropriate for other regions.

6.5.1.8 *The Future of the Net Biotic Flux from Land-Use Change*

Two studies addressed how the net biotic flux of carbon might change in the future as a result of changes in land use. Based on economic predictions of the cost of fuel wood versus other energy sources, Armentano (1984) suggested that increased harvest of U.S forests might increase the biotic flux to the atmosphere by 0.1 $\times$ 10^{15} to 0.2 $\times$ 10^{15} g C/year by the year 2010. If other countries outside the tropics increased their rates of harvest such that the U.S. proportion remained constant (~25% from Melillo et al., in press), the net flux from temperate and boreal systems would be 0.4 $\times$ 10^{15} to 0.8 $\times$ 10^{15} g C/year in 2010. This estimate does not include any flux of carbon from changes in agricultural area.

Under the assumption that the biotic flux in 1980 (2.6 $\times$ 10^{15} g C/year) was directly related to population growth in the tropics, which would continue to increase at current rates until the year 2000, Woodwell et al. (1983a) estimated that the annual biotic flux would reach between 6 $\times$ 10^{15} and 7 $\times$ 10^{15} g C annually. If rates of deforestation were allowed to increase beyond the year 2000, the biotic flux would reach almost 10 $\times$ 10^{15} g C/year before tropical forests disappeared by the mid-21st century. This rate of release of carbon is not a prediction, but an upper estimate of the annual rate that might result if all the remaining tropical forests were cleared in the next 50 years.

6.5.2 Biotic Fluxes of Carbon in Addition to Those Resulting from Land-Use Change

There is a growing awareness among ecologists that the net flux of carbon to the atmosphere resulting from changes in the use of land may be only a part of the total net biotic flux. An additional flux from ecosystems undisturbed by axes, fires, or ploughs is also possible, and the direction of such a flux could either increase or decrease the carbon content of the atmosphere. There are at least three potential causes for such a flux: CO_2 fertilization, steady-state assumptions, and changes in climate.

6.5.2.1 *CO_2 Fertilization*

Enhanced rates of photosynthesis and plant growth in response to elevated levels of CO_2 were observed

[3] J. F. Richards, personal communication, 1984.

under some conditions, and suggest that the world's biota might grow in proportion to the concentration of CO_2 in the atmosphere. Such an enhanced storage of carbon was included in early models of the global carbon cycle as the so-called β-factor (Bacastow and Keeling 1973; Oeschger et al. 1975), and received considerable attention in recent years. Whether an increase in growth will result in an increased storage of carbon on land is not yet clear (Kramer 1981; Houghton and Woodwell 1983; Chapters 1 and 10).

6.5.2.2 *The Assumption of Steady State*

The assumption of steady state has been used throughout analyses of the global carbon cycle, in part for convenience in modeling and in part through inability to detect low-level background changes in carbon storage. The accumulation of carbon in "mature" systems, not only in wetlands (Armentano et al. 1984) and boreal tundra (Miller 1981; Billings et al. 1982) but in some tropical forests (Lugo 1983), suggests that undisturbed systems may not be in steady state (also see Chapter 8). The counterargument is that such fluxes cannot be both large and long-term—large enough to be important annually (0.1×10^{15} g C/year) and long enough to have occurred since the last glacial maximum 10,000 B.P.—because there is not enough carbon in vegetation and soils to have sustained such a flux. But trends in carbon storage need not be on the same time scale as glacial periods. The possibility of changes over hundreds of years seems confirmed from recent analyses of isotopes in ice cores (see Chapter 3), and cycles of such frequency are sufficient to invalidate the assumption of steady state in models of the last 200 years. The difficulty is, of course, detection of a significant accumulation or loss of carbon from terrestrial systems at this time (see Section 6.5.3).

6.5.2.3 *Changes in Climate*

The effect of increased levels of atmospheric CO_2 on climate has received considerable attention, most recently in the National Academy of Sciences' assessment (NAS 1983). In contrast, the feedback of climate on terrestrial metabolism, and hence on the storage of carbon on land, has received little attention. The neglect seems strange, given the potential effect of a global warming on additional releases of CO_2 from biota (Chapter 8). Woodwell (1983) estimated that a 0.5°C temperature rise could increase total respiration by 10–15%, or approximately 10×10^{15} to 15×10^{15} g C/year, globally. The extent to which gross productivity also might be increased and might reduce the net flux of carbon is unknown. The potential effect in the short term, however, might exceed the release of carbon from combustion of fossil fuels. The subject is not well understood, yet records of temperature, precipitation, and atmospheric CO_2 concentrations from around the world offer a means of study. Several methods for quantifying the total net flux from biota are presented in the next section.

Table 6.7
Estimates of the Net Release of Carbon from Terrestrial Ecosystems to the Atmosphere Based on $^{13}C/^{12}C$ Records in Tree Rings

Net Release (10^{15} g)	Time Interval	Source
135-228[a]	1860-1980	Houghton et al. (1983)
62[a,b]	1860-1978	Richards et al. (1983)
120	1850-1950	Stuiver (1978)
170	1800-1935	Wagener (1978)
70	1860-1974	Freyer (1978)
157-195	1860-1974	Siegenthaler et al. (1978)
150	1850-1950	Tans (1978)
260	1830-1980	Peng et al. (1983)
150	1600-1975	Stuiver et al. (1984)
90	1860-1975	Stuiver et al. (1984)
140	1800-1980	Peng and Freyer (1986)

[a]Based on analyses of historical changes in the use of land, not on $^{13}C/^{12}C$ ratios.
[b]Based on changes in agricultural area only.

6.5.3 Methods for Determination of the Total Net Biotic Flux

6.5.3.1 *Carbon Isotopes in Tree Rings*

Estimates of the long-term biotic flux based on records of $^{13}C/^{12}C$ ratios in tree rings are similar to estimates based on changes in land use (Table 6.7). Most recent analyses based on both techniques suggest an upper estimate of 150×10^{15} to 180×10^{15} g C, that is, a somewhat narrower range than suggested in Table 6.7 (cf. preceding discussion on land-use analyses in Section 6.5 and Houghton et al. 1983; and three latest estimates in the table). The timing of the release is different, however. Most isotopic analyses suggest a decreasing biotic flux since about 1900. Changes in cultivated area, however, appear to be known well enough (Figure 6.4) to preclude the possibility of a continuously declining biotic flux from 1900 to 1980, although the flux may have declined in recent decades (Houghton 1986). Although historical changes in pasture area are much more difficult to determine than changes in cultivated area, there seems to be no reason to suspect that a large

expansion of grazing lands occurred about 1900. The greatest changes in grazing lands in recent centuries are believed to have occurred in the last 20 years in Central and South America.

A reconciliation of the isotopic and land-use results may be possible through two mechanisms. First, the isotopic ratios were shown to be affected by variations in local climate and other factors (Francey and Farquhar 1982; Farquhar et al. 1982; Chapters 3, 10, and 11). If these effects could be factored out of the isotopic record, the signal from changes in the atmospheric $^{13}C/^{12}C$ ratio alone might be refined. Of more fundamental significance is the possibility that the difference between the two methods is a measure of the worldwide biotic response to changes in climate, to the increase in atmospheric CO_2, or to changes in the availability of nutrients or toxins since industrialization. Changes in carbon storage in undisturbed systems are not considered in the analyses based on land use, but may be recorded in the $^{13}C/^{12}C$ ratios in tree rings. Thus, the difference in results might reveal another biotic flux or the existence of other carbon sources and sinks (but see Chapters 3, 10, and 11).

There is no direct evidence for a net flux to or from undisturbed systems, but direct measurement is difficult (see Section 6.5.3.3 on detection of change). The potential for a biotic response to various environmental changes is large (Houghton and Woodwell 1983; Woodwell 1983), and ecologists' views on the direction of the response are divided (e.g., Miller 1981; Billings et al. 1983). Thus, the suggestion of such a flux, even by difference, could provide clues for the important question of how biota will respond to climate change.

6.5.3.2 *Change in the Seasonal Amplitude of Atmospheric CO_2 Concentrations*

The seasonal oscillation of CO_2 concentrations in the atmosphere (see Figure 3.9 in Chapter 3) is determined primarily by the metabolism of the terrestrial biota. The lowest concentrations of CO_2 in the Northern Hemisphere occur at the end of the growing season when photosynthesis exceeds respiration, and the highest concentrations occur in early spring, at the end of the season when respiration exceeds photosynthesis. The amplitude of the oscillation is greater in the Northern than in the Southern Hemisphere, consistent with the distribution of temperate and boreal land masses; and greater at higher latitudes than at the equator in keeping with the greater seasonality of temperate versus tropical systems.

Because the seasonal oscillation of CO_2 concentrations is largely a function of terrestrial metabolism, changes in the seasonal pattern might be useful in determining changes in the biota. For example, the seasonality of photosynthesis and respiration might be different in a forest than in the cropland replacing it. The effect of the surface oceans (seasonal temperature and metabolic changes) on the seasonal oscillation of CO_2 concentration is presumably smaller than, and reversed in phase from, terrestrial metabolism (Fung et al. 1983; Fung 1986). Seasonal changes in the combustion of fossil fuels are not well known, but are presumably small in comparison to the effects of either terrestrial metabolism or sea surface change (Pearman and Hyson 1981). Gillette (1982) used 1 year's record of CO_2 levels measured at 14 stations distributed globally to determine which effects were greatest at different locations. He attributed 69% of the variance in the CO_2 patterns to the seasonal biotic effect, 13% to an oceanic effect, and 7.5% to a local, perhaps anomalous, biotic effect at high northern latitudes.

The first attempt to analyze changes in the amplitude of the seasonal oscillation of CO_2 concentrations revealed no change from 1957 to 1973 (Hall et al. 1975). Since then, with the longer atmospheric record and the use of various techniques, several studies have found an increase in the amplitude through time (Chapter 3; Pearman and Hyson 1980, 1981; Brown et al. 1981; Cleveland et al. 1983; Keeling 1983). Interpretation of the increase is difficult. Brown et al. (1981) and Cleveland et al. (1983) attributed the change to either larger or more photosynthetically active terrestrial biota. They dismissed the possibility that the change was caused by changes in either sea surface temperature or fossil fuel use. Changes in sea surface temperature apparently account for only a small fraction of the observed seasonal changes in CO_2 (Fung 1986); however, changes in atmospheric circulation are large enough to account for the observed change in the seasonal amplitude (Fung 1986). Thus, the interaction between atmospheric circulation and the increasing rate at which CO_2 has been added to the Northern Hemisphere from fossil fuels also may be important.

Even if the increased amplitude of the winter-summer difference in CO_2 concentrations could be attributed entirely to the terrestrial biota, it does not follow that there has been a net storage of carbon in the biota or soils of the world. Increased uptake is not equivalent to increased storage; there may be an increased release as well. Increased respiration during

a warmer winter would cause the same increase in amplitude as increased photosynthesis, or net storage, in summer. Change in the seasonal oscillation of CO_2 concentrations by itself, therefore, is not sufficient to infer changes in the worldwide storage of carbon on land. On the contrary, if the large terrestrial releases of carbon to the atmosphere are from deforestation in the humid tropics, where there is little seasonal change, such releases would have little effect on the seasonal oscillation of atmospheric CO_2 concentrations.

Despite pitfalls in the interpretation of long-term changes in the annual amplitude of CO_2 concentrations, records of seasonal variation in the concentrations both temporally and spatially will be extremely valuable when compared with other climate data. For example, the relationship between the annual pattern of CO_2 concentrations (measured at Point Barrow, Ocean Station PAPA, and Mauna Loa) and a severe drought in the Soviet Union in 1975 (Schnell and Harris 1981) suggests that continuous atmospheric monitoring at many sites will be useful in observing and predicting the biotic response to changes in climate (Chapters 3 and 11). Given the potential response of terrestrial biota to changes in temperature and precipitation, it is a mistake to attribute variations in the atmospheric concentrations of CO_2 correlated with the Southern Oscillation or with sea surface temperature anomalies to only oceanic fluxes of CO_2 (Chapter 3). Interpretation of the biotic response would be aided tremendously with a network of CO_2 flask stations across continents to match the existing network established at sites remote from terrestrial effects (see Chapters 3 and 11).

6.5.3.3 *Direct Detection of a Change in the Annual Flux*

The maximum difference between the net annual flux as determined by analyses based on changes in land use and changes in isotopic ratios in tree rings is about 2×10^{15} g C (see Figure 10.1 in Chapter 10). Direct detection of a change of 2×10^{15} g C/year in the world's biota and soils would be difficult. With the amount of terrestrial organic carbon estimated at 2050×10^{15} g ($\pm$ 20%), detection of an annual change of 2×10^{15} g (i.e., 0.1%) would require about 2 centuries.

On the other hand, direct measurement of a change in the flux of carbon, rather than in the standing stocks, might be achieved more readily. The estimated gross flux between terrestrial systems and the atmosphere is 90×10^{15} to 120×10^{15} g C/year (GPP; see Section 6.4). Assuming the error of this estimate to be ~20% (Table 6.2), detection of a 2×10^{15} g/year change would require only 10 to 20 years. In fact, the gross flux of carbon from the atmosphere to terrestrial ecosystems is less than GPP, because 30 to 40% of the CO_2 used in production is derived from respiratory sources from within the ecosystem (see Section 6.4). The gross flux of CO_2 from the atmosphere to the world's biota may be of the order of 60×10^{15} to 80×10^{15} g C/year, and detection of a 2×10^{15} g/year change might be possible in a decade.

The difficulty with direct measurements, whether of standing stocks or fluxes, is that they require a large number of locations throughout the world. Measurement of the global flux would require, in addition, almost continuous measurements during the course of a year. An alternative to the tens or hundreds of sites where such measurements would be required would be the development of a method to measure flux remotely (from satellites) or at an altitude where integration of a large area of the Earth's surface occurs.

The seasonal patterns of CO_2 concentrations observed at atmospheric monitoring stations are integrated measures. These data were used to calculate the net amount of carbon stored in terrestrial communities during the summer and released in winter. The seasonal storage of carbon in the ecosystems of the Northern Hemisphere is 4×10^{15} to 6×10^{15} g C according to several estimates (Bolin and Keeling 1963; Pearman and Hyson 1980). This storage of carbon is not an estimate of NPP, which for the Northern Hemisphere is estimated to be 13×10^{15} to 16×10^{15} g C (Whittaker and Likens 1973; Ajtay et al. 1979). The difference between the two estimates is mainly due to differences in the seasonality of primary production relative to respiration. Most of the annual heterotrophic respiration (release of CO_2) occurs during the growing season simultaneously with NPP (uptake of CO_2). Measurements in a forest in the eastern United States showed that about 60% of the annual respiration occurred during the growing season (Woodwell and Botkin 1970; Woodwell and Dykeman 1966; Woodwell and Whittaker 1968). If the same fraction applies to other ecosystems in the Northern Hemisphere, the excess of uptake minus the release during the summer would be 5×10^{15} to 6×10^{15} g C ($0.4 \times 13\text{-}16$), in agreement with estimates made from seasonal variations in CO_2 concentrations.

Measurement of the seasonal flux at a global or hemispheric scale is not sensitive enough to detect year-to-year changes in metabolism. Seasonal changes in the flux of CO_2 from the oceans in response to temperature or marine productivity and changes in atmospheric circulation (Fung 1986) confound the analysis. Measurement at the scale of an ecosystem, on the other hand, is more detailed than necessary for a global analysis. A network of CO_2-monitoring stations over continental areas would be useful in detecting changes in seasonal storage or metabolism, in relating these changes to observed environmental anomalies,

and in learning more about the relationships between climate change and changes in the storage of carbon on land (see Chapters 3 and 11).

5.6 RECOMMENDATIONS FOR FUTURE RESEARCH

5.6.1 Estimates of the Current Biotic Flux

The apparent agreement among ecologists on a net biotic release that, at the lower end of its range, is consistent with some estimates of oceanic uptake is gratifying, but does not support the conclusion that the global carbon cycle is well understood. The net biotic flux is small relative to the gross annual fluxes of 90 $\times$ 10^{15} to 120 $\times$ 10^{15} g C/year. Future changes in climate can be expected to change the gross fluxes from photosynthesis and respiration, such that the net flux may have less to do with changes in land use than with changes in climate. That is, the terrestrial biota is likely to exert a feedback on the atmospheric increase in CO_2. The net biotic flux due to climate change, including the direct effect of CO_2 concentration, is not predictable; the ecological community is far from agreement as to even the direction of such a flux (e.g., see Billings et al. 1982, 1983; Miller 1981; Chapters 1, 8, and 10).

The research needed to estimate the current biotic flux in the next few years includes:

1. Establishment of a network of CO_2-monitoring stations (over continents) similar to the current NOAA network of remote (oceanic) stations. Upwind and downwind stations for a temperate/boreal land mass (e.g., Canada) and for a tropical land mass (e.g., Peru and Brazil) would be the minimum distribution. The purpose of the network is to detect changes in the metabolism of major terrestrial systems and to relate those changes to changes in either climate or land use.

2. Initiation and continuation of a worldwide monitoring of land-use change using NOAA7 and LANDSAT imagery. The purpose is to document when, where, and how large an area of forest is being cleared and re-established. NOAA7 imagery could be used to identify areas of intense change, and LANDSAT to investigate those areas with higher resolution.

3. Measurement of the carbon stocks of forests in a sample of the stands surveyed for volume by the FAO/UNEP (FAO 1981) study. Site selection should be limited to areas currently being deforested or degraded.

4. Use of the geographic detail of deforestation (obtained from satellites) and carbon stocks to calculate geographically specific estimates of the biotic flux of carbon. The estimates of flux should be compared with observed atmospheric CO_2 concentrations at the same locality (No. 1 above). The questions are: To what extent do changes in land use explain the variation in CO_2 concentrations? To what extent are variations suggested in climate or some other non-steady-state process? (Note: Whereas historical analyses of land-use change are generally based on data of 10-year intervals, changes in the use of land since 1972 can be obtained through LANDSAT at yearly intervals or less.)

5. Inclusion in the geographic analysis (No. 4) of terrestrial metabolism and geographically specific estimates of the flux of CO_2 between the surface ocean and the atmosphere. The requirements of such a study include:
   - A map of the surface of the Earth including the oceans and types of vegetation and land use;
   - Estimates of metabolism for the vegetation and estimates of CO_2 exchange for the surface ocean;
   - A network of measurements of temperature and precipitation;
   - A network of measurements of atmospheric CO_2 concentrations;
   - A model of atmospheric mixing and air transport. Inconsistencies between modeled and observed concentrations of CO_2 would suggest where further research was necessary.

6. Testing of the feasibility of No. 5 above in an experiment of limited time and space. Such an experiment might be defined by the size, stability, and predictable trajectory of an air mass or by an anomalous event, such as the 1982/83 El Niño year or, more specifically, the 1983 burning of tropical forests in Borneo.

7. Field experiments to determine the effects of increased CO_2 concentrations, temperature, and moisture on the storage of carbon on land. Such experiments must consider the entire ecosystem, soil as well as vegetation.

8. Further experiments to test the validity of the hypothesis that so-called "mature" ecosystems are in steady state. Such experiments should include field measurements of rates of recovery of carbon pools in vegetation and soils, documentation of the frequency and magnitude of natural perturbations in ecosystems, and testing of models of the terrestrial biota with non-steady-state scenarios.

6.6.2 Estimates of the Historic Biotic Flux

To the extent that research on the current flux shows that changes in climate or CO_2 concentrations affect the storage of carbon in predictable ways, it should be clear that such knowledge will apply to reconstructions of the historic release of CO_2 from biota.

The research recommended to improve estimates of the historic flux because of changes in land use includes:

- Detailed reconstructions of the use of land on the scale of countries, states, or counties, based on original records available in archives;
- Establishment of relationships between the rates and locations of deforestation or degradation and such factors as population growth and dispersion, soil fertility, transportation routes, and climate.

The first approach is more direct if the data exist; however, it is not clear how well it will provide estimates of a reduction in carbon stocks. The second approach (e.g., Tosi 1980) is less precise, is probably more rapid, and is required in the absence of reliable data. Some combination of the two will be most effective. Either approach will provide geographically precise information readily compatible with the geographic detail required for analyses of the current flux. To the extent that a current map of land use can be constructed from satellite imagery to identify the areas used and not used by humans, the task of the historical research may be reduced to consider only those lands affected by use.

6.7 SUMMARY

The organic carbon content of the world's terrestrial systems (biota and soils) is thought to be $2000(\pm 400) \times 10^{15}$ g C. About 75% of this carbon is thought to be in soils and about 25% in vegetation. The natural exchange of carbon between terrestrial systems and the atmosphere is thought to be between 60×10^{15} and 120×10^{15} g C/year.

Human activities have altered this natural flux and caused a net release of carbon to the atmosphere. In the past 100 to 200 years this net release is estimated to have been 90×10^{15} to 180×10^{15} g C (Section 6.5.3). In 1980, the net release from changes in land use was probably within the range of 0.6×10^{15} to 2.6×10^{15} g C. Almost all of this net release was from the tropics; the northern temperate and boreal zones are thought to have contributed between 0.1×10^{15} and 0.2×10^{15} g C.

The major uncertainties in these values result from different assumptions about the types of vegetation converted to cropland or pasture, the extent of the change in carbon stocks within forests before versus after 1860, the current carbon stocks of tropical forests, and rate of tropical deforestation, including loss of fallow forests. There is also considerable uncertainty (and speculation) about whether the ranges given above represent the net flux of carbon between terrestrial systems and the atmosphere or whether there is another flux, either positive or negative, because of changes in systems unaffected by land-use change. There is no direct evidence for any such change, because a change as large as 2×10^{15} g C/year would require decades to centuries to detect in the large reservoirs of terrestrial carbon. Measurement of a change in exchange rates, rather than in standing stocks, appears more promising, especially if methods are developed to measure flux at a scale on the order of 10^4 km^2.

Several indirect methods have been used to estimate the net flux of carbon between terrestrial ecosystems and the atmosphere. The method reviewed most extensively here is that based on the history of land use and on changes in carbon associated with changes in land use. According to this method the cumulative net release since 1860 is in agreement with some of the values obtained from analyses based on the $^{13}C/^{12}C$ ratios recorded in tree rings. The timing of the net release, however, may be different. If techniques for interpretation of $^{13}C/^{12}C$ ratios in tree rings were to be perfected (however problematic this might be), differences between the two types of analyses might provide clues to one of the most important, yet unresolved, questions of the global carbon cycle: How do terrestrial ecosystems respond to changes in climate? Will they release or store carbon?

Another method used to suggest changes in the storage of carbon on land is analysis of temporal changes in the amplitude of the seasonal oscillation in atmospheric CO_2 concentrations. Evidence of such change is not sufficient to establish a net flux between the atmosphere and terrestrial ecosystems worldwide; but, together with other geographically specific information on variations in temperature and moisture, a change in the seasonal pattern of CO_2 concentrations may suggest how the world's biota respond to climate and land-use change.

Improved accuracy in the estimates of biotic flux, past and present, is possible from simultaneous consideration of the characteristics of vegetation, soil, and land use in greater geographic detail. The geographic specification of biotic fluxes, together with atmospheric circulation models, will allow more sensitive comparisons between metabolic changes on the ground and observed regional variations in CO_2 concentrations. Such comparisons may be one of the most fruitful approaches for recognizing positive or negative biotic feedback to climate change.

Finally, a map with geographic detail of the Earth's surface, including oceans as well as natural and managed terrestrial ecosystems, will have uses far beyond the important issues of the carbon cycle. Such a map will contain the information necessary to calculate exchanges of all types—gases important in the greenhouse effect (N_2O, CH_4), aerosols important in both the greenhouse effect and in the transport of acid precipitation, and other trace gases of both natural and human origin. The integration of information on surface fluxes and their controls, of satellite imagery, and of global circulation and atmospheric chemistry models suggests that the important questions concerning the Earth's resources can be addressed in powerful, new ways.

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7. MODELING THE GLOBAL CARBON CYCLE AND CHANGES IN THE ATMOSPHERIC CARBON DIOXIDE LEVELS

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7.1 INTRODUCTION AND BACKGROUND

From 1959 to 1980, the annual average concentration of atmospheric carbon dioxide (CO_2) at Mauna Loa Observatory, Hawaii, increased from approximately 315.66 ppm to 338.36 ppm (Keeling et al. 1982). This change corresponds to an increase of 48.4×10^{15} g in the carbon content of the atmosphere. Data compiled by the United Nations indicate that during this time period approximately 84×10^{15} g C were released to the atmosphere from fossil fuels and the manufacture of cement (Marland and Rotty 1984; Chapter 4; henceforth all chapter and section references are to this volume). The observed increase is less than the release because of interactions between the atmosphere and other reservoirs that collectively provide a sink for excess carbon from the atmosphere. The concentration of CO_2 is controlled by these interactions.

If reliable projections of future CO_2 changes are to be made, exchanges between the atmosphere and other carbon reservoirs as well as the dependence of exchange processes on physical, chemical, and biological factors must be thoroughly understood. In this chapter, we review the status of global carbon cycle modeling from the standpoint of developing this understanding.

First, general approaches are summarized. Primary models now in widespread use are then described, followed by a review of model tests and comparisons of attributes. At this writing, there is no entirely satisfactory model of the carbon cycle. We attempt to identify principal shortcomings and to outline work to refine existing models, develop more realistic models, and extend understanding of the carbon cycle through modeling studies.

7.1.1 Major Reservoirs and Fluxes

In modeling the carbon cycle, three major reservoirs, the atmosphere, the oceans, and the vegetation and soils in terrestrial ecosystems are considered (Figure 7.1). The oceans constitute the largest pool and likely are the primary sink (Baes et al. 1977; Bolin et al. 1979; Broecker et al. 1979). Because of disturbance by human activities, terrestrial ecosystems act as both sources and sinks at different times and in different regions (Bolin 1977; Woodwell et al. 1978, 1983; Olson et al. 1978).

The total inorganic carbon content of the oceanic reservoir is estimated to be approximately $38{,}000 \times 10^{15}$ g (Chapter 8). The average residence time of water in the deep sea is approximately 500 years (Broecker et al. 1979).

The exchange of CO_2 across the atmosphere-ocean interface is governed by the difference between the partial pressures of CO_2 (pCO_2) in the atmosphere and n surface waters. The exchange is generally assumed to be by molecular diffusion through a thin film and is controlled by temperature, wind stress, and processes such as turbulent mixing (see Chapter 5). The pCO_2 in the atmosphere is proportional to the ratio of the masses of CO_2 and dry air, whereas the pCO_2 in surface waters depends on the chemical equilibria among carbon compounds (Keeling 1973). As CO_2 is added to surface waters, chemical equilibration is rapid compared with transport and mixing processes.

Mixing and circulation remove carbon from surface waters into deeper layers where it may be sequestered for some time, making the ocean a more effective sink than is suggested by surface water chemistry alone.

Prior to impacts by human activities, carbon in the world's vegetation approached 1000×10^{15} g (Olson 1974, also see Chapter 8). Contemporary storage is estimated to be approximately 560×10^{15} g (Olson and Watts 1982; Olson et al. 1983; see also Chapters 6 and 8). There are about 1500×10^{15} g C in litter and soils (e.g., Zinke et al. 1984; Chapters 6 and 8), and human activities have also altered storage in these pools from natural levels.

In terrestrial ecosystems, the assimilation of carbon from the atmosphere is by photosynthesis. Net primary production (NPP), the difference between gross assimilation and the return of CO_2 to the atmosphere by plant respiration, is responsible for longer term accumulation in vegetation and ultimately in dead organic matter. Variations in gross photosynthesis and autotrophic respiration that occur relatively quickly (hours and days) are inconsequential to secular trends in CO_2 levels. Rather, changes in the relative magnitudes of cumulative net production and losses of organic matter from litter and soil pools due to decomposition, heterotrophic respiration, and fire effect longer term changes in atmospheric CO_2 levels.

The difference between these assimilation and loss fluxes, each of which is about 60×10^{15} g C/year (Olson et al. 1978; also see Chapters 6 and 8), dictates the magnitude of the net global exchange between the atmosphere and terrestrial pools. For reasonably undisturbed ecosystems, uptake and loss fluxes may balance each other closely if averaged over sufficiently long time periods (e.g., Olson 1963; see also Chapters 6 and 8). On shorter time scales, even under natural conditions, the balance certainly shifts with changes in nutrient availability, climate, and sporadic disturbances such as fires or anomalous weather events (e.g., Bormann and Likens 1979). Human disturbances can cause large changes in terrestrial pools and associated fluxes that in many instances mask responses to natural influences (Houghton et al. 1983; Woodwell et al. 1983), and as a result of human management of the landscape, balance between income and loss fluxes is approximated in only a few remaining undisturbed regions.

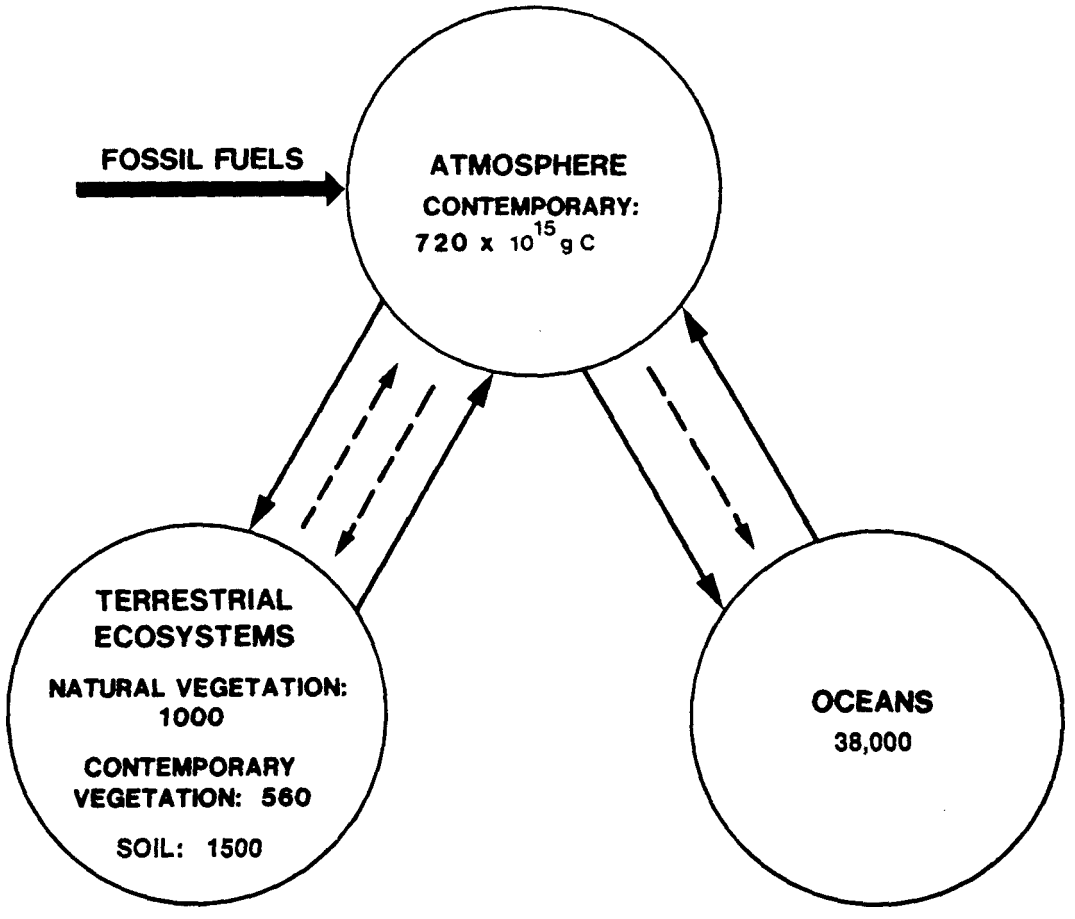


Figure 7.1. Major reservoirs and exchanges generally considered in modeling the global carbon cycle. The oceans, the largest pool of carbon, are the primary sink for excess carbon from the atmosphere. Because of disturbances by human activities, primarily land-use change, carbon storage in vegetation and soil has decreased. The timing of terrestrial releases is complex, and different regions have likely been net sources or sinks at different points in time. In models, this carbon cycle is viewed as closed except for the input from fossil fuels. Many models consider primarily the atmosphere and oceans and handle terrestrial sources as an additional input to the atmosphere. The inventory of carbon in each major reservoir ca. 1980 is given in 10^{15} g C.

Disturbance to terrestrial ecosystems by land-use change alters the standing levels of carbon and the dynamics of cycling (e.g., Houghton et al. 1983; Emanuel et al. 1984). With land-use change such as forest clearing, carbon is immediately released to the atmosphere by burning or, after a delay, by decomposition. In many instances, ground vegetation (e.g., annual crops, pasture, or other herbaceous types) rather than trees is established after clearing, and by management this altered land cover with lower carbon storage than in the original ecosystems may be maintained indefinitely. Following clearing, net productivity generally exceeds losses, and disturbed ecosystems may act as sinks for some time (Houghton et al. 1983; Emanuel et al. 1984; Chapter 6). Abandoned areas recover toward natural conditions and, during recovery, may also serve as carbon sinks. Human management of the landscape has created a complex pattern of sources and sinks in different regions that has changed through time.

If other constraints are not limiting, increasing atmospheric CO_2 levels may stimulate plant productivity, such that the accumulation of carbon exceeds losses, forcing a net sink in some ecosystems. In several modeling studies (e.g., Bacastow and Keeling 1973; Oeschger et al. 1975), assimilation of carbon from the atmosphere was assumed to increase with increasing atmospheric CO_2 concentration. This assumption, generally referred to as "CO_2 fertilization," is supported in part by results from experiments under controlled conditions. In reality, the effect of rising CO_2 concentrations on productivity is complicated by the

influence of other limiting factors (e.g., nutrient availability or climate) on communities composed of mixed species that respond in different ways to CO_2 increases (see Chapters 1 and 10).

In many ecosystems, there is a pronounced seasonal variation in the differences between assimilation and loss fluxes that is responsible in large part for the annual cycle in atmospheric CO_2 records. Seasonality in net carbon exchange between terrestrial pools and the atmosphere is due to variations in both production and loss fluxes and is associated with both temperature and precipitation as well as photo period. Although seasonality is most easily perceived in ecosystems composed primarily of deciduous plants, those made up of evergreen vegetation may also exhibit strong seasonal variations.

On 1000-year time scales, climate changes have altered terrestrial carbon storage substantially (see Chapter 8). It remains to be demonstrated that response to climate variations contributed significantly to changes in atmospheric CO_2 concentration during the period of fossil fuel use. However, the large climate changes projected to occur as CO_2 levels continue to rise may alter terrestrial carbon dynamics in major ways and thus feed back on atmospheric CO_2 levels (e.g., Emanuel et al. 1985; also see Chapters 1 and 10).

Historically, forest clearing and similar land-use changes were the primary causes of imbalances between assimilation and loss fluxes, and, on the whole, vegetation and soils were probably a net source of carbon to the atmosphere in addition to fossil fuels. The net flux caused by forest clearing may well have been larger than the release from fossil fuels until about 1950 (Houghton et al. 1983; Chapter 6). Accurate analysis of fluxes associated with forest clearing and related disturbances is currently the most important goal in modeling the terrestrial component of the carbon cycle.

If fossil fuel use continues to increase at about 1% per year (see Chapters 4 and 9), then the relative contribution to the atmospheric CO_2 increase from future terrestrial releases will likely be small in comparison with that from fossil fuels. The current standing stock of carbon in world vegetation ($\sim 560 \times 10^{15}$ g) is too small compared with the likely fossil fuel source (see Chapter 4). Terrestrial sources may be important if substantially lower rates of fossil fuel use occur or when changes over the next decade or so are analyzed from the perspective of detecting a CO_2-induced climate change. The relative importance of terrestrial and fossil fuel sources in projecting future CO_2 levels is discussed further in Chapter 10.

The reservoir of CO_2 in the atmosphere serves to couple the oceanic and terrestrial carbon pools. In analyzing long-term trends, the atmosphere can be treated as a well-mixed reservoir. Geographic and intra-annual variations in CO_2 concentration are less than 5% of the global annual mean value. However, spatial and temporal patterns apparent in CO_2 measurements can provide important additional insight into interactions between the atmosphere and other reservoirs (see also Chapter 3).

7.1.2 General Approaches to Carbon Cycle Modeling

Ideally, carbon cycle models should be formulated from basic physical, chemical, and biological principles. Model parameters should be associated with recognized processes, and rigorous procedures should be used to estimate parameter values. Parameter estimation should be independent of testing and evaluation.

In actuality, current data sets and understanding at the process level are inadequate to support this preferred approach. To make progress in global-scale geochemical modeling, important natural processes must be mimicked, using only those parameters whose values can be reliably estimated. Because our knowledge of nature's kinetics is still limited, the models we use are rather simple.

Submodels of the oceanic and terrestrial systems are formulated and coupled to models of the atmospheric reservoir. Depending on the emphasis of a particular investigation, a carbon cycle model may emphasize one reservoir more than others.

Because oceans are the primary sink for carbon from the atmosphere, most modeling efforts have been directed toward evaluating the rate of oceanic uptake. By comparison, disturbances to carbon storage in vegetation and soils have been treated in detail only recently.

7.1.2.1 *The Atmosphere*

The general circulation of the atmosphere is complicated both in time and space. In addition to causing the familiar changes in seasonal climate and day-to-day weather, these circulation patterns redistribute CO_2. This redistribution is responsible for the signature of fossil fuel CO_2 and atmosphere-land exchanges observed at remote and marine monitoring stations. Trajectory analyses demonstrate that CO_2 variations at monitoring sites are also related to the history of the air masses that reach stations (e.g., Ciattaglia 1983; Fraser et al. 1983; Gaudry et al. 1983; Halter and Harris 1983).

To investigate long-term trends in CO_2 levels, the atmosphere, with a mixing time of approximately 2 years, can be represented as a well-mixed reservoir, but in actuality atmospheric CO_2 is not well mixed. Its sources and sinks are not uniform in space or time.

Analyses of geographical, seasonal, and interannual variations in atmospheric CO_2 concentration yield large amounts of information on the interactions between the atmosphere and the other carbon reservoirs.

One- and two-dimensional diffusion models that resolve variations in latitude or in latitude and height have been used to investigate the geographical and temporal distributions of atmospheric CO_2 (Bolin and Keeling 1963; Junge and Czeplak 1968; Machta 1972; Machta et al. 1977; Azevedo 1982; Keeling and Heimann in press). These models represented atmospheric transport processes by diffusion equations. Pearman and Hyson (1980) employed a two-dimensional advection-diffusion model of the atmosphere in which observed seasonal variations in winds are coupled with a two-dimensional stress tensor to simulate atmospheric transport. The diffusion coefficients used in these models were calibrated by comparing simulations of other atmospheric tracers, such as chlorofluorocarbons and ozone, with data.

7.1.2.2 *The Atmosphere-Ocean Carbon System*

Three primary factors determine the net uptake of carbon from the atmosphere by the oceans: (1) the capacity of seawater in equilibrium with the atmosphere to incorporate excess CO_2, (2) the rate of CO_2 invasion into the surface ocean, and (3) the rate of carbon transport into deeper layers by mixing and circulation (e.g., Broecker et al. 1979).

The atmosphere and surface waters exchange CO_2 to equalize their respective partial pressures. The flux from the atmosphere to the ocean is proportional to atmospheric CO_2 concentration, but the flux from the surface ocean to the atmosphere depends on shifting chemical equilibria among carbon-bearing species as CO_2 is added to the surface waters. The effect is that the partial pressure of dissolved CO_2 in the surface ocean is a nonlinear function of the concentration of total inorganic carbon and therefore of the mass of carbon.

A thermodynamic calculation relates the partial pressure of dissolved CO_2 and the total mass of inorganic carbon in surface waters (e.g., Keeling 1973; also see Chapter 5). The reactions of primary concern are

$$\begin{aligned} CO_2(g) + H_2O &\rightleftharpoons H_2CO_3(aq) \\ H_2CO_3(aq) &\rightleftharpoons H^+ + HCO_3^- \\ HCO_3^- &\rightleftharpoons H^+ + CO_3^{2-}, \end{aligned} \tag{7.1}$$

where (g) and (aq) indicate gaseous and aqueous phases. From the equilibrium constants for these reactions and from the equations for total boron and titration alkalinity, the partial pressure of dissolved CO_2 and the total mass of carbon can be related to the hydrogen ion concentration $[H^+]$ and hence, implicitly, to each other (see Chapter 5 for a detailed explanation). The nonlinear equations specifying this relationship are solved numerically. The relationship between partial pressure and total inorganic carbon content of surface waters can be approximated by a buffer factor,

$$\beta = \partial \ln P_S / \partial \ln C_S, \tag{7.2}$$

assumed to be constant; 10 is a typical value (see Chapter 5). The variables P_S and C_S correspond to the pCO_2 and the mass of total inorganic carbon in surface waters, respectively.

Slow vertical mixing is the major aspect of ocean dynamics that dictates the rate of carbon uptake by the oceans (e.g., Broecker et al. 1980). To capture the dynamics of this system in a simple model with few parameters, Oeschger and co-workers (1975) represented vertical transport by a diffusion equation,

$$\frac{\partial c}{\partial t} = K_z \frac{\partial^2 c}{\partial z^2}, \tag{7.3}$$

where $c(z,t)$ is the concentration profile of carbon with depth z, and K_z is a constant coefficient of turbulent diffusivity. The resulting one-dimensional (depth) box-diffusion model is still the most widely studied ocean representation.

In actuality, fluxes such as the sinking of cold water at high latitudes are likely the most important mechanisms for transporting carbon from surface to deeper layers. A number of models have been proposed that either are structured around the major advective fluxes or suggest modifications to the assumption of diffusive transport by incorporating advective fluxes into box-diffusion-style models (e.g., Hoffert et al. 1981; Viecelli et al. 1981; Crane 1982; Enting and Pearman 1982; Peng et al. 1983; Sarmiento 1983; Siegenthaler 1983). However, these modeling studies suggest that even more spatially detailed ocean models are required.

7.1.2.3 *Terrestrial Ecosystems*

To date, cycling in terrestrial pools has been represented by a set of compartments corresponding to a functional division (e.g., plants and soil) or based on the distribution of turnover times in vegetation and soil. Carbon uptake from the atmosphere corresponds to net primary production (NPP); the relatively rapid exchange associated with gross primary production (GPP) and plant respiration usually is not treated. Releases are by heterotrophic respiration, decomposition, and fire. Terrestrial compartments may be interconnected by fluxes corresponding to translocation, litterfall, mortality, and soil formation processes; but in many instances, isolated compartments, each exchanging with the atmosphere, are used. Almost all models published

thus far aggregate carbon across the world land surface and across different types of ecosystems.

Changes in the carbon content of each compartment are related to fluxes by a first-order ordinary differential equation that expresses mass balance. Each flux term is expressed as a function of compartment contents and perhaps of atmospheric CO_2 levels. In actuality, fluxes also depend on exogenous variables (e.g., temperature) or other factors (e.g., soil type), but external or environmental conditions have not been considered in detail thus far.

In most cases, terrestrial fluxes can be assumed to be linearly proportional to the content of the donor compartment. The critical exception is the assimilation of carbon from the atmosphere by net production. In many ecosystems, this flux depends on the standing stock of carbon in vegetation; however, if a CO_2 fertilization effect is assumed, then the primary production flux also depends on the carbon content of the atmosphere. Net production also depends on the availability of nutrients. This linkage to other element cycles is just beginning to be explored.

General formulations for expressing net production in models of the terrestrial component of the carbon cycle have not been adopted. Several specific approaches are briefly reviewed in Section 7.3.

In equilibrium, assimilation and loss fluxes to and from terrestrial compartments exactly balance and there is no net exchange between atmospheric and terrestrial components. If such an equilibrium condition is maintained, terrestrial components do not influence long-term changes in atmospheric CO_2 levels. Changes in productivity, with increases in atmospheric CO_2 levels and perturbations associated with forest clearing and similar land-use changes, have been independently implemented in models and analyzed from several standpoints; but to date, complete modeling studies that consider both of these perturbations simultaneously have not been reported.

Forest clearing and other land-use changes are generally implemented as the compound result of changes in the areal extents of the ecosystem types involved and changes in the densities of carbon storage in vegetation and soils in the areas involved in the land-use change (e.g., Houghton et al. 1983; also see Chapter 6). Although changes in both areal extents and densities of carbon storage must be considered to determine the implications of changes in carbon storage in terrestrial pools, only the net flux from terrestrial reservoirs to the atmosphere or vice versa need be considered in analyzing the response of the atmosphere-ocean system because the dynamics of these major reservoirs are decoupled. In most studies of the terrestrial carbon fluxes in the context of the response of atmospheric CO_2 levels to fossil fuel use, terrestrial fluxes have simply been added as an additional term or source to the atmosphere. The reconstruction of the historical net terrestrial flux from changes in land use, areal extents of ecosystems, and associated impacts on carbon storage in vegetation and soils is analyzed in a separate data analysis and modeling framework (see Chapter 6).

7.2 MODELS OF THE ATMOSPHERE-OCEAN CARBON SYSTEM

7.2.1 Review of Atmosphere-Ocean Carbon Models

The oceans are recognized as the primary sink for excess concentrates on interactions between the atmospheric and oceanic reservoirs (see earlier discussion of this topic in Chapter 5).

In early work, Craig (1957) and Revelle and Suess (1957) used three-box models—atmosphere, surface ocean, and deep ocean—to analyze the distribution of radiocarbon and to assess the rate of oceanic CO_2 uptake from the atmosphere. In these models, four rate coefficients must be specified. Several approaches are used, but in general parameter values are computed for agreement between simulated and observed distributions of natural radiocarbon.

Keeling (1973) described the incorporation of surface-ocean chemistry in ocean carbon models. Similar to the models proposed by Craig, and Revelle and Suess, Keeling's model partitioned the ocean into a shallow surface layer and a deep portion, with the former exchanging carbon directly with a single-box atmosphere. The capacity of the surface ocean to incorporate excess CO_2 was evaluated by solving chemical equilibrium equations as described in the preceding section. Bacastow and Keeling (1973) suggested parameterizations of a similar model to project CO_2 concentrations under alternative scenarios for future fossil fuel use (Keeling and Bacastow 1977).

To better represent the vertical distribution of ^{14}C in the oceans, Oeschger et al. (1975) introduced a box-diffusion model. As described in the preceding section, the turnover of carbon below 75 m was represented by a diffusion equation. A constant coefficient of diffusivity was estimated to match an idealized profile of natural ^{14}C with measurements.

In the box-diffusion model the thickness of the surface layer is taken to be that of the average depth of the actual wind-mixed layer (average thickness, 75 m). However, if a 75-m surface layer is used in a three-box formulation, the simulated natural ^{14}C distribution does not agree with observations (Bacastow and Keeling 1979). In the three-box models, this difficulty is resolved by associating the surface box with waters to about 260 m. A rationale for adopting this thicker surface layer

is that the thermocline (from about 75 to 600 m) is a transition from surface to deep waters and the boundary in a two-compartment representation should not be at either extreme but rather somewhere near the middle of this zone (Bacastow and Keeling 1979).

A number of modifications to the basic box-diffusion model of Oeschger et al. (1975) have been proposed. These refinements are meant to increase the realism of ocean models by including features such as polar outcrop, biological activity, nutrient cycling, upwelling, and separate deep-ocean boxes for the Atlantic and Pacific.

Björkström (1979) proposed the first of a number of models whose structures are meant to be more realistic than the simple diffusive assumption but which are still basically one dimensional. He divided the surface ocean into cold and warm boxes (Figure 7.2). The subsurface water is represented by 10 horizontal layers, 2 of which correspond to intermediate waters. Water downwells directly from the cold surface layer into the intermediate and deep layers. There are direct return flows from intermediate waters to the cold surface layer. The only direct connection between adjacent subsurface layers is the upwelling flow that balances the sinking water. This model also allows a higher concentration of dissolved inorganic carbon in the cold surface layer; downwelling from this layer forms the deep water.

Hoffert et al. (1981) extended the box-diffusion model by incorporating upwelling in the conservation-of-mass equation for their deep-ocean box. A downwelling of polar water is assumed in order to balance water and carbon mass equations. The Hoffert et al. (1981) model also includes a carbon sink term associated with CO_2 removal by particulate outfall and photosynthesis.

Enting and Pearman (1982) divided the oceans into warm and cold components (Figure 7.3). Within each of these components, both turbulent mixing and upwelling are considered. A flux from the mixed layer to the deep ocean is present in the cold-ocean component. Warm and cold components exchange carbon to form a major circulation. Enting and Pearman (1982) modeled air-sea exchange in more detail than that found in most one-dimensional formulations.

Crane (1982) included a well-mixed deep polar reservoir in contact with the atmosphere and a split diffusive deep box subdivided into two reservoirs corresponding to the Atlantic and Pacific Oceans (Figure 7.4).

Peng et al. (1983) proposed a modified box-diffusion model (Figure 7.5) that incorporated most of the refinements cited above, including:

1. The oceanic photosynthesis-respiration cycle.

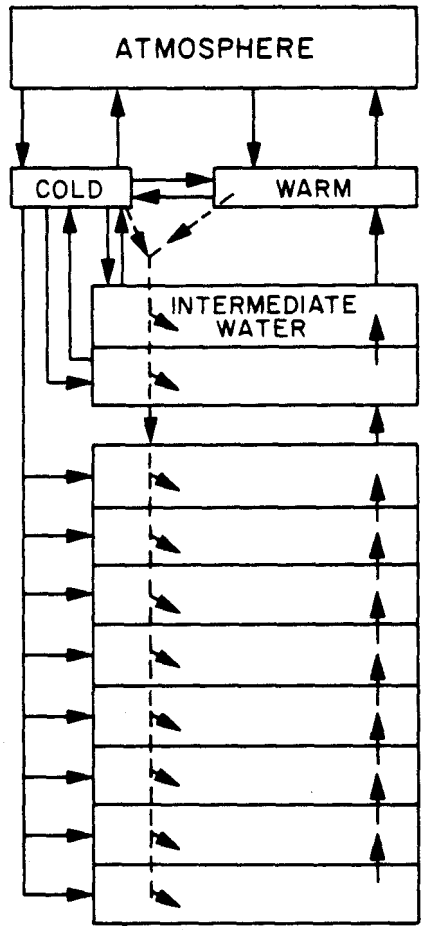


Figure 7.2. Box diagram for a modified box-diffusion model proposed by Björkström. The model divides cold and warm surface waters and explicitly treats advective fluxes. (Redrawn from Björkström 1979.)

2. The formation of deep water. A loop is inserted that brings intermediate water (from about 1000 m) to the surface in the polar region, increases its density, and sends it back to the bottom. Upwelling to intermediate waters completes the cycle. The eddy-diffusion coefficient below the level from which the source water is drawn is reduced to about one-half the value chosen by Oeschger et al. (1975) to account for the additional mixing provided by this cycle.

3. An eddy-diffusion coefficient for the main thermocline (i.e., above the depth of the deep water source) larger than that adopted by Oeschger et al. (1975), to be consistent with the penetration of bomb-produced tritium.

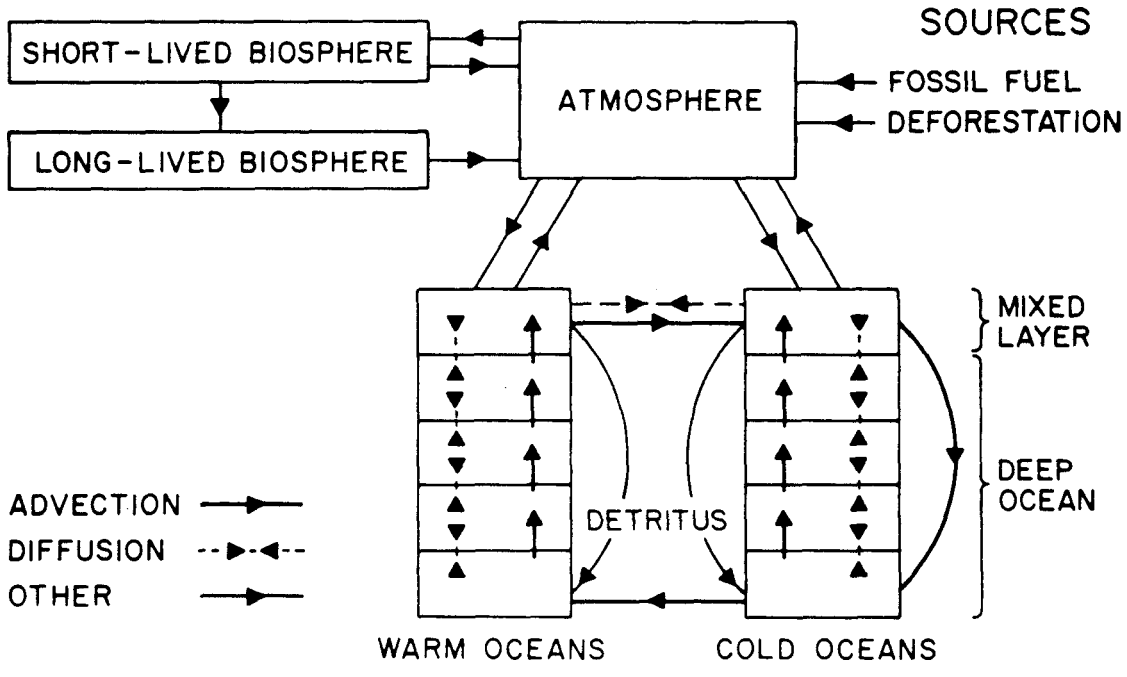
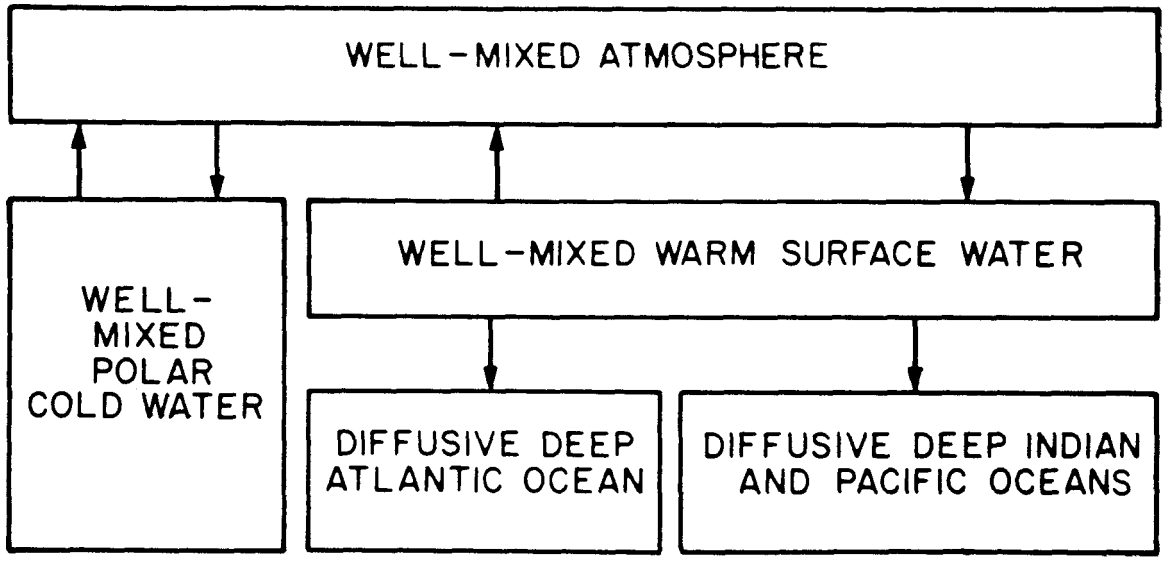


Figure 7.3. Model proposed by Enting and Pearman. Two major components of the oceans are considered, warm and cold. Within each component, cycling is by mixing and advection. (Redrawn from Enting and Pearman 1982.)

Figure 7.4. Box diagram for a regionalized ocean model proposed by Crane. (Redrawn from Crane 1982.)



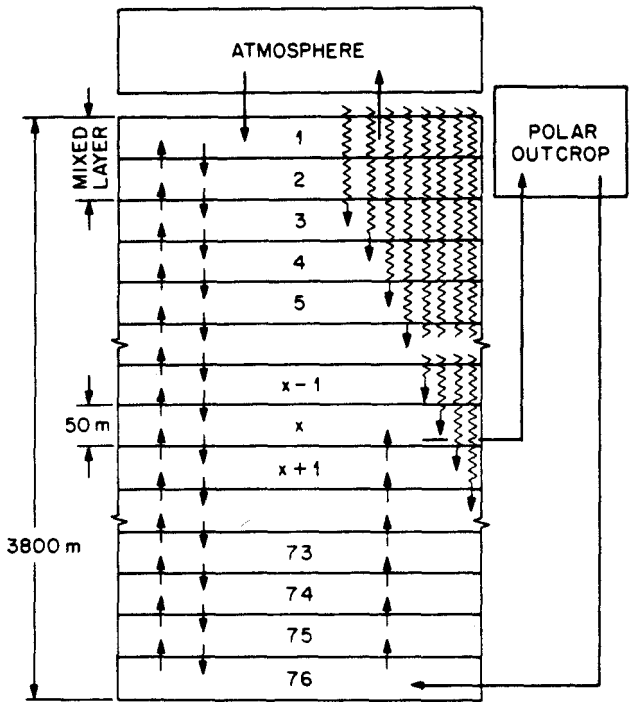


Figure 7.5. Diagram of a modified box-diffusion ocean model developed by Peng and co-workers. Wavy arrows are particle fluxes from the surface water, redissolved exponentially with depth. Deep water formation from intermediate water and subsequent sinking to the bottom and upwelling from the bottom are shown on the side loop with polar outcrop. Two different vertical eddy-diffusion coefficients are applied, with a higher value for the upper ocean. The major link between the atmosphere and the ocean is through gas exchange. (Source: Peng et al. 1983.)

Values of diffusivity for the thermocline and deep ocean, the downward advective flux, and the depth of the deep-water source were chosen for consistency with the observed mean penetration depth of bomb-produced tritium at the time of the GEOSECS survey and the observed mean depth distribution of natural radiocarbon. The representation of the particulate cycle is designed to reproduce the mean profile of phosphate (PO_4). All the phosphate reaching the surface ocean is incorporated into organic debris with a carbon:phosphorus ratio of 105. The debris falls to the thermocline and deep sea where it is destroyed, releasing PO_4.

Siegenthaler (1983) developed an outcrop-diffusion model that takes into account the direct ventilation of the intermediate and deep oceans at high latitudes (Figure 7.6). Outcrops for all sublayers are added to the box-diffusion formulation. The invasion of carbon into the deep water is controlled by gas exchange processes at the outcropping surface area. Once the carbon enters the outcrop, it is homogenized throughout the deep layer. The efficiency of the model in taking up excess CO_2 depends both on the assumed fraction of total ocean surface area assigned to outcrops for deep layers and the choice of ^{14}C tracer data used in calibration (see Chapter 5).

Broecker et al. (1980) described a model that emphasizes processes that operate on time scales of up to 30 years. These processes are most important in determining the rate of fossil fuel CO_2 uptake during the next 50 to 100 years. The model is structured around three main features of ocean circulation (Figure 7.7). It involves the upwelling of water that, at the time of the GEOSECS survey, was not yet contaminated with nuclear bomb products. On reaching the surface, this water flows poleward and downwells in the adjacent temperate zones. Vertical eddy mixing is superimposed on the upwelling in the tropical zone and the downwelling in the temperate zone.

The absolute rate of these processes is fixed from independent knowledge of the rate of CO_2 invasion (i.e., 228 g C m^{-2} $year^{-1}$ for the tropical zone and 264 g C m^{-2} $year^{-1}$ for temperate zone). A flux of 30 Sverdrups (30 $\times$ 10^6 m^3/s) is assumed for deep water formation in the North Atlantic and 40 Sverdrups (40 $\times$ 10^6 m^3/s) for the Antarctic. Both values are at the high end of the range of literature values. A vertical diffusion coefficient of 3 cm^2/s is assumed for high-latitude regions.

The model explains the distribution of bomb-produced ^{14}C in the tropical and north temperate Atlantic Ocean. The model-calculated time trend of ^{14}C in north temperate Atlantic surface water is comparable with measurements made by Druffel and Linick (1978) on a ring-dated coral from the Florida Straits.

Comparison of the ocean uptake estimated using the Oeschger et al. (1975) model with results from this regionalized model indicates that more fossil fuel CO_2 is taken up by the regional ocean model (53 $\times$ 10^{15} g C for 1973, versus 35 $\times$ 10^{15} to 42 $\times$ 10^{15} g C for the box-diffusion model).

Bolin et al. (1983) described a method for the simultaneous use of tracers to determine ocean circulation as well as biotic production and decomposition. Rates of exchange between oceanic regions (boxes in their model) due to advection and turbulent mixing and the rates of internal sources and sinks due to formation and dissolution of organic and inorganic compounds are determined in order to best reproduce a set of steady-state concentrations that satisfy boundary conditions and constraints (e.g., turbulent mixing is along the positive gradient; organic production can occur only in surface boxes).

This method for calibrating a model from tracer data was initially applied to a 12-box model of the world oceans: surface water, intermediate water, deep water, and bottom water of the Arctic, Atlantic, Antarctic, and

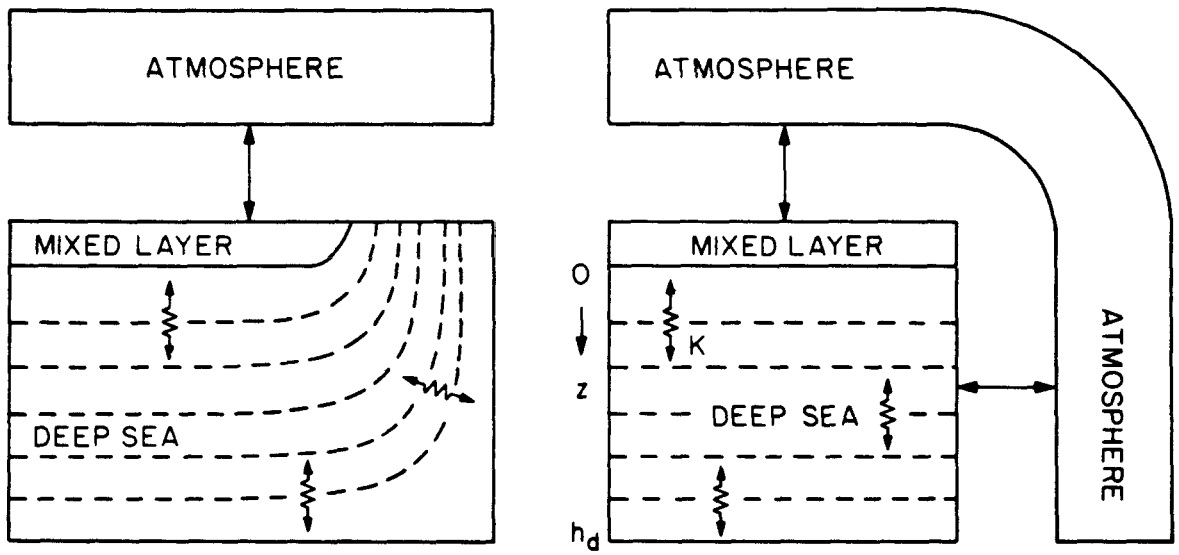
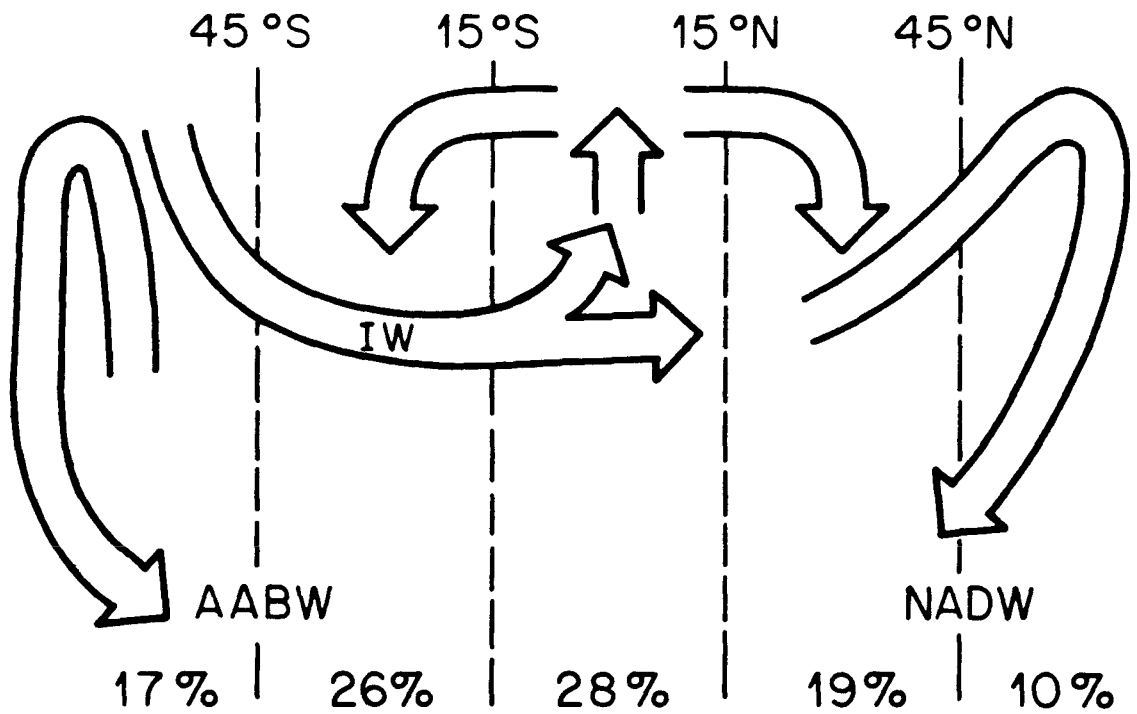


Figure 7.6. Diagram of box-diffusion models with polar outcrop. (Redrawn from Siegenthaler 1983.)

Figure 7.7. Patterns of flow in the ocean considered in a regional model by Broecker et al. (1980). Upwelling occurs in the equatorial zone; downwelling occurs in the temperate zone and polar zone. The ocean area in each zone is shown as a percentage of total ocean area.



Pacific/Indian Oceans, with some simplification for the polar seas. Total dissolved inorganic carbon, alkalinity, phosphorus, oxygen, and radiocarbon were used as tracers.

In a reference case model calibrated from these multiple-tracer data, a meridional circulation among cells is obtained (Bolin et al. 1983). Antarctic surface water penetrates into intermediate levels of the Atlantic and Pacific, where it lifts and is advected at the surface back toward the south. In the Atlantic, much of the penetrating intermediate water reaches sufficiently far north to join the Arctic deep water, which sinks and moves toward the deep Antarctic Ocean through the deep Atlantic. The total upwelling in the Antarctic Ocean is about 30 Sverdrups. Turbulent exchange prevails between surface and intermediate waters in both major oceans, with an average diffusivity of about 1 cm^2/s.

This 12-box model was used to estimate oceanic CO_2 uptake. The computed airborne fraction of fossil fuel CO_2 emissions during the period 1959 to 1980 was 74 to 75%, whereas the observed value was about 58% (see Table 5.3 in Chapter 5). The computed Suess effect (i.e., the change of $^{14}C/^{12}C$ ratio due to fossil fuel CO_2 emissions) in 1955 was between -3 and -7%, depending on model assumptions, whereas the observed value is about -2% (Stuiver and Quay 1981). Bolin et al. (1983) partially attributed these discrepancies to the fact that the exchange of ^{14}C with the terrestrial biosphere is not included in the model.

These researchers are also developing an 84-box model of the Atlantic Ocean (Bolin 1986; Moore and Björkström 1986), using calibration methods similar to those developed around their 12-box model. This Atlantic model considers 8 density layers and 12 oceanographic regions to define a total of 96 possible boxes. Because some regions do not contain water of each density category, this number is reduced to 84. Values of model parameters are estimated from data collected by the GEOSECS program. Concentrations for seven tracers (total carbon, ^{14}C, alkalinity, phosphorus, oxygen, salinity, and heat) are defined for each box. The model ocean exchanges material with the atmosphere; the Arctic, Pacific, and Indian Oceans; and the Mediterranean Sea.

In summary, one-dimensional or globally averaged models, calibrated against tracer data, are currently the main formulations available for analyzing the uptake of carbon from the atmosphere by the oceans. The basic box-diffusion concept, first proposed by Oeschger et al. (1975), has not been substantially improved by modifications or alternative formulations. The box-diffusion model has two major parameters that require estimation: CO_2 invasion rate in the surface ocean and eddy diffusivity in the subsurface ocean. Values of these parameters were estimated from several data sets (Broecker et al. 1980) but may be further refined as additional data become available.

The main criticism of the box-diffusion model is that it does not structurally match the large-scale features of ocean circulation, particularly with regard to advective fluxes at high latitudes. Further, the parameterization of all ocean physics into a single vertical diffusion coefficient does not realistically represent actual ocean-mixing processes (Brewer 1983). Diffusion is not a major contributor to vertical mixing in the real ocean. Although a number of essentially one-dimensional models that incorporate explicit treatment of advection or that divide the oceans according to major regions have been proposed, these models introduce additional parameters that are currently difficult to estimate from existing data.

The greatest uncertainty in applying one-dimensional models is whether explicit treatment of more realistic ocean physics might alter estimates of the rate of carbon uptake from the atmosphere. The potential importance of advective fluxes, as an example, highlights the need for better representations of oceanic processes in spatially differentiated models. As discussed below, preliminary two-dimensional models have been developed that make better use of available data (also see Chapter 5).

7.2.2 Testing and Comparison of Models of the Atmosphere-Ocean Carbon System

Accurate simulation of atmospheric CO_2 measurement records is the primary test of carbon cycle models used to relate changes in atmospheric CO_2 levels to fossil fuel use. Models and assumptions that do not satisfy this basic criterion are of only secondary interest in projecting responses to further fossil fuel use. However, this basic test is complicated by uncertainty about CO_2 concentrations before the beginning of accurate measurements in 1958.

It is difficult to set initial conditions corresponding to conditions at the beginning of CO_2 measurements in 1958. Data are inadequate to characterize the partitioning of anthropogenic CO_2 released prior to 1958 among carbon reservoirs represented in models. If fossil fuels are the only source of excess CO_2 considered, initial conditions at 1860, corresponding to the start of significant fossil fuel use, can be assumed. Inclusion of other significant anthropogenic or external natural sources requires a different set of initial conditions.

The carbon cycle is assumed to be initially in equilibrium before the onset of significant anthropogenic releases to the atmosphere. Parameter values are constrained by this assumption, and model simulations are referenced to the initial concentration of CO_2 in the atmosphere. However, the preanthropogenic CO_2

concentration is not accurately known, and these values must be determined by indirect means (Chapter 3).

Oeschger et al. (1975) stated the likely range on the preanthropogenic CO_2 concentration to be from 288 to 295 ppm and assumed a value of 292 ppm. Rust et al. (1979) formulated an exponential function for best agreement with the Mauna Loa CO_2 record and back-extrapolated to obtain a base concentration of 290 ppm. Ambiguity in deriving "best" values was noted. A number of other estimates of preanthropogenic CO_2 concentration are summarized by Elliott (1984). Chapter 3 provides a detailed review of premeasurement CO_2 levels.

The first test of atmosphere-ocean models against CO_2 measurement records is to assume an initial atmospheric concentration in 1860 of about 290 ppm and to assume that the only input of carbon to the atmosphere is the CO_2 released from fossil fuels. For reference case parameter values, Oeschger et al. (1975) found that their box-diffusion ocean model could not accommodate sufficient carbon to account for the difference between the estimated fossil fuel release and the observed increase in the carbon content of the atmosphere. Terrestrial releases can exacerbate this inconsistency.

Nuclear weapons testing between 1945 and 1974 produced about 1.3×10^6 g of radiocarbon (Killough and Till 1978). Large quantities of other isotopes were also released. The distribution of bomb ^{14}C, indicated by direct tropospheric measurements and the GEOSECS surveys of the oceans, provides a basis for testing models independent of their calibration against the natural ^{14}C distribution. Comparison against other isotopes, particularly tritium, is useful as well.

Broecker et al. (1980) reviewed existing models of oceanic CO_2 uptake and concluded that the model of Oeschger et al. (1975) gave a good fit to distributions of both natural and bomb ^{14}C. This conclusion was based on three tests. Using the observed atmospheric time history of ^{14}C as an exogenous function, they integrated the model equations to 1973 (the midpoint of the GEOSECS Atlantic and Pacific sampling programs) and compared an estimate of the excess ^{14}C inventory in the ocean, based on observations, with that predicted by the model. Data from the later (1977-1978) Indian Ocean sampling program were not considered in the analysis. The estimate derived from observations exceeded that predicted by the model by 8%, an amount less than the uncertainty in the former quantity.

The second test considered by Broecker et al. (1980) was based on a similar integration of the model equations to compare the predicted increase in surface-water ^{14}C with estimates of the excess based on measurements. The 1973 value predicted by the box-diffusion model exceeded the observed value by 12%. The third test sought to determine if the estimated uptake of ^{14}C by the ocean and biosphere could account for the measured decrease in atmospheric ^{14}C during the period 1965–1969.

Table 7.1
Summary of Oceanic Fossil Fuel CO_2 Inventories Obtained Through the Use of Various Sets of Parameters for the Box-Diffusion Model

	K (cm^2/s)	E (mol m^{-2} y^{-1})	ΣCO_2 (10^{14} mol)
Oeschger et al. (1975)			
Best Case	1.25	21	29.0
Radon-based E			
Tritium-based K	1.7	16	31.1
Bomb ^{14}C			
Upper Limit	3.3	24	42.2
Best Fit	2.6	21.5	38.1
Lower Limit	1.9	19	33.3
Best Estimate	2.2	19	35.1

Note: K is vertical diffusivity, E is CO_2 exchange.
Source: Broecker et al. (1980).

Table 7.1, extracted from Broecker et al. (1980), compares predicted CO_2 uptake for specified values of vertical diffusivity (K) and net exchange rate (E) estimated from different data and assumptions. The table indicates a difference of about 21% between the predicted inventory based on the parameter values of Oeschger et al. (1975) and that based on the "best" estimates by Broecker et al. (1980).

An important ocean model comparison was carried out by Bacastow and Björkström (1981). Four models were considered: two-box (2B), box-diffusion (BD), advection-diffusion (AD), and multibox (MB).

The AD model is an extension of the box-diffusion model. It features a direct flow or "pipeline" from the surface layer to the deep ocean, with advective upwelling to balance the sinking water. This advective cycle is superimposed on a diffusion model, and Bacastow and Björkström (1981) asserted that suitable choices of the coefficient of diffusion and the advective flow parameter will make the AD model imitate either of the simpler models (2B or BD). The MB model was based on the work of Björkström (1979) as described in the preceding section.

Bacastow and Björkström (1981) calibrated the 2B, BD, AD, and MB models to oceanic ^{14}C data in as consistent a manner as possible. A linearized buffer parameter, an exponential input function with characteristic time constant $1/\mu = 22.5$ years, and $\rho_{mo} = 1 + (\Delta^{14}C)_{mo}/1000 = 0.95$ (the steady-state activity ratio value for the surface water) were assumed. They plotted isopleths of ρ_{so}, the activity ratio fraction averaged over the entire ocean, and model-calculated airborne fractions (AF*s) versus values of a long- and a short-term parameter for each model.

Observed limits on the activity ratio fraction and AF* were translated into a region of the model-

parameter plane. Examples of short- and long-term model parameters are the depth of the surface layer (h_m) and the deep-to-surface transfer rate coefficient (K_{dm}) in the case of the 2B model; and (h_m) and the coefficient of diffusion K in the case of the BD model.

Applied to the 2B and BD models, this procedure yields values of h_m that are many times the values normally used (hundreds of meters versus the 75- to 100-m). Bacastow and Björkström (1981) emphasized the dependence of the AF* on the depth chosen for the surface layer of the ocean, pointing out that "the lowest ultimate AF* will be achieved by the model that has the most ocean water in the surface layer or closely coupled to it." Their comparison, plotted as a function of the exogenous function's time constant μ, shows that, by a substantial margin, the highest AF* (and the lowest CO_2 uptake) is associated with model 2B, with the surface layer depth $h_m = 75$ m. The lowest AF* is given by model BD with $h_m = 75$ m, for μ less than 1/(22.5 years), the point at which model 2B with $h_m = 340$ m overtakes it (this is not accidental; the value 340 m was calculated to give the same AF* as model BD for this value of μ). In this and other comparisons, their results indicate that model AD gives a higher AF* than model BD.

Bacastow and Björkström (1981) concluded that, for projections into the near future, the use of model BD with a somewhat higher value of the diffusivity parameter (K) than that adopted by Oeschger et al. (1975) might be satisfactory. But for longer term calculations, the deep water would not be adequately represented in such an arrangement, and they suggested the possibility of using a box-diffusion model with a deeper surface layer ($h_m = 201$ m) and a diffusivity parameter (K), recalculated to match average ^{14}C activity for the ocean, equal to 1.437 cm^2/s, comparable to that used in the Oeschger et al. (1975) formulation.

Killough and Emanuel (1981) compared five prototypes of one-dimensional models of carbon turnover in the ocean. All five were consistently calibrated to the distribution of natural radiocarbon versus depth. The principal measure for characterization of the models' behavior over time was the probability distribution of a carbon atom's transit time, but, as will be discussed here, direct computations of the cumulative uptake of carbon were also computed by integrating the differential equations with an exogenous history of the atmosphere.

The structures of the five models are shown in Figure 7.8. Model 1 consists of a surface layer and a deep-water reservoir. It is subdivided into models 1a and 1b according to the depth of the surface layer: 75 m in the case of model 1a and 260 m for model 1b.

The three remaining models consist of a 75-m surface layer and a deep reservoir subdivided into 18 horizontal layers. The layered models differ from each other only in the arrangement of fluxes among the boxes. Of the layered group, model 2 injects carbon from the surface layer directly into each of the subsurface layers, and each of the latter layers returns carbon directly to the surface layer. There is no direct mixing of carbon among the subsurface layers.

Models 3a and 3b move carbon upward through the subsurface water by transfers from layer to layer. Model 3b represents downward transport in the same way, with flows joining adjacent layers, whereas model 3a uses direct injection from the surface layer to the deep water layers.

All of the layered models take into account the variation of the ocean's horizontal cross-sectional area and inorganic carbon concentration with depth. The former consideration and the larger number of layers differentiate model 3b from the box-diffusion model of Oeschger et al. (1975).

In all time integrations with the five models, nonlinear buffer chemistry in the surface water was taken into account. Using a specified history of atmospheric CO_2 levels based on the derivation by Rust et al. (1979), Killough and Emanuel (1981) computed cumulative uptake of CO_2 by the ocean for each of the five models. The cumulative uptake curves versus time are shown in Figure 7.9. For the period from 1860 to 1975, these curves show a considerable range of model performance. By 1975, model 1 (two-box, $h_m = 75$ m) has taken up only 24×10^{15} g C, whereas model 3a has taken up 66×10^{15} g C, nearly a factor of 3. Model 3b, similar to a box-diffusion model, takes up slightly less carbon than model 3a, with 60×10^{15} g C. The performances of model 1b (two-box, $h_m = 260$ m) and model 2 (direct injection from surface water to deep layers followed by direct return to surface water) are quite similar and intermediate in character, with 43×10^{15} and 41×10^{15} g C uptake, respectively.

It is difficult to compare these results with the work of Bacastow and Björkström (1981). Their conclusions suggest greatest carbon uptake by the Killough and Emanuel (1981) model 2 because of the direct coupling of all subsurface layers with the surface layer, a condition they associated with the low AF*. The probable resolution to this inconsistency lies in the fact that the surface layer is in contact with boxes with a wide range of time constants, from very fast to very slow, depending on the depth of the layer. Only the boxes with fast turnover time contribute strongly to enhancing ocean CO_2 uptake and, thus, reduce the AF*. Because the boxes with slower turnover times are effectively in competition with the former in the exchange of carbon with the surface layer, they diminish the effectiveness of the direct interaction with the surface layer.

This comparison illustrates the importance of the interplay between structure and calibration. The

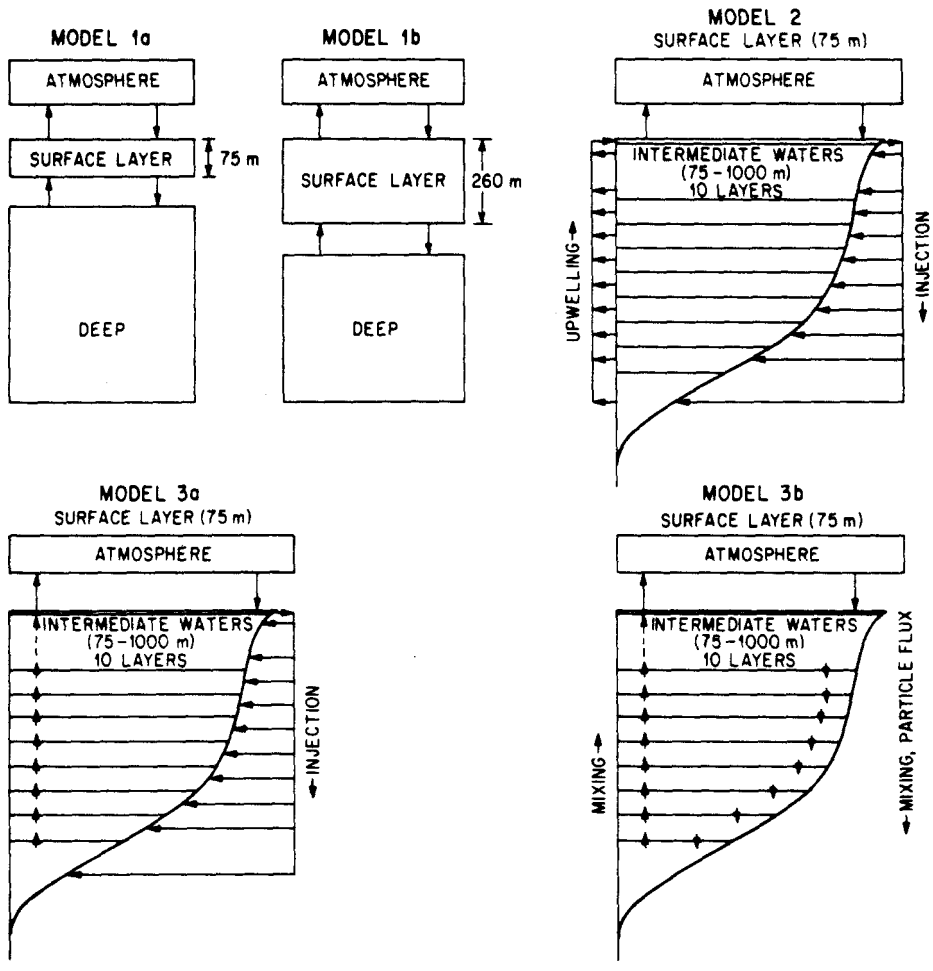


Figure 7.8. Alternate compartment models of carbon exchanges in the ocean. All may be represented by a surface layer which exchanges CO_2 with the atmosphere and a deep ocean. Areas of the layers are proportional to the respective water volumes.

advection model 3a of Killough and Emanuel (1981) is similar to a limiting case of the advection- diffusion model (AD) of Bacastow and Björkström (1981), yet model AD consistently produced a higher AF* than did model BD (Bacastow and Björkström 1981). This behavior is expected in light of the authors' remarks about the effects of the close coupling of large volumes of water with the surface layer, and in model AD, a further contribution is produced by diffusion in the subsurface layers. However, there may be a point at which the advective flow gains in importance as diffusion diminishes. Model 3a of Killough and Emanuel (1981) represents a case with virtual extinction of diffusion, with only the advective cycle active and calibrated in such a way as to preserve the required natural radiocarbon profile. Model 3a is marginally more effective than model 3b in taking up CO_2 from the atmosphere, whereas the results of Bacastow and Björkström (1981) suggest that the reverse should be the case. These comparisons suggest that one-dimensional models may not be as well characterized as is commonly believed.

The apparent inability of one-dimensional ocean models to accommodate enough fossil fuel CO_2 to balance the global carbon budget may be attributed, in part, to lack of both spatial detail and parameterizations of important physical processes. The development of two-dimensional ocean models with outcropping isopycnal horizons is an important step toward understanding the role of oceans in the global carbon cycle.

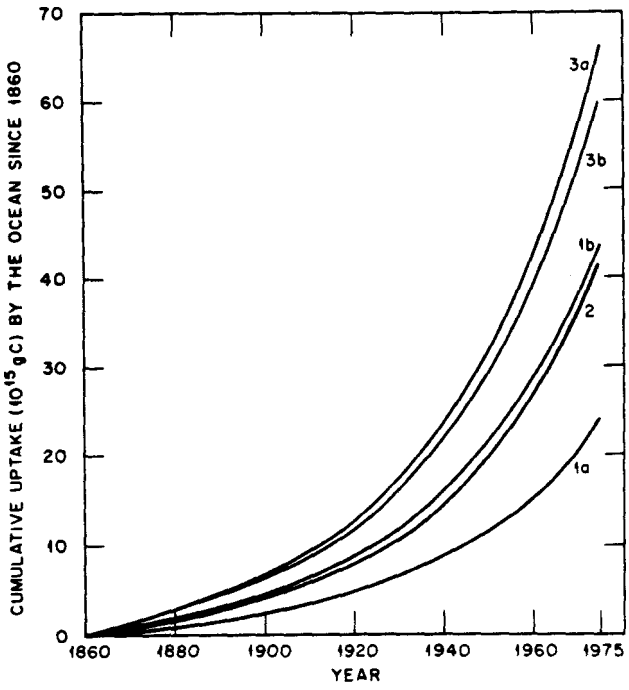


Figure 7.9. Cumulative net uptake of the excess CO_2 by the five ocean models (Figure 7.8) for the period 1860–1975.

Peng and Broecker (1985) compared the behavior of two-dimensional isopycnal thermocline ventilation models of the North Atlantic to evaluate the sensitivity of CO_2 uptake to circulation patterns. The distributions of anthropogenic ^3H, ^3He, ^{14}C, ^{85}Kr, and chlorofluorocarbons were used to calibrate these models. The value of using multiple tracers lies in the considerable differences in their temporal and geographic input functions and their geochemistries in the sea.

Tritium (^3H), for example, is carried to the ocean surface by rain and vapor exchange shortly after it reaches the troposphere. Thus, the delivery of this isotope was influenced strongly by the locations of the bomb tests, which released major amounts of tritium, and by characteristics of atmospheric dispersal. Once in the ocean the tritium remains there. By contrast, ^{85}Kr and the chlorofluorocarbons will remain primarily in the atmosphere. These gases are almost uniformly dispersed throughout the atmosphere, and their concentrations in the surface ocean are geographically symmetrical. Only differences related to the temperature dependence of their solubilities are to be expected.

Because of the large ratio of $HCO_3^- + CO_3^{2-}$ to CO_2 in sea waters, the $^{14}CO_2$ produced during nuclear bomb tests will ultimately be transferred largely to the ocean. However, the large concentration of inorganic carbon also dictates that the $^{14}C/^{12}C$ ratio in surface water is slow to achieve equilibrium with the atmosphere. Hence, the transfer of $^{14}CO_2$ across the air-sea interface constitutes a major resistance to its oceanic uptake. Helium-3 is produced by the decay of ^3H within the ocean and ultimately is irreversibly lost to the atmosphere. Thus, its distribution within the ocean is quite distinct from those of the other tracers.

Carbon dioxide differs from the other tracers in two important ways. First, its input extends over a much longer period of time. Second, its uptake characteristics lie geometrically between those of $^{14}CO_2$ and those of ^{85}Kr and the chlorofluorocarbons.

The equilibration time between surface water and the atmosphere for CO_2 ($\sim$1 year) is roughly an order of magnitude longer than that for ^{85}Kr and the chlorofluorocarbons and an order of magnitude shorter than that for $^{14}CO_2$. The chemical equilibration time for CO_2 and the isotopic equilibration time for $^{14}CO_2$ differ because chemical equilibration involves the adjustment of the CO_3^{2-} ion content of the water, whereas the isotopic equilibration involves all of the carbon in the water (Broecker et al. 1980). The carbonate ion makes up only about 10% of the carbon dissolved in surface ocean water.

The characteristic of the thermocline that makes these differences among tracers potentially important is that mixing is thought to occur mainly along horizons of constant potential density (i.e., isopycnal surfaces). These horizons reach the surface only during the peak winter period. The input characteristics of individual tracers to these outcrops differ significantly. Krypton-85 and the chlorofluorocarbons should achieve near equilibrium in surface water, even during the short winter exposure of waters at the isopycnal outcrops. Any tritium-derived ^3He brought to the surface during the period of outcropping will be lost in large part to the atmosphere.

On the other hand, little atmospheric tritium or ^{14}C will enter the outcrop region from the air during these brief periods. Rather, these isotopes are mainly transported laterally from the adjacent broad regions of the warm surface oceans. The situation for CO_2 lies between those for ^{14}C and ^{85}Kr.

Peng and Broecker (1985) depict a single hemispheric sector (15°N-54°N) by an 11-box model (Figure 7.10). Three levels of the thermocline (upper, middle, and lower) are each subdivided into two well-mixed interior reservoirs and one outcrop reservoir. The model thermocline is overlaid by two surface reservoirs (15°N-30°N and 30°N-45°N). The sidewelling, upwelling, backwelling, and downwelling model structures are considered to be isopycnal mixing models. Inputs into the thermocline are by outcrops of isopycnal horizons i high-latitude regions. The diffusion model is equivalen.

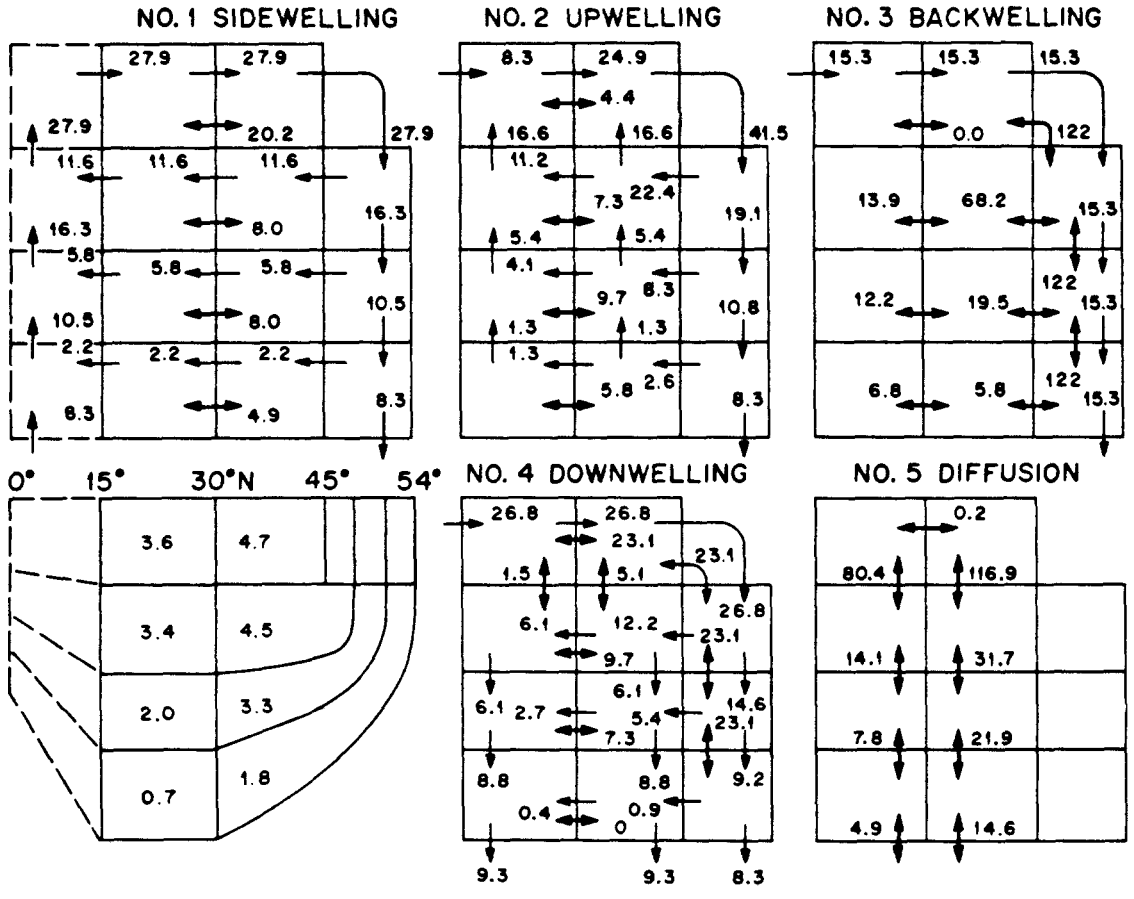


Figure 7.10. Circulation patterns of a set of two-dimensional temperate thermocline models of the North Atlantic Ocean. The isopycnal horizons and their outcrops are shown at the lower left of the diagram. The tritium distribution observed during GEOSECS is also shown. For convenience in presenting the patterns, the polar outcrops between 45°N and 54°N are moved to the side. The fluxes are in units of Sverdrups (10^6 m^3/s).

to the box-diffusion model in which inputs to the thermocline are only through overlaid surface boxes by vertical mixing. The observed distributions of ^3H, ^3He, ^{14}C, ^{85}Kr, and chlorofluorocarbons were analyzed to test the viability of several different advective flow patterns as shown in Figure 7.10. For each circulation pattern, absolute flow rates and diffusive mixing rates were adjusted for approximate agreement with the tritium distribution observed during the GEOSECS program. Simulations of other tracers were compared among models as well as between models and observed distributions.

The CO_2 exchange rates between the atmosphere and the ocean were calibrated for each model using GEOSECS ^{14}C data. The simulated inventories of excess CO_2 for 1981 are given in Table 7.2; the model-to-model differences are quite small. Peng and Broecker (1985) concluded that uptake of anthropogenic CO_2 by two-dimensional isopycnal models employing outcrops was not significantly different from that of one-dimensional box-diffusion models without outcrops.

Although the Peng and Broecker (1985) work suggests that, at the level of spatial detail considered, two-dimensional isopycnal ocean models do not imply a significantly larger sink for CO_2, the polar ocean north of 54°N and new North Atlantic Deep Water formation were not considered. Also, the impact of southern ocean circulation on the tracer distribution in the northern ocean was not included. Peng and Broecker (1985) noted that simulations of observed 1981 Transient Tracers in the Ocean (TTO) ^3H distribution indicate that these models do not generate enough ^3H penetration in the thermocline. Thus, until a large-scale two- or three-dimensional model is developed,

Table 7.2
Summary of Water-Column Excess CO_2 as of 1981

Fossil Fuel Only	Sidewelling No. 1	Upwelling No. 2	Backwelling No. 3	Downwelling No. 4	Diffusion No. 5
15°N-30°N	60.2	61.1	69.5	69.1	74.6
30°N-45°N	69.3	72.9	81.5	78.2	74.4
Total	129.5	134.0	151.0	147.3	149.0
Ratio[a]	0.87	0.90	1.01	0.99	1.00
Fossil Fuels Plus Biota					
15°N-30°N	138.4	138.6	153.9	155.5	148.3
30°N-45°N	150.8	154.5	169.4	166.8	143.8
Total	289.2	293.1	323.3	322.3	292.1
Ratio[a]	0.99	1.00	1.11	1.00	1.00

[a]Ratio of total excess CO_2 in each model to that in model No. 5.

which also simulates other aspects of ocean physics and circulation, the significance of multidimensional models for CO_2 uptake remains uncertain.

7.3 MODELING CHANGES IN TERRESTRIAL CARBON STORAGE

Initially, terrestrial carbon pools (vegetation and soils) were included in global carbon cycle models to account for their influence on the distribution of ^{14}C. In these geochemically oriented studies, terrestrial ecosystems were divided according to "live" and "dead" or "fast" and "slow" carbon pools (Figure 7.11). These pools were assumed to remain in equilibrium with respect to total carbon, with income and loss fluxes balancing exactly.

For fast and slow compartments, each pool exchanges carbon with the atmosphere independently. Equilibrium compartment contents and turnover times determine exchange rates. If carbon is divided between live and dead boxes, carbon is assimilated by the live compartment, transferred as a result of mortality and litterfall pools from live to dead, and lost from the dead pool. Loss fluxes are estimated from average standing levels of carbon storage and turnover times for each compartment. Net production fluxes depend on the standing levels of carbon in terrestrial pools and may also be influenced by atmospheric CO_2 concentration. The inadequacy of estimates of carbon uptake rates by the oceans in accounting for the difference in estimated releases from fossil fuels and the observed change in the carbon content of the atmosphere is one motivation for seeking an additional sink in vegetation and soils.

Bacastow and Keeling (1973) expressed the fluxes corresponding to assimilation by fast and slow terrestrial compartments by functions of the form

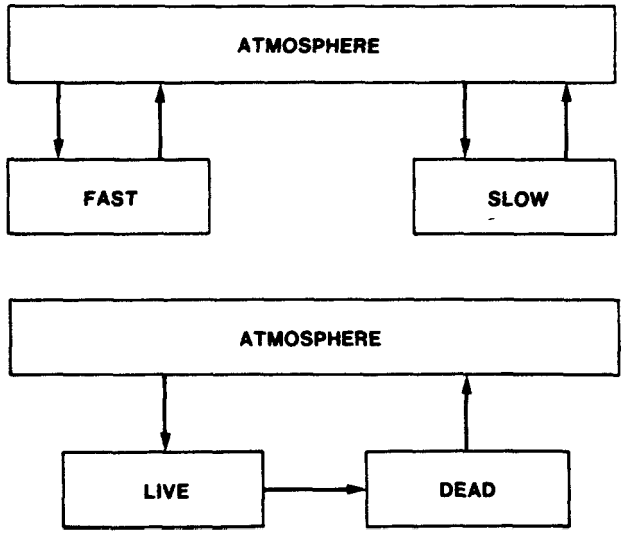
$$\mathrm{NPP} = \mathrm{NPP}^{\circ}[1 + \beta \ln(C_A/C^{\circ}_A)](C^{\circ}_{TS}/C^{\circ}_{TS}),$$

where NPP is the assimilation flux; C_A is the mass of carbon in the atmosphere; C_{TS} is the mass of carbon in the slow compartment; and superscript ° indicates an initial value. The parameter β dictates the dependence of the net production flux on the carbon content of the atmosphere. The relative increase in net production is approximately β times the relative increase in atmospheric CO_2 concentration.

Bacastow and Keeling (1973) adjusted the value of β for best agreement between model response and the atmospheric CO_2 record from Mauna Loa. The simulated Suess effect was also considered.

Setting β to zero, Bacastow and Keeling (1973) were unable to adjust other parameter values over reasonable ranges to yield satisfactory agreement with the Mauna Loa record. This result was taken to

Figure 7.11. Two-box model structures commonly used in geochemically oriented carbon cycle models to represent terrestrial pools.



imply an increase in terrestrial carbon storage, at least during the period of the record. Unambiguous "best" parameter values cannot be selected using the tests applied by Bacastow and Keeling; however, values of β between 0.4 and 0.5 give reasonable results—other parameters being set to nominal values. Bacastow and Keeling (1973) further noted that the standing levels of carbon and turnover times in terrestrial ecosystems are poorly known, but from the standpoint of simulated changes in atmospheric CO_2 levels, their own model's response is not significantly sensitive to these values.

Subsequent studies that focused on alternative representations of carbon turnover in the oceans (e.g., Oeschger et al. 1975; Björkström 1979) adopted similar formulations for incorporating a CO_2 fertilization effect. Clearly, more rapid oceanic uptake permits lower values of β, in other words a smaller CO_2 fertilization effect.

The hypothesis that terrestrial carbon storage increased during the period of the Mauna Loa record conflicts with data on forest clearing and similar land-use changes (e.g., Bolin 1977; Olson et al. 1978; Woodwell et al. 1978; Houghton et al. 1983; also see Chapter 6) that indicate a net decrease in terrestrial pools due to these human activities.

There are several ways to resolve this inconsistency. First, estimates of the rate of oceanic uptake may be too low, implying the need for modifications to ocean models. A second possibility for accommodating terrestrial releases follows from the uncertainty over the premeasurement history of CO_2 levels (see Chapter 3). The preindustrial CO_2 concentration was lower than is implied by simply extrapolating the measurement record backward in time. Although current estimates of oceanic carbon uptake are not consistent with releases from terrestrial ecosystems during the period of the Mauna Loa CO_2 record, substantial releases before the beginning of measurements may be accommodated if in fact early CO_2 concentrations were lower than those previously assumed. Time series of $^{13}C/^{12}C$ ratios in tree rings can be used to analyze this possibility.

A number of investigators have assembled tree-ring ^{13}C time series (e.g., Stuiver 1978; Freyer 1979; Tans and Mook 1980; Francey and Farquhar 1982; Freyer and Belacy 1983; Stuiver et al. 1984; Freyer 1986). Most records show a decreasing trend in the relative abundance of ^{13}C that in general cannot be accounted for by the release of carbon from fossil fuels depleted in ^{13}C relative to the atmosphere. Release of carbon by forest clearing and similar land-use changes may have been the additional ^{13}C-depleted source responsible for the decrease in atmospheric ^{13}C abundance.

Using a model to account for the influence of the oceans on ^{13}C geochemistry, the net flux of carbon (in addition to the fossil fuel source) required to match the change in ^{13}C abundance in any particular tree-ring record can be determined. Once the additional source is specified, the initial atmospheric CO_2 concentration can be adjusted for the best agreement between model response and the modern record. In this approach, the tree-ring ^{13}C time series serves as a constraint in place of the premeasurement CO_2 history.

Peng et al. (1983), using the modified box-diffusion model described in the preceding section, analyzed the ^{13}C record assembled by Freyer and Belacy (1983). The net flux of carbon, with $^{13}C/^{12}C$ at -25‰, required to produce the decreasing trend in the Freyer and Belacy (1983) record is shown in Figure 7.12. Setting the initial atmospheric CO_2 concentration to 245 ppm gives general agreement between the response of the modified box-diffusion model and the Mauna Loa CO_2 record as shown in Figure 7.13, but the curvature (trajectory) of the CO_2 trend is not simulated accurately.

In their analysis of the Freyer and Belacy (1983) record, Peng et al. (1983) adjusted the value of a net flux into the atmospheric compartment of their model to bring the simulated $^{13}C/^{12}C$ ratio into agreement with the tree-ring record. Although this additional net flux is assumed to be associated with terrestrial pools, the internal changes in the terrestrial components of the model were not investigated. Revised estimates of terrestrial fluxes have been made by Peng and Freyer (1986), following the substantial revisions to the Freyer and Belacy (1983) tree-ring record reported by Freyer (1986).

Emanuel et al. (1984) proposed a five-compartment model of carbon cycling in the world's terrestrial ecosystems that includes provision for direct transfers from plant compartments to the atmosphere as a result of land-use change. Forest clearing is emphasized in the model, and indirect releases from slash material and other dead organic matter are made available for decomposition by disturbance from human activities. The dynamics of ^{13}C are also modeled, including the discrimination against the heavier isotope in primary production. Changes in atmospheric ^{13}C abundance that result from changes in terrestrial pools and fluxes with land-use change can therefore be analyzed.

Carbon in plants is divided between "trees" and "ground vegetation," with a further distinction between "nonwoody parts of trees" (foliage, flowers, fruits, and transient roots) and "woody parts of trees" (branches, boles, and most root material) (Figure 7.14). Two compartments represent carbon in dead parts of terrestrial ecosystems and in their decomposers. The "detritus and decomposers" compartment corresponds to litter and its decomposers at the soil surface. This pool receives the input of carbon from decay of the aboveground parts of vegetation. The active soil carbon

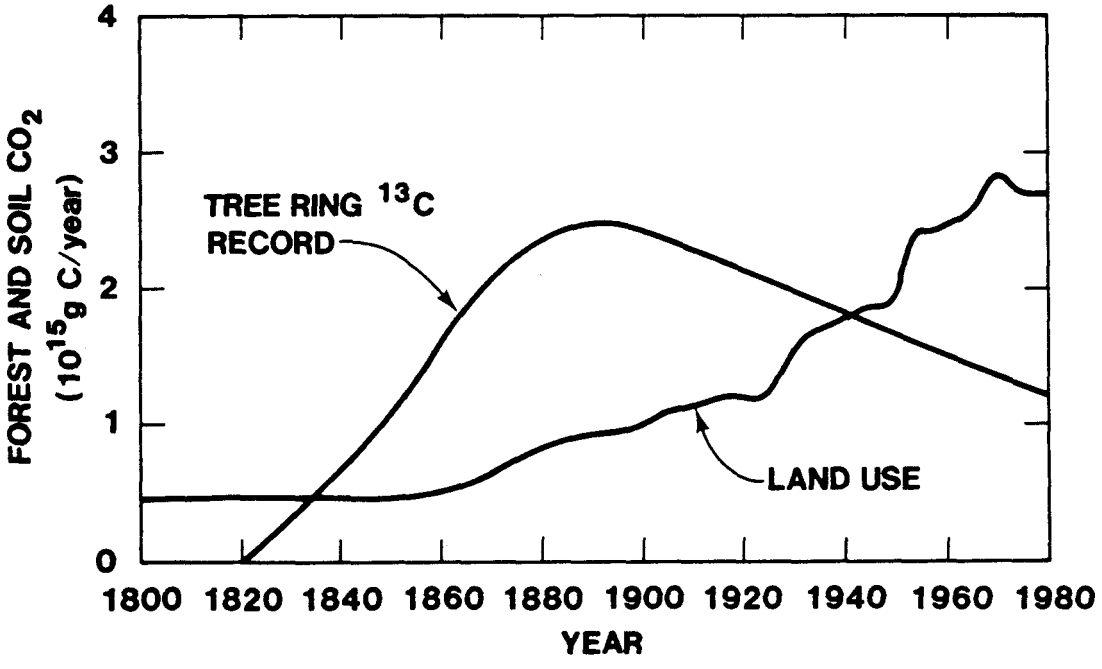
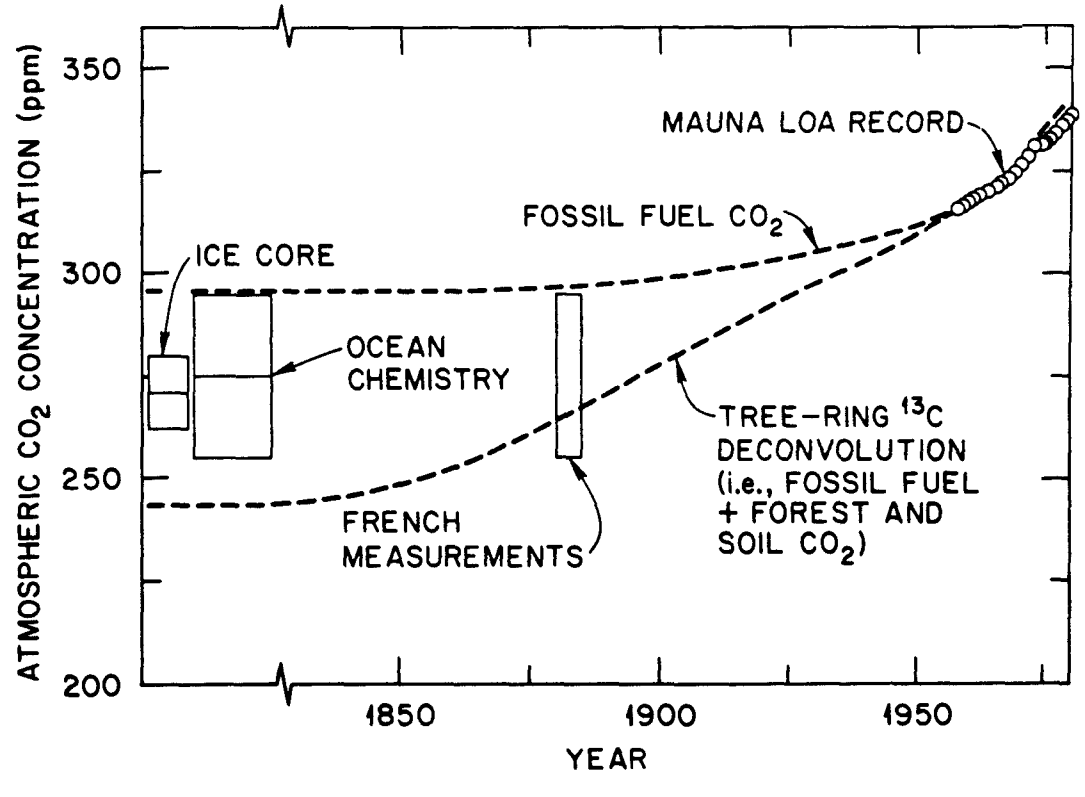


Figure 7.12. Terrestrial source reconstruction based on interpretation of ^{13}C tree-ring data (Peng et al. 1983).

Figure 7.13. Response of a modified box-diffusion model (Peng et al. 1983) to the combination of fossil fuel and terrestrial sources.



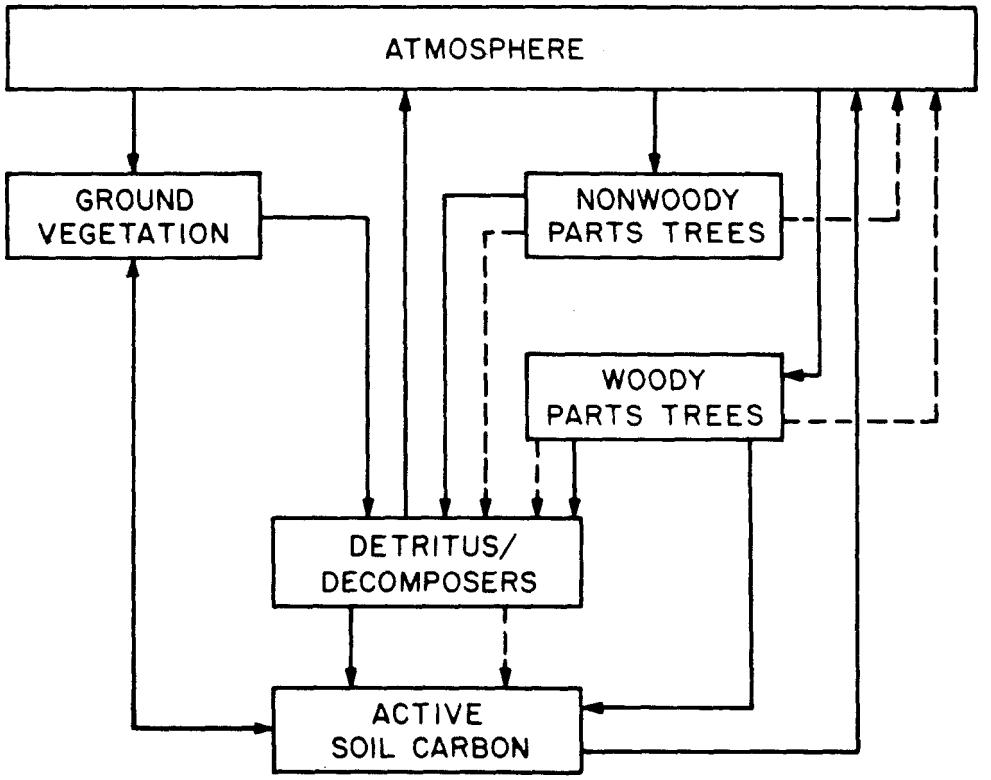


Figure 7.14. Five-compartment breakdown for terrestrial ecosystems. Carbon in living material is divided between tree and ground vegetation compartments, whereas carbon in dead organic matter is allocated either to detritus/decomposers or to active soil carbon. Solid arrows indicate natural fluxes from the atmosphere corresponding to net primary production and back to the atmosphere by decomposition. Dashed arrows indicate transfers of carbon because of forest clearing.

pool consists of carbon in soils and in its decomposers that is in relatively active exchange with the atmosphere. Carbon is transferred into this pool from the initial decomposition of belowground parts of vegetation and the transport of decomposed material from the actively decaying litter layer.

The primary effect of forest clearing in this model is the removal of carbon from both tree compartments on the basis of the ratio of their carbon masses. A fraction is transferred directly to the atmosphere; the remainder is transferred either to detritus and decomposers, where it is eventually released by decomposition or sequestered in active soil carbon, or out of the system, corresponding to transformation to an unreactive form such as charcoal. Because forest-clearing activities frequently result in a stirring or tilling of soil, making additional carbon amenable to rapid decomposition, a carbon transfer from active soil carbon to detritus and decomposers, given as a specified fraction of the forest-clearing flux, is allowed.

The secondary effect of forest clearing is land-use change, implemented in this model as a shift in the relative equilibrium levels or capacities for carbon storage in "ground vegetation" and "trees." When reduced below their equilibrium levels by disturbance, these pools recover according to nonlinear growth equations. Land-use change is modeled by allowing the parameters specifying equilibrium levels in these equations to vary as a result of forest clearing or reestablishment.

Emanuel et al. (1984) analyzed the tree-ring ^{13}C record assembled by Freyer (1979). The decreasing trend in this earlier record is slightly slower than that in the Freyer and Belacy (1983) time series. In contrast with the analysis by Peng et al. (1983), historical values of the removal of carbon from the tree compartments in the five-box terrestrial model were adjusted to yield the decrease in ^{13}C abundance. The fate of carbon removed from trees is specified in the model by a set of parameters set to nominal values. Carbon in the terrestrial boxes and assimilation by trees and ground vegetation change, and the net transfer of carbon to the atmosphere is calculated at each time step in the analysis as the difference in income and loss fluxes. The

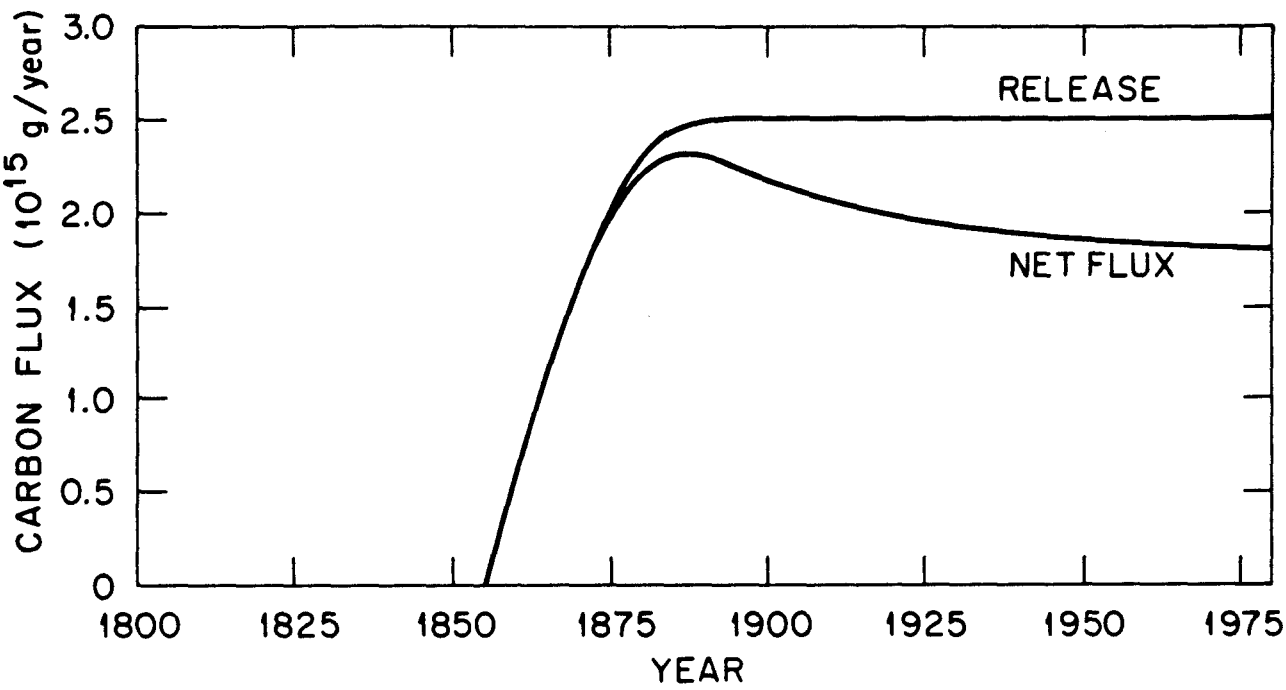


Figure 7.15. Total carbon in trees cleared globally and associated net exchange of carbon with the atmosphere that yield best agreement between model response and 13C tree-ring data. Positive values indicate flux to the atmosphere.

historical releases of carbon from the tree compartments and the corresponding net flux into the atmosphere are shown in Figure 7.15. The corresponding changes in terrestrial compartment levels are given in Figure 7.16.

There are many technical problems associated with using tree-ring ^{13}C data as a surrogate for direct atmospheric measurements. The isotope fractionation when carbon is incorporated into wood depends on many environmental and physiological variables and is not constant through time (Francey and Farquhar 1982). We further note that the interpretation of ^{13}C records is model dependent, and an internally consistent analysis of the sensitivity of results to model assumptions has not been completed (also see discussion in Chapters 3, 6, 10, and 11).

Perhaps the most promising possibility for resolving the discrepancy in modeling raised by substantial decreases in terrestrial carbon storage during the period of the Mauna Loa record is that analysis of data on land-use change may have led to inaccurate reconstructions of the historical terrestrial source of carbon to the atmosphere.

It is difficult to apply models that simulate changes in carbon storage, such as that proposed by Emanuel et al. (1984), to the interpretation of land-use data. Area is not explicitly treated. Chan et al. (1980) describe a model in which the areal extents of wooded and nonwooded ecosystem complexes and the densities of carbon storage in each are dynamic variables. The product of density and areal extent is terrestrial carbon storage. The utility of this model was demonstrated by application to scenarios of historical and future land-use change.

Houghton and co-workers developed a model that reconstructs yearly changes of carbon in terrestrial ecosystems (Moore et al. 1981; Houghton et al. 1983). Theoretically, the model can contain 10 geographic regions, with up to 14 types of ecosystems, including agroecosystems, within each region. In computing the net flux of carbon between terrestrial systems and the atmosphere, the model accounts for the changes over time that follow disturbance to vegetation and soil. Forest products are accumulated in delay pools. The rates at which each of these components regrows or decays as well as the rates of disturbance vary with geographic region and with ecosystem type. Time lags associated with growth and decay make an annual flux calculated in the model very different and more realistic than a flux calculated on the basis of 1 year's clearing and harvest.

Response curves (Figure 7.17) that describe changes in carbon stocks after disturbance are the most critica

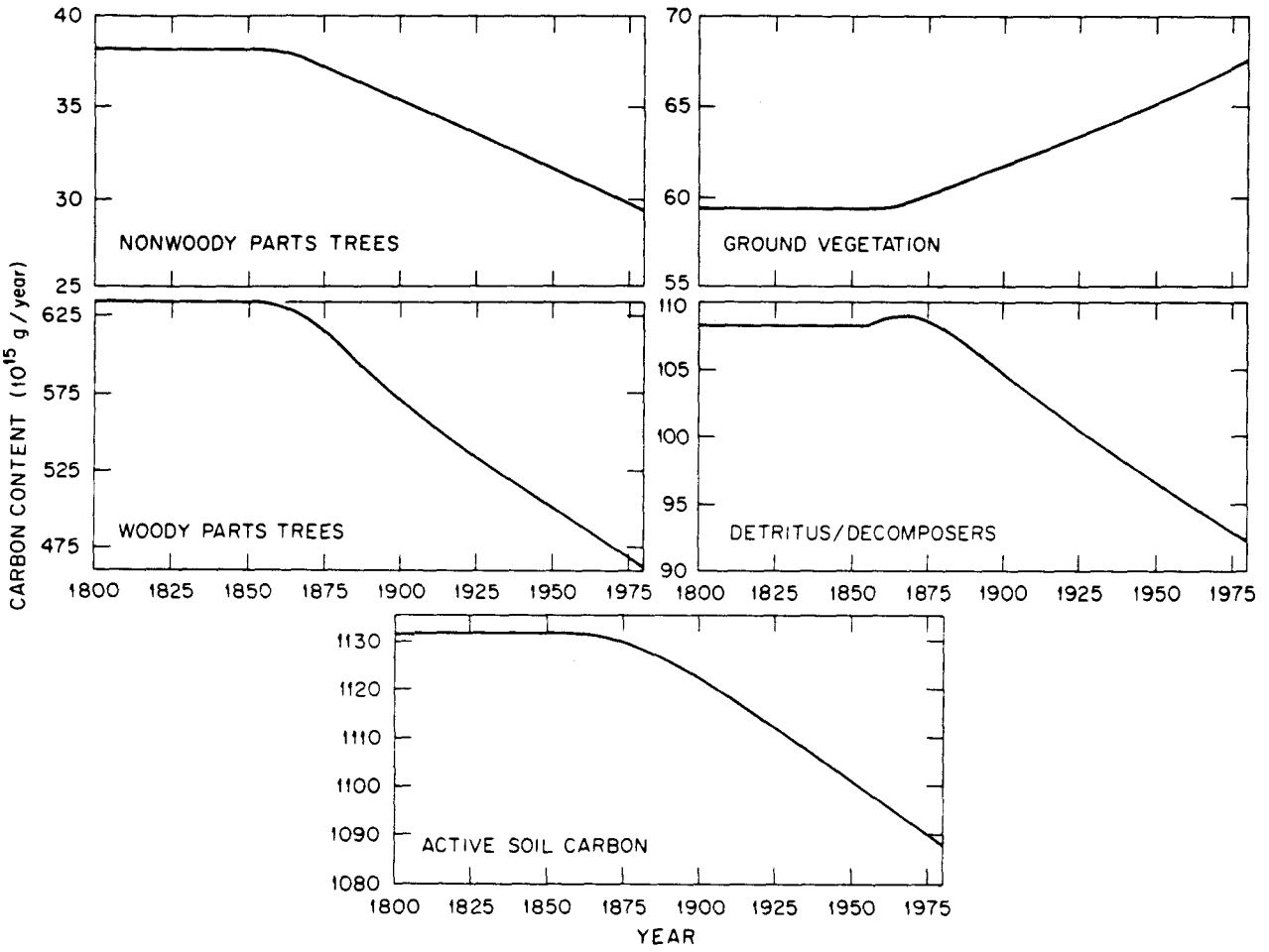


Figure 7.16. Changes in compartment contents in the terrestrial subsystem of the global carbon cycle model in response to forest-clearing releases adjusted for fit to tree-ring 13C data using five-box terrestrial model.

aspect of the model. In one form of these curves, harvest of a forest reduces forest biomass. Some of the carbon is transferred to domestic or commercial use where it decays at various rates, some is transferred to soil as slash and dead roots, and some remains as living material. Soil organic matter declines for several years after harvest until regrowth of vegetation results in net accumulation. Response curves for afforestation are similar. In another form, clearing of land for agriculture reduces carbon in both vegetation and soils, with no subsequent change unless the fields are abandoned.

Using these curves, the model keeps track of the area, age, and carbon content of each disturbed ecosystem. It also computes the oxidation of fuel wood and wood products. The computed net flux of carbon between the atmosphere and land for any year depends on the balance of the factors affecting the loss and accumulation of carbon on land.

A standard reconstruction with this model begins in 1700 A.D. Because of time lags, the computations are allowed to proceed for 160 years before considering the results.

Houghton et al. (1983) reported the annual net flux in 1980, the total net flux for the period of the atmospheric CO_2 record at Mauna Loa (1958-1980), and the total net flux for the 120-year period (1860-1980) for which estimates of fossil-fuel use exist. The estimate of the net release of carbon from vegetation and soils in 1980 ranged from 1.8×10^{15} to 4.7×10^{15} g (but see revisions to these estimates in Chapter 6). Over the 120-year period, the range of the net flux was 135×10^{15} to 228×10^{15} g. If vegetation only had been included in the Houghton et al. (1983) analyses, the net flux in 1980 would have been one-half as large. Thus, soils and wood products together contribute as much to the flux of carbon as does vegetation.

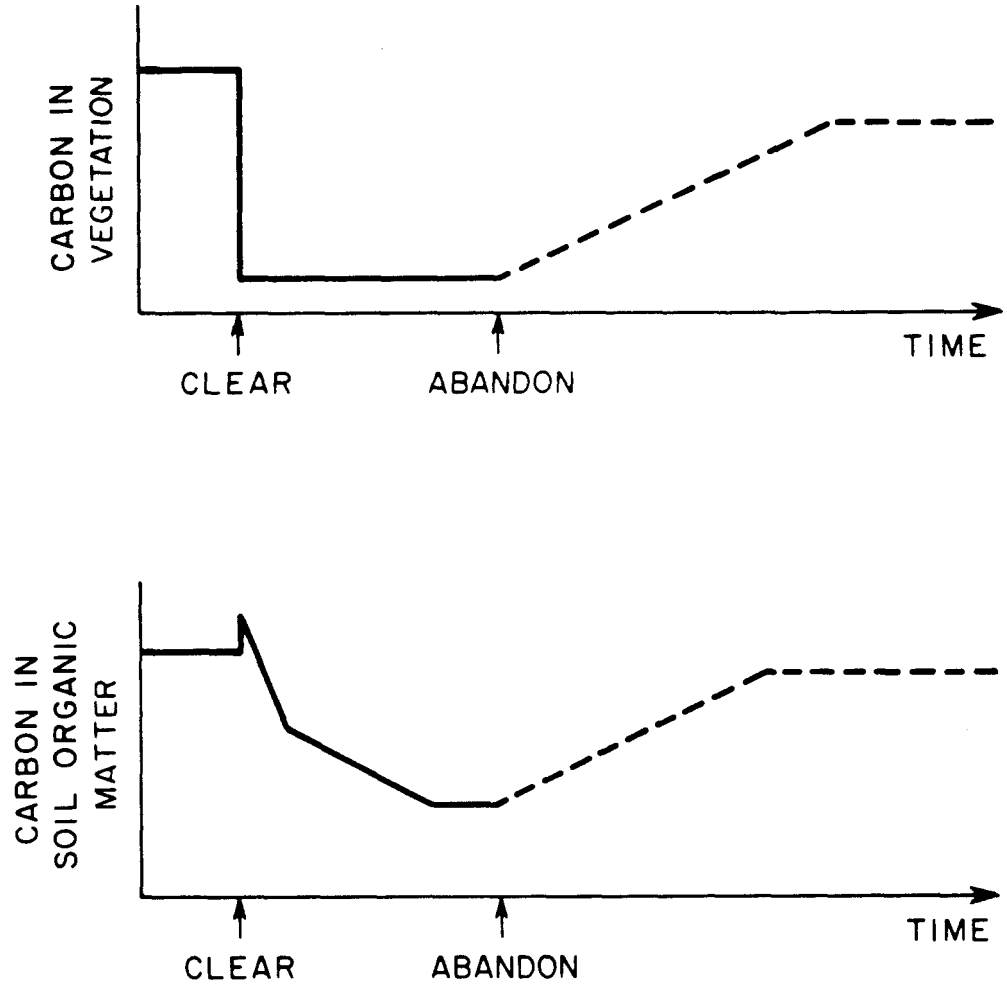


Figure 7.17. Response functions describing changes per unit area in the content of vegetation and soil following forest-clearing for agriculture. The dashed line shows changes that occur if agricultural land is abandoned.

The magnitudes of the immediate redistributions of carbon are determined by the densities of carbon storage in vegetation and soil assumed for predisturbance conditions and the assumed fractions of the affected pools that are transferred directly either to the atmosphere or to other pools such as litter or charcoal. Existing data are inadequate to evaluate these factors, and values in current use are by and large judgmental. Variability across different ecosystem types and for the same types in different regions necessitates large data sets if these assumptions are to be quantified.

A key uncertainty in studies thus far is the assumed magnitude of the decrease in soil carbon that accompanies the removal of vegetation. In the case of forest clearing, for instance, this secondary effect has been quantified in only a few field situations; however as indicated above, the general assumption of a 50% decrease in soil carbon is responsible for a large part of the estimated terrestrial release (also see Chapters 6, 10 and 11).

7.4 CO_2 VARIATIONS AND ATMOSPHERIC CIRCULATION

As discussed previously, the variations in atmospheric CO_2 levels contain information about interactions between the atmosphere and other carbon reservoirs.

Bolin and Keeling (1963), Junge and Czeplak (1968), Azevedo (1982), and Keeling and Heimann (in press) investigated latitudinal variations in atmospheric CO_2 with one-dimensional models that assume uniformity in the vertical and longitudinal directions. These one-dimensional models consist of a series of well-mixed boxes in the north-south direction. The exchange of CO_2 between the boxes is treated as turbulent diffusion.

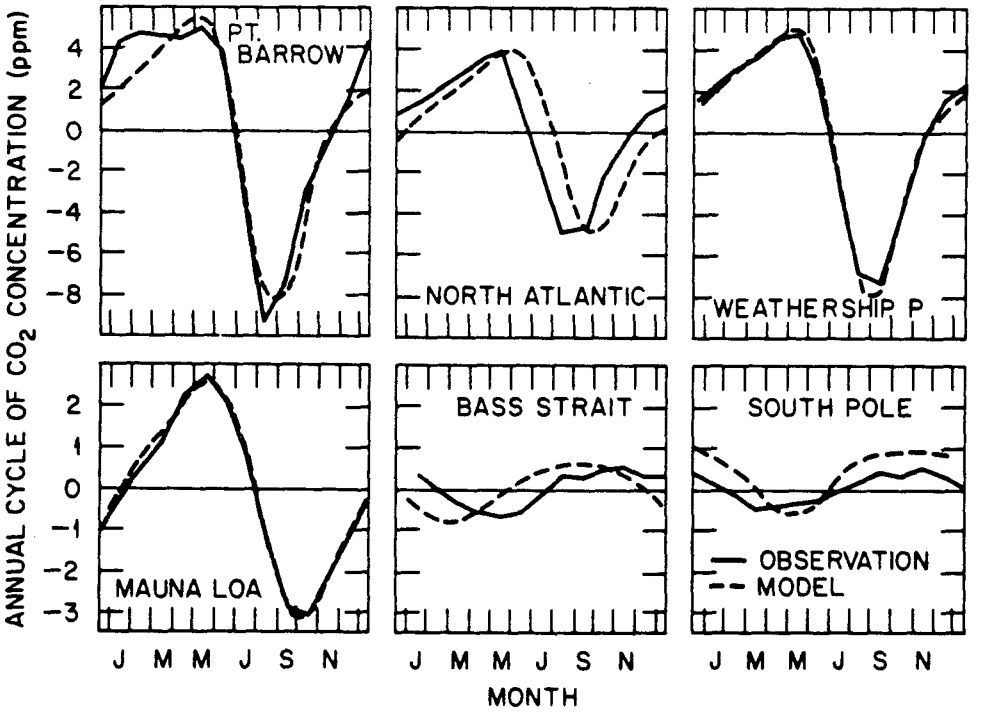
Bolin and Keeling (1963) used a diffusion coefficient that is constant throughout the troposphere; Junge and Czeplak (1968) and Keeling and Heimann (in press) varied the diffusion coefficient with latitude; and Azevedo (1982) introduced a latitudinally and seasonally dependent "restriction factor" to the otherwise constant diffusion coefficient to simulate the Intertropical Convergence Zone as a barrier to interhemispheric transport.

Machta (1972) and Pearman and Hyson (1980) used two-dimensional models to study the annual cycle of CO_2 in the atmosphere. Machta (1972) modeled atmospheric transport by constant diffusion coefficients in the meridional and vertical directions. Pearman and Hyson (1980) used observed monthly mean winds in the vertical and meridional directions and introduced a two-dimensional stress tensor to represent eddy transport.

Calibration of model parameters is a difficult task. Pearman et al. (1983; p. 3585) stated, "It is appreciated that while the model is tuned to cope with a monotonically increasing, nonseasonal tracer such as CCl_3F, it may not adequately distribute a seasonally varying compound with a more complicated surface exchange distribution."

The diffusion coefficients used in the one- and two-dimensional diffusion or advection-diffusion models cited above were calibrated against the distributions of atmospheric trace gases such as chlorofluorocarbons and ozone. Different model structures and tracers may require different model parameters to achieve satisfactory calibration. In turn, different models may require different source-sink functions to produce the same tracer distribution. For these reasons, although these one- and two-dimensional models have provided very instructive first analyses of the variations of CO_2 in the atmosphere, different conclusions have been reached about the magnitude of the sources and sinks of atmospheric CO_2.

Figure 7.18. Simulated and observed annual variations of CO_2 at six stations. Simulated values are from a model based on diffusion equations. (Source: Pearman and Hyson 1980.)



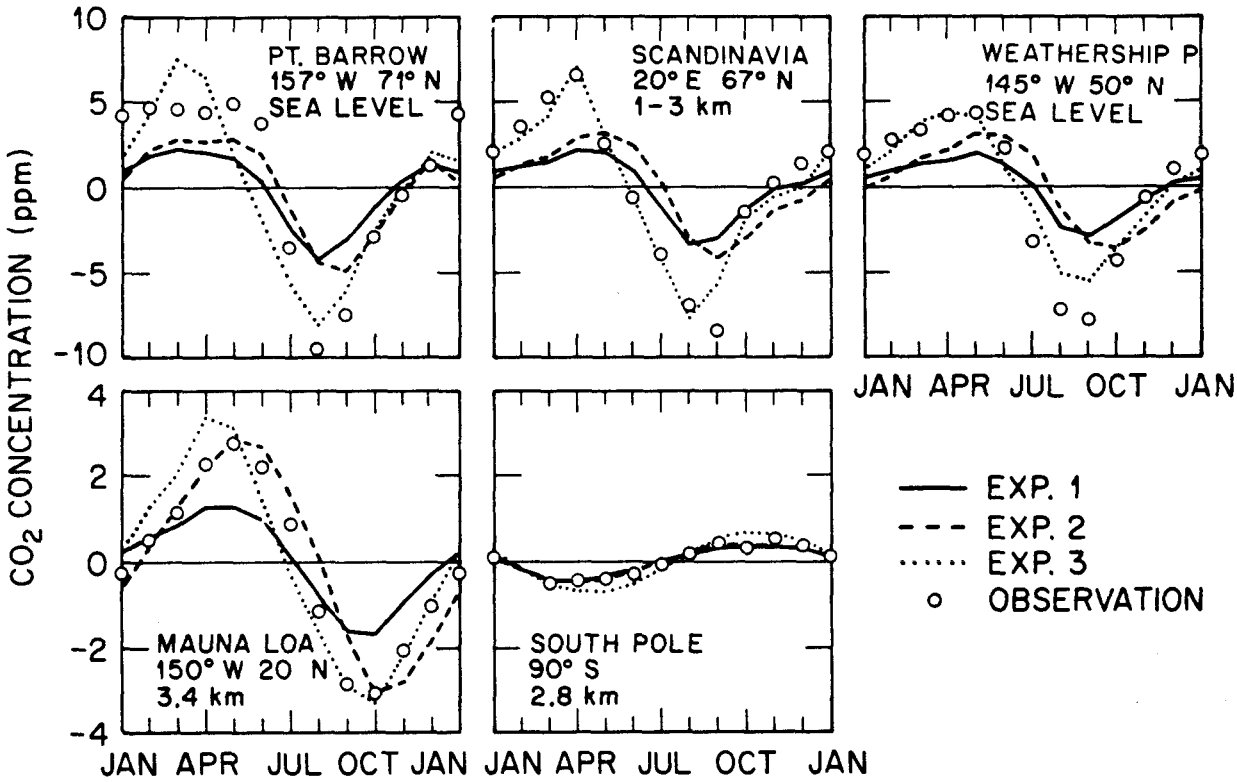


Figure 7.19. Annual cycles of CO_2 simulated under three alternate assumptions on terrestrial sources and sinks using a model that parameterizes wind fields generated by a general circulation model and CO_2 cycles observed at five monitoring stations. (Source: Fung et al. 1983.)

The first applications of atmospheric CO_2 models were to study the annual cycles in atmospheric CO_2 concentration. Junge and Czeplak (1968), Machta (1972), and Azevedo (1982) constructed biospheric exchange functions based on simple assumptions about seasonality. Bolin and Keeling (1963) and Pearman and Hyson (1980) derived sources and sinks necessary to simulate the observed CO_2 cycles. Reasonable agreement was found in all of these studies between simulated (longitudinally averaged) annual cycles and observations. The results simulated at six selected stations by Pearman and Hyson (1980) are shown in Figure 7.18.

Fung et al. (1983) performed three experiments with a three-dimensional transport model to investigate the seasonal exchange of CO_2 between the atmosphere and the terrestrial biota. These experiments specified seasonal biospheric exchange functions based on (1) Machta (1972), (2) Pearman and Hyson (1980), and (3) a global vegetation map (Matthews 1983) and Azevedo's (1982) definitions of seasonality in the uptake and releases of CO_2 by the biosphere. The simulated seasonal oscillations at model grid boxes corresponding to five observation sites are shown in Figure 7.19 together with the observed annual cycles.

Of the three biospheric exchange functions investigated, the one constructed from a global vegetation map and Azevedo's (1982) seasonal exchange curves produces CO_2 annual cycles most similar to observations. The exchange functions of Machta (1972) and of Pearman and Hyson (1980), which produced reasonable CO_2 oscillations in the respective two-dimensional models, underestimate the amplitudes of CO_2 oscillations in the three-dimensional model.

In all three experiments, substantial zonal gradients of CO_2 concentration exist between land and sea at midlatitudes in the Northern Hemisphere. Figure 7.20 shows the geographic distribution of the amplitude of the CO_2 oscillations at the surface, as simulated in Experiment 3 of Fung et al. (1983). The number and distribution of CO_2 monitoring stations neither permits the testing of Figure 7.20 nor checks to see if the CO_2 variations at the monitoring stations are representative of the zonal means. This uncertainty may be a reason

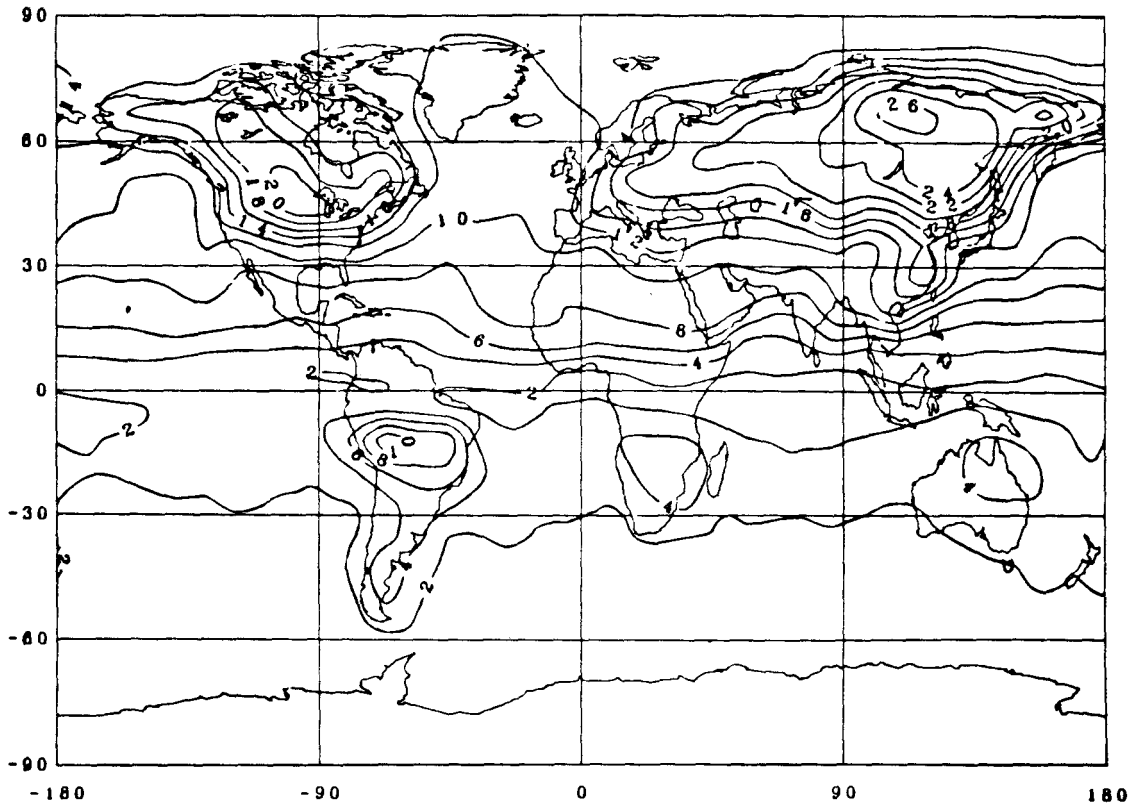


Figure 7.20. Geographic distribution of the amplitude of the annual CO_2 cycle as simulated by Fung et al. (1983).

for the difference between results based on two- and three-dimensional models.

The upper ocean exhibits very clear seasonal oscillations that may influence the annual oscillations of atmospheric CO_2. Sea surface temperatures (SST) have peak-to-peak amplitudes as large as 10°C at midlatitudes in the Northern Hemisphere. Indeed, observations in the Sargasso Sea (Takahashi et al. 1983) and in the north and south tropical Pacific (Weiss et al. 1982) reveal large seasonal CO_2 oscillations correlated with SST oscillations. The atmospheric response to SST-induced oscillations in oceanic CO_2 level has been investigated with the two-dimensional model of Pearman et al. (1983) as well as with the three-dimensional transport model of Fung et al. (1983). In these models, observed monthly SST and carbon chemistry are included to describe the seasonal oscillations of CO_2 in the surface waters. Both Pearman et al. (1983) and Fung (1986) found that; in the Northern Hemisphere, SST-induced oscillations in atmospheric CO_2 have amplitudes about 10% of those of the biosphere-induced oscillations and are out of phase with the biosphere-induced cycle. In the Southern Hemisphere, the SST-induced cycles have amplitudes comparable to the biosphere-induced cycles. Indeed, these oceanic contributions appear in $^{13}C/^{12}C$ data at a few land stations (e.g., Mook et al. 1983). Fung also showed that inclusion of the SST-induced cycles may introduce a phase shift to the total CO_2 response at the South Pole.

The models of Bolin and Keeling (1963), Azevedo (1982), Pearman et al. (1983), and Keeling and Heimann (in press) have also been used to study latitudinal gradients in annually averaged CO_2 concentrations. The Bolin and Keeling (1963) and the Keeling and Heimann (in press) models adjusted surface ocean fluxes to reproduce the annual mean gradients of atmospheric CO_2. The models of Azevedo (1982) and Pearman et al. (1983) were coupled to simple ocean models in which the latitudinal gradients of oceanic CO_2 were held fixed, close to the observed CO_2 gradients in the surface waters, but with magnitudes adjusted to simulate uptake of anthropogenic CO_2 by the oceans.

These simulations indicated that exchange with the upper ocean is responsible for the relative peak of

atmospheric CO_2 observed near the Equator. However, this net oceanic source at the Equator in 1960 is estimated to be 2.2×10^{15} g C/year by Pearman et al. (1983), 5.0×10^{15} g C/year by Keeling and Heimann (in press), and 10.9×10^{15} g C/year by Bolin and Keeling (1963). Again, the different efficiencies of atmospheric transport in their models could contribute to these discrepancies.

Azevedo (1982), Pearman et al. (1983), and Keeling and Heimann (in press) also analyzed sources and sinks associated with land-use change that would be consistent with observations of the changing latitudinal gradients of atmospheric CO_2. All studies concluded that, with the uncertainties in the atmospheric CO_2 data, the atmosphere could not have accommodated a tropical deforestation source of greater than 2×10^{15} g C/year since 1965, regardless of whether the carbon was taken up by the oceans of the world or by the temperate biosphere in the Northern Hemisphere.

To investigate the interannual variation of atmospheric CO_2, Fung (1983) used winds generated from the general circulation model (GCM) simulation of the El-Niño/Southern Oscillation (ENSO) to advect CO_2 in the three-dimensional transport model. To isolate the effect of circulation changes on the CO_2 distribution in the atmosphere, the vegetation and ocean sources/sinks were as in the "control" case, with winds from the climatological GCM simulation. The results show that changes in the simulated annual cycle of CO_2 are global. The magnitudes of the amplitude changes are largest at middle to high latitudes in the Northern Hemisphere, where the amplitudes of the CO_2 cycles are large, and decrease southward. The sign of the change does not show any smooth latitudinal or longitudinal progression, but is consistent with the anomalous circulation pattern. Amplitude decreases (increases) where the anomalous winds advect CO_2 away from (into) the region. In general, these amplitude changes represent 10 to 20% of the amplitudes simulated in the "control" case. Keeling (1983) found that the interannual variations in the amplitude of the CO_2 cycles at Mauna Loa and at Weathership P are 10 to 20% of the mean amplitude. However, this study demonstrates that variations at a single monitoring station may reflect changes in atmospheric redistribution processes in addition to fluctuations in CO_2 sources and sinks.

The results of the three-dimensional model show that midlatitudes in the Northern Hemisphere are a region with large seasonal changes in sources and sinks of CO_2, incomplete zonal mixing, and possibly the greatest sensitivity to changes in atmospheric circulation and sources and sinks. Although the modeling results confirm the dominant role of the terrestrial biota in the annual cycle of atmospheric CO_2, there is as yet no quantitative description, on a global basis, of the geographic and temporal distribution of the fluxes between the atmosphere and the biosphere. Furthermore, there is little information to evaluate the contributions of the biosphere to the interannual variations of the annual cycle. The pCO_2 in the surface waters is determined by SST as well as by total carbon concentration, total alkalinity, salinity, and the concentrations of minor constituents such as borates and silicates. Apart from SST, there are deplorably few observations of the seasonal behavior of CO_2 or of the factors that determine the pCO_2 in the oceans. The results of the modeling efforts so far suggest that an accurate description of the fluxes of CO_2 between the atmosphere and the terrestrial biosphere and between the atmosphere and the upper ocean requires additional observations in the atmosphere, in the terrestrial biota, and in the ocean.

By using a GCM as the basis for calculations, Fung and co-workers avoid the requirement of determining diffusion coefficients and parameterizing advection fluxes. However, parameter values in the diffusion model are estimated from independent data, and, as additional observations become available, the parameter set can be improved. Refinements to the GCMs, which express atmospheric physics, generally require either changes in the ways processes are included in the model or increased numerical resolution in implementing the basic equations.

Basic equations that express the physics of atmospheric circulation may be refined, but the greatest improvement in GCMs likely will come from changes in the numerical implementation of the models. Spatial resolution is a key problem as discussed above. Processes that are important at finer resolution may have been inadequately treated in simulations thus far. Because the results from models based on parameterized diffusion equations can be improved by refining parameter estimates as alternative data become available, there is potential for better simulations through changes quite different than those required in GCMs. However, diffusion equations are not a clear expression of the processes that dictate atmospheric circulation. Diagnosis and implementation of consistent changes in parameter values are restricted by this shortcoming.

7.5 SUMMARY AND RECOMMENDATIONS

7.5.1 Atmospheric Models

Initial attempts to model atmospheric CO_2 distribution, with the ocean and biosphere as seasonal sources and sinks, indicate that such analyses can lead to useful

quantitative constraints on CO_2 fluxes and changes in 'alues. Realization of this objective requires:

1. Continued improvement in the realism of three-dimensional global transport models;
2. Information about the geographic distribution and seasonal variations of the marine biota, of total alkalinity, and of total carbon concentrations in the surface waters;
3. Realistic models of seasonality in terrestrial ecosystems, including both live and dead carbon pools; and
4. Extended time lines of atmospheric CO_2 monitoring, with improved precision and improved definition of the uncertainties. Many potential applications depend on characterizing perturbations to CO_2 distributions and changes in time or in geographic distribution; hence, accurate calibration and intercalibration among different observing stations are important.

7.5.2 Refinements to Oceanic Models

As described in earlier sections, the one-dimensional box-diffusion model is not a realistic representation of the water mass structure and dynamics of the ocean. Realistic models of the oceans must account for actual geometry.

Early three-dimensional models of the deep oceans for chemical tracer studies include Arons and Stommel (1967), Kuo and Veronis (1970, 1973), Holland (1971), Fiadeiro and Craig (1978), and Fiadeiro (1982). Recently, Sarmiento (1983) reported a three-dimensional ocean circulation and mixing model for North Atlantic tritium distribution. However, these models are not designed on a global basis and are not directly applicable to analyzing the global carbon cycle. Maier-Reimer and co-workers (Universität Hamburg) are developing a three-dimensional global ocean circulation model for carbon cycle studies, but results are not yet available.

At present, a realistic, general circulation model of the entire world ocean does not exist. Although progress is expected with increased computing capabilities, global oceanography is insufficiently understood for many aspects of modeling. For example, the Antarctic Ocean plays a very important role in linking other oceans, but is poorly understood. The formation of North Atlantic deep water is very important with respect to global circulation, but we do not know how this water is formed nor where nor how fast.

As discussed previously, interim models will continue to be based on tracer data. The regional advection-diffusion model (Broecker et al. 1980) for e ventilation of the main thermocline and for the formation of new deep water, the isopycnal ventilation models for the North Atlantic thermocline by Peng and Broecker (1985), and the three-dimensional 84-box model of the Atlantic Ocean under development by Bolin's group are examples. Further, these models may be used as diagnostic tools for future GCMs of the ocean (Sarmiento 1983).

The distribution of bomb-produced ^3H and ^{14}C in the ocean has been widely used in determining the flow patterns of surface-ocean waters and the mixing rates in subsurface waters, as well as in calibrating a number of ocean models. The inventories of bomb ^{14}C in various major regions of the oceans are especially valuable in linking ocean circulation latitudinally. Results of the GEOSECS sampling program can be used to derive water column ^{14}C inventories. Because of the sparsity of data for the pre-bomb ^{14}C distribution in the upper ocean, the bomb ^{14}C inventories derived this way are unreliable. With the help of tritium profiles to indicate the penetration of bomb-produced nuclides, this drawback can be minimized. However, if the level of pre-bomb ^{14}C distribution can be established through coral head growth-ring ^{14}C measurements, more confidence can be placed in inventories and, hence, in the derivation of circulation patterns.

In addition to the effects of ocean circulation, the distribution of the bomb-^{14}C inventory also depends on gas exchange rates between the atmosphere and the ocean. To separate the imprints due to gas exchange processes from those due to lateral transport within the sea, it is necessary to obtain better estimates of the pattern of wind stress over the ocean and of the dependence of CO_2 invasion on wind stress.

The lateral transports of ^{14}C within the sea are very important to the distribution of water-column ^{14}C inventories, even on time scales as short as a decade. The ^{14}C anomalies will become even more apparent as the decades pass. Because these anomalies will set very important constraints on models of ocean mixing, it is imperative that we carefully document their evolution over the next several decades. Tritium is an equally important tracer, with its different input function and geochemistry. The water-column tritium inventory reflects more directly the effects of circulation and mixing processes.

As described in an earlier section, CO_2 differs from other tracers such as ^{85}Kr, chlorofluorocarbons, and $^{14}CO_2$ in a very important way. Its uptake characteristics lie geometrically between those of $^{14}CO_2$ and those of ^{85}Kr and the chlorofluorocarbons. To resolve the CO_2 problem, we must understand the distributions of these related tracers.

The use of multiple tracers can also improve the design of circulation patterns in models. For

example, ^3He is produced by the decay of tritium and is irreversibly lost to the atmosphere when exposed to the surface ocean. Therefore, its distribution within the sea is very different from that of other tracers and is the most model dependent. By using ^3He distribution, extra constraints can be placed on model construction.

The Antarctic Ocean plays a very important role in linking other major oceans. Study of tracer distributions in the Antarctic Ocean will certainly help our understanding of the origin of subsurface waters and their modes and rates of flow. The tracers of particular interest include preformed phosphate content, $^3He/^4He$ ratio, and bomb-^{14}C distributions in upper waters.

As two- or three-dimensional ocean models evolve to more spatially detailed formulations, calibration procedures to set a large number of parameters become quite complex. Major oceanographic programs such as GEOSECS and TTO have provided only partially required data to support general approaches. More data similar to those collected by the TTO program, but with coverage extending over the entire world oceans, are needed to make such multidimensional ocean modeling possible (also see Chapters 5 and 11).

7.5.3 Spatially Detailed Terrestrial Studies

The highest priority for further research on carbon cycling in terrestrial ecosystems is to improve the reliability of reconstructions of past and current changes in carbon storage in vegetation and soils that were caused by forest clearing and similar land-use changes.

Eventually, it may be possible to determine the historical role of terrestrial ecosystems in the carbon cycle by indirect means such as the analysis of ^{13}C data, but this inferential approach relies on models to account for the ocean's influence. Satisfactory resolution of the history of changes in the carbon cycle requires a direct accounting from independent data of the net exchange between the atmosphere and terrestrial ecosystems. The analysis of these data must be free from assumptions about the atmosphere-ocean system.

Refinements that will improve the reliability and applicability of reconstructions of terrestrial carbon storage fall into two areas: analysis of data that indicate the areal extent and intensity of land-use change and accounting of the changes in carbon storage in vegetation and soils that result from disturbance. Models are used in both of these areas.

Two kinds of direct data bear on the task of documenting the areal extents of past land-use change: (1) tabular data, such as timber yield, crop yield, or population increase, that can be translated into estimates of areas affected by land-use change over some period of time; and (2) maps of the distribution of ecosystem complexes or land cover at various points in time (maps of this sort can be derived from remote sensing or direct survey). Data in the first category correspond to differences between conditions on maps in the second category.

To date, world-scale estimates of terrestrial carbon fluxes have mainly used tabular data extracted from large institutional data bases such as the Food and Agriculture Organisation (FAO) yearbook data. Regional data of this type are likely more reliable and provide finer resolution. However, the character and quality of these data differ substantially from region to region. A modeling framework is needed to translate these regional data into consistent world summaries.

In a tabular format, the locations of land-use changes are usually not available beyond country or regional levels, and uniform assumptions as to impacts on carbon storage are invoked for all similarly tabulated areas. The geographic distribution of land use at arbitrary times in a reconstruction cannot be mapped for comparison with available land-cover maps or other geographically based data. The consistency of the history of land use implied by tabular data and maps of contemporary land cover or at intermediate stages of development has not been tested.

A major area for further research is the development of methods for relating data that indicate the character and magnitude of land-use change to maps of land cover at various points in development.

Models used to simulate the impacts of land-use change on carbon storage require substantial refinement. The greatest weakness in currently used approaches is the lack of treatment of mechanisms for compensating for decreases in storage. Potential CO_2 fertilization is perhaps the most important of these, although at present its global impact is extremely speculative (see Chapters 1 and 10). It is important to note that the likelihood of an enhancement in primary production by increasing CO_2 levels may be greater for disturbed ecosystems in which regrowth is occurring. Presumably such systems are below equilibrium standing stocks dictated by the complete set of limiting factors.

Linkages between the carbon cycle and other element cycles are very important in determining the response of terrestrial ecosystems to disturbances such as forest clearing. Nutrient limitations are in many instances the determining factors of the rate and degree of recovery. Multiple-element limitations on production are so important that terrestrial carbon models should now incorporate the dynamics of nitrogen, phosphorus, and other key nutrients.

Certainly, if climate change expected for a doubling of atmospheric CO_2 concentration occurs, the impact on terrestrial carbon cycling will be quite large (e.g., Emanuel et al. 1985). Climate-dependent models of terrestrial carbon cycling are needed. Two aspects of the

influence of climate must be considered. First the broad-scale distribution of ecosystem complexes depends on climate. To understand and analyze the changes that can occur because of this dependence requires that models express the dependence of particular fluxes on climatic variables. Current understanding requires that this be done at the ecosystem level. Second, models typical of each major ecosystem type must be constructed. The maps of land-cover or ecosystem distribution, derived from climate and other theoretical considerations for natural conditions and based on land-use data for historical and contemporary modifications by humans, are the basis for integrating ecosystem-level results to the global scale.

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8. THE NATURAL CARBON CYCLE

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8.1 HOW DOES (DID) THE NATURAL GLOBAL CARBON CYCLE OPERATE?

Industrialization and changes on the landscape have disturbed natural geochemical cycles of carbon and other elements significantly for several centuries (Clark 1982). Evaluating what happened requires understanding conditions that existed when there was less fuel burning, forest harvest, or farming (Hutchinson 1954, 1970; Craig 1957, 1958). Answering the question "How *did* the natural global carbon cycle operate?" also requires reaching back past millennia of human population expansion, past the glaciations of the last million years, and through the many millions of years that fossil carbon was being banked in the stratigraphic column (Sundquist 1985, 1986). To aid comparison, we summarize first the processes and pools, and then the rates of transfer or flux that constituted the carbon cycle before technological disturbances (Sections 8.2 and 8.3).

Dumas and Boussingault (1841) popularized an idea that may trace back to the discoverers of carbon, oxygen, and nitrogen; that is, conditions essential to life had to be maintained by Earth's natural cycling of elements (Figure 8.1, after Olson 1984). This natural balancing maintains a more or less steady level of several atmospheric constituents. "How steady" and "how maintained" are major questions for researchers. Sections 8.4 and 8.5 discuss some physical, chemical, and biological forces that tend to cause or to resist significant natural fluctuations in the carbon cycle over a very wide range of time scales (see also Sundquist 1986).

Finally, Section 8.6 summarizes points that need specific attention in order to understand subsystems of the cycle and their coupling. Uncertainties are identified, along with suggestions for research on one of life's most basic problems—maintaining a habitable planet. How do the lifeless geophysical processes, evolving ecosystems, and human cultures interact to modify Earth's carbon cycle and climate?

Are feedbacks from diverse living communities regulating planetary chemistry (Figure 8.1A) perhaps essential to maintain conditions suitable for higher forms of life? Interactions seem significant over most of the 3.8 billion years of the geologic record (Holland 1978, 1984). But can natural controls significantly help to limit the technological impact of humans over the next few centuries?

All these questions cannot be answered here. The key objectives of each section are to identify progress and unsolved problems in working toward answers. Integrating knowledge of the biosphere within that of the whole geosphere can hardly be made effective while specialists in biological and physical sciences work in ›lation. Our overall goal is to indicate why the answers about relatively short-term changes in the global carbon cycle (and climate) discussed in other chapters in this volume need to be placed in a longer term perspective relative to natural stability or change.

8.2 PRINCIPAL PROCESSES AND RESERVOIRS IN THE BIOGEOCHEMICAL CYCLE[1]

Direct measurement of the concentration of carbon dioxide (CO_2) in the atmospheric pool is possible only under current conditions. However, analyses of air trapped in glacier ice have helped refine estimates for atmospheric content in the several centuries before 1800 to within a range of 260 to 285 ppm, or about 550 $\times$ 10^{15} to 610 $\times$ 10^{15} g C (Chapter 3, henceforth, all chapter and section references are to this volume). The significant increase in CO_2 levels in recent centuries is attributed to the accelerated burning of fossil fuels plus oxidation of plant and humus carbon as a by-product of changes in land use (Chapters 4, 6, 7, and 10). The correct projection of future trends under any given set of assumptions about these sources of CO_2 requires improved understanding of exchanges with biotic and other pools.

Gross primary production (GPP) or photosynthesis (P_g) of plant organic matter from CO_2, water, and light energy is partly offset by total plant respiration (R_a) which recycles CO_2 to the air (Figure 8.1B) (also see Section 6.4). In most cases, subtraction of R_a leaves more or less half of the carbon assimilated as net primary production (NPP or P_n):

$$\text{photosynthetic fixation-respiration of autotrophic plants} = P_g - R_a = P_n \qquad (8.1)$$

Additional respiration by decomposers and consumers releases at least 90% of P_n to CO_2. Probably over 90% of this heterotrophic respiration is microbial, but where and how fast it takes place are influenced by many groups of animals. The precise role of animals remains one of the bigger uncertainties regarding the natural carbon cycle.

Natural fires caused mostly by lightning, or burning by early humans (Batchelder and Hirt 1966; Seiler and Crutzen 1980), hastened the return of most of the remaining ~10% of P_n in organic carbon to the atmosphere (Olson 1981b). During the last few centuries, more CO_2 has been released by respiration plus burning than has been reabsorbed via photosynthesis

[1]The preferred unit of mass will be the petagram (10^{15} g = 10^9 ton = 1 gigaton = Gt). Related metric units are teragram (10^{12} g) and metric ton (10^6 g). Dividing by the atomic mass number (12) for carbon, to give moles, is convenient when cycles of different elements are being compared.

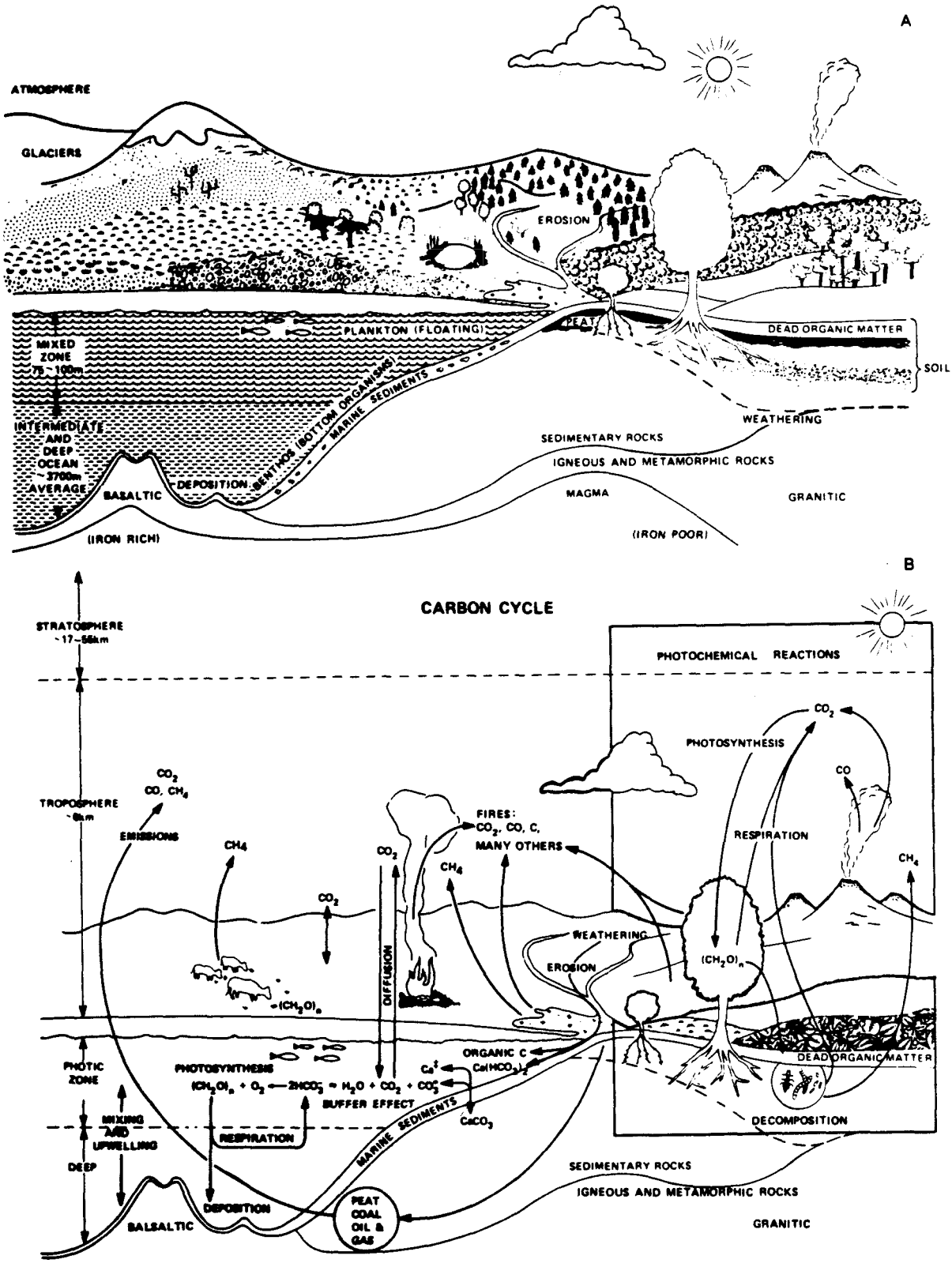

A
ATMOSPHERE
GLACIERS
EROSION
PLANKTON (FLOATING)
PEAT
DEAD ORGANIC MATTER
SOIL
MIXED ZONE 75 - 100m
INTERMEDIATE AND DEEP OCEAN ~3700m AVERAGE
BENTHOS (BOTTOM ORGANISMS)
MARINE SEDIMENTS
WEATHERING
SEDIMENTARY ROCKS
IGNEOUS AND METAMORPHIC ROCKS
BASALTIC
DEPOSITION
MAGMA
GRANITIC
(IRON RICH)
(IRON POOR)
B
CARBON CYCLE
STRATOSPHERE ~17-55km
TROPOSPHERE ~8km
PHOTOCHEMICAL REACTIONS
PHOTOSYNTHESIS
RESPIRATION
EMISSIONS
FIRES: CO2, CO, C, MANY OTHERS
DIFFUSION
WEATHERING
EROSION
DEAD ORGANIC MATTER
DECOMPOSITION
PHOTIC ZONE
DEEP
MIXING AND UPWELLING
ORGANIC C
BUFFER EFFECT
MARINE SEDIMENTS
SEDIMENTARY ROCKS
IGNEOUS AND METAMORPHIC ROCKS
BASALTIC
PEAT COAL OIL & GAS
GRANITIC

Figure 8.1. The main exchanges of carbon before major changes of natural landscapes by humans. (A) Diverse regions are conditioned by drying climates resulting from solar energy, atmospheric circulation, and redistribution of heat and water. (B) Research on atmospheric chemistry (upper right), plant physiology, and soil biology explains part of the balance of CO_2: by photosynthetic income of organic carbon and by its return to the atmosphere by respiration, decay, and fires; and also by photosynthetic fixation in the upper layers of lakes and oceans from dissolved carbon dioxide (mostly in the form of bicarbonate ions), and return by respiration or decay of detritus. (Source: Olson 1984.)

(see Chapter 6; Olson 1981a, 1982, 1985). An increase in temperature could hasten decay. A decrease in precipitation over large regions could increase the frequency and areal extent of burning, thereby further shortening the time required for recycling of organic carbon to CO_2 on a global scale. If this influence of climate on cycling became great enough to increase the atmospheric CO_2 concentration, to warm temperatures and increase decay rates, and to produce enough local drying to further enhance the probability of burning, the resulting positive feedback loop could destabilize the climatic system. Stabilizing effects or negative feedbacks (e.g., from cloud redistribution) may not be strong enough to counteract the climate change. Even if photosynthetic rates were increased directly by the so-called "CO_2 fertilization," despite conditions of limited nutrients and water the negative increment ΔC of organic carbon and other changes could add to the direct releases of CO_2 related to fuel burning. Understanding how this evolves will require that analyses of climate change and carbon cycle change be linked more closely than in the past (Olson 1980, Berner and Barron 1984).

Rate of change (i.e., net accumulation or loss) of organic carbon is given by the sum of income minus several loss rates:

$$\frac{\Delta C}{\Delta t} = P_n - R_h - F - E, \qquad (8.2)$$

where R_h is respiration of heterotrophic organisms (herbivores, predators, parasites, decomposers) to CO_2, CO (carbon monoxide), or fermentation to CH_4 (methane) and so forth; F is fire loss, mainly to CO_2, CO, and C; and E is other exports including erosion and harvest.

Cold, wet, acidic, or otherwise inhibitory chemical conditions can retard decay of dead organic matter (Swift et al. 1979; Bunnell and Tait 1974; Kononova 1966). Resultant pockets of peat and other organic debris become layers (as coal) or are dispersed (mostly in shales) through the stratigraphic column of consolidated sedimentary rocks. In zones where persistent deposition or earth movements sufficiently increase pressure and temperature, the sedimentary rocks are altered to metamorphic rocks, melted as magma or extruded as lava, and later recrystallized after cooling to become igneous rocks. Figure 8.1B suggests that volatile CO_2, CO, and CH_4 are released into the atmosphere directly (upper right and center) or via the ocean (lower left), either directly from organic residue decay or indirectly through the rock cycle of heated sediments (e.g., Burke et al. 1981, 1983; Senum and Gaffney 1985; Kroopnick 1971; Gwilliam 1982; Welhan and Craig 1984).

For modeling millennial or longer exchanges of carbon between the atmosphere, rocks (lithosphere), and organic matter (biosphere), the ocean (most of the hydrosphere) is often treated as one reservoir or pool. For treating changes over decades and centuries, the ocean is divided into regions (Chapters 5 and 7). The surface layer, heated by the sun and agitated by winds, hence relatively well mixed and on average near equilibrium with atmospheric CO_2, has often been modeled abstractly as a single box (Figure 8.1A, Table 8.1). The well-mixed surface zone is 75 to 100 m deep on average, but can be deeper locally, as in some polar or subpolar regions. Some analysts treat cool surface waters separately because they interchange more readily with deeper water (Wenk and Siegenthaler 1985; Ennever and McElroy 1985).

Under much of the warm ocean, the large amount of energy required to overturn the thermocline (warm layers over cold) makes the intermediate waters relatively stagnant (Table 8.1). Salinity gradients may further stabilize this thermal layering (Brass et al. 1982). The cold deep ocean below 1000 m is thus relatively isolated from surface waters and air. When surface waters are sufficiently cold and/or saline, however, the density gradient is weak; surface waters can sink (or mix) to various depths and then spread almost horizontally. A complex worldwide circulation pattern involves descending waters mostly in polar regions (e.g., Greenland and Antarctic seas and the North Pacific), slow gyration at all levels in major basins (Craig 1970, 1971; Broecker and Peng 1982, 1984), and upwelling of deep water elsewhere (Fine et al. 1981).

A small fraction of the water system, and of its carbon, is in freshwater lakes and streams (Kempe 1979, 1984). Their waters, sediments, and carbon (see Section 5.4.1) are mostly in transit toward the ocean, and in that way are important in regulating the linkage between landscapes and waters, which has commonly been ignored (Degens et al. 1983, in press). Some streams lead to dry basins having no drainage to the ocean, but they contribute materials to continental sediments.

Table 8.1
Estimated Oceanic Carbon Pools

Pool	Volume (10^9 km^3)	Inorganic Carbon (10^{15}g)		Organic Carbon (10^{15} g)	
Surface (0–75 m)	0.024[a]	580[a]	(560[b]–660[c])	50	(30[a]–64[d])[e]
Intermediate (75–1,000 m)	0.25[a]–0	37,542[c]	(6,000[a]–7,840[d])	675[f,g]	(125[h]–480[i])
Deep Water (1,000 m) (Polar, if separated)	1.1[a]		(25,000[d]–31,800[a]) (5,000[d]–0)		(440[h]–1,650[i])
Total	1.374	38,122		725[g]	(~600–2,200)

[a]Baes et al. 1976. Implies 6.2-year residence time if income and loss are near 93 × 10^{15} g/year as discussed in Section 5.3.3.
[b]Björkstrom 1979.
[c]Killough and Emanuel 1981.
[d]Munro and Olson 1982.
[e]If our intermediate estimate of 30 × 10^{15} g C/year input is used for the surface pool only, residence time of 1 to 2 years for living plus dead organic material follows from the top line. Alternative incomes of ±40% of these estimates would make correspondingly wider ranges of estimated turnover.
[f]For organic matter in the remaining layers together, the middle income estimate of 4 × 10^{15} g/year implies residence times of 142 to 535 years, assuming pools of 570 × 10^{15} and 2140 × 10^{15} g, respectively. For alternative inputs of 2 × 10^{15} or 6 × 10^{15} g/year, residence times become 50% shorter or longer, respectively.
[g]For world average of (5 mg/L) × (1350 × 10^9 km^3) below a 75-m mixed layer, a pool size of 675 × 10^{15} g implies a residence time estimate of 170 years for carbon in combined intermediate and deep water, near the lower range of note *f*. Taking the middle input estimate as in *f*, divided by 200 × 10^{15} g for intermediate waters only, implies a residence time of 50 years. If 1 × 10^{15} g C or less per year reaches the deep waters, and if these total near 440 × 10^{15} g of combined dissolved and particulate organic carbon, a residence time near 440 years would be implied for the world average ocean below 1000 m. Local examples noted in Section 8.2.3 suggest that this is an upper bound.
[h]Mopper and Degens 1979 (low range, wet oxidation).
[i]After Gordon and Sutcliffe 1973 (high range, dry combustion).

Within the ocean, organic biomass and especially productivity vary so widely that it is still difficult to have much confidence in weighted averages for the world: (1) Dense concentrations of plant life (seaweeds and other algae; e.g., see Smith 1981) and animals (coral reefs, Druffel 1985; Druffel and Suess 1983; Aharon 1983, 1985) mostly occupy small shallow areas (e.g., Smith and Kinsey 1976). (2) Other coastal and (3) upwelling waters that have sufficient nitrogen and phosphorus are responsible for much of the ocean biomass, production, and CO_2 release (Riley 1944; de Vooys 1979) within an area comprising perhaps 10% of the total. (4) Much wider areas of open ocean have long been known to have very low nutrients near the surface (Ryther 1969), and comparatively low production rates of organic matter (Deevey 1960; de Vooys 1979). Some of the midocean gyres, like the Sargasso Sea, may seem impressive because of accumulations of relatively inactive biomass. Yet the global total for marine organisms is still very low (1 × 10^{15} to 3 × 10^{15} g C, reviewed by Whittaker and Likens [1973] or Ajtay et al. [1979]) by comparison with that of land plants (Chapter 6). (See Chapter 5 and Baes and Killough [1986] for attempts to provide a generic treatment of marine production that formalizes the dependence of production on nutrients and [implicitly] on light or temperature, without much complication by regional geography.) Modern spatial patterns of floating algal (plankton) production have been mapped by Koblentz-Mischke et al. (1970).

The global pool of biospheric carbon is the sum of products (i.e., the area occupied by each water or landscape type multiplied by its carbon per unit area). In recent millennia, the areal extent of tundra has changed little; however, the area of desert may have increased as a result of subtropical climatic shifts and overgrazing. Grassland has increased because of increased livestock grazing in many parts of the world, but the prairies, pampas, and moister steppes, in turn, have provided much of the cropland now devoted to maize, wheat, and other cereals. Most other wild communities have been reduced in area by cropland, human settlements, and other modifications.

The amount of carbon per unit area has also diminished in most areas (see Section 6.5.2) because of human disturbances. Therefore, we cannot multiply representative modern carbon densities by the areas of modern (or even original) cover types and obtain valid estimates of primordial plant carbon; the product would be too low. Nor would it be valid to use data on mature stands and multiply by forest area; the estimate

would be too high for modern and even for virgin landscape complexes; the latter had significant areas still undergoing recovery from fires or other disturbances at any point of time. New data and reasoning have narrowed the bounds around the estimated values for plant, animal, microbe, litter, and soil carbon pools discussed below and in Chapter 6.

Using different working assumptions, several authors estimated plant carbon levels prior to agricultural disturbance. Beginning with the biomass calculations of Bazilevich et al. (1971), Olson (1974) estimated carbon levels in preagricultural vegetation to be 1070 × 10^{15} g. An estimate of 900 × 10^{15} g C now seems a more realistic upper bound, allowing for disturbances even before human populations expanded in recent millennia. A slightly lower estimate follows from Matthews (1983), using the potential vegetation types of the United Nations Educational, Scientific, and Cultural Organisation (UNESCO 1973). Olson's (1985) estimate of plant biomass shown in Figure 8.2A for about 6000 years B.P. (mid-Holocene time after glacial melting, but before much agricultural development) was 796 × 10^{15} g C for all live plants. A tentative breakdown gives ~72 × 10^{15} g each for nonwoody plants and for photosynthetic plus other short-lived parts of woody plants. That leaves 652 × 10^{15} g for woody parts of plants, and 756 × 10^{15} g if we include the upper bound suggested above.

A percent or less of the terrestrial carbon pool (~4 × 10^{15} g is suggested here) is represented by heterotrophic organisms: animals and nongreen plants. Animals may be a very small fraction of this mass, but they could influence 10 to 30% of the CO_2 recycling, including microbial respiration (Seastedt and Crossley 1984).

The humus within soil profiles was estimated by several authors (Chapter 6; Schlesinger 1977, 1986). Most estimates are in the range of 1400(±200) × 10^{15} g C (Buringh 1984; Post et al. 1982; Schlesinger 1986; Zinke et al. 1984). Schlesinger's (1977) estimate of 1450 × 10^{15} g included a litter estimate of 60 × 10^{15} g. Bohn's (1976, 1982) estimates are higher than others, principally because he emphasized organic soils (especially peats and mucks, called histosols).[2] The latter soils include part of the "subfossil" carbon that eventually becomes incorporated into geologic formations.

An estimate of the level of carbon currently stored in soil humus, based on recently expanded data (Zinke et al. 1984; Chapter 6) is 1580 × 10^{15} g C. This value is higher than the estimate by Zinke et al. (1984) based on the allocation of soil profiles according to climatic life zones because wetland soils were not included in the latter. This new estimate by Olson includes the surface meter of wetland soils, allocated by areas of ecosystem complexes according to Olson and Watts (1982) and Olson et al. (1983).

The allowance for increased oxidation of humus by anthropogenic activities such as cultivation (Richards et al. 1983) may increase estimates of the predisturbance carbon pool to 1700 × 10^{15} to 2000 × 10^{15} g C. The highest value may be questionable because Holocene soil profiles in some regions, especially those on recently deglaciated till or on glacial outwash or lake sediments, may not have had time to approach an assumed equilibrium condition (maximum carbon storage) before human activities modified humus (Jenny 1980).

These estimates do not include old soil horizons ("paleosols," reviewed by Yaalon 1972), where glaciers and especially windblown dust (loess) have buried soils. These and other subfossil pools, and peats at depths >1 m, are not generally included in the active short-term carbon cycle. The few published estimates of the size of this pool are very rough approximations, perhaps useful for modeling exercises (Chan et al. 1980).

Besides humus, the soil carbon pool also includes calcium carbonate in so-called calcic or caliche horizons: loose nodules and crusts and consolidated layers called calcretes. Two independent estimates are available: Yaalon (1981) estimated 1700 × 10^{15} g C, based on the carbonate content of soils in semiarid regions (13% of land surface). Schlesinger's (1982) lower values (800 × 10^{15} g C) started with a lower value of soil carbonate, but this was extrapolated to a larger total area (24% of land surface for mass and 31.5% for flux). Later estimates by both authors overlap near 1000 × 10^{15} to 1200 × 10^{15} g C (Chapter 6; Schlesinger 1986).

The relationship between rapid ecological and slow geologic cycling of carbon is illustrated in Figure 8.2. The "subfossil" category includes several transitions to the lithosphere with inputs from both land ecosystems and shallow aquatic systems (discussed further below). Over millennia, sediments from both systems can become interbedded with relatively minor shifts in river, lake, and ocean levels.

Sedimentary rocks (Figure 8.2B, Table 8.2) were first produced by the weathering and erosion of primeval igneous crust, whose minerals (such as silicates with calcium and magnesium) were attacked by water and acid gases released during the consolidation and chemical differentiation of the Earth (Rubey 1951; Holland 1978, 1984). This attack produced a mass of sediments described by the reaction:

[2] Bazilevich (1974) separately estimated recent peat as having a carbon equivalent of only 88 × 10^{15} g. Reiners et al. (1973) allocated only 700 × 10^{15} g to dead organic carbon on land, but assigned 330 × 10^{15} g in freshwater; we include many bogs, which are associated with small lakes and swamps. There also may be a significant amount additional carbon in buried peat which has a still slower turnover than that included here (see Sections 8.3.2, 8.3.3 and 8.5.3).

Table 8.2
Approximate Total Mass of Major "Spheres," and Estimated Carbon in Reservoirs of the Large-Scale Global Cycle of Carbon

	Total Mass (10^{24} g)	Approximate Carbon (10^{18} g)
Atmosphere		0.66
Ocean	1.4	
Live and Dead Organic Matter		0.72
Fresh Sediments		~0.5
Crust	25	
Total Sedimentary Rocks[a]	2.5	
Limestone (20%)	0.5	Carbonate 62,200
Sandstone (15%)	0.375	
Shale (65%)	1.6	Organic 10,300
Mass of Continents Currently Above Sea Level	0.3	
(Average through time perhaps 20% less)		

[a]See Figure 8.2. Estimates of present sediment mass range from 1.5×10^{24} g (Poldervaart 1955) to 2.5×10^{24} g (Garrels et al. 1975), and estimates of the total carbon in sediments range from 12.5×10^{21} to 72×10^{21} g. If the present crustal abundance of carbon of 200 ppm in igneous rock holds for the entire mass of the primordial Earth (6×10^{27} g), there are 1.2×10^{24} g of carbon in the Earth.

Igneous Rocks + Water and Acid Volatiles
= Sediments + Salty Oceans. (8.3)

Interactions of the cycles of silicate and sulfate in the rocks with the cycles of carbonate and organic carbon are described in Sections 8.3.1 and 8.5, which look farther into the geologic past to provide a context for returning to the fairly modern history of carbon cycling discussed in Sections 8.3.2 through 8.4.2. With a wide range of recycling rates (or residence times) and frequencies of driving forces in mind, we can later return to the longer term history and processes of the carbon cycle which are discussed in Sections 8.4.3 through 8.5.2.

This review of previous inventories of global carbon highlights one major problem—the transitional categories we refer to as subfossil carbon. Several kinds of slowly cycling organic matter, soil carbonate, and dissolved CO_2 (and bicarbonate) in groundwater, and CH_4 combined with water (as hydrate or clathrate in polar sediments) where temperature is low or pressure is high (Bell 1982), have turnover coefficients or residence times intermediate between those for very slow rock pools (mentioned above) and the comparatively rapid exchanges of ecological systems and oceans that have attracted most attention in the assessment of responses to anthropogenic CO_2. For example, the National Academy of Sciences (NRC 1983, Figure 17) tentatively identified a pool near $10{,}000 \times 10^{15}$ g for methane hydrate, with a range of uncertainty spanning more than an order of magnitude. Revelle (1983) and others recognize that releases to the atmosphere are likely to be a small fraction of 1×10^{15} g/year, even if releases are hastened by climatic warming at the rates anticipated early in the 21st century. However, the size of this and several other subfossil pools may be large enough that persistently warmed climate (e.g., Arctic melting as Flohn [1982] proposed) might contribute a larger fraction of the 5×10^{15} g C/year now released by burning of fossil fuel to CO_2. Improved perspective on the natural and modified releases therefore requires closer evaluation of the fluxes. This is outlined in the next section (8.3).

8.3 FLUX RATES, TURNOVER FRACTIONS, AND TIME FRAMES

This section relates the preceding overview of Earth's pools or aggregations of carbon to the inferred rates of exchange or flux between each major "sphere" of the carbon cycle. How promptly can each readjust when sudden imbalances occur due to natural or human-induced changes? The answer will depend partly on the fractional loss rates or turnover coefficients of the materials (loss rate/pool size). The pool size might approximate a steady state, but a temporary shift of the equilibrium level could occur over either historic, prehistoric, or geologic time when the income or loss rate changes.

A wide range of flux rates is indicated in Figure 8.2. Rapid exchanges and changes occur within the atmosphere and land ecosystems and upper oceans

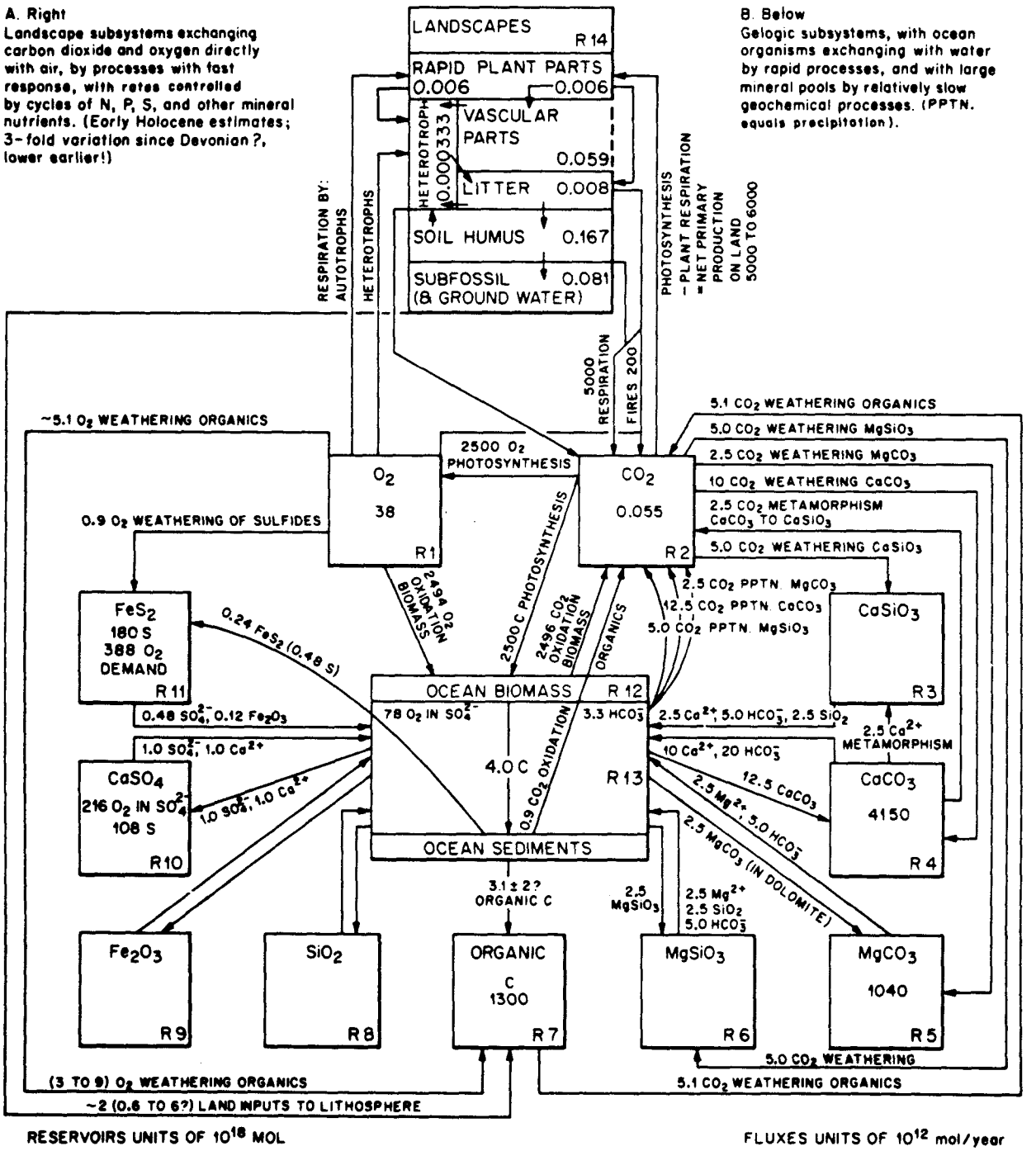


Figure 8.2. Links between cycling of carbon and other elements. The figure contrasts rapid ecological processes (A) with slow geologic processes (B) averaged over the latter portion of post-Cambrian time. (A) Landscapes in early Holocene (recent) times had roughly equal quantities of rapidly cycling (mostly photosynthetic) tissue for nonwoody and woody parts of plants. The latter probably were of negligible ass before the late Silurian Period about 400 million years ago. Estimated values, uncertainties, and some sources of variation are discussed the text and summarized in Table 8.3 and Figure 8.4. (B) Synthesis of information on oceanic and lithospheric cycles. Some details and estimated values are elaborated in Figure 8.3 for carbon and sulfur and in Figure 8.7 for calcium, magnesium, and oxygen.

slower changes are tempered by the deeper oceans (Chapters 5 and 7) and by the subfossil pools just mentioned. The slowest processes evolve along with the rock cycle. The discussion that follows will first briefly expand our view to the beginning of the long-term geologic time scale, and then narrow again to focus on a more contemporary picture.

8.3.1 Erosion, Deposition, and Geologic Residence Times

The cumulative contribution of CO_2 to air from volcanoes, fumaroles, hot springs, natural venting, and other recycling of carbon has been significant over the history of the Earth, but rates have varied considerably; change has been offset in part by the consumption of CO_2 in the weathering of silicate minerals (Figure 8.2B). The release of volatile compounds has been part of the general development of core, mantle, and crust. Interpretations of the history of the Earth currently favor a $\geq$80% degassing of the volatile compounds (H_2O, CO_2, CO, CH_4, COS, CS_2, F_2, HCl_2SO_2), perhaps from 4.5 to 4.3 billion years B.P. (Cloud 1978; Holland 1978, 1984; Sundquist 1986). In the following discussion it is assumed that the degassing of acid volatiles and water was nearly complete after a few hundred million years of Earth's history.

Table 8.2 lists some values for major pools of the cycle of erosion, deposition, and uplift termed the very long-term geologic (sometimes called exogenic) cycle. Most of the rock masses have been relatively constant over the last 3 billion years. Whether the fraction of organic carbon in this geologic cycle has changed, through altering of the biosphere's organic carbon cycle (Holland 1984), is an issue related to biological evolution.

The current rate of erosion of the land areas (and therefore the rate of deposition of sediments) is estimated at about 24×10^{15} g/year, divided between 20×10^{15} g/year of suspended load and 4×10^{15} g/year of dissolved materials. Various interpretations yield a prehuman average erosion rate of about one-third the present rate, with total transport to the oceans of 8×10^{15} to 10×10^{15} g/year, here taken as 3×10^{15} g/year dissolved load plus 5×10^{15} to 7×10^{15} g/year suspended load. The dissolved load of stream transport per year has apparently been less changeable with time than the suspended load that responds quickly to soil disturbance and reservoir trapping of particles. How human and ice age (Pleistocene) disturbances modified the longer term cycling of carbon and minerals (as by erosion) is another of the poorly understood issues of multielement cycling that requires additional research.

Several major conclusions can be derived by considering the amounts of various elements carried to the oceans each year, along with the amounts in the oceans and in sedimentary rocks. The land mass above sea level today is $\sim 0.3 \times 10^{24}$ g or only a fraction of the total sediment mass, and, at the average rate of erosion in the prehuman world, the continents should have survived only 30 million years. But the continents have a history of more than 3 billion years. The simple conclusion is that over geologic time whatever is removed from the continents is restored to them. What is removed from the top must be replaced beneath.

The rate of addition today of most elements to the ocean by streams would double their concentrations in the ocean in a few million years. On the time scale of the Earth's history, the ocean is not a storage place, it is a medium of transfer for elements from one sedimentary reservoir to another or for return to the original reservoir. The water of the ocean circulates from bottom to top each 500 years or so; its water is essentially renewed by streams each 40,000 years. Over more than 3 billion years of Earth's history the oceans have been stirred millions of times and the water could have been renewed 75,000 times, a truly active system if photographed over 3 billion years and replayed as a 2-hour cinema!

Not only is the ocean not a place of prolonged element storage over geologic time, but the total mass of existing sediments is small compared with the total mass that has been deposited. The existing mass of sediments from all ages is $\sim 2.5 \times 10^{24}$ g (Table 8.2). The average annual flux was about 10×10^{15} g/year; therefore, the total existing mass would be deposited in 250 million ($2.5 \times 10^{24}/10 \times 10^{15}$) years. Detailed analysis of such relationships indicates that the total sedimentary mass deposited over the last 3 billion years was 6 or 8 times greater than the mass now remaining. Therefore, if the total sedimentary mass at any given time is constant, the concept emerges that sediments must cycle—that there is a roughly constant mass eroded, deposited, uplifted, and eroded again (see Section 8.5.1). As igneous and metamorphic rocks get involved, the assumption that the initial sedimentary mass has simply been recycled is not adequate; but chemical models based on this simplifying assumption seem to produce results consistent with the geological record, whose interpretation is limited by the uncertainty of our knowledge.

Looking at Earth's history in its entirety, the main action has thus concerned the transfer of elements from one rock reservoir to another, with the ocean and the atmosphere remaining relatively constant. These relationships can be used in mathematical modeling of this large, long-term (exogenic) system. Transfers between sedimentary reservoirs are coupled; if the changes with time in a given reservoir are known, the changes in many others are fixed. The changes in various reservoirs required by the observed increases

in calcium sulfate ($CaSO_4$) deposition in gypsum, for example, have been expressed by the reaction:

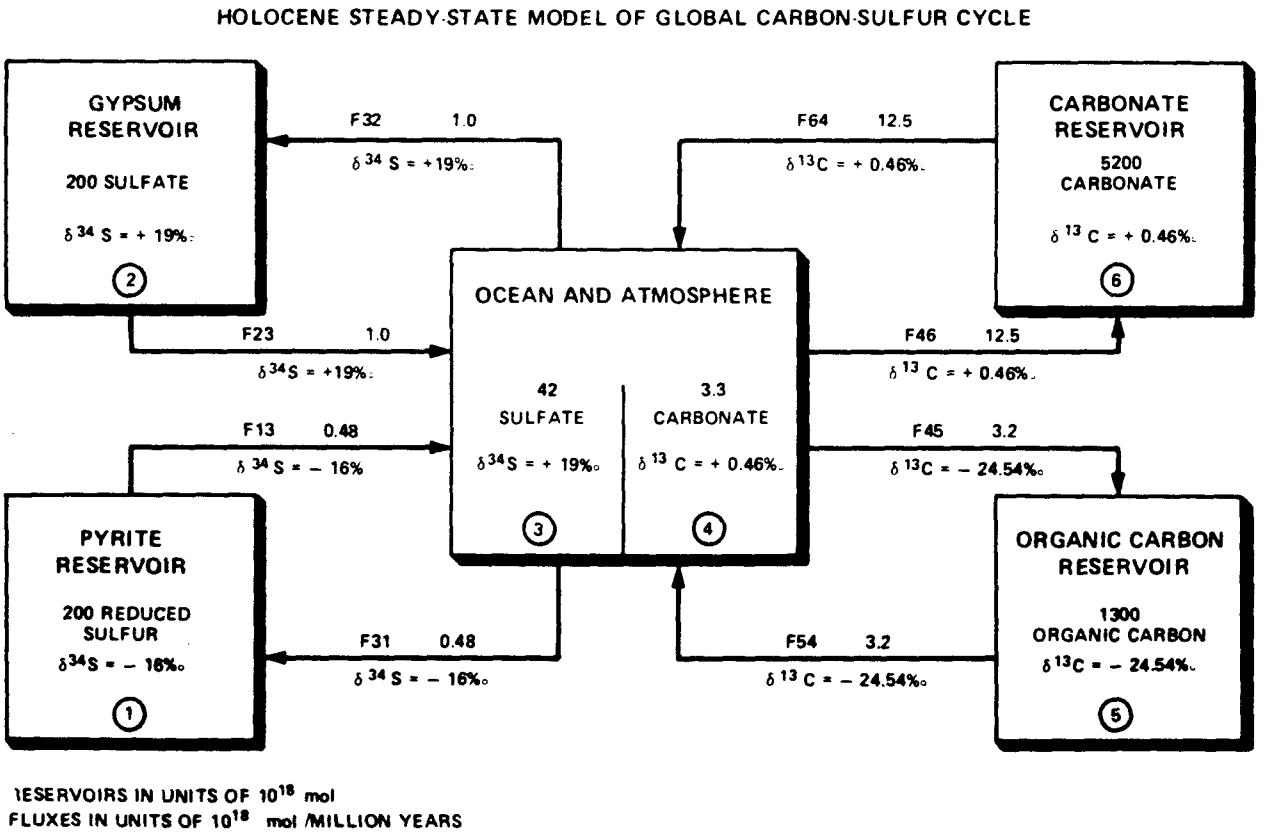
$$15H_2O + 4FeS_2 + 8CaCO_3 + 7MgCO_3 + 7SiO_2$$
$$15CH_2O + 2Fe_2O_3 + 8CaSO_4 + 7MgSiO_3 . \quad (8.4)$$

This seems to define uniquely the coefficients of the fluxes from one set of reservoirs to the other (see also Section 8.5). Because of this extensive coupling, it is possible to isolate the coupled major oxidation-reduction pairs: sulfide/sulfate and organic carbon/carbonate. Fortunately, the isotopic composition of sedimentary sulfate in marine evaporite deposits is close to that of the oceanic sulfate from which it precipitated, and the $\delta^{13}C$ of carbonate deposits is also close to the isotopic composition of the bicarbonate of the ocean from which it was derived. A mathematical model of a coupled organic carbon/inorganic carbon-sedimentary sulfate/sedimentary sulfide explores what equations like (8.3) and (8.4) mean.

Figure 8.3 is such a model coupling the present sulfur and carbon systems. A computer program drives the systems backwards in time (Garrels and Lerman, in press), forcing it to conform to the observed values of $\delta^{13}C$ of carbonate rocks or those of the $\delta^{34}S$ of evaporite deposits (gypsum, $CaSO_4 \times 2H_2O$, and anhydrite; $CaSO_4$). The basic assumption is that the fluxes related to the weathering of the sedimentary reservoirs are proportional to the sizes of the reservoirs and that the atmosphere-ocean system is constant in terms of the total concentrations of carbon and sulfur, but variable with respect to carbon and sulfur isotopic compositions of the reservoirs. Either isotope, $\delta^{13}C$ of calcium carbonate or the $\delta^{34}S$ of deposited gypsum, can be used to drive the system. The validity of this concept can be tested by the correspondence of the prediction for the other isotope ratio with the measured values. Currently, agreement is within the limits of knowledge of the actual values for the predicted isotope.

In terms of rates of erosion and deposition of carbonates and sulfates in the distant past, the model's validity corroborates the concept of uniformitarianism: the present is the key to past processes. Model estimates of the past fluxes of organic carbon from the ocean range only from 70% of modern values to 170% (but see Sections 8.4 to 8.6 for more estimates).

Figure 8.3. Linking of carbon and sulfur cycles (after Garrels and Lerman, in press), focusing part of Figure 8.2 on multimillion year averaged "modern" conditions. (See Sections 8.3.1, 8.4.3, and 8.5.2 for variations over time, including varied proportions of land and ocean sediments.)



8.3.2 Atmospheric/Biospheric Rates and Residence Times

Rapid changes in modern landscapes contrast even more sharply than do ocean trends with the geologic rates discussed in the previous section. Most animals and microbes metabolize and die so rapidly that heterotrophs have very short residence times. If a heterotroph pool size of $\sim 4 \times 10^{15}$ g C (see Section 8.1) were supported by unburned NPP (approaching 60×10^{15} g C), the mean residence time would be about 1/15 of a year, or 24 d. Any such value would be a weighted average of metabolisms, ranging from much faster for bacteria and small invertebrates to slower for large or long-lived vertebrates. The rates of consumption by humans (Deevey 1960) and roles of animals in regulation of productivity and cycling (Dyer 1979; Dyer et al. 1982; McNaughton 1976) deserve more attention.

Several estimates of land NPP (P_n) have been near 56×10^{15} g C/year (Deevey 1960; Olson 1970): slightly lower according to Lieth (1975) and Box (1975) or higher (Olson et al. 1978, 1983). An NPP value equal to 60×10^{15} g C is assumed here. Rates of GPP (P_g in Equation [8.1]) are probably higher than some current estimates, because the roles of grazing or harvesting tend to be minimized and because respiration studies have been feasible for relatively few field-scale ecosystems. If GPP averages approximately twice NPP (see Chapter 6), and if Holocene communities were as productive as Olson (1985) has estimated, annual income and loss could have averaged 110×10^{15} to 145×10^{15} g C/year. If annual losses (to respiration plus transfers to other pools) and income both averaged 130×10^{15} g C/year, the estimate for rapidly exchanging parts of plants (Figure 8.2) (144×10^{15} g C/year for early to mid-Holocene ($\sim$6000 years B.P.) gives a turnover rate of 0.9 per year and a residence time near 1.1 years. If essentially all of the NPP goes into the layer of litter and associated debris in the form of cast-off plant parts and animal ejecta and then is completely recycled to the atmosphere, the nominal 100×10^{15}-g litter pool in Table 8.3 implies turnover fractions averaging 0.6 per year and residence times near 1.67 years. A large margin of uncertainty exists concerning real variations of these numbers among different landscapes and climates, and in the estimated proportions of woody and other biomass components. However, much better estimates may be very difficult to obtain (Olson et al. 1978, 1983).

Seasonal variations in atmospheric pools of CO_2 related to biospheric exchange are clear (Chapters 3, 6, and 7; Section 8.4.1). Most of the annual oscillation seems related to exchanges with the rapidly growing but short-lived parts and their residues as just mentioned. Values for the mid-Holocene may not be different from modern ones, because historic removal of trees over wide areas has left room for more herbaceous ground cover and, in some places, rather small, young woody regrowth. It is still not clear how unequally the rapidly exchanging pool has changed with fluctuations in the occurrence of fire throughout history; this subject requires further analysis (Olson 1981b).

Allocations of photosynthetic fractions to woody plant or humus carbon on land are relatively poorly known. If the earlier pool estimates of 652×10^{15} to 756×10^{15} g C in woody parts of plants are divided by 15×10^{15} and 25×10^{15} g C/year, respectively, the ratios give residence times ranging from 43 to 30 years. The reciprocals imply that 0.023 to 0.033 (2.3 to 3.3%) per year of carbon fixed in woody parts is annually recycled to atmospheric CO_2. Actual values could have been quite different, and probably varied appreciably with climate changes. Because most losses of plant and litter carbon go to the atmosphere without contributing to humus formation, it seems unlikely that more than 5×10^{15} to 10×10^{15} g C/year could be contributed to or lost from the soil pool. If these two values are taken to correspond to the earlier estimates of 1700×10^{15} and 2000×10^{15} g C for the mid-Holocene soil pool, respectively, the ratios represent turnover times of 0.3 to 0.5% per year and residence times of 340 to 200 years. These estimates are quite uncertain because humus fractions are quite variable in composition and the turnover times vary widely among chemical fractions. Research on such phenomena needs to be expanded to define realistic estimates for actual humus and plant residue mixtures.

Before the advent of expanded agriculture and forest harvesting, the other slowly exchanging pool of organic matter other than woody plant parts was humus. Of the estimates for modern humus to 1 m depth in soil, at least 200×10^{15} g is a preliminary estimate for alluvium, shallow muck, and peat. An even larger fraction or total may be more appropriate. Perhaps 1×10^{15} g C of this total is considered en route to the subfossil deposits, and a few percent of that flux could persist each year to become incorporated into consolidated geologic sediments (see Sections 8.3.3 and 8.5.3).

The long-term mean annual flux to or from paleosol humus besides peat also may be only a small fraction of 1×10^{15} g C/year. Nevertheless, a cumulative release in some millennia may be significant, because this part of the subfossil pool would likely alternate between net positive and negative balances. For example, unresolved questions are whether glacial burial temporarily removes much humus and peat from the influence of atmospheric oxidation, and whether relatively sudden releases of CO_2 occur soon after thawing and exposure to oxygen after glacial retreat. Such a contribution to atmospheric changes may be

Table 8.3
Simplified Global Carbon Inventory and Budget Estimates for Recent (Early Holocene) Time

Major Carbon Pool	Annual Balance Sheet: Inferred Pool of C (10^{15} g/year)	Income (10^{15} g/year)	Loss (10^{15} g/year)	Difference[a] (10^{15} g/year)	Turnover = Loss ÷ Pool (year^{-1})	Residence Time = Pool ÷ Loss (year)	Comment (See Section 8.2 and other sources below)
Atmosphere	660	220	220	0(?)	0.333	3	Chapter 3
Ocean Exchange		90	90				
Landscapes		130	130(?)				
Ocean							Table 8.2
Inorganic	38,000	90	90	0(?)	~0.00237	422	Chapter 5
Organic	675	40	40		0.06	17	
	to 1,000	to 30	to 30	0(?)	0.03	to 33	Section 8.2
"Recent" Sedimentary Organic	~400	0.048	0.0048	0	0.00012	8,333	
Landscapes							Figure 8.2
Live							Chapter 6;
Heterotroph	4(?)	60	60	0	15.0	0.067	Sections 8.2, 8.4.4
Rapid plant	144	130	130	0	0.9	1.1	Section 8.3.3
Slow plant	652	15	15	−0.1(?)	0.023	43	Section 8.5.3
	to 756	to 25	to 25	to 0.02	to 0.033	to 30	
Litter	100	60	60	0	0.6	1.67	
Soil	1,700	5.2	5	−0.2	0.003	340	Chapter 6
Organic (1 m)	to 2,000	to 10.1	to 10	to 0.1	to 0.005	to 200	Section 8.2, 8.3
Subfossil	970	1	0.976	0.024	0.001	1,000	Section 8.2
Organic	to 3,000	to 0.33	to 0.3	to 0.03	to 0.0001	to 10,000	Section 8.5.3
Soil carbonate	1,100(±100)	0.03	0.029	~0.001(?)	0.0000263	37,900	Section 8.2
Rocks							Figure 8.2b
Carbonates	62,200,000	0.18	0.18	0	0.28×10^{-8}	345×10^{6}	
Organic	10,300,000	0.037	0.037	0	0.35×10^{-8}	278×10^{6}	
	to 15,600,000[b]	to 0.06	to 0.06		to 0.38×10^{-8}	to 260×10^{6}	

Note: Prior to major expansion of human populations, agriculture, and forestry, but after most postglacial forest accumulation had become well advanced.
[a]Negative estimates for "difference" indicate loss of carbon from the pool on the same line, which could be an input of CO_2 to the atmosphere or from landscapes to water bodies or to subfossil and rock pools. Positive estimates reflect plausible but unproven precivilization storage rates, neglecting such exports as erosion. Net values are more likely to be between these bounds than not, but could have changed significantly over prehistoric intervals.
[b]From Figure 8.2.

related to glacial-interglacial alternations in CO_2 and climate (see Section 3.5.2).

Figure 8.1 reminds us that CO and traces of CH_4 would have reached the atmosphere naturally from volcanoes and from anaerobic fermentation from animals and microorganisms (Zimmerman et al. 1982) and sediments (Burke et al. 1981, 1983). Provisional estimates of ambient CH_4 levels are near 1.65 ppm, compared with 343 ppm CO_2. Fluxes of CH_4 are at least $300(\pm 140) \times 10^{15}$ g C/year (Senum and Gaffney 1985). Recent upward trends are potentially influenced by human activities (e.g., rice farming and livestock herding). Atmospheric fluxes and sources of these carbon-containing gases have been reviewed by Crutzen et al. (1979), Ehhalt et al. (1983), Khalil and Rasmussen (1984), Logan et al. (1981), as well as Senum and Gaffney (1985). Present estimates of the contributions from CO and CH_4 to atmospheric CO_2 are quite small (NRC 1983). However, CO and CH_4 are radiatively active materials that may also have their own climatic significance (see Chapter 10).

8.3.3 Span of Rates and Residence Times

To recapitulate, our inquiry in Section 8.2 moved from the rapid exchange processes that can be measured on land to the oceanic exchanges damped by large water volumes and associated physical inertia. Still slower processes are associated with the poorly known subfossil pools, and finally with consolidated rocks. As we shifted attention from quantifying the pool inventories to rates of exchange between pools in Section 8.3, the lithosphere was covered first, analogous to the set of low-range gears of the Earth's cycling machinery which provides steadiness to the overall system of coupled cycles. The previous section returned to rapid biospheric processes and provided estimates (supplementary to those in Chapter 6) needed to reconstruct biospheric turnover rates in the pre-agricultural world (see also Olson 1974, 1975, 1985). The high-speed gears in the carbon cycle are principally associated with the atmosphere and biosphere of both the land and the ocean surface waters. Chapters 5 and 7 provide many more details on the contemporary oceanic fluxes. With the exception of rapid processes occurring at the air-sea interface and within the mixed layer, these fluxes and those involving subfossil pools may be likened to the intermediate-speed gears in the global machinery.

Comparing the turnover coefficients (like gear speeds) of all major pools, Table 8.3 draws together estimates for rather broadly averaged pools along with corresponding estimates of income and loss. The turnover ratios (loss/pool) are also given along with their reciprocals, the residence times implied for averaged pools in an idealized world. Such numbers and the lines on the left side of Figure 8.4B provide perspective for additional remarks about pools of intermediate residence time.

The bottom lines of Table 8.3 summarize the pools and inferred income and loss rates (from Figures 8.2 and 8.4) for carbonates and organic carbon in the lithosphere; the two annual turnover fractions correspond to residence times of 345 million and 260 million to 278 million years, respectively. The last value, derived from Figure 8.2B, uses the sum of land (subfossil) and marine inputs. The input of 0.037×10^{15} g C/year is based solely on the marine input, combined with a more conservative estimate of organic matter dispersed in sediments.

Gardner et al. (1983) estimate particulate organic matter falling through the water column to have residence times ranging from 0.3 to 3 years. The organic ocean totals in Table 8.3 imply mean residence times of 17 to 33 years. Much longer times seem possible for the bulk of organic (including resistant dissolved) material in the large mass of cold water in the deep ocean (see footnotes f and g in Table 8.1). At deep Pacific Ocean sites, Emerson et al. (in press) infer residence times of 15 to 150 years in surface sediments, allowing adequate time for bioturbation to leave a small fraction of carbon in sediment zones where decay is even slower. The inorganic carbon pools (from Table 8.1) in water average to an overall residence time of over 410 years, again dominated by the nearly stagnant deep layers rather than surface or even intermediate waters (see Chapters 5 and 7 for more details on this subject).

There was probably a small positive difference between income and loss, reflecting buildup of forest mass during slow ecological successions (e.g., as newly deposited parent material for soil after the last ice age). However, the question marks and footnote in Table 8.3 beside the loss rate from landscapes to the atmosphere is there to remind us of the need for better background for understanding our contemporary dilemma about the role of the biosphere in the carbon cycle (Chapters 6 and 10). Temporal factors of primary succession and accumulated nutrients (especially nitrogen, phosphorus, and calcium in soils and litter) limit the carrying capacity for life and require attention in new research.

One could suggest that only those natural pools in Table 8.3 with residence times less than 100 years have the potential to affect atmospheric CO_2 significantly over the next century. One could argue that ensuring that our carbon cycle models contain accurate simulations for the marine and terrestrial biota ought to be sufficient. However, this is an oversimplified view. The other pools with longer residence times (e.g., landscape subfossil humus, and rock outcrop or soil carbonate carbon, in Table 8.3) become very large before their small annual fractional losses approach

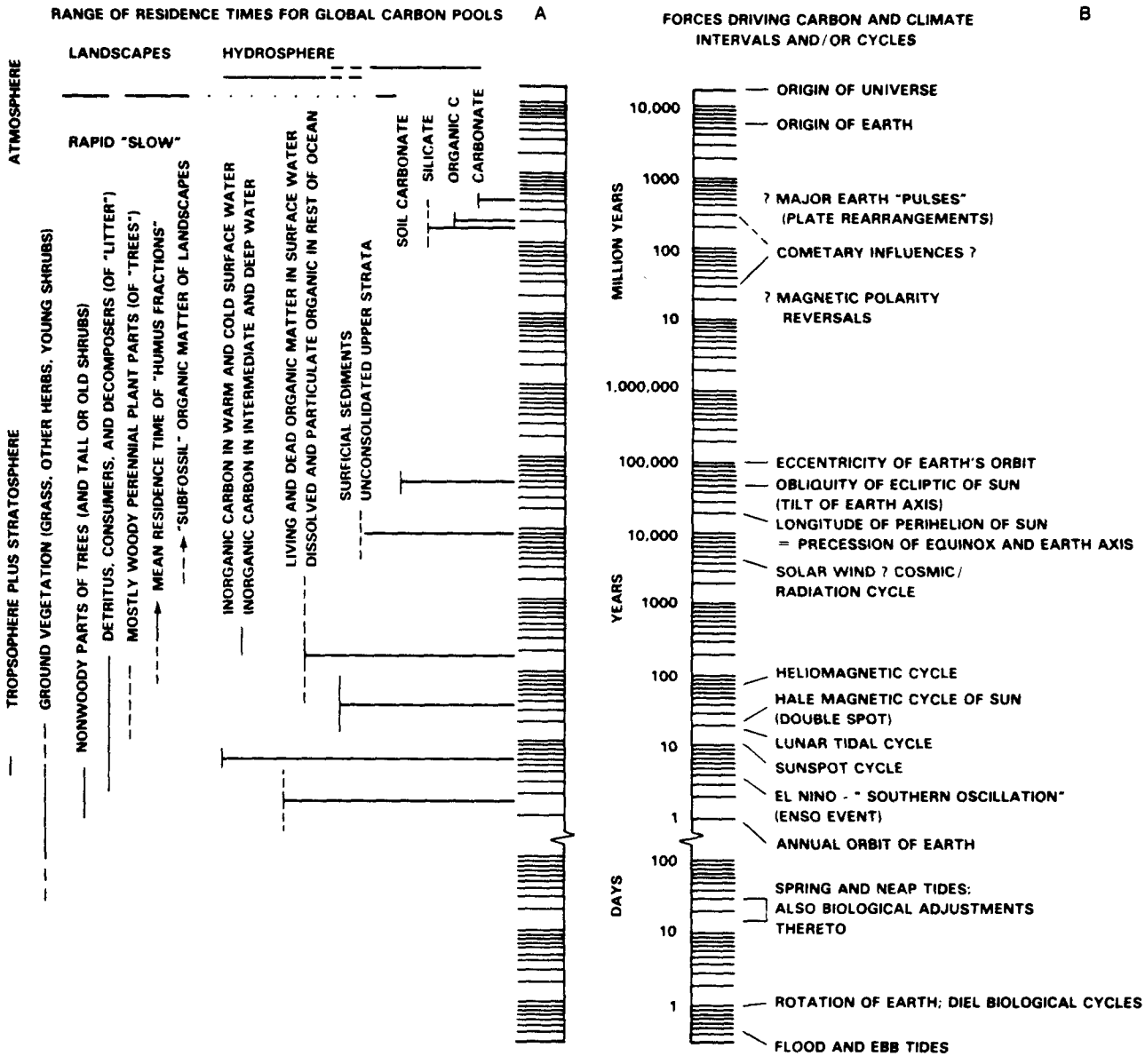


Figure 8.4. Time scales for carbon cycling: (A) Mean turnover of major pools; (B) Periodic oscillation and irregular fluctuations that may drive carbon cycling or climate changes.

balance with income rates. A plausible increase in one rate and corresponding decrease in another could produce a net atmospheric flux that could be significant as one contribution to the imbalance in the contemporary system.

Further, we must reserve judgment about how closely the early Holocene system may have been balanced, until we have explored the potential effects of natural climatic fluctuations occurring over a wide span of time scales (Section 8.4; Olson 1985; Solomon and Tharp 1985) and from variations in the geochemical cycles of other elements coupled to carbon (Section 8.5). The discussion which follows in those sections should clarify whether the gears (turnover rates) of the global carbon system components have meshed smoothly and predictably with those of other geophysical and

geochemical phenomena. For simplicity, we will continue to treat the global biosphere in an artificially averaged way, but we must remember the contrasting relationships between the different kinds of organisms and landscapes of Figure 8.1A and their interactions with climate and an atmosphere of changing CO_2.

8.4 FLUCTUATIONS OF DIVERSE FREQUENCIES

Some changes in atmospheric chemistry, physiological ecology, and marine biology can be measured on very short time scales (seconds, hours, days; Figure 8.4). At the other extreme, geologic changes occur on scales measured in millions of years. Processes involved in the global carbon cycle are thus spread across this enormous range of time scales. In this section we will survey these time scales to provide perspective on the natural oscillations and external perturbations that affect the steady states and turnover rates presented in the previous sections.

The shifts in income and loss rates seen on the shortest time scales may lead to trends that persist over a few months, years, or decades (e.g., weather and human disturbance) (Section 8.3.2). Postglacial and glacial-interglacial climatic fluctuations are related to alternations of landscape and marine carbon ranging from centuries to hundreds of thousands of years (Imbrie and Imbrie 1980; Shackleton and Pisias 1985). There have been times when atmospheric CO_2 level changes have been relatively fast, at least 10% of the recent rate of anthropogenic change, because of a rebalancing of the Earth's storage after glacial and ocean shifts (see Section 3.2). One of the current challenges is to understand the prehistoric shifts between balanced and imbalanced conditions of carbon storage on time scales of a century or less (Section 8.4.2).

Geochemical trends over longer periods in Earth's history have also had some interesting fluctuations around those long-term average conditions presented so simply in Figures 8.2 and 8.3. Alternations of periods of major peat and coal deposition with those of marine organic and carbonate deposition have had an impact on the storage and return of organic matter to CO_2. Section 8.4.3 stretches our geologic perspective back to Earth's Archean history.

The rapid increments from short-term contributions add algebraically to progressively slower net changes in total storage of carbon, as we move from small carbon pools to large ones. Figure 8.4A summarizes essential aspects relating pools to fluxes from Sections 8.2 and 8.3, and from Chapters 5 through 7, for carbon pools with turnover coefficients ranging from fast to slow and with residence times ranging from short to long. The right side of Figure 8.4B then outlines some of the forces and alternations that drive these processes on different time scales. Reading from the bottom upward, tides and other changes of the day, month, and year influence finely tuned physical cycles and predictable, if less precise, biological responses. Alternations over years and decades may be less perfectly periodic; yet they may be of special importance for climate or element cycling when peaks or troughs of slightly different frequency become superimposed—creating exceptional departures ("beats") from the average condition. Many controversies about significant geophysical or astrophysical alternations affecting our planet are beyond the scope of this chapter (see Sections 3.2 and 3.3; Sundquist 1985, 1986).

8.4.1 Alternations on Daily, Seasonal, and Multiyear Scales

8.4.1.1 *Patterns of Day/Night Metabolism: C_3, and C_4, CAM Plants*

Respiration that is measured in the dark continues during the day, even though masked by the net intake of CO_2 in photosynthesis. Hence, the apparent photosynthetic rate (Monsi and Saeki 1953; Olson 1964, 1975) is less than the GPP rate by the amount of daytime respiration.

In plants with carbon fixation products having three carbon atoms per molecule (C_3 plants; Foyer 1984) there may also be significant additional respiration stimulated by light (Zelitch 1971; Olson 1964). Metabolism involving 4-carbon organic molecules, which does not incur this "photorespiration," occurs in many warm-season grasses and a few other plants (C_4 plants; Hatch 1976). Brown (1978) also suggests that C_4 plants, compared with C_3 plants, have evolved to gain higher carbon fixation and biomass production rates per unit of nitrogen. His inference presumed a smaller investment of nitrogen in the photosynthetic carboxylation enzymes in the C_4 metabolism, so that the evolutionary advantage would have been particularly favorable in competition for establishment and reproduction on sites of low nitrogen content, or where there was low availability of the nitrogen that was present. Yet C_3 plants tend to have the highest quantum yield in converting solar energy to food and an amazing variety of adaptations to severe environments (Ehleringer 1978). All things considered, however, the special adaptations of C_4 metabolism to heat drought are of particular interest in a world of regionally warmer and drier climate, perhaps expanded by increasing greenhouse gases.

For a similar reason, there is particular interest in a third mode of metabolism, the so-called CAM pathway of photosynthesis, an adaptation for fixing CO_2 through

pores (stomates) that open during the night when there is less risk of excessive moisture loss (Kluge and Ting 1978). The temporary storage products (organic acids) are then converted by solar energy during daylight hours into sugars and other molecules. These details and other physiological reasons suggest that internal adaptations to climate change may occur within existing ecosystem complexes. Thus the effects of past and future changes in climate are difficult to estimate, and are not simply reducible to a gross empirical correlation with annual temperature or moisture averages, or with global averages of carbon inventory per unit area. The more significant challenge, discussed further in Section 8.6.1, will be to interpret, separately, the control of the rate terms of Equations (8.1), (8.2), (8.3), and (8.4) and the elaborations of the latter (in Equations [8.5] through [8.7]), as a function of changeable climate.

8.4.1.2 *Seasons: Cold or Dry Alternating with Seasons of Maximum Growth*

Photosynthesis occurring in terrestrial vegetation during Northern Hemisphere summers contributes most of the annual drawdown of the atmospheric reservoir of CO_2. Also, melting of sea ice and heating of surface waters during warm seasons allow greater ocean uptake from the atmosphere to replenish CO_2 used in marine photosynthesis (Baes 1982). Respiration of organisms, like the combustion of fossil and nonfossil fuels, occurs at varying rates throughout the year, perhaps over much longer periods than the main season of photosynthesis. Subtracting these loss rates from the photosynthetic fixation of CO_2 for each day and month gives the seasonal net exchange of CO_2 between the atmosphere and biosphere for each climatic zone (discussed further in Chapters 3 and 6).

Measurements of atmospheric CO_2 at stations from 73°N to 90°S show that seasonal amplitudes are greatest at high northern latitudes and diminish southward, being very small south of the equator (Chapter 3). The observed pattern reflects the fact that the bulk of the seasonal vegetation is found in the Northern Hemisphere and a much larger fraction of the area in the Southern Hemisphere is covered by oceans. Whether seasonal amplitudes are increasing as a result of human disturbance, and what that would mean about the effects of changes in vegetation carbon storage or climate on the carbon cycle, are important issues discussed in Chapters 3 and 6.

8.4.1.3 *Multiyear Alternations of Carbon Cycling*

Mathematical models of the carbon cycle commonly start by assuming a time of balanced incomes and losses. This simplifies calculating unknown transfer rates as differences among better known rates. Some imbalances are adjusted out within a few years. One example is the El Niño-Southern Oscillation (ENSO), a recurring alternation averaging 3 to 5 years in duration, but of varying intensity (Smith 1983). Suppressing upwelling along the eastern Pacific Ocean (Cane 1983) significantly affects mineral nutrients, marine photosynthesis, and ocean CO_2 uptake (Chapters 3 and 5). A further unresolved question is whether droughts and other climatic effects on land, responding to the same global geophysical phenomenon (Rasmusson and Wallace 1983), may diminish photosynthesis on large regional scales and increase plant mortality and subsequent decay, along with prompt or delayed wildfires.

More evident on land are alternations of longer term dry episodes (recorded by narrow tree rings). For example, droughts in Iowa and Illinois have commonly recurred on a roughly 20-year interval that might be related to the 22-year Hale solar magnetic cycle or an 18-year lunar tidal cycle (Figure 8.4); at some time the coincident interaction of these phenomena may serve to reinforce or to partially cancel each other's effects (Blasing and Duvick 1984).

8.4.2 Centuries to Hundreds of Thousands of Years

Postglacial and glacial-interglacial climatic fluctuations (Sections 3.3 and 3.2; Williams 1985; Nilsson 1983; Ennever and McElroy 1985; Peterson and Prell 1985) and earlier geologic variations have repeatedly disrupted any balance in the carbon cycle that might have occurred for several long intervals in the last million years. Sediments are deposited episodically, and (except for annual varve layers in very quiet water) most are stirred after they are deposited, thereby causing the geologic record in sediment cores to be smoothed over hundreds or thousands of years. That smoothing hides records of minor fluctuations in atmospheric CO_2 and carbon storage which should have been as typical of the distant past as of the present.

Interest in geologic variations of CO_2 increased greatly after publication of the ice-core CO_2 data (Chapter 3). These data show rising CO_2 levels within centuries near the end of the last glacial period (see Figure 3.5). Broecker (1982a, 1982b) and Broecker and Takahashi (1984) pointed out that direct chemical effects of warming the sea surface could explain only a fraction of this observed atmospheric CO_2 increase. However, effects of climate changes may have provided a mechanism to move CO_2 from the oceans to the atmosphere by altering the oceans' nutrient budgets. Removing approximately 30% of the oceans' dissolved phosphate (e.g., via increased sedimentation of organic matter) might have reduced algal productivity in the ocean surface layer, causing more dissolved CO_2 in upwelling waters to be released to the atmosphere.

Continental shelves and estuaries are significant

sites of organic matter sedimentation. The rise in sea level as glaciers melted must have expanded the shelf area available for marine organic deposition. However, the time lags inherent in the sea-level rise itself, and in the time required for significant peat (or carbonate) accumulation, now leave doubt that these slow shifts hold the key to finer fluctuations in CO_2 over a few centuries as found in ice cores (Chapter 3; Wenk and Siegenthaler 1985). Some carbon as well as phosphorus would be removed by shelf sedimentation. McElroy (1983a, 1983b) suggests that nitrate may be more important than phosphate. Yet the atmospheric effect of organic carbon removal may be negligible compared with the effect of decreased productivity on the sea's overall metabolism (cf. Baes 1982; Berger and Keir 1984).

Explanations involving a decrease in the ratio of carbon to phosphorus fixed by plankton requires fundamental biogeochemical changes in response to climate-related ecological changes. These changes could have caused dissolved CO_2 in upwelling waters to be released to the atmosphere if algae were less efficient assimilators of carbon. A related mechanism, proposed by Broecker and Peng (1984), is an increase in the ratio of planktonic carbonate to organic carbon production caused by rising sea levels due to glacier recession.

This change would have affected atmospheric CO_2 in the same way as the shallow water carbonate mechanism suggested by Berger (1982). He proposed that deposition of shallow water carbonates during the period of rising sea level would have decreased the alkalinity and increased the pCO_2 of ocean surface water, causing the release of CO_2 into the atmosphere. This excess atmospheric CO_2 eventually would be taken up by the ocean, mixed into the deep waters, and reacted there with carbonate sediments. The lag time between the release of the CO_2 and its "neutralization" in the deep sea could be a time of higher atmospheric CO_2 levels.

Broecker and Takahashi (1984) proposed that changes in ocean mixing because of glacial recession also may have contributed to the atmospheric CO_2 increase. They emphasized the importance of deep sea ventilation, defined as the extent of equilibrium of polar surface waters (viewed as deep water outcrops) with atmospheric CO_2. This process competes with ocean mixing and the photosynthesis/respiration cycle across the thermocline for control of the distribution of carbon within the atmospheric-ocean system. A shift in the balance among these processes—such as a shift in the fraction of nitrate and phosphate utilized by plankton in polar surface waters—might have contributed to the increase of atmospheric CO_2 levels. This possibility has been considered more explicitly by Knox and McElroy (1984), Sarmiento and Toggweiler (1984), and Siegenthaler and Wenk (1984). Other conflicting judgments and many further citations concerned with the history and mechanisms of change in Pleistocene and earlier times were discussed earlier in Section 3.2 and in several contributions in the monograph edited by Sundquist and Broecker (1985).

Glaciation disturbed Pleistocene geography repeatedly on land as well as on the continental shelf and slope and in the deep ocean. Some of the geologic weathering rates of significance to carbon dynamics are likely to be altered directly by coverage of the soils by Pleistocene ice, and also by temperature and moisture changes affecting ecosystem budgets far beyond glacial limits. Such changes in weathering rates need to be compared with those averaged over millions of years. The latter are considered in the next section.

8.4.3 Multimillion-Year (and Longer) Changes

A sedimentary mass nearly comparable in size and composition to the present mass (Table 8.2) probably existed at least 3.5 billion years ago, as well as an ocean of the present size and of similar composition. The atmosphere was of more debated composition. Oxygen levels increased in the atmosphere in Proterozoic time (Cloud 1978; Holland 1978, 1984), and simpler (prokaryotic) organisms were replaced by those with complex (eucaryotic) cells (i.e., those having separate nuclei and organelles for aerobic respiration and other functions). Multicellular animals of early, noncalcareous types thrived by 680 million years ago (the Ediacarian fauna: Cloud 1978; Holland 1984). Fossils with carbonate and phosphate shells became common in most of the major groups (phyla) of animals shortly after 600 million years ago (the Cambrian period).

Garrels et al. (1975), Budyko and Ronov (1979), and Berner et al. (1983) suggest that, over time scales longer than a million years, the relationship among atmospheric CO_2 levels, oceanic carbon, and carbonate sediments can be approximated by being near equilibrium between seawater, CO_2, calcite, and dissolved calcium and bicarbonate (Sections 8.3.1, 8.5.1, 8.5.2). The Chapman symposium proceedings (Sundquist and Broecker 1985) treat many aspects of these longer time scales (Worsley et al. 1985; Hay 1985; Sandberg 1985) as well as the shorter ones noted above. According to the model of Garrels and Lerman (in press), burial rates of organic (reduced) carbon and of carbonate (oxidized) carbon over 600 million years can be calculated where isotopic data are available. Using the compilation of $\delta^{13}C$ data of Lindh (1983), other values for rates of organic carbon burial have been calculated and are shown in Figure 8.5.

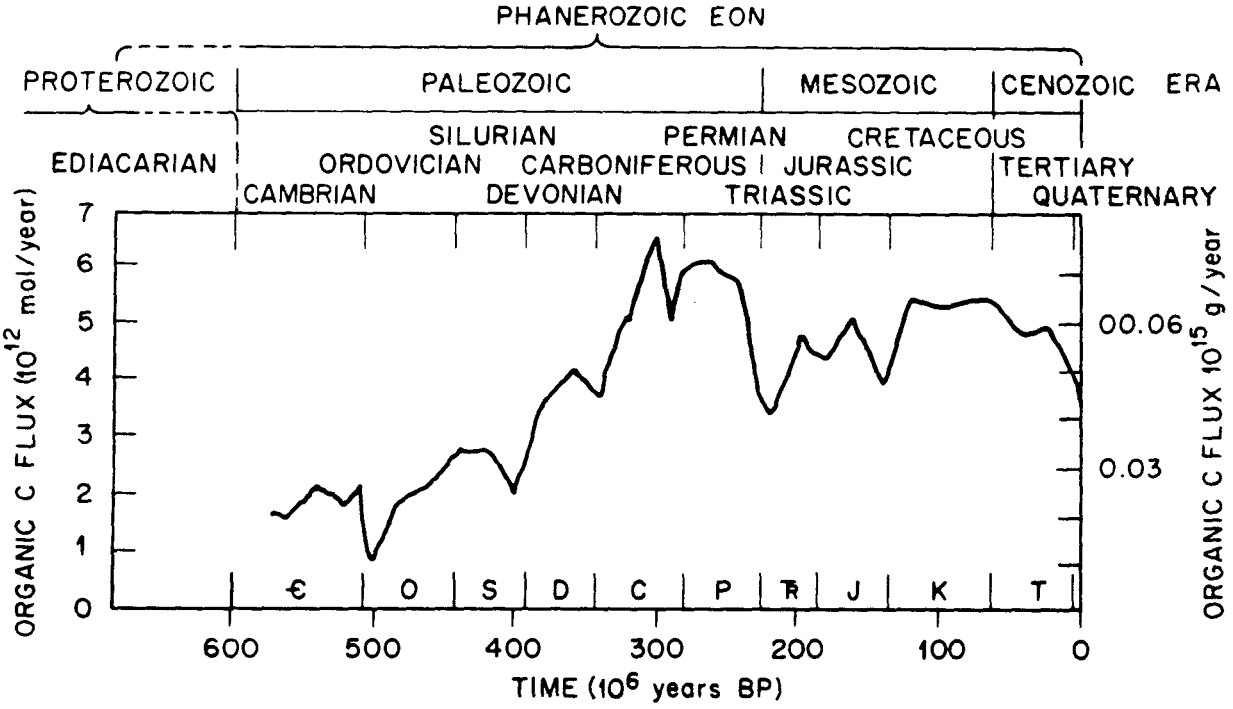


Figure 8.5. Burial rates of organic carbon (calculated from $\delta^{13}C$ data according to the model of Garrels and Lerman [1981]) at the Earth's surface over the past 600 million years ($\delta^{13}C$ data from Lindh 1983). ["Phanerozoic" time since other higher animal fossils were preserved under favorable conditions is now taken to be near 700 million years, according to Cloud (1978).]

Little dissolved carbon is stored in the ocean as compared with the amount present as organic matter and carbonate minerals in sedimentary rocks (Table 8.3; Figure 8.2). If input rates to the ocean and atmosphere had varied little over 10-million-year intervals, one might assume that changes in rates of burial of carbonate minerals form mirror images of the rate changes for organic carbon burial shown in Figure 8.5. Thus, carbonate burial should have been higher than at present during the early Paleozoic Era (organic burial was lower) and lower than at present during the Permo-Carboniferous (Permian, Pennsylvanian, Mississippian) periods.

It is possible, however, that input rates of carbon to the atmosphere-ocean system, especially those caused by releases of CO_2 during metamorphism and volcanism, may have varied appreciably over Paleozoic, Mesozoic, and Cenozoic Eras (see Section 8.5.1). Reasons for inferring this have been emphasized by Budyko and Ronov (1979), Mackenzie and Pigott (1981), and Berner et al. (1983). If this is true, then the simple assumption of a sole input by semiconstant weathering rates for organic carbon and carbonates may be inadequate. Thus, times of high volcanic CO_2 input could have been ›llowed by increased burial rates of calcium carbonate independent of the amount of organic carbon being buried. This concept is in contrast to shifts in burial between organic carbon and carbonate carbon. Such shifts require corresponding shifts in the rates of sulfur burial, or large changes in atmospheric oxygen would have occurred—beyond the range to which multicellular life has become adapted (See Sections 8.3.1 and 8.5.2).

8.5 INTERPLAY OF DRIVING AND STATE VARIABLES: EXAMPLES OF CHANGE

The cycling of carbon and other elements, on global and smaller scales, highlights a more general issue for science. When we focus on rapid changes, especially in small subsystems of the world, many slower and larger changes are neglected. But when we take a larger view, the "constants" or boundary conditions outside the initial small system may become interesting in their own right. Fluctuations governed outside the small system (the driving variables in a mathematical model of that system) become state variables within the definition of some larger system operating over longer time periods. The fluctuations just reviewed, and some of the processes linking the natural cycles of carbon with other chemical elements, further illustrate

this point. Many linkages among elements affect carbon biogeochemistry. Most are poorly understood and are excellent candidates for expanded research on global cycling.

Primordial inputs of carbon into the atmosphere have continued since early in Earth's history (Rubey 1951; Kasting 1985). Rates may have varied as major continental plates of relatively light rocks (of granitic composition) were moved over the underlying layers of heavier, iron-rich rocks. Local mountain folding and thrust faults occurred, partly where these continental plates were squeezed by the lateral motions of the plates (Cloud 1980). The gravity balance of the crust, as a whole, was readjusted by uplifts (e.g., where oceanic plates were thrust down beneath continents [Anderson 1984], their margins, or island arcs). Volcanism, mountain building, and more general uplifts of sediments or recrystallized rocks renewed the slow recycling of old organic carbon and mineral nutrients (Lasaga et al. 1985).

If the landscape contains available mineral nutrient elements, there will be vegetation—species selected from among flora available (Jenny 1980) in a given region and geologic time. Animals, fungi, and microbes will expedite the recycling of CO_2 (Figure 8.1). The structure of Figure 8.2B implies that if there is erosion, exposing fresh minerals for weathering from the rocks, there will be a demand for oxygen. If there is an oxygen demand, some of it is met by burial of organic carbon (Berner in press). To put it another way, for the long geologic time scale, the flux of organic carbon and sulfide to storage (along with nutrient elements) tends to equal the flux of old organic carbon and sulfide that were oxidized.

However, if significant variations in the organic carbon flux occur, it will be necessary to understand how the local or temporary departures from a balanced condition tend to be evened out or eliminated in the global system as a whole. Perspective, data, and models relating the alternations on several of the time scales outlined above are required, starting with the longest views.

In Proterozoic time, the Earth's crust already had a varying distribution of rocks of many different ages (Veizer 1985; Worsley et al. 1985). A highly schematic graph of the modern sedimentary mass in Figure 8.6 is a breakdown of the relative proportions of strata of various age classes. The dashed line restates the idealized (exponential) trend expected if the generation of new sediments (e.g., in every 10-million-year interval) were just balanced by the losses of older rocks. The peaks of the solid line show, alternatively, that surviving

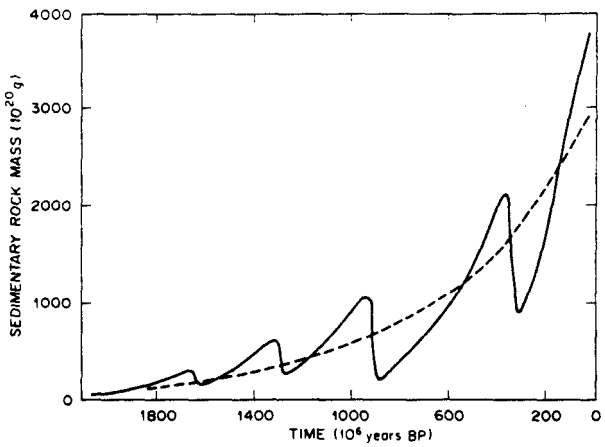


Figure 8.6. Highly schematic diagram of proportions of sedimentary rock mass of different ages. Fluctuations are driven by episodic renewals of continental plate movement and mountain building over geologic time, even after total sediment mass was nearly stabilized several billion years ago (after Ronov [1976] and Garrels [this chapter]). See text for discussion of differences between the solid and dashed lines.

sediments of a given time period may be rarer than those found from earlier or later intervals of equal length. Fewer sediments may have been preserved from periods when the continents were high and when new mountain chains were also higher. Some land or freshwater sediments were then washed off those highlands, partly into intermountain basins; but many of the freshwater or saline basin sediments would again be eroded. Therefore, there were fewer remaining samples than found for some other intervals, for example, when deeply down-warping basins stored large volumes of rock that were less exposed to new erosion or subduction cycles.

8.5.1 Carbonate and Silicate Cycles

The geochemical cycle of carbon is intimately intertwined with the cycle of silicon (Figure 8.2). In fact, it is this interrelationship that provides the main control on atmospheric CO_2 over long, million-year time scales (Budyko and Ronov 1979; Berner et al. 1983). Atmospheric CO_2 is absorbed by land plants and, upon respiration and decomposition, contributes to soil CO_2. This soil CO_2 then reacts with silicate minerals to form bicarbonate solutions with calcium (or other) cations. For example, the weathering of calcium-silicate minerals (here generalized by the surrogate formula $CaSiO_3$) can be written as:

$$H_2O + 2CO_2 + CaSiO_3 \longrightarrow Ca^{2+} + 2HCO_3^- + SiO_2 . \quad (8.5)$$

The overall effect is that atmospheric CO_2 is transformed to HCO_3^- in continental waters. After leaving the soil in groundwater, the bicarbonate is carried, via rivers, to the ocean (Schlesinger and Melack 1980). To retain the carbon balance, this HCO_3^- is removed from the oceans, mostly as calcium carbonate, $CaCO_3$ (Broecker and Peng 1982):

$$Ca^{2+} + 2HCO_3^- \longrightarrow CaCO_3 + CO_2 + H_2O . \quad (8.6)$$

The same reaction occurs within soil horizons during prolonged drying, especially under the grass and shrub landscapes of semiarid and arid regions. The net result of reactions (Equations [8.5] and [8.6]) is the removal of CO_2 from the atmosphere and its burial as $CaCO_3$:

$$CO_2 + CaSiO_3 \rightleftharpoons CaCO_3 + SiO_2 . \quad (8.7)$$

Constant drawdown of atmospheric CO_2 is opposed by the reverse of Equation (8.7) by metamorphism at depth. That is, upon heating via metamorphism or melting during burial, the CO_2 is liberated into the atmosphere and $CaSiO_3$ (or more complex silicate) is regenerated. Besides reactions shown in Equations (8.7) (with reverse arrow), (8.5), and (8.6), there is also the weathering uptake of CO_2 by carbonate minerals and the release of CO_2 on regeneration of carbonates and other minerals in the ocean (reverse of Equation [8.6]; Figure 8.1; Chapter 5). Reactions affecting CO_2, either directly or indirectly, in the geochemical cycles of carbon, silica, calcium, and magnesium are summarized in Figure 8.7.

The values for fluxes shown in Figure 8.7 are for a cycle balanced with respect to CO_2. A steady-state composition of the oceans and atmosphere was assumed. Perturbations of this steady state were quite likely over recent centuries because of the small amounts of carbon in the atmosphere or oceans. Such perturbations have probably occurred in the geologic past. Small changes in fluxes over the past 100 million years resulted in large changes in atmospheric CO_2, according to the model calculations of Berner et al. (1983). CO_2 levels appear to have been higher during the Cretaceous period (65–100 million years B.P.) than they are today, perhaps several times higher. Barron et al. (1981), Barron and Washington (1985), and Lloyd (1982) attempted to relate the connections between the climate during this period of warmer temperature to the infrared radiation balance in an atmosphere of higher CO_2 levels. The changes in carbon cycling that maintained such unusually high atmospheric values in Cretaceous time still need to be related to biogeochemical models for that period (see Arthur et al. 1985; Weissert et al. 1985; Barron and Washington 1985; Schneider et al. 1985). The recent difference in opinion about the degree and causes of climatic and biological changes since Cretaceous time was noted earlier in Section 3.2 (Alvarez et al. 1984; Hsu and McKenzie 1985; McLean 1985; Miller and Fairbanks 1985; Pisias and Prell 1985; Shackleton 1985). CO_2 was one of several factors that could help to explain such a difference. Varied rates of input (from volcanism) and patterns of recycling (in ecological and rock cycles) could well have regulated some of the differences in climate and biological process (Berner et al. 1983; Berner and Barron 1984; Lasaga et al. 1985).

Major changes in CO_2 levels on a sufficiently short time scale (e.g., less than 10^5 years) are rarely attributed to changes in the carbonate-silicate fluxes, as shown in Figure 8.7, because of the small values of the rock fluxes compared with those of biogenic fluxes and fluxes between the ocean and the atmosphere. Geologic fluxes tend to be smoothed out over longer time scales because the mass of carbon present in rocks is much larger than that bound up in the atmosphere-hydrosphere-biosphere system. However, it remains to be clarified how suddenly the weathering of soil and fresh rock parent materials may be even temporarily upset by certain shorter term changes, perhaps including some rates that have been affected by human behavior and by rapid Quaternary changes in temperature (Nilsson 1983).

An important long-term carbon flux is the one accompanying the burial of organic matter in sedimentary rocks. Flux values for this process are within an order of magnitude of those for the carbonate-silicate cycle, and changes in flux over geologic time could have contributed to changes in atmospheric CO_2 levels.

The estimate of 0.0382×10^{15} g C/year incorporated in rocks ($3.1[\pm 2]) \times 10^{12}$ mol C [Figure 8.2B] from consolidating ocean sediments) might be criticized as small. Clarifying uncertainty about the overall global flux is important for understanding the basic geochemistry of the lithosphere and biosphere. Part of the latter uncertainty is related to questions about the sedimentary flux from the Earth's landscape system (Figure 8.2A). Specifically the loop from subfossil carbon (including groundwater) into stratigraphic carbon assumes a conservative estimate of 2×10^{12} mol/year. If 3.1×10^{12} mol C/year is retained for the marine cycle, the combined deposition would be equivalent to $\sim 5 \times 10^{12}$ mol C, or 0.06×10^{15} g C, fixed per year. If this could be balanced by a steady-state release from weathering, then 0.14×10^{15} g O_2 would be consumed. This result produces an order of magnitude excess in the oxygen consumption of the carbon cycle compared with that of the sulfur cycle. Refinements of links between cycles will require further evaluation of the carbon- and sulfur-linked cycles (see Sections 8.5.2 and 8.6.2).

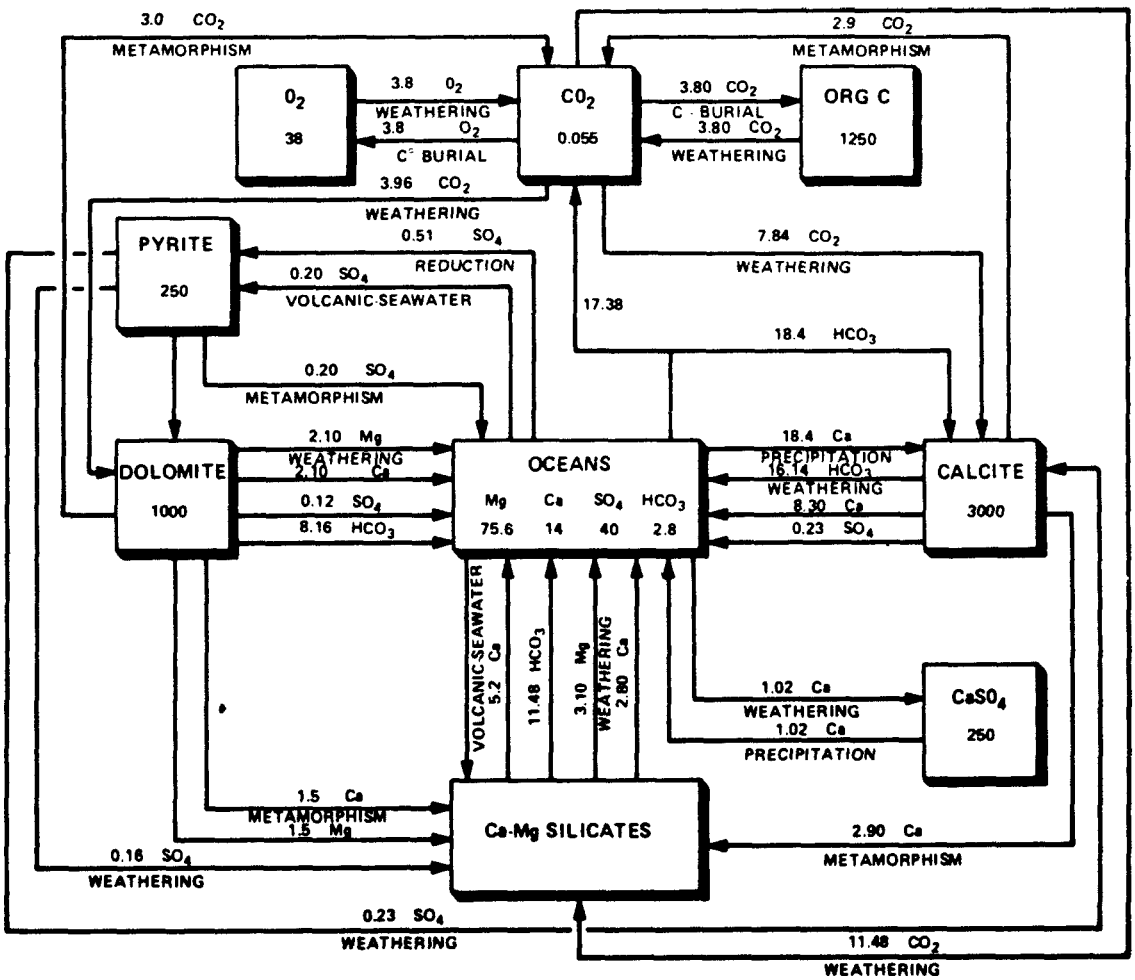


Figure 8.7. Larger pools and long-term (geologic) fluxes of carbon inferred by balancing silicate, carbonate, and organic carbon cycles (after Lasaga et al. 1985). Pools in 10^{18} mol (10^{18} g per atomic weight of element concerned). Fluxes in 10^{12} mol/year averaged over recent millennia but before major perturbations by humans from fossil fuel burning and most land clearing.

8.5.2 Organic Carbon and Sulfur Cycles

One way organic carbon burial is calculated for times in the geologic past is through the use of stable isotopes. We noted (Section 8.3.1) how changes in the ^{13}C level of the oceans are recorded in ancient limestone, and these shifts mainly reflect changes in the relative outputs of organic (reduced) carbon versus carbonate (oxidized) carbon over geologic time (Schidlowski et al. 1977). A model has been constructed for calculating organic carbon burial rates (Garrels and Lerman, in press; Lasaga et al. 1985). Some preliminary results were used in Figure 8.3. Through stoichiometric coupling (as shown by Equation [8.4] of Section 8.3.1), changes in burial rates of pyrite and $CaSO_4$-sulfur can be converted to changes in burial rates of organic carbon and calcium carbonate carbon using changes in $\delta^{34}S$ versus time.

The coupling of the geochemical cycles of carbon and sulfur suggests additional interesting results regarding the nature of organic carbon burial. If we plot the ratio between the rates of burial of organic carbon and pyrite sulfur versus time, the plot shown in Figure 8.8 is obtained. The observed fluctuations in the carbon:sulfur (C:S) ratio have been explained by Berner and Raiswell (1983) and Berner (1984, in press) by the burial of organic carbon in different environments. The peak of a high C:S ratio found for the Permo-Carboniferous periods represents a time (320–250 million years B.P.) when a much larger fraction of organic matter was buried on the continents, in freshwater swamps and

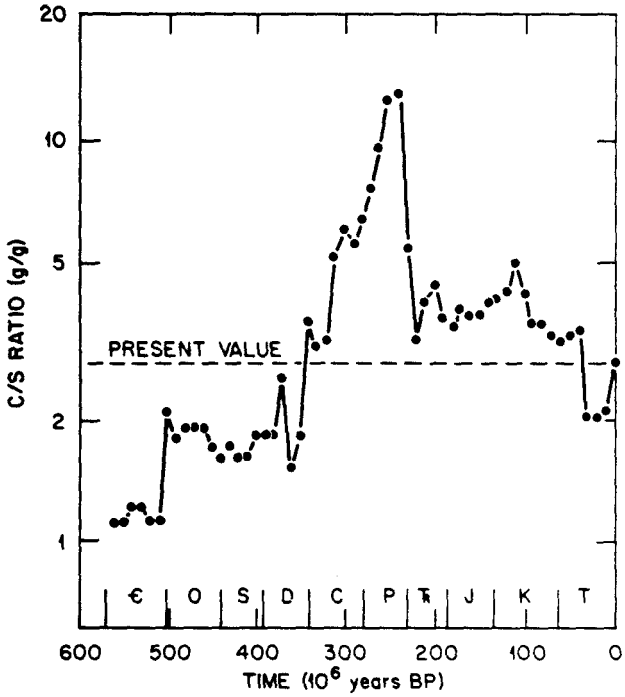


Figure 8.8. Calculated weight ratio of organic carbon to pyrite sulfur (C:S) buried in sediments over the past 700 million years (after Berner and Raiswell 1983). Geologic time intervals are also indicated on the abscissa. The coding follows that depicted on Figure 8.5.

lakes, than at present (Figure 8.8). (Low pyrite sulfur burial is explained as being due to the very low sulfate concentration in fresh waters as compared with seawater.) The values of low C:S ratio during Cambrian to Silurian time (600–400 million years B.P.), as compared with today, have been explained by the same authors in terms of increased burial of pyrite in portions of the ocean where hydrogen sulfide (H_2S) is present in bottom waters (euxinic basins, named after the Black Sea).

8.5.3 Carbon from Wetland Ecosystems

Wetlands are highlighted for special attention here because of their potential for recent and future net storage of subfossil carbon. Delayed release of excess CO_2 is also occurring (e.g., after soil drainage). Episodic deposition or oxidation can be modified by sea-level change or shore erosion. Many upland ecosystems lose organic material downslope, contributing inputs of carbon to floodplains, wetlands, and freshwater systems. Streams transport a fractional amount of this transitional ›l of carbon to the ocean (see Chapter 5).

Organic soils from wetland ecosystems were distinguished from those of other terrestrial systems in their function as long-term natural carbon sinks. Worldwide, carbon has been sequestered in peats and mucks, especially in poorly drained natural regions, exposed by melting ice after the end of major and minor continental glaciations. The total pool thus amassed is at least 165 $\times$ 10^{15} g C, based on a conservative estimate of world areas (Moore and Bellamy 1974), but may be as high as 500 $\times$ 10^{15} g C, given the latest estimates of peatland areas (Kivinen 1980) and carbon densities (Armentano et al. 1984).

Although present estimates of carbon sequestering rates fall between 0.14 $\times$ 10^{15} and 0.22 $\times$ 10^{15} g/year throughout the world (Armentano and Verhoeven 1985), these rates may not apply to most of Holocene time (~10,000 years after the last major glaciation). Some evidence indicates that present rates of peat input are significantly higher than earlier Holocene rates and may represent a midpoint in a transient of increasing accumulation (Durno 1961; Heinselman 1970). The basis for this hypothesis needs further investigation, particularly in regard to geographic differences and climate control, including the possibility that certain types of boreal peat lands accumulate carbon discontinuously.

Rates of vertical growth in peatland surface and hence in carbon sequestering rates are linked closely to prevailing climate and, in coastal areas, to eustatic responses of the Earth's surface to deglaciation. More locally, fire frequency and intensity can modify long-term rates of accumulation. Resolving the relative importance of these factors is critical for improving the reliability of rate estimates for wetland Holocene carbon sequestering. To understand the pattern of sequestering over time requires recognition of considerable geographic variability. Thus, the start of bog formation has been delayed up to several thousand years in some boreal regions such as around the Hudson Bay (Dionne 1979). Elsewhere in northern regions, lateral expansion of swamps by "creeping" water-logged peat (so-called paludification) covering areas up to 10^4 to 10^5 ha appears to have become increasingly significant in Siberia (Neishtadt 1977), possibly Alaska, and Canada. (In some areas impacted by acidic deposition, extra peat storage could be expected from slower decay and mineral recycling of organic residues in the future.) Farther south along several coastal plain regions, rates of organic production and sediment input are higher, but wetland areas are smaller. Sea-level changes mainly in the past 4,000 to 5,000 years have determined a subtle fluctuation of wetland sequestering roles (McDowell et al. 1969).

Subdividing the Holocene into climatic intervals reveals that during the past 2000 years (sub-Atlantic portion of pollen-climate profiles), peat accumulation rates have markedly increased in many regions—both

associated with and in addition to increased paludification (Durno 1961; Heinselman 1970; Neishtadt 1977). Boreal and perhaps many montane regions of the world have become more active, starting new sites of carbon accumulation as in blanket bogs (Gore and Olson 1967). Rapid rates of carbon accumulation also appear to have occurred in tropical coastal regions such as along the Sunda Shelf in Southeast Asia, in response to sea-level changes (Chambers 1980; Hanson and Koesoebiono 1979). At other sites within the same region, however, accumulation rates in bogs now 4,000 to 5,000 thousand years old are declining with maturation (Driessen 1978). Gore and Olson (1967; see also Figure 2, Olson 1981a) predicted such a trend for blanket bogs when loss rates increase along with total accumulations of peat, thus gradually overtaking the rate of carbon input to the top of the peat column. The net balance between these contrasting processes is not known, although a significant accumulation rate now seems most likely. The so-called recurrence (Grenz) horizons in many bogs indicate when deposition was interrupted by net oxidation, and these have usually been interpreted as evidence of how sensitively climatic fluctuations can alter the carbon balance of a region—examples of the kind of feedback noted in Section 8.2.

A shifting difference between opposed carbon flux rates also is indicated for some tropical and subtropical coastal regions where new sites of rapid carbon accumulation accompany land-building processes. In other sites marine transgression over previous land areas appears to be reducing or interrupting land-building processes (Gleason et al. 1984). Again, the net imbalance of these opposed effects is not known, but it may change over intervals as short as a century or so.

In addition to peat accumulation phenomena, rapid sedimentation and burial of organic material in coastal deltas, marshes, and bays may be sequestering carbon at even higher rates than interior wetlands. The present rates may be as high as 0.55×10^{15} g C/year in major deltaic areas, a rate which may be characteristic of the last 10^4 years (Section 5.4; after Mulholland 1981). Accumulation in shallow zones of lakes is an additional sink of about the same magnitude as deltaic accumulation (Mulholland and Elwood 1982).

Including bogs, carbon accumulation throughout all wetlands could be on the order of 1×10^{15} g/year at present. However, rates since the beginning of the Holocene and earlier geologic times may have varied significantly from this rate. Thus, historical reconstruction is needed to understand the links with climate shifts and with changing atmospheric CO_2 levels, as well as to define the wetlands' role in the biotic component of the global carbon cycle.

The preceding estimates of carbon accumulation could be a valid input to subfossil carbon and still be construed as substantially higher than the net output that becomes a long-term input to the geologic colum of sedimentary rocks. The reason for this simply involves the continued anaerobic decay that becomes manifested by the release of CH_4 as well as CO_2 and CO. Substantially more carbon may be released as CO_2 than as CH_4, even from saturated bogs; together, releases of $CO_2 + CH_4$ may exceed the carbon accumulated as peat. The subfossil summary line of Table 8.3 schematically uses the difference between the long-term geologic input of 0.024×10^{15} g/year (equivalent to that of 2×10^1 mol of C/year in Figure 8.2B) and the round number of 1×10^{15} g/year for input until a more refined estimate for the latter can be derived. If the overall subfossil pool of organic carbon is only 10^{18} g (about half as much as organic humus in the more active soil pool before widespread farming), the residence time near 1000 years gives a minimal value for definition of what is considered subfossil. The estimate would increase to several thousand years if the pool size proves to be larger, using a nominal input estimate of 1×10^{15} g C/year. Simply redefining the subfossil fraction as a larger part of the vast organic sediment pool (bottom line of Table 8.3) or allocating a smaller throughput to this fraction would also increase the calculated residence time.

Despite current uncertainties, the annual flux of carbon into wetlands and their sediments is not small when compared with current estimates of contemporary biotic and fossil fluxes to the atmosphere ($\sim 5 \times 10$ g C/year). It thus becomes important to determine whether the assumption of a near balance between income and loss in Table 8.3 and Figure 8.2 is a reasonable approximation or not. Perhaps there is a lag before income is overtaken by loss rates. The difference between these rates may be substantially larger than the 0.024×10^{15} g C/year estimate. The steady accumulation of carbon over a few centuries and millennia clearly could be very significant to the modern carbon cycle. This may have offset some of the carbon releases from other land ecosystems in later Holocene time.

Wetlands of previous geologic ages undoubtedly had a great impact on the global carbon cycle (Berner in press; Olson 1985). The coal, oil, and gas deposits which are extracted today and contribute, with human help, to increasing atmospheric CO_2 levels, represent a large amount of carbon removed from the atmosphere by wetlands, especially during the Permo-Carboniferous periods and early Tertiary time.

8.5.4 Other Ecosystems in the Evolving Biosphere

The kinds of land-clearing activities discussed in Chapter 6 reduced live vegetation to the range of 460×10^{15} g

660 × 10^{15} g C for the last century or two; probably in the lower half of this range in recent decades and the upper half around the year 1800. If Olson (1985) is correct in these allocations, and in his estimate of near 800 × 10^{15} g C for live plants 6000 years ago, then biospheric sources of CO_2 that averaged near or above 1 × 10^{15} g C/year to the atmosphere in recent decades (Olson 1982; Chapter 6) were an order of magnitude or more lower over earlier millennia of agriculture. The Dark Ages of Europe and times of depopulation in some early Asian civilizations could even have had significant periods of rebuilding of carbon in woods and soils (net CO_2 sink).

There is much uncertainty about whether the plant carbon could have been as high as 900 × 10^{15} to 1100 × 10^{15} g C in early Holocene or late glacial time (i.e., if plant migration and succession had been rapid enough, and if soils that had little time for the accumulation of humus and nutrients following the last deglaciation had been productive enough). During maximum glaciation, there was less land area for any kind of vegetation because ice and snow fields eliminated more available land area than was added to land on the continental shelf by the lowering of sea level. Diminished area in the tropics and expanded areas of desert where we now have Great Plains grassland probably did more to deplete plant mass during glacial episodes than the lessening of drying rates in some other cooler climates contributed to increase woody biomass. However, it is not yet clear whether global vegetation during major glaciations had similar or possibly lower plant carbon than does modern vegetation after its depletion by human exploitation. Presumably many soils were adjusting to altered levels of equilibrium between humus production and decomposition. New modeling studies will be required to determine whether lower production (affected mostly by dry climate) was more than offset by lower rates of decomposition (conditioned by cold and by waterlogging, especially on poorly drained terrain).

The potential role of heterotrophic organisms in regulating global carbon cycles has yet to be intensively investigated. The importance of microorganisms in soil building and in other processes is of course well known (Swift et al. 1979; Kononova 1966), but much less is known about animals. Many treat herbivores only as simple input-output systems, with the only impact being that of the amount of carbon stored in their tissues. This view overlooks an important aspect of system function (e.g., types and rates of turnover). However, even documenting the amounts of stored carbon, as apparent as it is, tends to be omitted from global equations, mostly because regional averages seem relatively insignificant when compared with biomass of utotrophs and humus. Indeed, in a static view, the estimates for amounts of carbon in herbivore consumers, and especially in their predators and parasites, are probably smaller than standard error measurements for each major kind of plant; animals hardly affect the summations in the live or total landscape pool of carbon.

However, more than one kind of contribution of heterotrophs is involved in regulating turnover rates. These fractional loss rates can be relatively high in grassland systems (McNaughton 1976, 1979) or can be rather low, as in large old-growth coniferous forests (Waring and Franklin 1979). Thus, the simple measure of turnover rates per se might not make the global contribution of heterotrophs to the carbon cycle seem significant. Yet animals also hasten the conditioning of grazed and detrital plant material so that microbial decay is accomplished at a faster rate. Bacteria or fungi (or fire) would eventually dispose of most residues even if such preconditioning were absent. But, without processing by animals, the growing medium for many decomposer organisms would be different—generally more resistant to decay than are excreta which are preconditioned by animals and their gut microbes (Mattson and Addy 1975). Insect and fungus pests also may hasten the destruction of tree stands, thus further shortening the mean residence time for live plus dead material, as averaged over wide regions (Mattson and Addy 1975). The rate at which and the manner by which carbon turnover is accomplished has a corollary, perhaps limiting residence times and mixtures of available nutrients besides carbon. These can limit ecosystem regeneration and the rates of primary production (Equations [8.1], [8.2]).

Heterotrophs of the carbon cycle may contribute more directly in other ways to the potential growth and storage rates of green plant biomass. It has been hypothesized that herbivores may contribute either positively or negatively to plant growth, depending on their grazing intensity. This biotic-control potential, which is known for grasslands (McNaughton 1979; Dyer et al. 1982), has never been assessed adequately for other systems, nor has it been applied to global problems. It has always been assumed that measures of standing crop biomass implicitly incorporate this potential series of positive or negative feedbacks into regional or global equations. However, the assumptions of zero flux and of zero influence from animals have to be wrong, even if the true mean and its error of estimate are small when compared with that for some other uncertainties. Ecologists need to determine where herbivore feedback can be positive or negative. Possibly for important cases the dynamics modify the measure and models of the global ecosystem more than anyone had considered previously.

Therefore, a first step is to assess heterotrophic influence more accurately in a variety of ecosystem types, starting with groups covering wide areas or

showing drastic changes. Then new models can incorporate parameter values into the global estimates of standing crop and seasonal changes in these crops, compared with models having flow-through but no control. Based on initial estimates from native grasslands, it is estimated that heterotrophs can increase such turnover rates by factors of up to 200% and in some instances they increase production of standing crop biomass, events not now considered in the global carbon cycle.

The emergence of higher (vascular) plants contributed a large, mid-Paleozoic increase in the pool of organic carbon (about half of the live biomass and humus of the actively cycling soils). A two- or threefold post-Devonian variation around the mean pool in living biomass (800×10^{15} to 1000×10^{15} g in early to mid-Holocene time) could represent a credible envelope of variation over geologic time. High organic carbon production on land occurred several times when increased continental areas, and exposure to fresh nutrient minerals in young soils, coincided with favorable climates.

As completely new Paleozoic and Mesozoic land communities emerged, heterotrophic processes must have played a major part in their development, based on what we know about coevolutionary patterns (e.g., between plants and animals; Gilbert and Raven 1975). Little has been published about plant-animal coevolution in aquatic communities, so it cannot be determined if the same theory holds true for them. By the time the Cenozoic era arrived, it is probable that many of the general interactions between families of plants and their grazers were already well in place. Coevolutionary associations since then have likely been more or less of a "fine-tuning nature" rather than of major consequence as far as global carbon patterns or other biogeochemical cycling phenomena are concerned. Many general associations of plant and animal communities likely had evolved to modern status, with mammals and flowering plants by the Paleocene or Eocene. Adaptation of higher organisms to extensive grasslands by late Miocene (Wolfe 1985; Olson 1985) and to deserts and tundra even later are exceptions. New coevolutionary adjustments in the world's biota may have had local or regional consequences, but probably little impact on global cycles of elements.

It is heuristically interesting to consider the impact of heterotrophic processes earlier in the history of the Earth, for tendencies for modern heterotroph control over primary production could have had early biochemical controls. Details of these associations are lost in the past, but we can speculate about their nature. The most likely candidates for modern evaluation involve the well-known association of herbivores and the production of a large series of plant secondary compounds (Rosenthal and Janzen 1979). These secondary compounds not only deter herbivore feeding, thus adding to the potential for plant growth in a community where herbivore influence is lessened; but they also detract from plant growth, because these compounds are very energy expensive (Mooney and Gulmon 1982). Up to 30% or more of annual plant productivity can be shunted into such herbivore defense, thus figuring heavily in both the amounts and types of resources a plant can develop.

No one yet knows how to develop the full equation of plant growth and herbivore effects. Herbivores can sometimes induce increased photosynthesis (Dyer et al. 1982), but plant energy and resource partitioning can be grossly affected by the reduced ability to complete life cycles. Compounds manufactured and sequestered in various plant parts can deter herbivore feeding but might also inhibit decomposition, and affect nutrient recycling. Thus, there is a link in plant-heterotrophic associations.

These associations can come to steady state, probably within a reasonably short period of time, certainly as far as the long-term global carbon cycle is concerned. After steady state is reached, then the only global effect that we may observe is heavily mediated by time lags and becomes damped after local or regional perturbation, such as shifts in climate, or perhaps major outbreaks of the herbivores themselves.

However, of particular interest might be the break in such steady-state conditions that could occur with major global events. We consider events such as extraterrestrial bodies striking the Earth (Alvarez et al. 1984) or the relatively rapid development and disappearance of continental ice sheets as potentially major perturbers of such autotrophic-heterotrophic associations. Behavioral evolution of great migrations might have affected these associations and led to new nutrient cycling patterns. It not clear at this time, based on the level of current information, how well the effects of humans and their livestock can be compared with earlier major episodes of balance or imbalance. The next stages of research can determine this, in the context of global atmospheric effects resulting from any great expansion of a single dominant species.

8.6 CONCLUSIONS AND RECOMMENDATIONS

We have reviewed carbon pools and flux rates (Sections 8.2, 8.3) and the tentative simplifying idea of steady-state balance. Some assumptions or estimates about the steadiness of the natural cycle could be misleading in the face of shifting forces (Sections 8.4, 8.5). Estimating and interpreting the rates of changes for the next century require longer time perspectives; allowing for imbalances could influence modeling, possibly enough

to alter inferences about climate change or feedback from climate to an altered world. Ecological impacts, human decisions, and possible actions may be so important that they must be evaluated long before our information on cycling and climate is complete.

The preceding sections suggest ways of reaching out for new connections between the formerly more isolated physical and biological views of our world. Several problems are ripe for faster interplay between the patient information gathering and imaginative activities of science. Even our partial knowledge can aid contingency planning. Lingering uncertainties about the penalties of assuming "no change" versus others about preparing for some alternative outcomes that differ from the real future can now be faced—without wishfully thinking that no planning is needed (NRC 1983; Olson 1980, 1984).

8.6.1 Conclusions and Uncertainties

Our planet's natural history has been controlled by a system of checks and balances. Departures from the equality of income and loss for carbon have tended in the past to be restored by a variety of mechanisms. Now the unprecedented impact of human technologies seems to be interfering with natural restoration. Natural fluctuations still make it difficult to recognize global changes that may have occurred as a by-product of trace gases altered by humans.

8.6.1.1 *Dominance of Photosynthesis and Respiration in the Fast Carbon Cycle*

Every year the biosphere exchanges about 100 times as much CO_2 as the small annual atmospheric increase observed in recent upward trends or as the annual input to subfossil sediments. As a result of technical difficulties encountered in measuring CO_2 gas exchange, ecophysiologists have relatively few records of respiration that differentiate between light and dark processes (Zelitch 1971; Foyer 1984) or between processes proportional to the photosynthetic rate and processes more nearly proportional to net growth or to accumulated carbon storage (Olson 1964). Such gaps need to be remedied for modeling ecosystems dominated by different proportions of C_3, C_4, and CAM plants (see Section 8.4.1). The common C_3 pathway of photosynthesis has been favored in many environments by superior quantum yield in photosynthetic conversion of solar energy. But some of this advantage has been offset by the natural and (recent) cultural spread of C_4 plants that adapted to the grassland and desert (and some agricultural) environments of hot, sometimes dry, growing seasons (Ehleringer 1978; Brown 1978). How carbon allocation changes over whole years or life cycles (see Section 8.4.1) is another problem area that is particularly significant for interpreting the coexistence of annual, perennial, and woody vegetation in many ecosystem types (Figure 8.1a).

The logistics facilities required by gas exchange methods limit the method to calibrations or validations for a few key reference conditions. Field measurements involving elaborate approaches, such as eddy flux correlation measurement of CO_2 gradients from aircraft and towers, may play a role in testing whether it is even possible to approximate the balance of net primary production, carbon storage, and recycling (Equations [8.1] and [8.2]) under particularly favorable air flow conditions.

As a result of such limitations on gas exchange methodology, there is no substitute for decades of regional-scale studies of sequential dry matter harvests, seasonal growth measurements and indices (phenology), cumulative growth histories (e.g., tree rings), allometric regressions among dimensions, and allowances for rates of fall of soft and woody litter. Ingenious approaches (some using remote sensing) are required to relate such productivity indicators across landscapes with spatial differences in substrates, topography, and climate. For example, element and energy budgets of agricultural sites need to be compared with those of equivalent sites lacking intensive, expensive, technological inputs and also with wild-land sites of widely disparate site quality. Equations (8.1) and (8.2) can and must be solved for diverse boundary conditions to quantify better the biosphere's inputs and net changes of carbon, respectively.

8.6.1.2 *Roles of Decomposers, Consumers, and Soils*

Net change (Equation [8.2]) is merely a difference between incomes and losses. Most organic residues are rapidly recycled to CO_2 or CH_4 by bacteria and fungi. The details of decomposition are poorly understood but they influence the quality as well as quantity of forest floor litter or grassland mulch. We understand the short-term impact of decomposers (simply as a flux) far better than we understand their influence on nutrients, their recycling rates, and their feedbacks to future productivity.

We can assume that consumers have divergent roles in grazing and metabolizing the carbon fixed by plants. But their roles in hastening remineralization of organic residues and otherwise stimulating growth seem likely to be more significant than if they were simply catalysts for oxidizing the carbon that was first reduced by plants. All of these processes can shorten the residence time for organic carbon, and also that for nitrogen, phosphorus, and many nutrient elements.

Ecosystem processes, such as decomposition and grazing, can respond rapidly to changes such as shifts

in climate or in land management. Yet the levels of more refractory humus compounds that dominate carbon budgets in soils may take centuries or millennia to adjust to a new steady state under altered conditions. A combination of detailed soil analyses that examine humus fractions and modeling and that take into account their very different residence times will be required to resolve questions such as those raised in Sections 8.2 and 8.4.2: What is the impact of the burial of active humus and ecosystems by advancing glaciers? What is the subsequent impact of the oxidation of these substrates after glacial retreat? What time lags are introduced into the process by the succession of pioneer and more stabilized ecosystems with each postglacial or interglacial retreat, or with recessions of shorelines? Questions of ecosystem migration, primary succession on raw substrates, and soil genesis over millennia must be brought into perspective for understanding the complex, transient alterations in carbon cycling that can result from changes in land use or climate.

8.6.1.3 *Carbon Flux from Forest, Grass, and Other Fires*

Seiler and Crutzen (1980) and Olson (1981b, 1982) agree that natural fires approach, but probably do not exceed, the annual release of nearly 5×10^{15} g C from the burning of fossil fuels (Chapter 4). However, natural fires may not occur with such regular seasonal and annual periodicities and trends as burning of fossil fuels. Droughts encourage more frequent, closely distributed, fires. The concurrence of extended drought with oceanic oscillations (see Section 8.4.1) could lead to disturbances of atmospheric CO_2, perhaps in excess of annual average oxidation by natural fires. Such a concurrence, combined with potential climate changes in the next century, could even set off a chain of positive feedbacks: increased CO_2 release, speeded warming, and possibly wider drought would encourage even more extensive burning. Controlled burning could be used more widely to reduce the impact of this feedback, but normally hastens carbon recycling from the landscape.

The potential for such positive feedbacks (Olson 1980, 1984) has recently received increased attention in the context of the effects of a nuclear war. It has been pointed out that sudden conflagrations, along with darkness, sudden cold, and displacement of biotic patterns, could curtail photosynthetic fixation of CO_2 and the very survival of many species for a period of uncertain duration. Such effects have never been considered in global cycling models such as those discussed in Chapters 7 and 10. Even studies of gradual readjustments of production or turnover rate as related to climate changes have not reached past preliminary stages.

The extent of burning and its changes in Pleistocene and Tertiary time (Herring 1985; Olson 1985) requires more attention. Local fires along coal outcrops and microbial oxidation of sediment carbon may have increased in uplifted sedimentary outcrops, new soils, and old soils (paleosols, buried or not). The drying climates of late Tertiary time, and the expansion of open woodlands, shrublands, and especially grasslands, may have increased the importance of fires on wild landscapes. Several times such fires may well have exceeded the 5×10^{15} g C/year burning rate now attributed to fossil fuel burning.

Whether the modern technical capability to limit wildfires could be used as a technique to store large amounts of forest and humus carbon, to help offset the releases from fossil fuels, is an issue that deserves imaginative scientific attention—combined with very cautious societal evaluation. Fire prevention, with the aid of aircraft and chemicals and public awareness, may have contributed to some of the regrowth of temperate forest biomass, which partly compensates for the drastic reduction of biomass and carbon in some tropical regions, thereby confusing the issues of worldwide change that have been discussed in Chapter 6 and by Hutchinson (1954), Olson (1980, 1981a, 1981b, 1982), and Armentano et al. (1984). Armentano and coworkers show how changes in the use of biomass for cooking and heating could have shifted rapidly enough to alter overall atmospheric budgets of CO_2.

Baes et al. (1976, 1977) noted that such substitution of nonfossil fuels might "buy time" for longer term resolutions of human energy problems. But shortened residence times of woody organic carbon thereby draw down the so-called "growing stock" volumes of wood that might have other higher quality uses; the areas of forest having many values for protection of catchments, gene pools, and numerous amenities (Olson 1981a); and the inputs of litter and soil carbon that counteract the ever-continuing decay of humus and nutrients (Jenny 1980; Section 8.2).

8.6.1.4 *Volcanic Releases of Oxides and Methane in Air and Oceans (Geochemical and Climatic Importance)*

Other processes tend to compensate for the slow release of CO_2 through volcanism. Injections of large quantities of sulfur dioxide (SO_2) sometimes lead to stratospheric sulfate; direct volcanic dust appears to have short-term effects on climate (Mitchell 1973; Hansen et al. 1981). Yet fumaroles and midocean contacts with fresh lavas have importance in sulfur cycling, the weathering of silicates, and the redistribution of bicarbonate solutions and calcite dust.

Atmospheric sulfuric acid and combustion pollutants like ozone (O_3) have direct biological influences

on vegetation. Effects are complicated by the interactions of ozone with nitrogen oxides that are also generated from pollution, compared with others that were part of the natural nitrogen cycle. Humans interfere with the latter cycle in other ways: directly with fertilizers to compensate for the slow rates of natural nitrogen fertilization of plants and indirectly through the runoff to streams, freshwaters, and coastal ocean. A further indirect influence of the enrichment (eutrophication) of water bodies involves local shifts toward more anaerobic bottom waters and sediments than would have been expected naturally. Are such effects already widespread enough to account for the observed shifts of sulfur cycling or the increase of methane? Is the latter a more direct by-product of hydrocarbon technology and exploration? Other aspects of pollution that affect global cycling could be important, but are beyond the scope of this chapter.

Another global question about methane, in addition to its recycling from fresh or consolidated sediments (Burke et al. 1981, 1983), concerns the possibility of deeper sources. Hunt (1979, 1984) reviewed the evidence that CH_4 in natural gas and ethane in petroleum can migrate to a limited degree even through fine-grained rocks like shale from source rocks having biological residues; but fractures, faults, bedding planes, unconformities, and sheet sands were the main avenues of migration of larger molecules of hydrocarbons. The concept of recycling of whole sediment masses, as in zones of subduction, raises interesting possibilities: that some fraction of the methane cycling depicted in the lower left corner of Figure 8.1 could reach midocean ridge lavas and basalts following remelting of sediments with reduced carbon (Welhan and Craig 1984; Welhan 1984). An alternative hypothesis, that a significant part of the natural gas is primordial (from deep crust and mantle rocks), has been reviewed in recent workshops (Gwilliam 1982; Marland 1984) because it bears on estimating ultimate bounds of hydrocarbon fuel source. Studies using isotope ratios (Craig 1984) suggest that the latter sources from lavas and crustal or mantle rocks are real, but quantitatively very small. Clarification should bear not only on natural gas but on the vast literature on the origins and migration of petroleum and its relations to the natural rock cycles.

8.6.1.5 *Rock Cycles and Internal Forces of the Earth Causing Long-Term Alterations*

Cloud (1978), Degens (1979), Berner et al. (1983), and Lasaga et al. (1985) discuss major changes in the element cycles since early eras in geologic history. We illustrated the linking of the carbon cycle with the sulfur cycles, according to Garrels and Lerman (in press), for the last 600 million years and with silicate weathering, according to Berner et al. (1983), for about the last 100 million years. A breakdown relating cycles of mineral carbonates and silicates is shown in Figures 8.2B and 8.7.

The current estimates of stratigraphic pools of calcium and magnesium carbonates seem as satisfactory as historic geology and sedimentology can give at this time. However, we are open to further revision of the fluxes of CO_2 involved in the weathering of these carbonates. The silicates are represented by $MgSiO_3$ and $CaSiO_3$ to balance the global rock cycle (also see Section 8.5.2). The steady-state approximations seem to be in better agreement with the model of early outgassing of the Earth's mantle than it is with older ideas that such deep sources of CO_2, CO, and CH_4 might have continued at steady rates throughout most of the Earth's history.

8.6.1.6 *Other Linked Cycles: Carbon, Nitrogen, Phosphorus*

The role of nitrogen cycling rates in controlling the pace of the carbon cycle, and possibly in influencing climate via nitrogen oxides, is one of the larger unknowns in comprehending linked element cycles. Nitrogen is generally given less attention by geochemists than by physiologists, ecologists, and soil scientists. More linkage between these disciplines is important, because terrestrial and marine organisms depend on nitrogen and also on phosphorus as rate-limiting factors affecting their growth and the turnover of their residues. How the internal forces modify nitrogen and phosphorus cycles remains to be clarified.

8.6.2 Priority Needs for Research on Global Cycling

Emphatic statements about gaps in existing knowledge and requirements for extending present knowledge have been rendered at dozens of places above, through either reasoning or literature sources to back up the authors' conclusions. Naturally we vary in sensitivity to disciplinary specialties. Neither this review nor the many outside opinions received need to be smoothed out into one bland endorsement. Many worthwhile things must be done sooner or later. In the following sections, we highlight just a few thrusts made toward understanding our geosphere and biosphere that will require strenuous persistence in work already begun or shifts in emphasis.

8.6.2.1 *Isotope Sampling of Geochemical History*

Our current knowledge of the concentrations of carbon and sulfur isotopes in rocks spanning the last 700 million years is based on a few thousand carbon isotope determinations. The carbon and sulfur isotope concentrations quantify the movement of materials from one sedimentary reservoir to another, the average

rates of production and storage of oxygen and carbon. On this slow-breathing system the shorter term perturbations of humans are imposed. Limits on oxygen concentration fluctuation (i.e., between 15 and 25% of the atmospheric gas content) have kept the Earth habitable for higher forms of life (eukaryotes). Before the natural carbon cycle itself can be fully understood, the linkage to silicate weathering (Section 8.5.1) and even "reverse weathering" will involve the rock cycles of calcium and other metallic nutrients, soil formation, erosion, and marine processes.

It should become feasible to explore how the carbon cycle changes between glacial and interglacial history, over tens or hundreds of thousands of years and in diverse global climates, might compare with the shifts over millions of years of a slowly cooling average climate since the warm Cretaceous period. Doing this, however, will require more attention to rate coefficients and temperature, and transient conditions of several kinds, because the conditions of interest may change even before equilibrium (i.e., balancing around the whole cycle) can be attained. The large organic pools of vegetation and soils will require treatment not only as conduits without relevance for carbon storage but as state variables and storage parameters that modify physical and chemical rates.

8.6.2.2 *Ocean Tracer Studies and Alternations*

Radioisotope tracers aid in the study of natural CO_2-climate relationships because the climate problems involve similar scales of space and time: decades, centuries, or longer. This provides an understanding of the past history of the cycle and the potential perturbations by climate. Comprehensive tracer studies on the world oceans, like GEOSECS (Geochemical Ocean Section Study) in 1972–1978 and the TTO (Transient Tracers in the Ocean) in 1981, provided excellent but limited data. The TTO program was to expand tracer studies to all oceans, but completion of the program may prove impossible.

A worldwide tracer program like the TTO concept is important for modeling the carbon cycle, along with climate and ocean interactions. Experimental worldwide data are needed to test the models and for credible predictions. The TTO or another similar program will have to be carried out for a few decades, starting as soon as possible, particularly while the tracer transients initiated by nuclear weapons testing are still developing (see Chapters 5, 7, and 11).

8.6.2.3 *Landscape Cycling Research*

Physical research is needed on the planet as a whole: atmosphere, lithosphere, and hydrosphere. Data from certain mountains, geologic strata, or ship cruises are merely increments to collective global knowledge. Ecologists want to integrate findings into a global picture. But global problems scale very differently than the limited issues about communities, food webs, and habitats. The principles gained from local studies are essential to improve current averages on a biospheric scale. Mixing of CO_2 from local sources into the rapidly stirred atmosphere may aid in testing whether seasonal trends can be resolved in the income and loss rates for landscape mosaics.

The changes in land use since human interference go back earlier than the 1860 date most used for estimating fossil fuel releases (Olson 1985). Yet we can only interpret the resultant changes by studying age sequences and carbon time series for vegetation and soil recovering after disturbance. We need to understand the incomes and losses of the major ecosystems' budgets for carbon and for nutrients represented by Equations (8.1) and (8.2). The role of heterotrophs, the consumer and decomposer of organic food, in modifying the annually recycled flux of over 50×10^{15} g/year of carbon remains a basic issue of ecosystem biology.

Fluxes of carbon, nitrogen, sulfur, and other mineral nutrients could be altered when climate becomes modified by CO_2, CH_4, or other atmospheric gases. Whether primary production could be altered much by feedbacks from animals and decomposers, as well as by CO_2 itself, is a question requiring the combination of global ecological and physiological approaches.

8.6.2.4 *Models for Nested Systems*

The differences in scales of space and time in which various processes of the lithosphere, hydrosphere, atmosphere, and biosphere operate must become a major focus in future research on the geosphere. There are problems inherent in modeling each of these scales that have been discussed in this chapter. Future research will also require additional attention to linking the widely different space-time intervals relevant to global cycling (Sundquist 1985). To attempt to answer this chapter's major questions requires a globally averaged scale to complement other subsystem studies. Yet such modeling provisionally allows simplification so as to look at one aspect of the system after another. More effort in the field has been applied to smaller scale problems, using a single geochemical species. Thus, there remains a critical need (Olson 1971) for more modeling across a hierarchy of space-time scales.

One suggestion from Olson (1971) is to start from intermediate levels such as major landscape complexes (drainage basins and biomes) or ocean gyres. These can be modeled now with fuller insight than ever before, using what has been discovered from local data or specialized modeling and from field and laborator research. Yet, without waiting for inductive syntheses t

be perfected, modelers must readdress the much larger problem of global coupling of land and oceans via rivers, lakes, and estuaries and coupling of carbon cycling with that of other elements. We need to quantify which rate-limiting factors affect the ranges of response times and quasi-periodic forces affecting such cycling.

8.6.2.5 *Role of the Biota in Climatic and Geological Processes*

Isolating discussions of geochemistry, ecology, and climatology in a review helps to simplify communication. However, these disciplines must be reconsidered together to address the feedbacks that link the carbon cycle to climate and the biosphere. Provocative insights emerged from pioneers like Dumas and Boussingault (1841), Vernadsky (1926), Riley (1944), and Hutchinson (1954) who dared to seek connections between operation of the planet and its whole zone of life, called its biosphere. Stretching perspective over successively longer intervals of historic, prehistoric, and geologic time is not so much a digression from the questions of only passing interest of current human activity, as it is a potential guide for allocating precious attention and time to issues of more than passing interest.

For example, changing atmospheric composition, during the early history of the Earth, might have involved greenhouse effects by methane (Lovelock and Whitfield 1982). The oxidation of this organic gas at least 2.2 billion years B.P., when oxygen appeared in the atmosphere, coincides with early glaciations. Ways in which higher CO_2 levels could have interacted with subsequent geologic events were suggested in previous sections and in Sundquist and Broecker (1985). The question remains as to whether current changes in atmospheric carbon (and other oxides) can cause drastic changes in rates discernible relative to those for glacial and earlier geologic times.

Other research questions can be formulated around the feedbacks that link the specialized disciplines. The system comprising the Earth's surface, atmosphere, and biota may have represented a long-term, although evolutionary, homeostasis. The climate and the chemical composition of the atmosphere have varied, but both were regulated by organisms as well as by physical processes. Regulation can arise simply through survival of those organisms and processes that can improve the odds for maintaining a life-support system. This, recently restated as the Gaia hypothesis (Hitchcock and Lovelock 1967; Lovelock and Whitfield 1982; Whitfield 1981), offers a view of the natural cycles different from that assuming a system of physical equilibria. It suggests that both atmospheric oxygen and carbon dioxide have remained within different ranges acceptable for life by ecosystems regulated by organisms changing their own steady states over billions of years.

This planetary life-support system has been remarkably effective, despite occasional impacts including random or perhaps periodic impacts of large asteroids or comet nuclei (Alvarez et al. 1984). Yet mass extinctions, including those near the Cretaceous-Tertiary transition, need not be restricted to one particular school's scenario for cause of disruption and recovery in order to make a general point; we have much to learn from and about long-term biospheric stability to understand the human-modified zone of life which Vernadsky (1926) called the noösphere. Human effects on atmospheric composition and on the size and operations of the terrestrial ecosystems represent departures in a life-support system that had evolved over billions of years. We should welcome more evidence of the biosphere's self-regulation. Could a kind of cybernetic control moderate the human cost of regulation or adaptation to change?

Le Chatelier's principle reminds us that, for closed systems, disturbances of the mixed components or of the system's environment (e.g., temperature) yield adjustments leading to a new equilibrium that partly compensates for the perturbation. In complex open systems, we need deeper inquiry to see how much analogy can be found in the real, living world. Oxygen apparently has not varied outside the range of 15 to 25% of the atmosphere in view of the relationship between metabolism and the flammability of residues to oxygen concentration. More satisfactory control of oxygen excursions in models of Lasaga et al. (1985) requires further refinement, along with simulation over longer periods of geologic time—600 million years instead of 100 million. Carbon dioxide is known to have varied manyfold, and its range has apparently been great enough to modify photosynthesis, carbon storage, and the planetary radiation balance affecting climate.

Ecological systems have evolved on different continental plates, slowly drifting across the latitudinal zones of a planet whose overall climate has been changing. Some organisms migrate very rapidly, and many (not all) groups adapt fairly rapidly compared with the slower geochemical and geophysical trends of the Earth (Olson 1985). It is not clear whether we must look mainly to chemical transfer between landscapes (via the atmosphere or streams) or to some other kinds of interactions among species to clarify their roles in "controlling" the long-term habitability of the world.

It therefore seems necessary to develop models for cycles of elements and climate that contain explicit formulations for more active participation by the organic agents. We need to compare such models' predictions with those of the geochemical cycles that allow only a passive role for the biota. As with the working hypotheses of steady state, the adequacy or limitations of model simplifications, made early for aiding entree

into a complex inquiry, may be seen more clearly. Theories that persist may become robust enough to be shared by the different models, designed for quite contrasting domains of time and space.

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9. FUTURE GLOBAL ENERGY AND CARBON DIOXIDE EMISSIONS

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9.1 INTRODUCTION

The principal uncertainty governing future concentrations of atmospheric carbon dioxide (CO_2) is manmade emissions of CO_2 from the burning of fossil fuels. Because the terrestrial carbon pool potentially exhaustible to the atmosphere is large (see Chapter 4, this volume; henceforth all chapter and section callouts are to this volume), future energy choices are a matter of immediate concern. Because CO_2 mixes rapidly throughout the global atmosphere, a global study of future energy choices and CO_2 emissions is essential. And, because CO_2 has a relatively long atmospheric residence time, the issue is long term.

This chapter reviews work to date on future CO_2 emissions from the combustion of fossil fuels. This report summarizes the state of the art in several key areas: forecasts of future fossil fuel CO_2 emissions (Section 9.2), uncertainty surrounding future emissions (Section 9.3), key factors determining future energy and CO_2 futures (Section 9.4), the analysis of policy impacts on future emissions (Section 9.5), the methodological techniques applied to the forecast and analysis of future CO_2 emissions (Section 9.6), the influence of the forecast and analysis validation process on forecasts (Section 9.7), the major programs that have helped shape thinking in the field of emissions (Section 9.8), and the study of emissions of other non-CO_2 radiatively active trace gases (Section 9.9). This chapter is written in light of, and builds upon, several other summary reports on future energy-CO_2 emissions including Perry (1982), Ausubel and Nordhaus (1983), Rose et al. (1983), Jarrett and Coates (1984a), and Keepin (in press).

9.2 FUTURE CO_2 EMISSIONS FORECASTS

This literature review focuses on those forecasts that are formal in nature. That is, they employ a structured approach to the development of scenarios which are reproducible at least in principle. We will define that reproducible process as a model. This definition is not without its risks because users are wont to define models more narrowly as that subset of models that have been reduced to computer code. This is far too narrow a definition. Several rather sophisticated models of CO_2 emissions are not computer models (e.g., Perry and Lansberg 1977; Allen et al. 1981; Williams et al. 1984). Although this limits their portability and accessibility to other users, they are nonetheless models.

This definition of modeling is also sufficiently broad to include models that consist of nothing more than a time trend term (JASON 1979; Bacastow and Keeling 1981). It excludes forecasts based solely on forecasters' judgments, and are therefore not open to formal inspection.

The results of 25 CO_2 emissions forecasting exercises are given in Table 9.1 and Figure 9.1. The coverage of this table and figure is not intended to be exhaustive but representative of a cross-section of work in the field published outside the centrally planned economies.

Table 9.1 summarizes forecasts of CO_2 emissions based on the assumption that no explicit policy action is taken to reduce CO_2 emissions. The policy issue is taken up explicitly in Section 9.6. No attempt has been made to standardize forecasts except to include only studies of global emissions and to use conventional units (10^{15}/g of carbon). Figure 9.1, on the other hand, does attempt to make forecasts comparable by standardizing emissions forecasts to a rate of growth over the period 1975 to 2050. Because all studies do not provide forecasts for the year 2050, some points that cannot be compared are glossed over by calculating the rate for the period whose length is most nearly comparable. Figure 9.1 displays both the "best-guess" scenario and the range of "business-as-usual" scenarios examined. In some reports the best-guess scenario is the only scenario. In others it is termed the "base case," "reference case," or the "median" scenario. In some reports no such scenario is designated and arbitrary calculations have been made (i.e., averaging the high and low growth rates).

Emissions in the year 2050 range from over 100×10^{15} g C/year (JASON 1979) to less than 1×10^{15} g C/year (Lovins et al. 1981). Recent emissions forecasts have included best-guess scenarios (base case, reference case, median scenario) as well as a range of alternative scenarios. The best-guess scenarios of major recent studies have ranged from 15×10^{15} g C/year in the year 2050 (Nordhaus and Yohe 1983) to 8×10^{15} g C/year (Edmonds et al. 1985), lower emissions forecasts than such earlier studies as JASON 1979. They admit a wide range of possible business-as-usual scenarios within the bonds of their 95% confidence intervals. Nordhaus and Yohe (1983) put the 95th percentile 2050 emissions rate as high as 26×10^{15} g C/year, while Edmonds et al. (1985) admit up to 64×10^{15} g C/year; 5th percentile emissions are 5×10^{15} g C/year and 1.5×10^{15} g C/year, respectively.

Figure 9.1 graphically represents the distribution of forecasts over the period 1975 to 2050 in the form of average annual growth rates. This 75-year period was selected to reflect the long-term nature of the emissions

Table 9.1
Forecasts of Carbon Emissions from Fossil Fuel Use by Study (units = 10^{15} g C)

	Study		Model[a]	1975[b]	Forecast Year 1980	2000	2020	2025	2030	2040	2050	2060	2075	2100
1977														
1.	Keeling-Bacastow	(H)[c]	[Own]	4.7	–	16	–	50	–	–	86	–	50	46
		Mid		4.7	–	13	–	26	–	–	38	–	39	30
		(L)[c]		4.7	–	11	–	18	–	–	25	–	26	25
2.	Perry-Lansberg	(H)	[Own]	4.7	–	–	–	27	–	–	–	–	–	–
		(L)		4.7	–	–	–	13	–	–	–	–	–	–
1978														
3.	Siegenthaler-Oeschger	(H)	[Log]	4.7	–	15.5	–	44.7	–	–	90.2	–	93.3	48.5
		(L)		4.7	–	4.7	–	4.7	–	–	4.7	–	4.7	4.7
1979														
4.	Rotty	(H)	[Log]	4.9	–	11	–	27	–	–	58	–	87	78
		(L)		4.7	–	10	–	19	–	–	28	–	30	27
5.	Niehaus-Williams (5 cases)	(H)	[IIASA]	–	5.5	13.2	28.8	–	–	33.1	–	38.6	–	–
		Base			5.5	13.0	13.0	–	–	9.0	–	7.0	–	–
		(L)			5.5	8.2	4.0	–	–	1.5	–	1.0	–	–
6.	Nordhaus		[Own]		6.9	10.7	18.3	–	–	40.1	–	–	–	64.0
7.	JASON		[Own]	4.7	–	14.1	–	42.5	–	–	127.6	–	383.5	–
8.	Colombo-Bernadini		[IIASA]	4.7	–	6.3	–	–	6.8	–	–	–	–	–
1980														
9.	Rotty-Marland		[Log]	4.8	–	9.0	–	13.6	–	–	15.0	–	13.2	10.8
1981														
10.	Bacastow-Keeling		[Own]	4.7	–	13.5	–	38.6	–	–	110.5	–	316.6	407.1
11.	Sundquist-Plummer		[Own]	4.7	–	12.2	–	–	31.6	–	82.0	–	212.7	551.7
12.	Haefele	(H)	[IIASA]	4.7	–	9.7	–	–	16.8	–	–	–	–	–
		(L)		4.7	–	7.5	–	–	9.8	–	–	–	–	–
13.	Lovins et al.		[IIASA]	4.7	–	2.9	–	1.0	–	–	1.0	–	–	–

Table 9.1 (Continued)

	Study		Model[a]	Forecast Year 1975[b]	1980	2000	2020	2025	2030	2040	2050	2060	2075	2100
1983														
14.	Edmonds-Reilly (1983b)		[ER]	4.7	–	6.9	–	12.3	–	–	26.3	–	–	–
15.	Nordhaus-Yohe	(95%)[d]	[Own]	4.6	–	8	–	17	–	–	26	–	–	55
		(75%)		4.6	–	7	–	13	–	–	17	–	–	27
		(50%)		4.6	–	5	–	10	–	–	15	–	–	20
		(25%)		4.6	–	4	–	7	–	–	8	–	–	12
		(5%)		4.6	–	2	–	4	–	–	5	–	–	7
16.	Rose et al. (11 cases)	(H)	[ER]	4.7	–	7.3	–	10.4	–	–	15.1	–	–	–
	Case H			4.7	–	4.1	–	5.1	–	–	7.4	–	–	–
		(L)		4.7	–	3.6	–	3.1	–	–	2.8	–	–	–
17.	Seidel-Keyes (6 cases)	(H)	[ER]	4.7	–	7	–	11	–	–	17	–	35	60
		Base		4.7	–	7	–	10	–	–	15	–	31	50
		(L)		4.7	–	7	–	9	–	–	10	–	19	30
1984														
18.	Edmonds et al.													
	A		[ER]	4.7	–	8.4	–	19.0	–	–	47.4	–	91.1	–
	B			4.7	–	7.2	–	10.3	–	–	14.5	–	18.8	–
	C			4.7	–	6.2	–	6.7	–	–	6.8	–	6.8	–
19.	Reister (1984a)													
	(GRI)	(H)	[ER]	4.7	–	8.1	–	15.3	–	–	27.1	–	43.7	62.2
		Base		4.7	–	7.3	–	9.1	–	–	10.7	–	10.6	10.6
		(L)		4.7	–	6.9	–	8.5	–	–	9.7	–	9.8	9.9
20.	Reister (1984b)													
	(EPRI)	(H)	[ER]	4.7	–	7.3	–	11.5	–	–	18.5	–	27.9	37.1
		Reference		4.7	–	7.3	–	10.8	–	–	15.8	–	15.1	13.7
		(L)		4.7	–	7.2	–	10.1	–	–	14.2	–	12.7	11.2
21.	Manne	(H)	[own]	4.7	–	7.4	–	10.7	–	–	14.8	–	–	–
		(L)		4.7	–	7.4	–	10.1	–	–	11.4	–	–	–
22.	Williams et al.													
		(H)	[IIASA]	4.7	–	5.4								
		Base		4.7	–	–	4.6							

(Continued)

Table 9.1 (Continued)

Study		Model[a]	1975[b]	1980	2000	2020	2025	2030	2040	2050	2060	2075	2100
			Forecast Year										
1985													
23. Edmonds et al.[d,e]		[ER]											
	(95%)		4.7	–	11.6	–	33.4	–	–	63.9	–	90.1	
	(75%)		4.7	–	8.0	–	15.3	–	–	23.5	–	34.5	
(a)	(50%)		4.7	–	6.2	–	8.7	–	–	10.7	–	12.6	
	(25%)		4.7	–	4.6	–	4.7	–	–	4.6	–	4.3	
	(5%)		4.7	–	3.1	–	1.9	–	–	1.5	–	1.1	
	(95%)		4.7	–	10.9	–	29.8	–	–	58.1	–	86.9	
	(75%)		4.7	–	7.7	–	13.0	–	–	18.7	–	27.1	
(b)	(50%)		4.7	–	5.8	–	6.9	–	–	7.7	–	8.5	
	(25%)		4.7	–	4.6	–	4.5	–	–	4.2	–	3.9	
	(5%)		4.7	–	3.2	–	2.4	–	–	2.3	–	1.8	
24. Chandler	Base	[ER]	4.7	–	7.2	–	10.3	–	–	–	–	–	–
	(L)		4.7	–	5.8	–	7.0	–	–	–	–	–	–
25. Rogner[f]		[IIASA]	4.7	5.4	7.0	–	9.0	–	–	10.0	–	–	7.8

[a]Several of the studies use either a common model or are tied to a common modeling framework. Several studies use the Edmonds-Reilly model developed for DOE at the Institute for Energy Analysis. These studies are noted by [ER]. Other studies are tied to the conventions, some assumptions, and/or model outputs of the Energy Project of the International Institute for Applied Systems Analysis [IIASA], though no other study actually employed the IIASA model itself. Some of the earlier studies used variants of a logistic curve [Log] to project emissions. Other studies developed projections using their own models [Own].
[b]Base year.
[c](H) highest CO_2 emissions case considered, (L) lowest CO_2 emissions case considered.
[d](95%) 95th percentile value in an uncertainty analysis, (75%) 75th percentile value in an uncertainty analysis, (50%) 50th percentile value in an uncertainty analysis, (25%) 25th percentile value in an uncertainty analysis, and (5%) 5th percentile value in an uncertainty analysis.
[e]Results in (a) and (b) categories were obtained from model runs with and without correlations, respectively, among certain key parameters.
[f]Source: Rogner (1986)

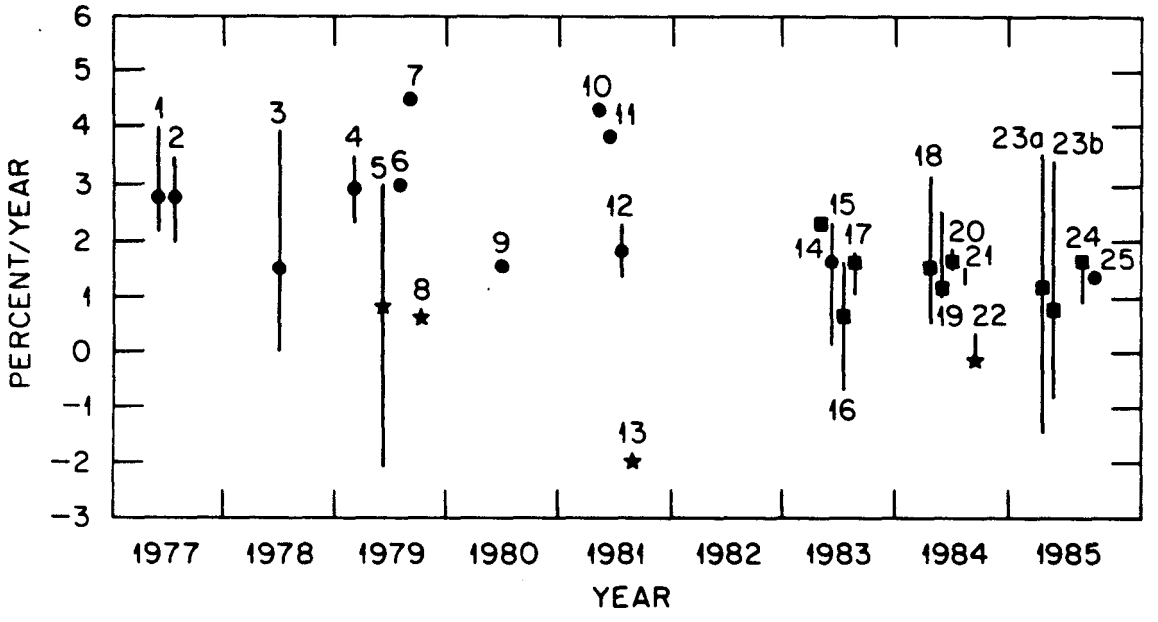


Figure 9.1. Rates of projected growth of carbon emissions over the period 1975 to 2050 grouped by study and date of publication: (*) base case or average of cases for feasibility-type studies, (□) base case or most likely values for study using the IEA/ORAU model, and (·) base case, average, or most likely values for other studies. Vertical lines represent uncertainty estimates (e.g., either high and low scenarios, or 5th and 95th percentile bounds, were applicable). Numbers refer to those in Table 9.1.

question (most contemporary analyses would place the date at which atmospheric concentrations of CO_2 reach 600 ppm after the year 2050), but short enough to capture the essential elements of most studies. In some cases, other time frames were used. Forecasts by Haefele (1981) and by Williams et al. (1984) terminate in the years 2030 and 2020, respectively. Although the growth rate is a convenient common denominator, allowing direct comparison of studies with different base and terminal dates, it does imply that a simple exponential growth model can be used as a first approximation of the analyses. In most cases this is reasonable, but in some instances (e.g., the Niehaus and Williams [1979] low case), such a model provides an extremely poor summary of the analysis.

If we divide the studies into two groups, those published before 1982 and those published after, there are some distinct trends that become apparent. Studies produced after 1982 are more likely to explore a range of scenarios than those produced before. There is a decline in the best guess (defined as base case, reference case, or an average between high and low values) from 1.5 to 1.2%/year. There is also a decline in the variance of the best guess. That is, there is a much closer agreement among studies as to expected emissions. Except for Edmonds and Reilly (1983b), all of the best-guess estimates published after 1982 involve CO_2 emissions growth rates between 0.0 and 1.6%/year. The opposite side of that observation is that there is greater CO_2 forecast variation in studies conducted prior to 1982. Within 2 years, both the JASON (1979) study with its estimated growth rate of 4.5%/year and the Lovins et al. (1981) study with its −2.0%/year decline rate were published.

Some, but not all, of the convergence in emissions forecasting is counterbalanced by a greater emphasis on the uncertainty surrounding estimations (see Section 9.4). Both Nordhaus and Yohe (1983) and Edmonds et al. (1985) develop uncertainty analyses that systematically explore the bounds of feasible emissions scenarios. Niehaus and Williams (1979) and Siegenthaler and Oeschger (1978) both explored broad ranges of emissions trajectories with conventional scenario analysis.

9.3 UNCERTAINTY: THE UNAVOIDABLE ISSUE

Uncertainty in future fossil fuel CO_2 emissions forecasting is unavoidable. It has two sources: (1) our inability to describe adequately the behavior of our socio-political-geophysical-economic system and (2) the

choices available to policymakers. Although research can reduce uncertainty of the first type, it cannot reduce uncertainty of the second type.

Behavioral forecast modelers have tended to lump both of these types of uncertainty together, whereas feasibility studies have attempted to explore the options available to policymakers (see Sections 9.5 and 9.6). Some forecast studies have addressed the issue directly either by constructing alternative scenarios or by using the more sophisticated techniques of Monte Carlo analysis. Both of these techniques focus on uncertainty in the assumptions that go into models. Neither is capable of addressing uncertainty that arises from the model structure itself (i.e., the basic description of the energy-economic system). This is a deficiency, but one which appears to have no clear remedy.

We next examine three measures of uncertainty in future CO_2 emissions, none of which is entirely adequate: variations among forecast scenarios, scenario analysis, and formal uncertainty analysis.

9.3.1 Variation Among Forecast Scenarios

One way to measure forecast uncertainty is to measure the variation among forecasts. Graphically this is done in Figure 9.1, where not only the best guess but also the extreme scenarios for 25 forecasts are drawn. There is a clear downward drift over time, with scenarios forecasting emissions rate growth in excess of 4%/year disappearing entirely after 1982, and scenarios that include actual declines in the emissions rate becoming slightly more frequent.

These distinctions are small, however, compared with the overall uncertainty described by the various scenarios. Pre-1982 annual growth rate scenarios ranged from 4.5 to -2.0%. Post-1982 scenarios ranged from 3.5 to -1.5%.

Forecast emissions rates for the year 2050 vary from 127 x 10^{15} g C/year (JASON 1979) to 1 × 10^{15} g C/year (Lovins et al. 1981) before 1982. In the post-1982 period, these rates vary between 64 × 10^{15} g C/year (Edmonds et al. 1984) and 1.5 × 10^{15} g C/year (Edmonds et al. 1985).

Best guesses, as was noted earlier, show a decline in variance over the period, but variation in best-guess forecasts is not a good measure of the variation in the actual population. The best guess is more akin to the mean of a sample of future events than to a sample future event itself.

Furthermore, there are several reasons why these estimates may lead to a greater sense of confidence than is warranted by our actual ability to predict the future. These are discussed in Section 9.9.

Thus, we find the conclusion reached by Ausubel and Nordhaus (1983) that, "based on a review of past projections, . . . CO_2 emissions [50 years hence] show a range of a factor of 3 or more," (p. 155) and the conclusions of Keepin (in press) that "an upper bound for carbon emissions in 2050 of approximately 20 [× 10^{15} g C/year]" (p. 89) and "a lower bound for 2050 of approximately 2 [× 10^{15} g C/year]" (p. 91) may be somewhat optimistic.

9.3.2 Scenario Analysis

The multiple scenario approach to the exploration of uncertainty results in a variety of scenario bounds based on a variety of criteria. Some show extreme ranges; for example, Siegenthaler and Oeschger (1978) examine emissions rate estimates ranging from 5 × 10^{15} to 90 × 10^{15} g C/year in 2050 and Edmonds et al. (1984) produce emissions scenarios ranging from 7 × 10^{15} to 47 × 10^{15} g C/year. Others (e.g., Haefele 1981) restrict their attention to a relatively narrow range of futures, 10 × 10^{15} to 17 × 10^{15} g C/year in the year 2030. In some cases, explicit subjective probability assessments accompany the studies; for example, Haefele (1981, p. 424) states that "the range was selected such that the expected future value is more likely within the range than outside it"; and Edmonds et al. (1984, p. 5) hold that "the likelihood that carbon emissions from cases A and C represent satisfactory upper and lower bounds is high in our judgment—certainly greater than a 50% probability and perhaps a level as high as 95%."

The diversity of ranges is sufficiently broad that in some cases there is no overlap at all. For example, the Keeling and Bacastow (1977) range of 18 × 10^{15} to 50 × 10^{15} g C/year for the year 2025 does not coincide at all with the Haefele range for the year 2030, or the Rose et al. (1983) range for 2025 of 3 × 10^{15} to 10 × 10^{15} g C/year. Some of these differences are rooted in divergent objectives. For example, Rose et al. (1983, p. 35) state that "we wanted to test the sensitivity of CO_2 outcomes to uncertainties in the price and availability over time of a variety of fossil and nonfossil supply technologies." Other differences represent a real divergence of perceptions about the future.

In general, individual scenario exercises do not reflect nearly as much uncertainty as do the entire set of scenarios or as that reflected in formal uncertainty analysis studies.

9.3.3 Formal Uncertainty Analysis

Two recent studies, Nordhaus and Yohe (1983) and Edmonds et al. (1985), explicitly examined the issue of uncertainty in future fossil fuel CO_2 emissions arising from uncertainty in model parameters and assumptions. Nordhaus and Yohe (1983) describe the method as follows:

> The analysis presented here attempts to recognize the intrinsic uncertainty about future economic, energy, and carbon cycle developments . . . by specifying the most important uncertain parameters of the model, by examining current knowledge and disagreement about these parameters, and then by specifying a range of possible outcomes for each uncertain variable or parameter. The emphasis is not to resolve uncertainties but to represent current uncertainties as accurately as possible and integrate them into the structure in a consistent fashion. The result of the entire process is the generation of a range of paths and uncertainties for major economic, energy, and carbon dioxide variables—projections of not only a "best guess" of the future paths of important variables but also a set of alternative trajectories and associated probabilities that quantify the range of possible outcomes on the basis of the current state of knowledge about the underlying uncertainty of the parameters. (p. 88)

The 5th and 95th probability bounds on future CO_2 emissions for the year 2050 obtained by Nordhaus and Yohe were 5×10^{15} and 26×10^{15} g C/year, respectively.

Edmonds et al. (1985) use essentially the same technique but with a different model and hence a different parameterization. In addition, the review of uncertainty developed for this study resulted in different descriptions of uncertainty for model parameters. Edmonds et al. obtain 5 and 95% probability bounds for year 2050 CO_2 emissions of 1.5×10^{15} and 64×10^{15} g C/year, respectively. In an exercise to test the effect of parameter interactions not captured by the model, a set of parameter interrelationships was assumed. These had the effect of changing the bounds to 2.3×10^{15} and 58×10^{15} g C/year, respectively, as well as of reducing the median emissions rate from 10.7×10^{15} to 7.7×10^{15} g C/year. This compares with a median of 15×10^{15} g C/year for Nordhaus and Yohe (1983).

Thus, these two studies indicate that uncertainty in future CO_2 emissions stemming from uncertainty in assumptions alone is between a factor of 5 and 40 by the year 2050. Both of these studies are predicated on the assumption that the underlying model is an accurate representation of the system and in fact is known with certainty. Each study uses a different model, and in Section 9.4 it is shown that model differences lead to markedly different conclusions regarding the importance of assumptions.

No methodology currently exists that allows an analyst to incorporate both uncertainty about model assumptions and uncertainty about model structure simultaneously. Assessments of uncertainty in model structure have been conducted by the Energy Modeling Forum. Their approach is to run each model with a common set of assumptions and to compare outputs. Differences in outputs are attributed to differences in model structure. This technique is not entirely satisfactory, however. First, model structures vary as do the input definitions. Thus, getting a common set of input assumptions may require some translations to the specific frames of reference used by various modelers. To the extent that truly different model structures are included, common terms of reference may disappear altogether. Furthermore, the exercise is conducted around a specific base scenario and, therefore, is indicative only of output variation resulting from model variation at that particular set of assumptions. A different set of base assumptions could yield very different results. Finally, there is no reason to believe that any model is correct or that the variation across models accurately reflects uncertainty in the underlying future structure of the energy-economy system.

9.4 FORCING FUNCTIONS

Forcing function is a term used to refer to the key factors of analysis—those elements whose influences on the numerical outputs of an analysis are so powerful as to override all but the most extreme values of other factors. Such overriding factors are said to force the results. Robert Gardner[1] has observed from his work in analyzing numerous models in diverse disciplines that even larger highly parameterized models generally contain only a small number (perhaps three or four) forcing functions. Until recently the identification of forcing functions was difficult, relying on subjective assessments. A few studies conducted sensitivity analyses (e.g., Edmonds et al. 1984), and a few, notably Nordhaus and Yohe (1983) and Edmonds et al. (1985), conducted formal uncertainty analyses. In other cases, users are left unsure as to exactly what drives a given analysis result. Although this observation seems to indicate that not much is known about forcing functions, the fact that nine studies used the IEA/ORAU model and eight studies used either a straight time-trend or

[1]Robert Gardner, Oak Ridge National Laboratory, personal communication, 1985.

Table 9.2
Indices of Sensitivity of Atmospheric Concentration in 2100 to Uncertainty About Key Parameters

Parameter[a]	Marginal Variance from Full Sample[b] (2)	Marginal Variance from Most Likely Outcome[c] (2)
Ease of Substitution Between Fossil and Nonfossil Fuels [1/(r - 1)]	100	100
General Productivity Growth [A(t)]	76	79
Ease of Substitution Between Energy and Labor [(1/q − 1)]	56	70
Extraction Costs for Fossil Fuels [g1]	50	56
Trends in Real Costs of Producing Energy [$h_1(t)$]	48	73
Airborne Fraction for CO_2 Emission [AF(s)]	44	62
Fuel Mix Among Fossil Fuels [Z(t)]	31	24
Population Growth [L(t)]	22	26
Trends in Relative Costs of Fossil and Nonfossil Fuels [$h_2(t)$]	(3)[d]	21
Total Resources of Fossil Fuels [R]	(50)[d]	5

Note: 100 = level of effect of most important variable. Value of sensitivity is scaled at 100 for the variable that has the highest marginal variance.

[a]Notations in brackets refer to the variable definitions in the model presented formally in Section 2.2. of Nordhaus and Yohe (1983).

[b]Marginal variance from full sample equals (1) the variance in the base case (i.e., with all varying according to their full range of uncertainty) minus (2) the variance with listed variable set at its most likely value (but all nine other variables varying according to their full uncertainty). Note that no resampling occurs.

[c]Marginal variance from most likely outcome calculated as the variance when the listed variable assumes its full range of uncertainty and all nine other variables are set equal to their most likely value.

[d]Parentheses indicate that the marginal variance is negative.

Source: Nordhaus and Yohe (1983), p. 96.

logistics curve model means that a total of 18 (including Nordhaus and Yohe [1983]) of the 25 studies cited quantitatively identified forcing functions.

9.4.1 A Comparison of the Nordhaus-Yohe and Edmonds et al. Results

The only two quantitative analyses conducted to date are Nordhaus and Yohe (1983) (NY) and Edmonds et al. (1985) (ERGB). Results of the uncertainty analyses of NY and ERGB are shown in Tables 9.2 and Table 9.3, respectively. Because the models and associated taxonomies and structure are different, direct comparison of results is difficult; however, one can make some observations.

First, labor productivity in the developing regions is first in importance in the ERGB analysis, with productivity in developed regions ranking third or fourth. The model in NY has a single world region; thus, one might expect productivity to rank first. However, it ranks second, and two factors appear likely to lead to this result: (1) this study has a considerably broader distribution on labor productivity than ERGB (see Edmonds et al. 1985, pp. 64-69) and (2) the NY model formulation implicitly aggregates labor and energy productivity growth in the labor productivity parameter. Energy productivity is similar in concept to energy efficiency in end use in the ERGB model. The impact of combining these two concepts in one parameter is equivalent to assuming perfect correlation between them. Because greater labor productivity alone would tend to increase output and energy demand (and CO_2 emissions) and improved energy productivity would tend to reduce energy demand (and CO_2 emissions), perfect correlation between these would clearly narrow the distribution from the no-correlation case.

The parameter that ranks first in the NY study is the price-induced interfuel substitution parameter. Here, the model specifications are actually quite similar in most applications. In addition, the Monte Carlo analysis in this study is conducted such that perfect correlation among regional values of the parameter i-

Table 9.3
Comparison of Parameter Rankings in Explaining CO_2 Emissions, Monte Carlo Versus Sensitivity Analysis

Year 2000			Year 2025		
Parameter	Monte Carlo Rank	Sensitivity Rank	Parameter	Monte Carlo Rank	Sensitivity Rank
32	1 (.21)	1 (.19)	32	1 (.27)	2 (.20)
12	2 (.17)	5 (.08)	12	2 (.17)	4 (.08)
31	3 (.17)	3 (.15)	9	3 (.14)	1 (.31)
9	4 (.07)	2 (.19)	31	4 (.12)	5 (.06)
2	5 (.04)	0	11	5 (.02)	8 (.02)
62	6 (.03)	0	62	6 (.02)	12 (.01)
7	7 (.02)	4 (.13)	7	7 (.02)	6 (0.06)
11	8 (.02)	8 (.03)	2	8 (.01)	0
8	9 (.01)	6 (.04)	60	9 (.01)	0
61	10 (.01)	0	8	10 (.01)	10 (.01)
1	0	7 (.04)	14	0	3 (.11)
79	0	9 (.02)	26	0	7 (.06)
3	0	10 (.01)	1	0	9 (.02)
Year 2050			**Year 2075**		
Parameter	Monte Carlo Rank	Sensitivity Rank	Parameter	Monte Carlo Rank	Sensitivity Rank
32	1 (.23)	2 (.23)	32	1 (.20)	2 (.29)
12	2 (.19)	3 (.10)	12	2 (.16)	3 (.10)
31	3 (.14)	4 (.08)	31	3 (.15)	4 (.06)
9	4 (.12)	1 (.36)	9	4 (.08)	1 (.04)
62	5 (.02)	0	62	5 (.04)	11 (.01)
11	6 (.02)	8 (.02)	60	6 (.02)	0
7	7 (.02)	6 (.05)	2	7 (.02)	0
2	8 (.01)	0	7	8 (.01)	5 (.04)
60	9 (.01)	0	11	9 (.01)	7 (.02)
4	10 (.01)	0	59	10 (.01)	0
15	0	5 (.05)	4	11 (.01)	0
27	0	7 (.04)	16	0	6 (.02)
77	0	9 (.01)	77	0	8 (.02)
1	0	10 (.01)	28	0	9 (.01)
			22	0	10 (.01)

Note: Results are based on log-transformed data because they provided a better fit (as indicated by R^2). Ranking of variables remained unchanged among the most important variables between the nontransformed and log-transformed data. Relative partial sums of squares in parentheses.

Key to Table 9.3

Parameter	Definition
1	Nonfuel cost of electric
2	Biomass cost
3	Interfuel substitution (final demand)
4, 5, 6	Energy/GNP feedback OECD, other, MIDEAST
7, 8, 9	Income elasticity OECD, USSR, other
10	Interfuel substitution (utilities)
11	Aggregate energy price elasticity
12	Energy efficiency in end use
13	[a]
14, 15, 16, 17, 18	Population OECD & USSR, 1975-2075
19	[a]
20, 21, 22, 23, 24	Population China, 1975-2075
25	[a]
26, 27, 28, 29, 30	Population all other regions, 1975-2075
31, 32	Labor productivity OECD + USSR, all other regions
33	Oil supply elasticity
34, 35, 36, 37, 38	Conventional oil resource, OECD, grades = 1,5
39, 40, 41, 42, 43	Conventional oil resource other, grades = 1,5
44, 45, 46 47, 48	Conventional oil resource MIDEAST, grades = 1,5
49, 50, 51, 52	Natural gas resource global, grades = 1,4
53	Natural gas supply elasticity
54, 55, 56, 57, 58	Coal resource, grades = 1,5
59	Coal supply elasticity
60	Technological change, coal
61, 62	Environmental costs, coal
63, 64, 65, 66	Shale oil resource, grades = 1,4
67	Shale oil supply elasticity
68	Technological change, shale oil
69	Environmental costs, shale oil
70, 71, 72	Nuclear power resources, grades = 1,3
73	Environmental costs, nuclear
74	Technological change, nuclear
75	Solar electric cost, year 2000
76	Technological change, solar[a]
77, 78	Synfuel costs
79	Hydroelectric resource base

[a]Not model implemented.
Source: Edmonds et al. (1985).

assumed, thus making it identical (for Monte Carlo analysis purposes) to assuming a single-world region. These similarities make it surprising that the ERGB substitution parameter does not rank at all in any year. The major reason for this result is probably the fact that four secondary fuels (electricity, gases, liquids, and solids) as well as new primary energy types and two major conversion processes exist in the IEA/ORAU model. In contrast, in the NY study, there are two fuels, fossil and nonfossil. Suppose the price of fossil energy rises in the NY model; obviously the ability to substitute nonfossil energy (as represented by elasticity) will significantly affect carbon emissions. If substitution is very easy, demand will fall greatly for fossil and rise for nonfossil. If substitution is difficult, a considerably smaller change in fossil use will occur. Of course, reductions in output (GNP) will somewhat counter these direct effects, but it is not surprising that the substitution effects significantly outweigh the output effects.

Suppose the price of a fossil fuel (e.g., coal) increases in the IEA/ORAU model. First, this shifts the mix of solids toward biomass before the interfuel substitution elasticity (final demand) can have any impact. Next, consumers respond by switching to fuels other than solids (i.e., liquids, gases, or electricity). But even if end-use consumers switch completely from solids, the impact on carbon emission is indeterminate. They could switch to liquids, possibly shale oil or coal liquids, which would increase carbon emissions, or possibly to biomass liquids, which would decrease emissions. Consumers also might switch to gas, but for reasons similar to those for oil, this switch could increase or decrease emissions. Similarly, electricity is generated by both fossil and nonfossil sources. Of course, the rise in coal prices would tend to reduce the average emissions ratio per unit of electricity, but this ratio might still be higher than direct use of solids. Thus, because it is likely that the average emissions rates for end-use fuels are not that different, it is not surprising that whether fuel substitution is easy or not makes little difference.

Presumably it would make a difference in ERGB if all fossil fuel prices were strongly correlated and all nonfossil fuel prices were strongly correlated, but prices for fossil and nonfossil were uncorrelated. This would essentially be the assumption regarding the two fuels in the NY model. If fossil fuel was the paradigmatic exhaustible fuel and nonfossil fuel the inexhaustible fuel, this might lead to the necessary set of correlations. ERGB, by explicitly including exhaustibility, provided a test of the importance of exhaustion in generating such a correlation and, as is obvious by the findings, the test was negative.

The other ERGB parameter that is important is the income elasticity of energy demand in developing countries. The NY model does not include a comparable concept. In the NY production function formulation, an income elasticity would be similar to a return-to-scale parameter; by omission, it is implicitly constant at unity.

Another relatively important parameter in the NY results is the labor-energy tradeoff. This is very similar to the ERGB aggregate energy elasticity parameter, where the relative partial sum of squares is fairly small, and the parameter was clearly less important than in NY. This is probably a result of the relative nature of the measures of importance rather than one of an absolute difference. Because ERGB gave a broader range of uncertainty to labor productivity, all other parameters became relatively less important. Similarly, income elasticity is a significant source of variability in ERGB, but is omitted in NY, thus reducing the relative importance of other parameters in ERGB.

9.4.2 Other Models

The International Institute of Applied Systems Analysis (IIASA) model was assessed by Keepin (1983). In his critique, he identifies LP constraints as the forcing functions of the IIASA models. Although at first blush, this would appear to be helpful, it leaves totally unanswered the underlying analysis that went into determining the value of the constraints and therefore the true forcing functions.

The logistic curve models are similarly frustrating. They are extremely simple models for which either total carbon emissions over all time or total fossil fuel consumption over all time appear as a model input. This is also true of CO_2-ceiling models–models in which a ceiling is placed on total CO_2 concentrations as a condition of the analysis. For such models, the forcing function on total carbon emissions is total carbon emissions. Where fossil fuel consumption is used, the problem is the same, because the fossil fuel consumption concept is different from the resource exhaustion concept. As Edmonds and Reilly (1985) demonstrate, the total fossil fuel resource base is simply too enormous to be consumed in its entirety. Arbitrary chunks are therefore eliminated on the basis that the price of energy will not rise sufficiently to make them economical. Thus, although resource exhaustion is the underlying paradigm, considerable statistical liberties must be taken to make the model operational. Thus, as models of total CO_2 release they are not very interesting.

The timing of CO_2 release in logistics curve models is another matter. In the simplest specification, this is determined by the initial rate of emission and the date of peak emissions. The pure time-trend analysis model is similarly unenlightening. Here the assumed rate of growth of the rate of carbon emissions and elapsed time uniquely determines the rate of emissions.

9.5 POLICY ISSUES

In most studies, policy has not been the principal focus of analysis. Assessments of the likely course of business-as-usual energy use and carbon emissions are the featured elements of most analyses. The lack of concentration on policy issues is, in part, due to the great uncertainty surrounding all of the basic issues in emissions. The few studies conducted on market equilibrium divide nicely into either feasibility studies or traditional behavioral forecasts (see Section 9.6).

Feasibility studies take a policy objective as given and study the feasibility of meeting that policy objective. Behavioral forecast models begin with a policy tool and examine the consequences of that tool's implementation.

9.5.1 Behavioral Models and Policy

Several studies fall into this category (Edmonds and Reilly 1983b; Seidel and Keyes 1983; Nordhaus and Yohe 1983). In some ways this type of analysis makes fewer demands on the researcher than do feasibility studies because the difficult part of the analysis is already accomplished (i.e., the specification of the model [the relationship between assumptions and CO_2 emissions] and the baseline set of input assumptions). All three studies explore the consequences of punitive policy actions (taxes) designed to shift the relative costs of fossil and nonfossil fuel production and consumption.

Table 9.4
Effects of CO_2 Taxes on Emissions

Study	Forecast Year	Tax[a] Rate	Note	CO_2 Emissions	Growth[b] Rate (%)
Edmonds and Reilly (1983)[c]	2050	0%	Base case	26	2.2
	2050	50%	U.S. only	26	2.2
	2050	100%	U.S. only; Coal export ban	22	2.0
	2050	100%	Global; coal Export ban	16	1.6
Seidel and Keyes (1983)[d]	2100	0%	Base case	50	1.9
	2100	100%	U.S. only; coal export ban	43	1.7
	2100	100%	OECD only; coal export ban	40	1.7
	2100	100%	Global; coal export	28	1.4
	2100	n.a.	Global coal ban	22	1.2
	2100	n.a.	Global shale ban	20	1.1
	2100	n.a.	Global shale and synfuel ban	17	1.0
	2100	n.a.	Global coal and shale oil ban	0	-100.0
Nordhaus and Yohe (1983)	2100	0	Base case	20	1.1
	2100	$20[e]	1980 global pulse tax	19	1.1
	2100	$20[e]	2025 global pulse tax	20	1.1
	2100	$10[e]	1980 permanent tax	20	
	2100	$10[e]	2025 permanent tax	19	1.1
	2100	$90[e]	Rises from year 2000 to 90 by 2050; it is equivalent to an approximately 60% tax	14	0.8

[a]Taxes are applied to carbon with coal as the reference fuel.
[b]Growth rate of CO_2 emissions 1975 to forecast year.
[c]Policies assumed to take effect in the year 1985.
[d]Policies assumed to take effect in the year 2000.
[e]Per metric ton coal equivalent.

All three show remarkably similar results (note that Seidel and Keyes [1983] use the same model as Edmonds and Reilly [1983b]; see Table 9.4).

In general, taxes appear to be a relatively poor method for curtailing CO_2 emissions. To be effective, they must be high (above 100%) and globally administered (Edmonds and Reilly 1983b; Seidel and Keyes 1983). No single region accounts for a sufficient share of either present or forecast CO_2 emissions to effectively control the global total. In addition, the "free-rider" problem dilutes the effects of individual regions acting separately because the absence from consumption by one region reduces the demand for carbon-based fuels, thus lowering the world price and making these fuels more attractive to other consumers.

Only stringent actions taken in concert by the global community have significant effects on atmospheric carbon accumulation. This is due partly to the fact that even if the rate of emission is not growing, there remains a steady stream of carbon into the atmosphere.

Seidel and Keyes (1983) examined the effect of a global prohibition of fossil fuel use, implemented in the year 2000. Such policies are extreme and equivalent to high fuel-specific tax rates. However, only the outright banning of both coal and shale oil fuel was sufficient to terminate CO_2 emissions by the year 2100. The continued significant emissions of CO_2 result from the fact that interfuel substitution occurs between fossil fuels as well as between fossil and nonfossil fuels.

Rose et al. (1983) examine a pluralistic set of punitive and positive policy options using the Edmonds and Reilly (1983a) IEA/ORAU model. They emerge optimistic about the prospects for policy intervention and emphasize:

1. early action;
2. efforts to accelerate improvements in end-use energy efficiency;
3. international cooperation; and
4. support for research and development of nonfossil fuel energy alternatives, in particular, solar photovoltaic systems.

The focus on end-use energy efficiency is a qualitatively different recommendation than that for other behavioral forecast studies. The potency of this recommendation is supported by the sensitivity and uncertainty analysis studies cited earlier, particularly Edmonds et al. (1985).

9.5.2 Feasibility Studies

Many more studies fall into the feasibility category and include Nordhaus (1979), Lovins et al. (1981), Perry et al. (1982), Kosobud et al. (1983), and Williams et al. (1984). These in turn can be subdivided into two additional categories: optimization studies that explore optimal strategies for meeting an exogenous but specified constraint (Nordhaus 1979; Kosobud et al. 1983), and feasibility studies that explore the potential of low CO_2 emissions futures but which make no claim that the path they put forward is necessarily optimal.

All feasibility studies that put a ceiling on eventual CO_2 accumulation in the atmosphere are forced to specify the link between CO_2 emissions and atmospheric carbon concentrations; that is, they require the addition of a carbon cycle model of some sort to the analysis. All of the carbon cycle models used in these studies are primitive by current standards. They range from simple constant airborne fraction models (Williams et al. 1984) to a seven-layer, well-mixed box model (Nordhaus 1979). This additional layer of analysis adds to overall uncertainty.

Kosobud et al. (1983) and Nordhaus (1979) use LP models to assess the consequences of various constraints on future CO_2 concentrations and to derive "optimal" tax rates.

The Nordhaus (1979) model is one of the oldest economic-energy-CO_2 emissions models in use. In this model the rate of emissions peaks in all constrained scenarios with tighter constraints, implying earlier and lower peaks.

Constraint (Percentage Increase Over 1975)	Year of Peak Emissions Rate
50	2000
100	2020
200	2040

The optimal tax rate for CO_2 is displayed in Table 9.5. In all cases, the tax rate rises to a peak and then either plateaus or declines slightly, and taxes peak at a later date than do the emissions. Only in the 50% increase case does the rate of emissions diverge from the control case before the year 2000, implying that policy actions, if necessary, need not be implemented immediately.

Table 9.5
Optimal Tax Rates for CO_2 Emissions (units = 1975 dollars per 10^9 g C)

	Cases							
	Nordhaus (1979)				Kosobud et al.[a] (1983)			
Year	Uncontrolled	200% Increase[b]	100% Increase[b]	50% Decrease[b]	Uncontrolled	100% Increase[b,c]	100% Increase[b,c] Plus Nuclear Ban
1980	0	0.00	0.14	1.65	0	0.59	1.18
2000	0	0.07	1.02	12.90	0	7.86	15.33
2020	0	0.52	8.04	109.00	0	84.90	162.13
2040	0	4.07	67.90	123.60	0	749.93	1344.22
2060	0	34.47	94.40	200.00	n.a.[d]	n.a.	n.a.[d]
2080	0	42.00	94.40	200.00	n.a.	n.a.	n.a.
2100	0	42.04	87.20	198.20	n.a.	n.a.	n.a.
2120	0	41.91	87.10	198.50	n.a.	n.a.	n.a.
2140	0	42.92	86.90	188.40	n.a.	n.a.	n.a.

[a]Kosobud et al. appear to assume that their ceiling is reached by the year 2040 and that there are either no further emissions or the ceiling is lifted.
[b]Increase assumed relative to a 1974 base of approximately 330 ppm atmospheric concentration.
[c]Costs in Kosobud et al. are computed from the tax on coal; prices are deflated from an assumed 1980 base to 1975 constant dollars.
[d]n.a. = not applicable.

Although also interested in analyzing optimal global tax strategies for controlling atmospheric CO_2 accumulations, the Kosobud et al. (1983) paper is less ambitious than that of Nordhaus (1979). Their model is extremely crude. Only the U.S. economy is modeled in any detail, using the ETA-MACRO model described in Manne (1981). In the United States, CO_2 emissions were constrained to be no more than 20% (140×10^{15} g C) of the global total additions ($\sim 700 \times 10^{15}$ g C). The assumption of 20% was adopted with reference to Haefele (1981), and implies that the rest of the world acts appropriately to maintain its net addition to the carbon pool below 560×10^{15} g C. No account is taken of the interaction between energy production and use in the United States and similar activities in the rest of the world. The relationship between emissions and atmospheric retention is not specified in the report so it is impossible to comment on their carbon cycle model assumptions. Finally, it is not clear whether the entire budget of 700×10^{15} g C is used up over the planning horizon, with world carbon release then going to zero, or whether post-2040 emissions are assumed. In any case, no discussion of either this point or the transition problem appears in the report.

Before the year 2040, the Kosobud et al. (1983) optimal tax rates for CO_2 (Table 9.5) for the 100% increase compare well with the Nordhaus (1979) 50% increase, raising the possibility that the airborne fraction used might be too high, perhaps by a factor of 2. Year 2040 taxes are enormous, however, raising the question of whether this might be a result of the frequently documented end-period problems of LP models.

We identified three other studies that explore (1) the feasibility of changing global energy systems to satisfy alternative ceilings on atmospheric CO_2 concentrations (Lovins et al. 1981; Perry et al. 1982; Williams et al. 1984) and (2) the feasibility (though not optimality) of meeting exogenous CO_2 accumulation constraints.

The earliest study (Lovins et al. 1981) is based on a detailed study from the Federal Republic of Germany, undertaken to assess the potential for energy conservation in the year 1973. Results were generalized to the remainder of the world and extrapolated forward to the year 2030 using population and economic assumptions borrowed from the IIASA scenario reported by Colombo and Bernadini (1979). Lovins et al. (1981) conclude that not only would it be possible to reduce CO_2 emissions to almost zero within 30 to 50 years, but there are strong economic pressures that, if market imperfections are removed, will cause it. They discuss the importance of end-use energy in the determination

of future energy consumption and CO_2 emissions, and argue for the adaptation of policies that would be undertaken regardless of the outcome of research on CO_2. Lovins et al. (1981) foresee energy efficiency improvements as being capable of lowering energy demand to a level which could be supplied by small decentralized energy supply systems with CO_2 benign technologies (such as hydroelectric power).

Perry et al. (1982) explore the interaction of ceilings on atmospheric CO_2 concentrations, business-as-usual forecast scenarios, and the timing of deviations from the forecasts to meet a constraint. They refer to the latter concept as Action Initiation Time (AIT). Two constraints (500 and 600 ppm) and two base cases (the IIASA high and low scenarios; Haefele 1981) are discussed. The construction of fossil fuel use scenarios is simple, employing time-trend and logistic-curve functions to sketch out the time path. They conclude that fossil fuel use will need to peak between the years 2010 and 2040. Rigorous actions must begin within the next 10 to 20 years to meet a 500-ppm ceiling; a 20- to 30-year grace period is consistent with the 600-ppm ceiling. They comment on the potential role of efficiency in energy use in extending the AIT and reducing CO_2 emissions.

Williams et al. (1984) examine two cases, a high and a low (base), ranging in energy consumption from 337 $\times 10^{18}$ to 467 $\times 10^{18}$ J in the year 2020. They set a target constraint of 490 ppm on the global system and transform that to a constraint on fossil fuels by assuming an airborne fraction of 0.50. Assuming biomass to be the swing fuel, they explore the consequences of alternative population assumptions (low, mid, and high United Nations population projections), less energy-efficiency improvement than in the base case, and CO_2 ceilings as low as 400 ppm and as high as 575 ppm. They conclude that a high population growth rate and reduced rates of energy-efficiency improvements would place greater pressure on biomass supply (the only sector allowed to respond). Only the lowest CO_2 ceiling examined, 400 ppm, caused any significant increase in pressure on the biomass sector and that was relatively minor (comparable to the midrange population assumption). Nuclear energy was considered to be a supply source of last resort. The general conclusion reached was that low CO_2 ceilings could be satisfied if the end-use energy efficiencies improved sufficiently. The latter was deemed possible, given currently available technologies.

9.5.3 The Cost of Abatement Policies

The economic costs of CO_2 abatement policies hav only been partially analyzed at this time. Edmonds and Reilly (1983b), Kosobud et al. (1983), and Nordhaus (1979), each using a different model, indicate that the reduction in aggregate GNP associated with even stringent punitive strategies is not large, usually only a few percentage points. Lovins et al. (1981) argued that the costs might actually be negative.

Seidel and Keyes (1983) took the analysis further to assess the impacts of abatement strategies on subaggregate factors. The costs and benefits of hypothesized policies were found to be spread disproportionately; regional and subregional differences might arise between the developed and developing regions and within resource-producing and resource-consuming regions. Fossil energy producers would find the value of their resource base and capital stock either reduced or wiped out completely. Laborers would have to shift from fossil to nonfossil energy sectors. Fossil-fuel-using capital equipment would have to be either gradually retired or scrapped, depending on the program.

9.6 METHODOLOGIES

The forecasts documented in Table 9.1 and Figure 9.1 were developed from a diversity of modeling techniques. Two major methodological classes, behavioral and feasibility studies, are distinguished, and these are subclassified by the particular modeling strategy employed in Table 9.6.

We define a behavioral study as one that explicitly seeks to forecast the future as a consequence of assumptions. Behavioral models are by nature deterministic (though two have been used in conjunction with Monte Carlo analysis which assumes that *inputs* are uncertain). We appended the adjective, behavioral, to the term to distinguish the group and perspective from the feasibility study approach to forecasting.

Feasibility studies take another perspective. They ask, given a set of exogenously specified conditions (e.g., global population, economic growth), is it possible to reach a goal (e.g., keep the atmospheric CO_2 concentration below an assumed level)? This method by definition is nondeterministic. In principle, it is not well suited to forecasting the future. But practitioners of the method argue otherwise, and the distinction between forecast and assumed constraint can become a murky one.

Table 9.6
Fossil Fuel CO_2 Emissions: Forecast Methodologies

Models Class[a]	Subclass	Modelers[b]	Methodology	Advantages	Drawbacks	Use	
						Studies[c]	Forecasts[d]
B	Time trend	JASON (1979)	1. Simple exponential time trend extrapolated to the future. 2. Equation: $y = ae^{rt}$ or $y = a(1+r)^t$, where y is rate of global CO_2 emissions at time t; a is base rate of emissions at time $t = 0$; r is rate of growth of CO_2 emissions. 3. Though generally based on recent trends, it is a parametric equation allowing examination of alternative rates. It is typically used as an input to complex natural science models.	1. Simplicity. Time is the only variable. 2. Useful as a gross summary of a more detailed analysis.	1. Not a tool of analysis. No geographic detail. No energy detail. No economic, demographic, technical, or policy detail. 2. Cannot replicate complex time paths for emissions.	JASON (1979) Bacastow-Keeling (1981) Sundquist-Plummer (1981)	4.3 4.3 3.9
B	Logistics	Keeling-Bacastow (1977) Rotty (1979)	1. Deterministic model that allocates a pool of carbon over time. The distribution is characterized by an initial period of increasing emissions that peak at a predetermined date and then steadily decline thereafter. It is symmetric with respect to time about the peak date. 2. Key assumptions: • Initial rate of growth of CO_2 emissions • Date at which the rate peaks	1. Simplicity. 2. Allows more complex time paths to be described than simple time trends. 3. Useful as a gross summary of a more detailed analysis.	1. Not an analytical tool: • No geographical detail • Unable to analyze or describe fundamental demographic, social, economic, technical, or policy phenomena governing future global fossil fuel CO_2 emissions.	Keeling-Bacastow (1977) Siegenthaler-Oeschger (1978) Rotty (1979) Rotty-Marland (1980)	2.8 2.0 2.9 1.6

(Continued)

Table 9.6 (continued)

Models Class[a]	Subclass	Modelers[b]	Methodology	Advantages	Drawbacks	Use: Studies[c]	Use: Forecasts[d]
			• Total pool of carbon • Some specifications add a skewness variable to create asymmetric distributions.		2. Cannot replicate complex time paths of emissions.		
B	Gap studies	Perry-Lansberg (1978) Allen et al. (1981)	1. Energy-economic analysis: • Disaggregation by region and fuel • Forecasts for energy production and use are based on analysis of key factors, including population, economic activity (GNP), and available energy resources. In some cases technological, engineering, and end-use energy disaggregations are included. • Differences between projected production and use are reconciled by assuming a "swing" fuel (e.g., coal or solar energy) provides residual production.	1. Provides more detailed analytical foundation of forecasts than B/time-trend or B/logistics methodologies, including explicit links between GNP, population, energy resources, and future fossil fuel CO_2 emissions. 2. Allows for multiple independent analyses to be brought together (e.g., can take advantage of regional expertise).	1. Does not include energy and fuel price effects. 2. Generally not computer-based and therefore difficult to perform re-evaluations using alternative assumptions.	Perry-Lansberg (1978) Rotty-Marland (1980)	2.8 1.6

Table 9.6 (continued)

Models Class[a]	Subclass	Modelers[b]	Methodology	Advantages	Drawbacks	Use	
						Studies[c]	Forecasts[d]
B	Linear/ nonlinear program-ming[e]	Nordhaus (1979) Haefele (1981) Manne (1981, 1984)	1. Computer-based simulation framework used to model the energy-economic system. 2. The energy-economic system is modeled as a set of mathematical equations describing activities (e.g., energy production, transformation, and use, and constraints or international trade restrictions). An objective, such as maximation of cost, is assumed. The computer solves for optimum rates of energy production and use consistent with the assumed energy-economy equations.	1. The entire energy-economic system and its interactions can be analyzed simultaneously. 2. Gaps between energy production and use are reconciled. 3. Energy costs and prices are explicitly considered. 4. Alternative assumptions can be examined with relative ease. 5. Geographical aggregation is possible.	1. The methodology is extremely sensitive to exogenously specified constraints, thus shifting much of the most interesting analysis outside the formal model. 2. The methodology requires extreme detail. All future techno-logical options and constraints must be explicitly described, an impossible task for forecasts of 50 years into the future. 3. The models are expensive and time-consuming to run and difficult for others to use. An exception to this is Manne's ETA-MACRO model.	Nordhaus (1979) Haefele (1981) Manne (1984) Rogner (1986)* *Rogner used the same IIASA model developed by Haefele (1981).	3.0 1.8 1.4 1.3

(Continued)

Table 9.6 (continued)

Models Class[a]	Subclass	Modelers[b]	Methodology	Advantages	Drawbacks	Use: Studies[c]	Use: Forecasts[d]
B	Market equilibrium	Edmonds-Reilly (1983a) Nordhaus-Yohe (1983)	1. Computer-based simulation framework used to model energy-economic behavior. 2. The energy-economic system is modeled as a set of mathematical equations describing activities (e.g., energy production, transformation, and use). The model describes energy market behavior given exogenous assumption (e.g., population, resource base for energy, technology).	1. The entire energy-economic system and its interactions can be analyzed simultaneously. 2. Gaps between energy production and use are reconciled. 3. Energy costs and prices are explicitly considered. 4. Alternative assumptions can be examined with relative ease. 5. Geographic disaggregation. 6. Models are small enough to be run extensively. 7. Models are transferable to other users.	1. The models tend to be governed by highly aggregated parameters that have no direct physical interpretation (e.g., the rate of interfuel substitution or the rate of labor productivity growth). 2. Little emphasis on linkages between energy and nonenergy activities.	Nordhaus-Yohe (1983)	1.6
						Edmonds-Reilly (1983b)[f]	2.3
						Rose et al. (1983)[f,g]	0.6
						Seidel-Keyes (1983)[e,f]	1.6
						Edmonds et al. (1984)[f]	1.5
						Reister (1984a)[f]	1.1
						Reister (1984b)[f]	1.6
						Chandler (1985)[f,g]	1.6
						Edmonds et al. (1985)[f]	0.7

Table 9.6 (continued)

Models Class[a]	Subclass	Modelers[b]	Methodology	Advantages	Drawbacks	Use	
						Studies[c]	Forecasts[d]
F	Gap studies	Colombo-Bernadini (1979) Lovins et al. (1981) Perry et al. (1982) Williams et al. (1984)	1. Explore the feasibility of reaching a target constraint on CO_2 emissions within the confines of a specific global setting (e.g., population, economic growth). 2. Use same techniques as B/gap studies. 3. Include regional and end-use disaggregations. 4. Differences between projected energy use and constrained energy production are reconciled by assuming a "swing" fuel (e.g., biomass or conservation) will provide the residual.	1. Provides a detailed analysis of the global energy system with regional and end-use detail. 2. Explicit policy orientation. 3. Allows for multiple independent analyses to be brought together (e.g., can take advantage of regional expertise). 4. Provides a challenge to, and a reference for, behavioral forecasters.	1. Studies only the feasibility and not the probability of scenarios. 2. Because each analysis is unique (despite the usual reference to a published scenario), it is difficult to make direct interstudy study comparisons. 3. Blurred distinction between policy goals and future expectation.	Colombo-Bernadini (1979)[h] Niehaus-Williams (1979) Lovins et al. (1981)[h] Williams et al. (1984)[h]	0.7 −2.0 −2.0 0.2

(Continued)

Table 9.6 (continued)

Models Class[a]	Subclass	Modelers[b]	Methodology	Advantages	Drawbacks	Use: Studies[c]	Use: Forecasts[d]
F	Market equilibrium	Rose et al. (1983) Chandler (1985)	1. Examine the impact of alternative, feasible, technological, social, or economic developments on future CO_2 emissions. 2. Use market equilibrium modeling techniques (see B/ market equilibrium).	1. Explicitly incorporates policy/ technical/ social options in the analysis. 2. More clearly distinguishes between policy goals and future expectations than F/gap studies. 3. Incorporates system interactions more clearly. 4. Using established formal models, clarifies area of debate.	1. Studies only the feasibility and not the probability of scenarios. 2. The most important elements of analysis lie outside the bounds of the models used. 3. Because of the predictive nature of the market equilibrium model, it can be difficult or time-consuming to meet an assumed constraint. 4. *Constrained* to the model framework adopted.	Rose et al. (1983)[g] Chandler (1985)[g]	–0.7 0.8

[a]B is behavioral studies; F is feasibility studies.
[b]Researchers who have developed models to forecast future global fossil fuel CO_2 emissions of this class and subclass.
[c]Studies that have employed a model of this class and subclass.
[d]Best-guess forecast. See notes to Figure 9.1. Units = none (annual percentage rate).
[e]This methodology assumes that as an aggregate the energy-economic system operates to ensure some options (the exact optimization criterion is specified by the modeler). See "Methodology" and Section 9.3.3. It, therefore, has a strong conceptual overlap with the feasibility study literature. In practice, however, these studies are much more closely aligned with the behavioral modelers and are therefore included under this category.
[f]Use the IEA/ORAU model (Edmonds and Reilly 1983a) with their assumptions.
[g]Use the IEA/ORAU model as an analytical tool to explore the feasibility of low energy-CO_2 emissions futures; their low case is therefore included under the discussion of feasibility studies.
[h]Use the IIASA forecasts (Haefele 1981) as a base.

To date, research in the area of future CO_2 emissions has focused primarily on the forecast issue. What are future emissions likely to be in the absence of any feedback effects or CO_2-oriented policy initiatives?

Within the ranks of forecasters there has been one notable shift in methodology, the abandonment of time-trend and logistics curve techniques as tools of analysis. These techniques are now used solely as summary devices to capture the conclusions of more complex analysis.

The simple logistics curve is symmetric and asymptotically zero. Rotty (1977) added an additional parameter that allows nonsymmetric exhaustion paths. The curve can be fit to a historical series and, as such, is a time-trend extrapolation. The added feature is a fixed resource defined as the total "economically recoverable" resource amount. These methods, while straightforward and easy to use, are incapable of explicitly incorporating and assessing many of the key issues generally assumed to be critical to carbon emissions (e.g., technological alternatives, population, and GNP growth rates).

While the logistic curve incorporates resource constraints, the general principle—as exhaustion of "economic" grade resources occurs, prices rise and make previously noneconomic resources "economic"—cannot be explicitly incorporated. To define the amount of resources that will be economic over all time, one must define the price path over all time, and therefore one must describe the demand for fuels, which depends on a variety of factors including the availability of substitute energy forms such as nuclear, solar, and hydro power. In general, these methods suffer from being noncausal, and therefore the effects of various elements thought to affect the time-trend rate must be handled subjectively. The problems inherent in this and other extrapolation techniques were emphasized by the radically changed rate of growth in carbon emissions after 1973. Nevertheless, Reister and Rotty (1983b) showed that under special conditions a model based on economic principles yields a logistic CO_2 emissions pattern.

More sophisticated methods have been substituted for the logistics curve and time-trend analysis. These include gap studies, linear and nonlinear programming analysis, and market equilibrium models. All three explicitly link the socio-technical-economic system to energy use. The advantages and disadvantages of these methods are summarized in Table 9.6.

It is interesting to note that all three of these techniques have been applied to both behavioral forecasting or feasibility analysis, a fact that argues for a commonality between the two approaches which transcends their different perspectives.

Feasibility analysis offers a different perspective that has resulted in different views of the future. Feasibility studies have consistently examined lower CO_2 emissions futures than other studies (see Figure 9.1). But the difference between the forecasts of these two future views is diminishing. This is partly the result of the cross-fertilization of ideas within this relatively small community of researchers, or, put less flatteringly, the herd effect.

Lansberg (1982) advocated the feasibility approach as a means of avoiding some of the problems associated with forecasting future events on a long-term global scale. He suggested that:

> One could pitch the inquiry to the question of what levels of CO_2 concentration can be considered critical (e.g., 500, 700, 900 ppm). One could then test what combination of economic growth rates, energy efficiency, and energy source mix would result in these concentrations and in what number of years these combinations might be reached. (p. 365)

This methodology, however, is not as much a way out of the forecasting problems as it is another way in. For a feasibility study one knows that the course to be followed leads from A to B, with the question, "Can you get there from here and if so, how?" But as the Perry et al. (1982) paper demonstrates, it is very hard to answer either of these two questions in a meaningful way without becoming involved with forecasting. To begin with, neither question has any relevance outside the context of some future forecasts. One needs to forecast something just to get started. What populations and GNP are likely to prevail? What in turn do these underpinnings imply about energy service demands?

Once you ask how best to accomplish a task, you are a very short step away from the formal optimization that models of Nordhaus (1979) and Manne (1984) employ. Furthermore, it is unlikely indeed that one set of assumptions will satisfy a target trajectory uniquely. Higher growth rates in population or GNP growth could be offset by either a greater reliance on nonfossil energy sources or increased energy efficiencies, and the distribution of these combinations over geography and time are endless. In the end the task is no less difficult than forecasting.

9.7 THE VALIDATION PROCESS

Researchers in the fossil fuel CO_2 emissions field have not operated in a vacuum. The number of active researchers is small by comparison to those in the other areas of the carbon cycle research, and there are few if any objective validation criteria available to test forecasters' performance. Thus, to a greater extent than in other areas of the carbon cycle, peer judgment is the dominant validation criterion and human interactions play an unavoidable role not only in determining the direction of research but also in determining results. This gives rise to the question, "How does this process affect research results?"

We have identified two areas in which this may be the case. Forecasters may move toward a consensual best-guess forecast simply as an outcome of the validation process available to them, the herd effect. We also note that 75- to 100-year forecasts are strongly tied to experiences in the recent past. Finally, we explore some of the problems found at the interface between models and forecasters, which were highlighted by the recent discussions concerning the IIASA model.

9.7.1 The Herd Effect

The herd effect grows out of the validation process. This process provides checks and balances to both the modelers and the forecast users. Unusual results and assumptions are challenged. Out of this clash of ideas, proponents modify their positions and revise their forecasts. The benefits of this process are obvious. Errors are found and eliminated. New ideas are introduced and vie for acceptance. Unfortunately, none of this has anything to do with objectively validating the accuracy of 75- to 100-year forecasts of global fossil fuel CO_2 emissions. And because this is a consensus-building process, there are professional penalties for being an "outlier." As a consequence, the decline in variance of best-guess forecasts after 1981 may be a side effect of the validation process as much as an indication of improved understanding.

It is also apparent from Table 9.1 and Figure 9.1 that many forecasters have adopted either the IIASA scenarios (Haefele 1981) as a base case for analysis or the IEA/ORAU model (Edmonds and Reilly 1983a). This shift toward common tools of analysis allows researchers to know where they share assumptions and where assumptions differ. Much of the time-consuming process of identifying the real issues for debate is eliminated. At the same time, this expedites the validation process (which is consensus-building), and the common modeling framework eliminates one source of forecast diversity, heterogeneous modeling structures.

9.7.2 Sophisticated Time Trends

The recent downturn in the rate of growth of global fossil fuel CO_2 emissions (Chapter 4) has been accompanied by a downward trend in the best-guess forecast rate of growth of future CO_2 emissions (Figure 9.1; note the exception of feasibility studies from this generalization). Although there is little evidence from which to draw firm conclusions, 75-year forecasts seem strongly influenced by the relatively near-term past.

Feasibility studies have not mirrored the recent past in their forecasts.

9.7.3 The Model-Modeler Interface

Models are built by analysts and used by forecasters. They contain three elements—the model itself, the parameters of the model, and assumptions. The model consisting of its equation structure alone is usually accepted as given. It is the analyst's view of the way the world operates, and is either accepted or rejected. The model's parameters are what ground the model to reality. They are treated as if they were part of the model. This leaves the model's assumptions, which are considered open to general inspection and are discussed openly.

Any but the simplest models, however, have sufficiently many equations, parameters, and assumptions that only a very few of them are ever subjected to a serious external review. Nevertheless, models and forecasters have taken on an aura of magic as if they represented a lens through which to see the future clearly. This occurs both because those who use the forecasts wish it to be so and because the modelers sell themselves as seers. Occasionally, someone looks twice and sees models and forecasts for what they are, a systematized characterization of the way the modeler and forecaster see the world. This can cause great shock to all concerned.

It is therefore important to remember that models are a systematic way of analyzing the future but are no more accurate than the analysis that went into them, are open to bad assumptions and poor parameterization, and are subject to recall if the characterization of the phenomenon of interest proves inaccurate or goes out of vogue.

The IIASA modeling system is a case in point. Built as part of the IIASA energy program, the model and forecasts acquired the prestige and notoriety associated with the effort. The models and forecasts (Haefele 1981) were reviewed both internally and externally (Meadows 1981; Ausubel and Nordhaus 1983; Keepin 1983; Keepin and Wynne 1984; and Keepin, in press). The system is extremely time consuming in operation. Its chief advantage (and disadvantage) is its extreme level of detail. Weeks were required to set up a run, and then to go through the delicate balancing process of bringing the global regional system into equilibrium. Meadows (1981) criticized the models for failing to adequately incorporate economic variables. For example, MEDEE-2 does not take prices into account. As a modeling effort the IIASA system is little more than an accounting tool for maintaining internal consistency.

The models' toughest and most vocal critics have been two IIASA staff members, Bill Keepin and Bryan Wynne (Keepin and Wynne 1984; Keepin 1983, 1984; and Wynne 1984). They argue that, in fact, "quantitative analysis behind the scenarios does not support the conclusion drawn from them" (Keepin 1984, p. 201). Keepin and Wynne's charges stem primarily from an analysis of MESSAGE, the LP model. Keepin documents the fact that by merely plotting out the value of constraints, entered as exogenous model inputs, one can very closely approximate the IIASA model results. That is, Keepin charges that the "results are essentially carbon copies of various unsubstantiated assumptions and arbitrary projections that were supplied as inputs to the mathematical analysis" (Keepin 1984, p. 201). Keepin and Wynne also charge that the IIASA models are extremely sensitive to model assumptions. That is, if critical cost assumptions are changed, the model changes its results dramatically.

Much of Keepin and Wynne's criticisms of the IIASA models could as easily have been levied against any LP model, because LP models are notoriously dominated by constraints. They are optimization models, not behavioral models. Constraints are necessary to exclude impossible or, in the modeler's mind, unreasonable choices. Linear programming models are also notoriously prone to high sensitivity, and sensitivity analysis is a method for assessing this tendency in a model. Keepin and Wynne charge that a formal sensitivity analysis was never conducted and that preliminary work which indicated such instability was abandoned.

Haefele and Rogner (1984) address these criticisms by arguing that the IIASA models were meant to serve primarily to force self-consistency in scenario development. Thus, the heavy intervention of judgment in the value selection process is not unexpected. They also argue that the model's results were not as sensitive to changing model assumptions as Keepin charged.

The intensity of the debate is in keeping with the magnitude of the entire IIASA energy program. It raises issues surrounding CO_2 emissions forecasts that are substantive. The Keepin-Wynne attack on the foundation and robustness of the two IIASA scenarios is fundamentally a skepticism about mankind's present ability to forecast. What are appropriate confidence intervals on energy forecasts and CO_2 emissions? Their critique is also a call for simplicity. If in the end the model structure was no more than a cumbersome and elaborate sidecar, why bother to continue to use it? And how do we go about the process of establishing and testing our own analyses? Is there no objective validation process for long-term forecast models? Keepin and Wynne argue for fuller documentation and external peer review as a place to start.

9.8 FOSSIL FUEL CO_2 EMISSIONS FORECAST PROGRAMS

Three programs have dominated this area of research: the energy program at IIASA, the U.S. National Academy of Sciences (NAS) climate change program, and the U.S. Department of Energy (DOE) carbon dioxide program.

9.8.1 The IIASA Program

The IIASA energy project was a large and diverse program of research involving 140 scientists from 19 countries over a period beginning in 1973 and ending in 1979. Though greatly reduced in size and scope, the energy program at IIASA continues to do global energy research.

The best known product of the IIASA energy project has been termed simply the IIASA model. The IIASA model is in fact two European models (developed outside of IIASA)—MEDEE-2 and MESSAGE—but brought in and used together with a third model, IMPACT, brought in from the U.S.S.R. to perform side calculations. The IIASA model became a central focus of the program, though it comprised a much smaller part of the overall effort. The program itself studied a wide range of energy-related questions including those concerning resources, technologies, demographics, and socioeconomic phenomena, the results of which are reported in Haefele (1981).

The models MEDEE-2 and MESSAGE were different in structure, but both required large data

bases, including inputs on economic structure, life-style variables, technical efficiencies, maximum buildup rates for supply technologies, production costs, resources, energy supply, production limitations, interregional energy trade, population, and economic growth.

MEDEE-2 modeled energy use and is an accounting-type model. That is, it ensures consistency among the components and aggregates of definitional equations, but few if any behavioral equations are employed. Energy uses are disaggregated to a high degree (e.g., individual appliance types are differentiated). Data bases were accumulated for each of the seven IIASA world regions.

MESSAGE, on the other hand, was the supply-side model. It determined energy production and transformation to end-use fuels. The model followed energy from final energy demands through refining and transport to the resource stage. Unlike MEDEE-2, MESSAGE is an LP model that seeks minimum-cost solutions to meet a specified final demand.

The models were run using inputs whose values were from the distillation (by Haefele and his staff) of research throughout the energy project. Judgment was an essential part of the value selection process, especially in light of the size of the model's input data base, the long forecast horizon (50 years, 1980–2030), and the uncertainty surrounding all of the data elements.

The two models were not linked together directly. Rather, one model would be run and then the other. The outputs of each (though primarily MEDEE-2) were assessed and modified before becoming the inputs of the other. IMPACT was run using inputs from both as part of this interface and evaluation process.

The end product of the analysis was a pair of global energy (and CO_2 emissions) scenarios referred to as the IIASA high and IIASA low scenarios. These two scenarios are reference points in the literature. Haefele (1981, p. 424) argues that "the range was selected such that the expected future value is more likely within the range than outside it," although IIASA staff members frequently stated that they felt that the "business-as-usual" future lay closer to the low scenario than to the high.

An important research element of the IIASA program was the work of Marchetti (1980). Marchetti's model was never formally incorporated into the IIASA model, but the research he conducted had an important influence on the formulation of constraints that proved to be important in the IIASA model. Marchetti's research focuses on interfuel substitution and market penetration over time. At its center is a logistic equation. The model is deterministic. Fossil fuel use depends on an exogenously specified total energy use and the parameters of the logistic equation.

Market penetration of different fuel types depende. on initial conditions (market penetration) and time. The mechanism was monodirectional. Once a fuel's use declines, there was no return. Because a resurgence of demand for coal was expected, model users were forced to create two different kinds of coal, old technology coal and new technology coal, the sum of which is coal. The distinction between old and new fuels seems arbitrary, especially at the extremely high level of aggregation—one region and one primary energy fuel type—used by the model. Economic factors are totally absent in the simplest versions of the model and are only marginally utilized in the more sophisticated theoretical expressions of the analysis.

Other IIASA researchers who contributed to the literature on future fossil fuel CO_2 emissions include Williams, Niehaus (Niehaus and Williams 1979), Lovins (Lovins et al. 1981), A. M. Perry (Perry et al. 1982), and Nordhaus (whose most prominent contributions have been associated with the NAS program).

9.8.2 The National Academy of Sciences

The now dormant NAS program is responsible for the publication of two major studies on the issue of CO_2: *Energy and Climate* (NRC 1977) and *Changing Climate* (NRC 1983). The earlier study forecasts CO_2 emissions using two methodologies—a logistic curve forecast (Keeling and Bacastow 1977) and a gap study (Perry and Lansberg 1977). Both methodologies were superseded by the analysis of Nordhaus and Yohe (1983).

The Nordhaus-Yohe study is interesting on a number of counts. Nordhaus abandoned the linear programming model of CO_2 emissions (Nordhaus 1979) in favor of a highly aggregated (one region, two fuels), 10-variable, parametric energy-economic model. Although the model is not particularly remarkable (see Reilly and Edmonds in press), it is encased within a Monte Carlo exercise that represents a break with previous technique of analysis. They termed their approach "probabilistic scenario analysis." In addition to the model, they developed a data base including (1) probability distributions for all input variables and (2) simulation of a "sample" of scenarios based on random selection of input values (Monte Carlo analysis). They use the results to make probabilistic predictions as to the timing of significant buildup of atmospheric carbon. The analysis also allows a ranking of parameters

in terms of their contribution to the uncertainty described in the simulation that Nordhaus and Yohe suggest as a guide for the direction for future research.

Previous studies employed scenario analysis in which considered assumptions were applied to a model to produce a future emissions path. In some cases (e.g., Haefele 1981; Edmonds et al. 1984; Rose et al. 1983; Seidel and Keyes 1983), multiple scenarios were constructed. In a few cases (e.g., Haefele 1981; Edmonds et al. 1984; Siegenthaler and Oeschger 1978), explicit statements are made about the subjective probability of a realized business-as-usual scenario falling within the bounds of the highest and lowest scenarios. But the assessments are purely subjective and do not have a rigorous base.

The Nordhaus and Yohe work made the link between uncertainty about inputs and the uncertainty of outputs explicit. Rather than an assumed value for each model input, they used an entire probability density function (normal distributions were used). By using Monte Carlo techniques with the assumption that the model was known with certainty, they could derive a distribution of scenarios that were interpreted as random draws of the future. Explicit probabilistic statements could then be made about future CO_2 emissions.

It must be remembered that those probability statements are conditional on the validity of the model and the input distributions. If either of these assumptions is violated, the uncertainty description must be redrawn.

9.8.3 The U.S. Department of Energy (DOE) Program

The third major program is sponsored by DOE's Carbon Dioxide Research Program. It has maintained support for research since the early 1970s. Most of this work has been conducted by the Institute for Energy Analysis, Oak Ridge Associated Universities. In 1979 the current research program was begun. The program consisted of research into the key issues surrounding future global energy and CO_2 emissions as well as the development of a model.

The model is less detailed than the IIASA model although it is more detailed than the model used by Nordhaus and Yohe. It has nine regions; nine primary energy categories; energy transformation for solids, electricity, and refinery sectors; final demand sectors; and an energy-GNP feedback mechanism. Unlike the IIASA model, it is entirely self-contained. That is, once a set of assumptions is specified, the model generates a global energy balance scenario through an iteration procedure without further human intervention. Two versions of the model now exist, version B.82 and version A.84. Version B.82 was used extensively by DOE, which sponsored its development, and by other organizations such as the MIT-Energy Laboratory (Rose et al. 1983), the U.S. Environmental Protection Agency (Seidel and Keyes 1983), the Electric Power Research Institute (Reister 1984b), the Gas Research Institute (Reister 1984a), the Oak Ridge National Laboratory (Edmonds et al. 1984), and the Worldwatch Institute (Chandler 1985).

The work is extensively documented. The theoretical structure of the basic model (version B.82) was published in the open literature (Edmonds and Reilly 1983a). Extensive documentation of the computer code and its use is documented in Edmonds and Reilly (1984) and is available from the Carbon Dioxide Information Center at the Oak Ridge National Laboratory. Documentation of revisions in the model and code, which were implemented in FY 1984, is provided in Edmonds et al. (1985).

The principal strengths of the model are its midlevel of detail, which enables it to address questions at the regional and technological levels; its extensive analysis of the underlying issues and their relationship to model structure (Edmonds and Reilly 1985; Edmonds et al. 1985, especially Appendix A); and its extensive documentation that allows other users to explore alternative scenarios.

Model version B.82 was criticized by Ausubel and Nordhaus (1983) for lacking full energy-economy interactions, for the use of logistic curves to model fossil fuel supply, and for its size and complexity, which are "somewhat forbidding for a casual user." Rose et al. (1983) critique the model for being insufficiently detailed in its end-use categories, having an inflexible supply schedule for unconventional oil and gas and coal, and for its lack of inclusion of indirect effects of CO_2 accumulation in the atmosphere.

Some of these criticisms have been addressed in recent modeling revisions that created version A.84 (Edmonds et al. 1985, Appendix B). For example, the logistic curves and "inflexible" supply schedules have been dropped in favor of a more flexible, resource-based description of energy supply. No additional detail was added to the energy end-use categories, and the energy-economy interaction remains modeled in an elementary way.

9.8.4 Other Programs

As Table 9.1 indicates, many studies have been conducted outside the confines of these major programs. The most prominent are the studies sponsored by the United Nations Environmental Program, the World Meteorological Organization, the Federal Republic of Germany, The Netherlands, the U.S. Environmental Protection Agency, the U.S. National Science Foundation, the Electric Power Research Institute, the Gas Research Institute, and the World Resources Institute.

9.9 OTHER RADIATIVELY ACTIVE ATMOSPHERIC CONSTITUENTS

While CO_2 is the most frequently discussed "greenhouse" gas, other gases may be important in determining the atmosphere's thermal structure (e.g., methane, ozone, nitrous oxide, and chlorofluorocarbons). The role of non-CO_2 greenhouse gases has been a matter of concern for some time (e.g., see Wang et al. 1976). Major reviews of the greenhouse problem (e.g., Kellogg and Schware 1981; Clark 1982; NRC 1983; and Jaeger 1983) have repeatedly emphasized the potential importance of these greenhouse gases in influencing future climate, either through enhancement of the effect of CO_2 or through interaction with other atmospheric elements. There is a consensus of opinion as to the importance of studying these gases, and Clark (1982) states:

> The greatest uncertainty in the trace gas question concerns the rates at which these gases are added to and removed from the atmosphere. (p. 254)

Despite this, there are no studies to date that have taken an integrative approach to the forecast of all greenhouse gases. Studies of the greenhouse issue treat non-CO_2 gases separately from CO_2. Emissions are not even considered as an issue except by Perry et al. (1982), Seidel and Keyes (1983), and Wuebbles et al. (1984). Even the review report (Jarrett and Coates 1984b) omits any discussion of the emissions question. Perry et al. (1982) use a multiplicative factor despite the statement:

> The effects and probable future concentrations of these other IR [infrared] gases are not well known, but they need not be proportional to that of CO_2. (p. 993)

Seidel and Keyes (1983) based their assessment of future climate impacts on the joint effects of CO_2 and non-CO_2 greenhouse gases. They included separate emissions for nitrous oxide, methane, and chlorofluorocarbons. Although the role of non-CO_2 gases was not a major focus of their report, subsequent analysis of the study reveals that non-CO_2 trace gases have a nearly equal role in controlling global warming and were crucial in shaping the overall policy conclusions of the report.

The most thorough discussion of non-CO_2 trace gases and factors that could influence future emissions is contained in the Wuebbles et al. (1984) report. Emissions forecasts for CO_2 plus four gas categories (methane, carbon monoxide and nonmethane hydrocarbons, chlorocarbons, and nitrous oxide) are developed. Even here, the forecasts are not systematically integrated and the analysis is rudimentary in comparison to CO_2 emissions forecasting techniques.

9.10 SUMMARY AND CONCLUSIONS

9.10.1 Energy CO_2 Emissions Programs

Three energy-CO_2 emissions programs dominate the field of fossil fuel CO_2 emissions forecasting: the IIASA energy program, the National Research Council program, and the DOE program. Of the three, only the DOE program remains active. There is a clear need for continuing support of programs in the area of CO_2 emissions analysis. It is also important to maintain a plurality of independent analysis capabilities. Research recommendations are provided in Chapter 11.

9.10.2 CO_2 Emissions Forecasts

Expectations about future emissions has shifted downward. Researchers now expect slower but continued growth in the rate of emissions over the next 50 to 125 years. Post–1982 studies project an average growth rate of 1.2% per year. Variation in best-guess scenarios has diminished with research, although it is not clear whether this represents a convergence in independent expert opinions or is the result of interaction.

9.10.3 Uncertainty

Uncertainty in future fossil fuel CO_2 emissions remains large. In the year 2050, it is an order of magnitude or more. Some of this uncertainty can be reduced throug

further research, some cannot and some represents the area of influence open to decision makers.

9.10.4 Forcing Functions

Only two studies systematically examined the determinants of uncertainty in business-as-usual forecasts, and these studies are not in agreement as to the most important determinants of uncertainty. Nordhaus and Yohe (1983) rank the top three as:

1. Ease of substitution between fossil and nonfossil fuels,
2. General productivity growth, and
3. Ease of substitution between energy and labor.

Edmonds et al. (1985) rank the top three as:

1. Labor productivity growth,
2. End-use energy efficiency improvements, and
3. Income elasticity of energy demand in the developing regions.

The two studies use different models and variables. It is therefore difficult to draw solid conclusions. Nevertheless, it should be observed that Nordhaus and Yohe's general productivity growth is similar to a combination of labor productivity growth and end-use energy-efficiency improvements.

Other studies assess the importance of various factors informally. End-use efficiency improvements rank high on many studies' lists of important factors (e.g., Rose et al. 1983; Edmonds et al. 1984; Chandler 1985; Williams et al. 1984; Perry et al. 1982; and Lovins et al. 1981). Other studies (e.g., Perry et al. 1982; Niehaus and Williams 1979) argue the importance of interfuel substitution.

No consensus exists on forcing functions for fossil fuel CO_2 emissions.

9.10.5 Policy Analysis

The field of policy analysis remains largely unexplored. This is due in part to a feeling that the underlying elements affecting future energy use and CO_2 emissions need to be better understood before policy implications can be drawn reasonably.

From the handful of policy experiments conducted, a general consensus exists that the United States, acting alone, cannot control global CO_2 emissions. Tax policy measures appear to be a poor tool to reduce emissions. To be effective, tax constraints would have to be administered on a global basis, all countries acting in concert.

Several studies explored the feasibility of low emissions scenarios. There is a sense that policy actions, if eventually needed, will not have to be undertaken before the year 2000. Furthermore, the technology is currently available for reduced CO_2 emissions, but the analysis has become clouded by the soft- versus hard-energy-path debate.

Even less attention has been paid to assessing the costs of various policies than to the policies themselves. Thus, no firm conclusions are available at this time.

9.10.6 Other Radiatively Active Atmospheric Constituents

There is a general consensus that a group of trace gases in the atmosphere that share the radioactive properties of CO_2 but which appear in quantities an order of magnitude more dilute, may, as a group, play an important role in determining global climate change. Despite their acknowledged importance in every major review of the greenhouse issue, their sources and rates of emission remain poorly characterized. Emissions forecasting is in its infancy, perhaps a decade behind in sophistication compared with CO_2 emissions analysis. No studies have systematically integrated forecasts for CO_2 and other trace gases.

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10. HUMAN ALTERATIONS OF THE GLOBAL CARBON CYCLE AND THE PROJECTED FUTURE

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10.1 SIMULATING THE GLOBAL CARBON CYCLE RESPONSE TO HUMAN PERTURBATIONS

A major objective of this chapter is to attempt to project the level of carbon dioxide (CO_2) in the atmosphere over the next century (to the year 2075), that is, to address the second question implied by the title. Projections of future atmospheric CO_2 concentrations are necessary to establish likely significant climatic and biological consequences of accelerated fossil fuel combustion. However, to estimate the effects of projected fossil fuel emissions on future atmospheric CO_2 levels, it is first necessary to know how this and other sources of CO_2 have already affected the natural biogeochemical cycle of carbon—a complex and dynamic set of processes linking the atmosphere, the oceans, and the terrestrial environment.

Most of the background material needed to address this topic, the first part of the chapter title, is presented in Chapters 3 to 8 (this volume; henceforth, all chapter references apply to this volume only). This information is synthesized in Section 10.2 as a necessary prelude to the following sections, which deal with other topics relevant to making atmospheric CO_2 projections. Fossil fuel and biospheric CO_2 emissions projections are covered in Sections 10.3.1 and 10.3.2, respectively. A set of reference scenarios for future fossil fuel CO_2 emissions was developed for this volume (Edmonds et al. 1984; Chapter 9), and is presented in Section 10.3.1.1.

The complicated and changing nature of anthropogenic sources of CO_2 and the carbon cycle requires two very different classes of simulation models to integrate the necessary quantitative data and to project future atmospheric changes. The first class deals with relationships among economic patterns, energy use, and CO_2 emissions, and the second with the major components of the global biogeochemical cycle. Models of the carbon cycle (the second class) should be consistent with the natural carbon fluxes among atmospheric, oceanic, and terrestrial pools; they should accurately reflect the impact of fossil and biospheric releases over past centuries; and they should satisfactorily account for atmospheric measurements of recent decades (e.g., 316 parts per million by volume [ppm] in 1958 rising to 342 ppm in 1983). However, this ideal model will be unachievable until we can reconcile incongruous estimates of anthropogenic CO_2 releases and ocean CO_2 uptake capacity (Chapters 3, 6, and 7).

How do such unresolved issues affect atmospheric CO_2 projections? Are the impacts significant enough to warrant the use of empirical approaches such as the airborne fraction instead of carbon cycle models? Can we clearly identify and separately quantify individual sources of projection uncertainties at the current state of the art? How time-dependent are errors in projections? These questions are addressed in a variety of ways in this chapter, in Section 10.2.8 and Sections 10.3.3 to 10.3.6. A reference set of atmospheric CO_2 projections is presented in Section 10.3.4, and is compared with other recent projections in Section 10.3.6.

The reader seeking a brief summary of the status of research and a discussion of "where we go from here" is advised to proceed directly to the concluding section (10.4) and Chapter 11, "Research Requirements."

10.2 FROM THE PAST TO THE PRESENT

10.2.1 Background

With some allowance for both natural annual variability and measurement uncertainties in atmospheric CO_2 levels, a satisfactory mathematical model of the global carbon cycle should reproduce the trajectory of the contemporary atmospheric record, while balancing all past fluxes from fossil fuels with the terrestrial landscape and with the oceans. Because existing models cannot accomplish this (Chapter 7), modelers currently must sacrifice agreement with the atmospheric CO_2 record (i.e., with both the historical values *and* the trajectory of the contemporary data).

This problem is illustrated by model-reconstructed atmospheric CO_2 concentrations with different assumed terrestrial release time series (Table 10.1). These types of calculations were detailed in Section 7.3. Table 10.1 includes reference values for historical concentrations at three different dates (1800, 1860, and 1880). There are only three cases in this table where the model-calculated atmospheric CO_2 level for the year 1800 falls within the probable range of 260 to 285 ppm for the natural baseline value (Chapter 3); only one (284 ppm) falls close to the new estimate of 280 ppm from the Siple ice core.

The lack of agreement between model response and the atmospheric record indicates that flaws exist in global carbon cycle models (i.e., because significant biogeochemical and physical processes have either been oversimplified or ignored) or in the atmospheric measurements. One or most probably some combination of the following is true: (1) estimates of contemporary oceanic CO_2 uptake are too low; (2) historical terrestrial biospheric CO_2 release estimates are too high or are masked in the atmospheric record by ignored or unknown sinks; (3) historical atmospheric CO_2 concentration estimates are inaccurate or are subject to considerable unrecorded natural variation;

Table 10.1
Model-Calculated Atmospheric CO_2 Concentrations at Key Historical Dates (1800, 1860, and 1880)

Reference Terrestrial History	Year	CO_2 Concentration (ppm) Globally Averaged Carbon Cycle Model			
		KE3B[a]	BJMO[a]	PEBR[b]	STBQ[c]
Neutral Biosphere[d]	1800	292	292	295	295
	1860	292	292	295	295
	1880	293	293	296	296
$\delta^{13}C$ Record, Freyer and Belacy (1983)	1800	245	245	243	233
	1860	254	255	252	242
	1880	266	269	263	256
$\delta^{13}C$ Record, Stuiver et al. (1984)	1800	–	–	–	284
	1860	–	–	–	290
	1880	–	–	–	298
$\delta^{13}C$ Record, Freyer (1986); Peng and Freyer (1986)	1800	–	–	266	–
	1860	–	–	273	–
	1880	–	–	278	–
MBL Land-use Conversion[e]	1800	245	245	260	257
	1860	253	253	267	258
	1880	257	258	272	263

[a]KE3B is model 3b in Killough and Emanuel (1981); BJMO is variant of Björkström (1979) ocean model described in Edmonds et al. (1984). [b]Peng et al. (1983).
[c]Stuiver et al. (1984).
[d]Assumes zero net atmosphere-biosphere flux from 1800 to present.
[e]Reconstruction based on Marine Biological Laboratory (MBL) population-growth reference case (Houghton et al. 1983).

(4) the historical fossil fuel CO_2 flux time-series contains large errors; or (5) the contemporary atmospheric CO_2 record (e.g., Mauna Loa) is not a suitable reference data set. Sections 10.2.2 through 10.2.9 review potential contributions from these phenomena to errors in the global carbon balance.

10.2.2 Contemporary Atmospheric Record

The Mauna Loa record of atmospheric CO_2 concentration (Keeling et al. 1982) and the companion records being developed by the National Oceanic and Atmospheric Administration/Geophysical Monitoring for Climate Change (NOAA/GMCC) CO_2 sampling network and others (Chapter 3) represent one of the best "knowns" in the global carbon cycle (see Chapter 11). The Mauna Loa record appears to be a satisfactory reference for modeling purposes, and it is doubtful that measurement uncertainties in the annual average values or the increase from 1958 to the present are larger than ±1 ppm (Chapter 3; also Bacastow and Keeling 1981; Keeling et al. 1982; Keeling 1983). Small additional errors of about the same magnitude could have been introduced both by anomalous interannual fluctuations and by long-term changes in sea surface temperature (Chapter 3; Oeschger and Heimann 1983). Thus there is precious little uncertainty here!

10.2.3 Historical Fossil Fuel Releases

Estimates of errors in the fossil fuel CO_2 emissions time-series vary significantly with time (Chapter 4): pre-1950 data "could well be in error by 20% or more" (Keeling 1973) and post-1950 values are believed to be in error by 6 to 13%, depending on the time period selected and the methods used to calculate errors (Chapter 4). Owing to the nature of the data, error assignments are not precise, and subjective judgments are very much a part of the estimation process. Authors have recognized, for example, that some conversion factors used are dominated by data from the most recent decade and that time-dependent biases may have been introduced by employing constant multipliers for the entire post-1950 period (Marland and Rotty 1983). These errors, about ±10% during the Mauna Loa record, are comparable to the sum of potential errors in the atmospheric CO_2 increase. The aggregate is potentially significant (see Section 10.2.8).

10.2.4 Historical Atmospheric CO_2

Serious questions do exist about both the magnitude and variability of preanthropogenic values of atmospheric CO_2 concentrations, particularly during the past millennium (Chapters 3 and 8; but also see Section 10.4.5).

The related uncertainty about the magnitude and timing of nonfossil CO_2 sources suggested by current ice-core CO_2 records is also a major concern. Terrestrial ecosystems, particularly the forests, are potentially a larger historical source of CO_2 to the atmosphere than fossil fuels because it is estimated that nearly half of the living terrestrial biomass was eliminated by humans during the late Holocene (Chapters 6 and 8).

The principal controversies are associated with the magnitude and timing of carbon losses from terrestrial ecosystems. Did a large fraction of the terrestrial CO_2 release occur over the past 200 years when human population was increasing exponentially (McEvedy and Jones 1978), and increased exploitation of both fossil fuel and biospheric carbon resources occurred throughout the world? Significant population growth is still occurring in tropical regions, along with associated exploitation of the forests, and is projected to continue for decades (Myers 1984). What are the implications of these releases today or in the future? Estimates of the rate of carbon uptake by ocean models are too low to accommodate even moderate CO_2 releases from vegetation and soils over the past 200 years without invoking additional contemporary carbon sinks (Killough and Emanuel 1981; Enting and Pearman 1986; Chapters 5 and 7). Thus, current estimates of significant terrestrial releases (Chapter 6) simply exacerbate the problem.

Uncertainty about the history of atmospheric CO_2 concentrations and potential effects of climatic variations on oceanic CO_2 uptake has made it difficult even to define this problem clearly (World Meteorological Organization [WMO] 1983). However, recent results (Friedli et al. 1984; Neftel et al. 1985; Chapter 3) have improved the prospects for resolution (see Section 10.2.7).

10.2.5 The Uncertain History of the Terrestrial Biospheric CO_2 Flux

Differences in the rates and timing of past biospheric carbon fluxes that correspond to the reconstructed atmospheric concentrations in Table 10.1 are shown in Figure 10.1. The differences between the reconstructed time series of terrestrial biospheric fluxes in Figure 10.1 are striking to say the least. The neutral biospheric case from Table 10.1 is represented by the zero-flux line in the figure. This case is not thought to be realistic, but it could be compatible with interpretations of some $\delta^{13}C$ records in tree rings. It nonetheless represents an interpretation of the role of the biosphere taken in past carbon cycle modeling exercises and thus serves as a point of reference. Ignoring the zero-flux case, the range of integrated biospheric fluxes represented by data in Table 10.1 and Figure 10.1 is 90×10^{15} to 360×10^{15} g C from 1800 to the present; the corresponding annual average fluxes for the same period range from 0.5×10^{15} to 2×10^{15} g C/year.

There are striking differences between reconstructed post-1960 biospheric fluxes in Figure 10.1, referenced to the same tree-ring data set (Freyer and Belacy 1983), which require further discussion. The differences are attributable to the manner of ^{13}C data representation for modeling purposes. In one case, a step function corresponding to 10-year averages of $\delta^{13}C$ data was analyzed to provide a direct estimate of the biospheric contribution (Stuiver et al. 1984), whereas in the other, a specified curve was fit to the data (Peng et al. 1983). Freyer (1986) shows that interpretation of post-1950 data is very sensitive to the degree of the polynomial employed to represent the $\delta^{13}C$ data; the biosphere could change from a source to a sink, based on the curve fit alone! Large residual uncertainties in the $\delta^{13}C$ records from tree rings attributable to physiological variability in isotope fractionation are mainly responsible for the poor agreement among tree-ring-based reconstructions of the biotic flux (Chapters 3 and 11).

If we ignore the reconstructions based on the older Freyer and Belacy (1983) tree-ring data, the greatest discrepancies among reconstructions in Figure 10.1 occur between 1900 and 1920 and after 1930. The lack of agreement is worst for recent decades, that is, in the period in which the incompatibility between terrestrial carbon releases and state-of-the-art ocean models is most acutely demonstrated (Chapter 7). Current estimates of uncertainty in the recent annual terrestrial carbon flux (e.g., 1970–1980) derived from contemporary land-use conversion data (0.6×10^{15} to 2.6×10^{15} g; Chapter 6) fall within the range of 0.2×10^{15} to 2.6×10^{15} g obtained from historical reconstructions as shown in Figure 10.1.

10.2.6 The Role of the Oceans

The oceans are the principal sink for excess CO_2 from human activities (Chapters 5 and 7). The magnitude of the net flux of carbon into the oceans is believed to vary with the atmospheric CO_2 content, past and present (see Figure 5.11), and also appears to exhibit significant variability on a seasonal and annual basis (Chapter 3). A range of $2.4(\pm 0.9) \times 10^{15}$ g C for the ocean sink in 1980 contains all recently published estimates and appears to be representative of existing uncertainty, including that contributed by historical atmospheric CO_2 variations and the supplemental coastal marine carbon sinks shown

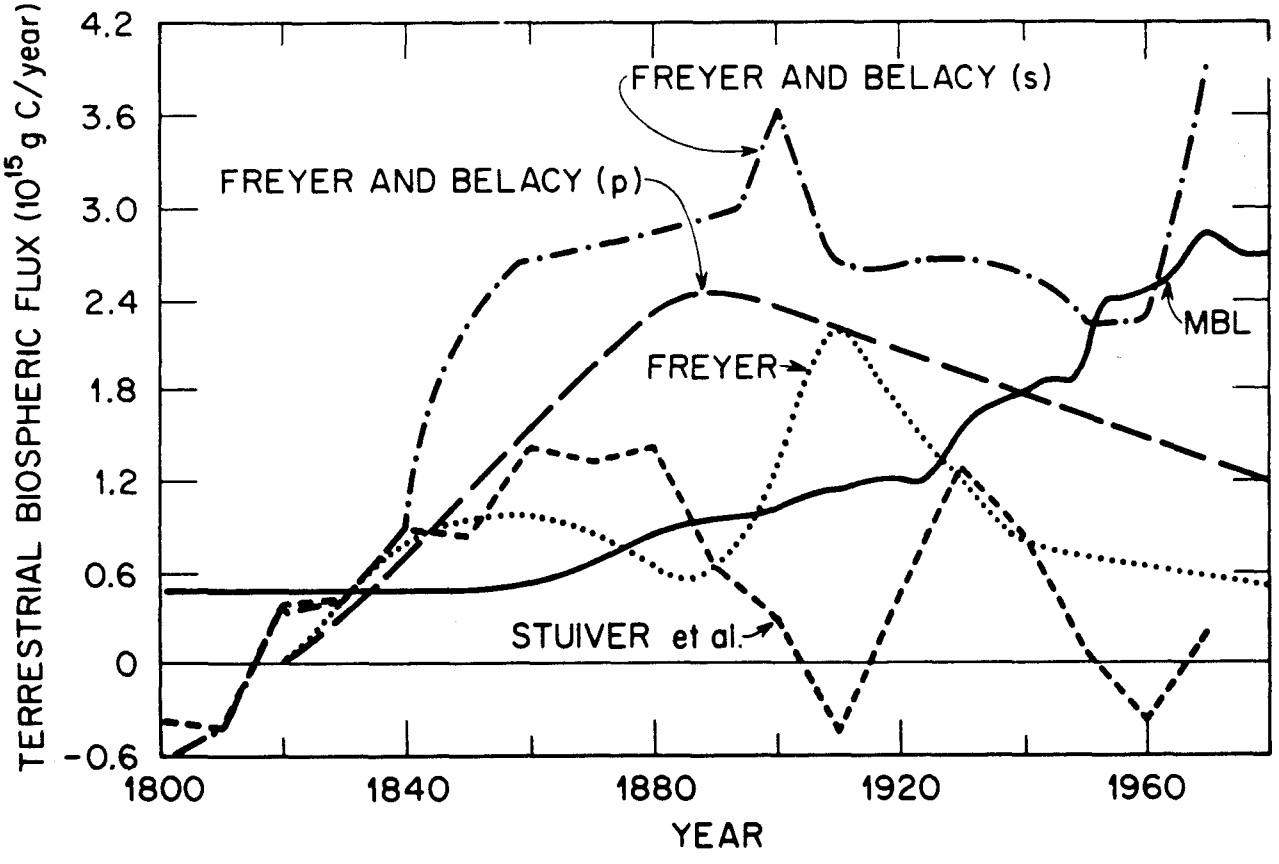


Figure 10.1. Reconstructions of the post-1800 carbon flux from the terrestrial biosphere with the atmosphere based on back-calculations with globally averaged carbon cycle models. Freyer and Belacy (s), from Freyer and Belacy (1983) $\delta^{13}C$ data by Stuiver et al. (1984); Freyer and Belacy (p), from Freyer and Belacy (1983) $\delta^{13}C$ data by Peng et al. (1983); Freyer, from Freyer (1986) $\delta^{13}C$ data by Peng and Freyer (1986); Marine Biological Laboratory (MBL), from Houghton et al. (1983) population growth reference case by Peng et al. (1983); Stuiver et al. , from Stuiver et al. (1984) $\delta^{13}C$ data by Stuiver et al. (1984).

in Table 5.2, unless thermocline vertical mixing rates have been *grossly* underestimated using tracer data (Chapters 5 and 7). The annual sink represented by current ocean models is only $1.9(\pm 0.6) \times 10^{15}$ g C (Table 5.4; Siegenthaler 1983) because coastal sinks are ill-defined and have yet to be incorporated into models in any substantive fashion. It also should be noted that the highest value in the preceding range (2.5×10^{15} g C) is thought to be an extreme upper limit.

10.2.7 Missing Sinks

It is possible to define the time series of atmospheric CO_2 sources and sinks needed to "balance the global carbon cycle" through relatively straightforward calculations with ocean models. The result of one such calculation is shown as the curve labeled $A(B_0)$ in Figure 10.2(a). (The reason for the dual label is explained below.) A specified history of atmospheric CO_2 concentration (i.e., the Siple ice-core and Mauna Loa records [Chapter 3] and the historical fossil fuel CO_2 emissions record [Chapter 4]) was used to constrain a state-of-the-art ocean carbon cycle model (Peng et al. 1983) for this particular exercise. Any one of several ocean models could have been employed in such a calculation, with similar results. Similarly, a smoothed version of the Stuiver et al. (1984) atmospheric CO_2 reconstruction based on Pacific Coast tree-ring $\delta^{13}C$ data could have been substituted for the Siple ice-core data.

The principal characteristics of curve $A(B_0)$ are that (1) a net source of CO_2 to the atmosphere ($\sim 60 \times 10^{15}$ g C) is needed to "balance" the carbon cycle before 1940; however, (2) a flux in the reverse direction, a net sink growing to 0.8×10^{15} g C/year by 1980, is required after 1930 (also see Enting and Pearman 1986; Killough and Emanuel 1981). (Two very recent reconstructions of nonfossil CO_2 fluxes based on the same data, but using other carbon cycle models, yielded similar pattern

of 19th century sources and contemporary sinks.[1,2] However, in both cases, the contemporary sink [1958–1980] required to balance the carbon cycle was negligible [$\leq 0.1 \times 10^{15}$ g C/year].) If the terrestrial biosphere is assumed to be the only active subsystem in the global carbon cycle other than the open ocean, then this same curve $A(B_0)$ also defines the source/sink status of the biosphere (i.e., the reason for the label B_0, representing the "baseline" biospheric case in Figure 10.2). Based on what we learned earlier in Chapters 5 and 8, the above assumption is just that, and a questionable one at present. The same assumption underlies *all* reconstructions of biotic fluxes based on tree-ring ^{13}C data (e.g., Figure 10.1).

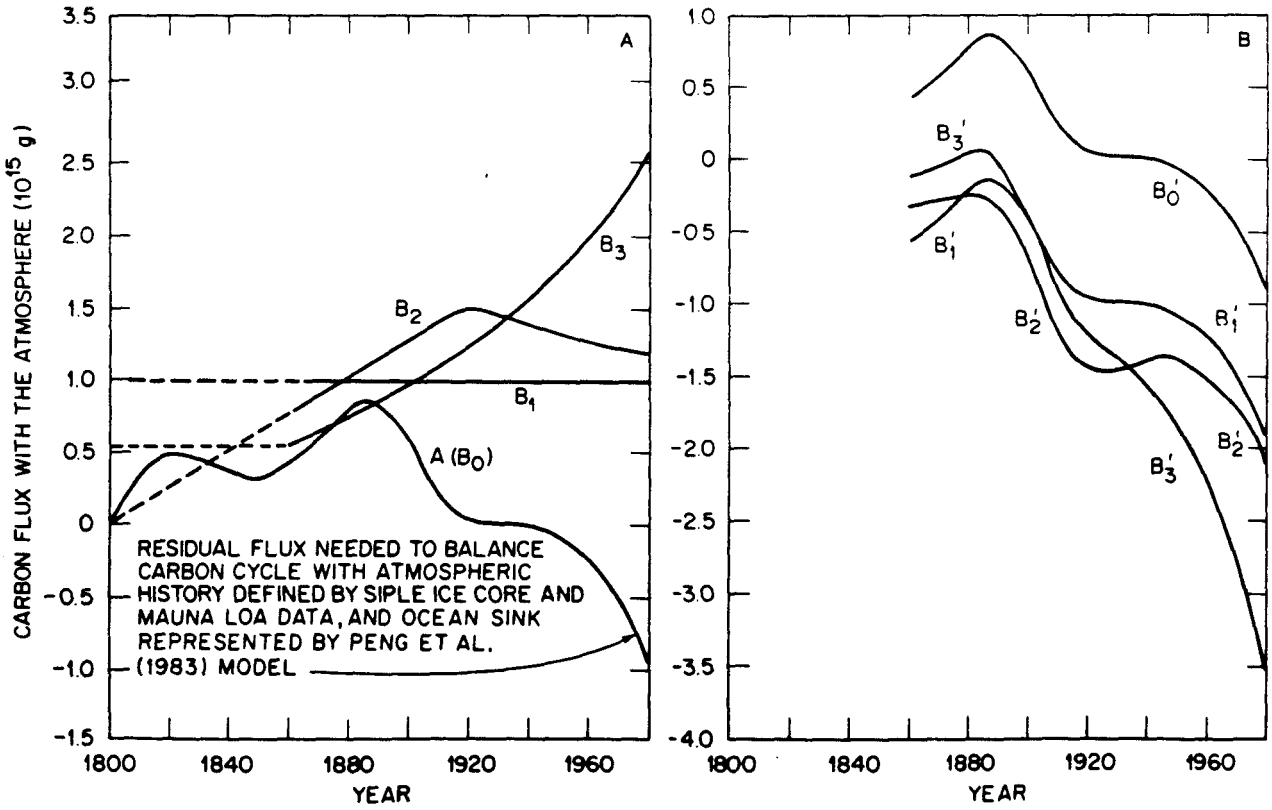
If other currently plausible time series are used to represent the history of the terrestrial biotic flux (curves B_1, B_2, and B_3 in Figure 10.2(a)), then the pattern of historical sources and sinks needed to balance the carbon cycle is generated by subtracting each of the latter three curves from curve $A(B_0)$. Curves B_1 and B_3 are idealized versions of two potential biospheric flux histories discussed by Houghton (1986), and curve B_2 was drawn from Trabalka et al. (1986, also see Figure 10.8 in Section 10.3.2 of this chapter). Many others could be substituted. The results are shown as curves B'_1, B'_2, and B'_3, respectively, in Figure 10.2(b), along with curve B'_0 [identical to curve B_{0_A} (B_0 - zero) in Figure 10.2(a)]. For purposes of simplification, we elected to present the patterns of sources and sinks after 1860 only in Figure 10.2(b). The objective was to present a potential range of reasonable source/sink time series to illustrate what is missing from our current representations of the carbon cycle.

Based on the best current estimates for the net biotic flux (Section 10.2.5) and the results discussed above, we currently need a "missing sink (or sinks)" of 0.6×10^{15} to 3.4×10^{15} g C/year around 1980 to balance the global cycle. Candidates for this missing sink(s) include the biosphere acting alone (but not as represented in current models; see Chapter 11) or in combination with other potential sinks. These could include the ill-defined coastal marine sedimentary carbon sinks (e.g., Table 5.2 in Chapter 5), other even less well-defined

[1]Siegenthaler (1983) box-diffusion model calibrated with bomb-^{14}C; H. Oeschger, University of Bern, personal communication, 1985.

[2]Killough and Emanuel (1981) model 3b; G. G. Killough, Oak Ridge National Laboratory, personal communication, 1985.

Figure 10.2. (a) Four potential histories of the terrestrial biospheric CO_2 flux with the atmosphere [$A(B_0)$, B_1, B_2, B_3], and (b) the corresponding patterns of sources and sinks of CO_2 needed to balance the carbon cycle since 1860 (B'_0, B'_1, B'_2, B'_3, see text for details). Positive and negative fluxes represent sources and sinks, respectively.



natural "subfossil" or anthropogenic sinks in freshwater wetlands and soils (Section 8.5.3; Chapter 11), and the often-cited higher uptake potential of the open ocean (currently masked by oversimplified models as one simple explanation). The challenge of future carbon cycle research will be to define the characteristics of these "missing sinks" and to quantify their flux rates with the atmosphere.

However, another challenge is to estimate the impact of these uncertainties at the present time on projections of future levels of atmospheric CO_2. In this context, the characteristics of all curves in Figure 10.2(b), particularly during the Mauna Loa record, are intriguing if not unsettling. The post-1880 downward trends (toward an increasing sink after 1958) could reflect either purely transient phenomena (e.g., opposed effects in carbon cycle terms or chance summation of errors) or perhaps even the growing response of feedback mechanisms attempting to restore a global equilibrium state (e.g., Gaia hypothesis, Section 8.6.2). The impact of the former can be at least partially described (Section 10.2.8), but if the latter explanation is correct, the ability to make future projections may be seriously impaired.

10.2.8 The Effective Airborne Fraction

The impacts of potential errors in the (1) atmospheric record, (2) fossil fuel time trends, (3) historical atmospheric CO_2 levels, (4) past terrestrial biospheric fluxes, (5) ocean model CO_2 uptake simulations, and (6) other contemporary sources or sinks on contemporary estimates of atmospheric CO_2 retention can be approximated using techniques described by Oeschger and Heimann (1983), and using data presented by these authors and by Siegenthaler (1983). Because it is generally agreed that future fossil emissions represent the most important potential CO_2 source (Chapters 4 and 9), the most important concept in Oeschger and Heimann's approach is that by correcting the atmospheric CO_2 record for all interferences that obscure the contemporary ability of the global carbon system to absorb fossil fuel CO_2, we can obtain a clearer understanding of the potential variability in future atmospheric retention of CO_2 emissions.

The index of comparative atmospheric retention is the fossil fuel airborne fraction (AF), which should "be regarded only as a convenient parameter" (p. 20, Pearman and Beardsmore 1984) for purposes of illustration. We expanded the analysis to include additional sources of error not specifically considered by Oeschger and Heimann and further updated the results to the year 1982 based on data in Figure 3.15 and Table 4.4.

Briefly put, the notation introduced by Oeschger and Heimann (1983) makes it possible to separate ar quantify the effects of historic and contemporary CC sources and sinks on the contemporary atmospheric CO_2 increase. If the ocean-atmosphere interaction can be represented as a linear system, then sources prior tc 1958 introduce contemporary concentration gradients opposite to the increasing trend. That is, if there were no additional CO_2 releases after 1958, atmospheric CO_2 content should have declined at rates determined by the effective rate of ocean uptake and by the magnitude and timing of past CO_2 releases. In theory, at least, if the past sources had been of sufficient magnitude, these could compensate for low ocean model CO_2 uptake, and perhaps could even allow moderate biospheric releases during the Mauna Loa record.

However, there are obvious limits to past biospheric sources. High net estimates of past releases (e.g., over 180×10^{15} g C from 1860 to 1958) would be incompatible with current estimates of historic CO_2 levels during the 19th century even if uptake by the ocean and other sinks were stretched to the limits. Furthermore, ocean models with higher CO_2 uptake induce the need for correspondingly larger and earlier past CO_2 sources, that is, at times when fossil fuel releases are relatively small (see Figure 10 in Killough and Emanuel 1981). These additional sources may be required at times that may not be compatible with other existing information, including historic CO_2 estimates (see previous section and Chapter 6).

The notation introduced by Oeschger and Heimann (1983) was expanded slightly to accommodate differences in ocean model CO_2 uptake and the presence of potential candidates for the missing sinks, as well as effects of historical variability in fossil fuel releases. The resultant equation is:

$$AF^* = [N_a(1982) - N_a(1959) + B_p + B_h + \Delta O_c + \Delta T_w + \Delta a_n + \Delta P_h](\Delta Q_f)^{-1} . \quad (10.1)$$

The various quantities are described in Table 10.2, and the results of calculations are shown in Tables 10.2 and 10.3. The values in Table 10.2 were chosen to produce a range of scenarios, including potential extremes, which were consistent with the observed atmospheric CO_2 increase and current estimates of carbon cycle uncertainties. This included the constraints that past releases not require an initial atmospheric CO_2 level in 1800 to 1860 less than 260 ppm or greater than 285 ppm, and that contemporary biospheric releases be completely accommodated by potential errors in other factors and/or by ocean sinks. However, the read should also recognize that there exists uncertainty .

Table 10.2
Summary of Estimates for Different Terms Used to Determine the Effective Airborne Fraction

Parameter	Value (ppm)	Description
N_a (1982)–N_a (1959)	25.5±1	Observed atmospheric increase (Figure 3.15; also see text)
ΔQ_f	43.7±4.4 (±10%)	Fossil fuel CO_2 input (92.7 × 10^{15} g C, 1959–1982; Section 4.2.4)
AF	0.58	Apparent airborne fraction
AF*	0.67[a] 0.52[b]	Effective airborne fractions calculated with ocean models (Siegenthaler 1983)
B_p	−5[a,b]	Change in atmospheric increase from constant biospheric source (0.6 × 10^{15} g C/year, 1959–1982)
	−21[b]	Change in atmospheric increase from constant biospheric source (2.6 × 10^{15} g C/year), 1959–1982
B_h	+4	Change in atmospheric increase from constant biospheric source (1.0[a] or 1.5[b] × 10^{15} g C/year, 1860–1958)
ΔO_c	+2 to +9	Change in atmospheric increase from enhanced sedimentary CO_2 sink (Table 5.2; also see Section 8.5.3)
ΔT_w	0 to +4	Change in atmospheric increase from freshwater wetland sink (Bramryd 1980; Mulholland and Elwood 1982; Armentano et al. 1984; Kempe 1984).
Δa_n	−2 to +2	Change in atmospheric increase from ocean temperature anomalies (Oeschger and Heimann 1983; also see text)
ΔP_h	−1 to +1	Change in atmospheric increase from variations in fossil fuel emissions from 1860 to 1958 (see text)

[a]Nominal estimate, box-diffusion model (Oeschger and Heimann 1983).
[b]Model with increased eddy-diffusion coefficient and 10% polar outcrop to maximize potential ocean CO_2 uptake (Siegenthaler 1983; also see Viecelli et al. 1981).

atmospheric sources and sinks that is not quantifiable at the current state of knowledge (see Section 10.2.9). Thus the results in Tables 10.2 and 10.3 represent a compromise. The choices of constant biospheric release rates and a starting date in 1860 were merely for convenience, and one could choose other alternatives, including more varied biospheric terms (as described by Oeschger and Heimann [1983]). However, these results are not seriously compromised by the choice of a simple biospheric release function, and the advantages to be gained do not appear to warrant, nor do data currently exist to support, more complex biospheric scenarios for this type of analysis.

The effective airborne fractions (AF*) in Table 10.3 represent the potential effects of carbon cycle uncertainties on estimated atmospheric retention of contemporary fossil fuel releases. The range in values, rounded off to one significant figure, 0.4–1, is believed effectively cover the potential extremes (i.e., it is unlikely that all terms will covary exactly as suggested in the table). The value in column (5) is the nominal estimate of the airborne fraction based on atmospheric measurements and the fossil fuel record. The results in columns (2), (3), (6), (7), and (8) reflect only adjustments for a small terrestrial CO_2 source, a small to moderate sedimentary CO_2 sink in coastal marine wetlands[3] (Table 5.2), natural atmospheric CO_2 variability that could have produced anomalies during the Mauna Loa record, and potential extreme errors in atmospheric measurements and fossil fuel release estimates. The corresponding range in AF* is substantial, 0.43–0.85 (Table 10.3). The range induced by potential errors in atmospheric CO_2 measurements and contemporary fossil fuel CO_2 releases alone is 0.51–0.67 (identical to range generated using values in columns [3] and [6] of Table 10.3). A range of 0.54–0.62 in AF* would result if variability in these same

[3]Also a moderate freshwater wetland sink in columns (2) and (8).

Table 10.3
Potential Range for Effective Airborne Fraction Based on Estimates from Table 10.2

Parameter	Range of Scenarios (1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
	Value (ppm)							
Observed Atmospheric Increase	26.5	26.5	26.5	26.5	25.5	24.5	24.5	24.5
B_p	0[a]	−5	−5	−21	–	−5	0[a]	−5
B_h	+4	+4	+4	+4	–	0[b]	0[b]	0[b]
ΔO_c	+3[c]	+3	+2	+9	–	+2	+2	+3
ΔT_w	+2[c]	+2	0	+4	–	0	0	+2
Δa_n	+2	+2	−2	+2	–	+2	−2	−2
ΔP_h	+1	+1	+1	+1	–	+1	−1	−1
Adjusted Atmospheric Increase	38.5	33.5	26.5	25.5	25.5	24.5	23.5	21.5
ΔQ_f	39.3	39.3	39.3	39.3	43.7	48.1	48.1	48.1
AF*	0.98	0.85	0.67	0.65	0.58	0.51	0.49	0.43

Note: AF* = (adjusted atmospheric increase)$(\Delta Q_f)^{-1}$, where (adjusted atmospheric increase) = (observed atmospheric increase + $B_p + B_h + \Delta O_c + \Delta T_w + \Delta a_n + \Delta P$).

[a]Contemporary biospheric source absent.

[b]Historical biospheric source absent.

[c]Representing a transient sink, present during the Mauna Loa record only.

two factors was restricted to 50% of the potential ranges from Table 10.2.

The highest AF* (0.98) is produced with the combination of (1) no contemporary biospheric release, (2) a release over the century prior to the Mauna Loa record of 1×10^{15} g C/year, (3) a nominal estimate of ocean CO_2 uptake, (4) small transient CO_2 sinks in freshwater and coastal marine sediments (occurring during the Mauna Loa record only), (5) other variations of data at their upper extremes, and (6) the contemporary fossil fuel release at its potential minimum value (column [1] in Table 10.3). If the coastal marine and freshwater sinks are not transient and counterbalance a small contemporary biospheric source (0.6×10^{15} g C/year), the AF* is reduced to 0.85 (column [2]).

The highest contemporary biospheric release (2.6×10^{15} g C/year) is accommodated only if both ocean model CO_2 uptake and other potential sinks are maximal (column [4] of Table 10.3) and if a large past biospheric source (1.5×10^{15} g C/year) was present in the century before the Mauna Loa record was initiated. Maximal extremes of other terms and the minimum fossil fuel release estimate are also required (i.e., the "chance compensation of opposed effects" [p. 417, Broecker et al. 1979]), in order to attempt a contemporary global carbon balance in the face of significant biospheric releases, and yet we are still barely able to accommodate the upper estimates of the latter. Note also that if the maximal value for the fossil fuel release (48.1 ppm) had been chosen for column (4), the magnitudes of the biospheric releases would have to be reduced (and AF* increased) because the ocean model uptake reserve, used to absorb part of the excess biospheric CO_2, would then be required to handle the excess from the higher fossil fuel release.

The compensation of the terrestrial CO_2 release in column (4) of Table 10.3 by upper extremes of error terms or other effects is responsible for the fact that the effective airborne fraction (0.65) is actually higher than both the nominal estimate of 0.58 in column (5), and the lower estimates in columns (6), (7), and (8), even though the latter are based on lower contemporary biospheric release rates. It would appear that we should be much more concerned than we have been up to now about the varying impacts of missing sinks (such as those represented here by sediments in freshwater and coastal marine systems) and historical terrestrial biospheric releases (cf. AF* in columns [1], [2], and [8] with the others), in terms of potential impacts on future projections of fossil fuel CO_2 retention by the atmosphere.

The key conclusion implied in Oeschger and Heimann's (1983) analysis, and strongly seconded here, is that our current carbon cycle uncertainties are large enough that we *cannot* make model agreement with the Mauna Loa record (trajectory) an absolute precondition for long-term atmospheric CO_2 projections. Rather, we must make allowances for such potential errors by employing a range of model simulations that provide reasonable coverage of the range of uncertainty in atmospheric retention (i.e., airborne fraction). A range of 0.5–0.8 for the effective airborne fraction is obtained by restricting parameter variability to 50% o

the ranges shown in Table 10.2, and this appears to ·s to be a reasonable statement of current uncertainty . the effective airborne fraction. Without a better understanding of why we are currently unable to balance the carbon cycle, it is not possible to state with certainty whether even this wide range may be too restrictive.

10.2.9 Potential Roadblocks to Improved Understanding

There still exist large potential conflicts between estimated CO_2 sources and sinks, whether derived from current attempts to balance the carbon cycle with exercises involving global models or from other detailed analyses of atmospheric CO_2 distributions and time trends (Table 10.3; cf. Houghton et al. 1983; Chapter 6 with Chapter 3 and with Elliott et al. 1985; Oeschger and Heimann 1983; Pearman et al. 1983; Keeling 1983; Keeling et al. 1984; Enting and Pearman 1986). The apparent discrepancy between the contemporary roles of the terrestrial biosphere and the ocean remains unresolved.

To resolve existing uncertainties, terrestrial research is now focused on geographically detailed analyses of historical data on changes in ecosystems brought about by the massive expansion of human populations. The role of the terrestrial biosphere or, better said, of terrestrial systems in the carbon cycle is so poorly constrained that there exists an obvious tendency to focus only on the paucity and quality of specific data on the timing and geographic distribution of land-use changes (such as forest clearing) and on carbon inventories in major ecosystems as the principal roadblocks to improved understanding. At first glance, it may seem that these factors each contribute approximately half of the existing uncertainties (Chapter 6). However, one should also realize that major sources of error in modeling also exist in the form of potentially important processes that may have been either overlooked (e.g., soil erosion and carbon storage in local alluvium or reservoir sediments [Kempe 1984]) or simply poorly parameterized (e.g., soil carbon responses to forest harvest and agricultural conversion [Schlesinger 1986; Pastor and Post, in press] and charcoal formation [Chapter 6]).

Background fluctuations induced by natural disturbance also may be significant (Sanford et al. 1985). Were disturbed ecosystems, many of which are now in successional stages (e.g., back to forest following clear cutting or abandonment of agriculture), pristine diverse systems at the time of first *presumed* disturbance? Or were they mainly lower in diversity, species content, or ;e distributions, with a lowered carbon content (e.g., ,rown and Lugo 1984) because of previous disturbance either by humans or by natural factors such as climate change (e.g., during the historical period called the Little Ice Age, 1430–1850; Gribbin and Lamb 1978), fire (Section 8.6.1; Sanford et al. 1985), windstorms (e.g., Canham and Loucks 1984), or pests (Section 8.5.4)? If the latter has been the rule rather than the exception, the history of the terrestrial carbon flux to the atmosphere may be quite different than that inferred from current models (See Chapter 11).

Soil and vegetation carbon losses to the atmosphere may also have been overestimated, in part because of soil erosion and deposition of the eroded material (humus, litter) within catchments and river systems. The bulk of this material will probably not be exported to the sea in most regions because of extensive reservoir development (e.g., Trimble 1975; Kempe 1984). Human-enhanced alluvial and colluvial storage represents a potentially significant but very poorly characterized sink, particularly if deposition is accompanied by additional biological production from large nutrient inputs resulting from human activities and by low rates of reoxidation of stored organic carbon in sedimentary deposits (Degens et al. 1984; Kempe 1984). Many factors may be significant (e.g., the fundamental questions implied by potential variations in input rates to the subfossil carbon pool [Section 8.5.3]).

Oceanic CO_2 uptake has been estimated with models that are essentially globally averaged (Chapter 7), generally represented by coupled systems of first-order linear differential equations. Typically there is little or no treatment of the large seasonal and geographic variability in key processes (e.g., biology, circulation) that determine CO_2 uptake (Chapter 5). The assumption of linearity (an exception is the Revelle or buffer effect; Brewer 1983) and the relative simplicity of the models may introduce unnecessary error into the results (Chapter 11).

Incorporation of improved geographic resolution and better representations for key processes in ocean models (air-sea exchange; thermocline vertical mixing; deep water formation and more realistic physics, in general; climate feedbacks on oceanic productivity/circulation; carbonate dissolution; and interactions of carbon with other element cycles) could change our understanding of the role of the oceans and provide a better definition of ocean CO_2 uptake. It is doubtful, however, that a satisfactory global representation incorporating all of these processes can be attempted at present.

Observation of a climatic freshening of the North Atlantic deep waters over the past 20 years also implies a relatively rapid response of deep water formation to climatic perturbation (Brewer et al. 1983). Attempts to incorporate high-latitude, deep water injection into existing models (e.g., Siegenthaler 1983) have shown

that this process can affect carbon uptake. However, the potential ramifications for the carbon cycle of a rapid change in overall ocean circulation and nutrient cycling induced by climate change via an effect on deep water formation are much more profound (see Section 8.4.2).

Evidence for the episodic nature of deep water formation and mixing in the North Atlantic (e.g., in the Norwegian Sea) was also provided by recent chlorofluoromethane tracer measurements (Bullister and Weiss 1983). However, episodic phenomena are not confined to the North Atlantic. A strong decrease in sea ice in the Weddell Sea, another important deep water formation area, from 1973 to 1977 was concurrent with the formation of the Weddell polynya, a major oceanographic anomaly (Zwally et al. 1983). These observations are only a few that illustrate that "We should not be too complacent (about our apparent knowledge of the oceans). Nature has vast resources with which to fool us. . . ." (p. 211, Brewer 1983).

Changes in global carbon storage induced by interannual climatic variations associated with El Niño-Southern Oscillation events (ENSO) may provide us with a preliminary indication of the magnitude of potential CO_2-climate feedbacks. One of the two extremes in the response of the carbon cycle to the 1982/83 ENSO event (see Chapter 3) may have been a partial embodiment of the latter. However, at present, our limited knowledge of climate feedbacks on oceanic or terrestrial processes prevents us from determining either the direction (sign) or the magnitude of the potential change (Chapter 11).

Resolution of all of the issues raised above may hinge on the answer to a more fundamental question: What are the principal processes involved in the Gaia[4] mechanism? The processes that operate on million-year time scales were well described in Chapter 8, but what of the shorter term (centuries or less) adjustments suggested by the ice-core records (Chapters 3 or 8)? Is *changing* ocean CO_2 uptake (e.g., via the interplay between biological processes, circulation, and climate) the dominant feature (Chapters 3, 5, and 8)? Or, have human-induced changes in the terrestrial landscape triggered other feedback processes that previously responded only to natural disturbance? Have humans unwittingly enhanced CO_2 uptake in soils and sediments by contributing to massive increases in soil erosion and rock/mineral weathering? Do some agricultural practices promote increased carbon sequestering? Does the presence of successional vegetation generally act to enhance CO_2 uptake by increased rates of organic carbon accumulation and (related) inorganic weathering reactions in soils? Is our missing sink represented by substantial, increasing soil and sedimentary carbon storage in recent decades masked by bomb ^{14}C contamination (e.g., of the soil column) as one simple explanation? To answer these questions, we must develop improved models for cycles of carbon and climate that contain explicit formulations for more active participation by the organic agents, treating these not as (in the past) conduits irrelevant to carbon storage but rather as state variables and parameters that may change and modify physical rates.

[4]The mythical symbol for biological feedback controls on global habitability (atmospheric CO_2); see Section 8.6.2.

10.3 THE FUTURE

10.3.1 Fossil Fuel CO_2 Emissions

The most important potential source of CO_2 emissions is the burning of fossil fuels (Chapters 4 and 9). There are many factors that can shape the future production and use of fossil energy resources. These include economic, technological, social, demographic, environmental, and political factors such as global population growth, gross national product (GNP) growth, changes in energy-use efficiencies, emergence of new energy supply technologies, energy-resource constraints, and the relationships among energy, income, and price. At some point, the effects of climate change on society may also affect fossil energy use, but such effects cannot be addressed quantitatively or qualitatively at the current state of the art (Chapter 1). Chapter 9 provides an extensive review of major fossil fuel CO_2 projection exercises and methods involved in making projections of fossil emissions.

10.3.1.1 *Development of Reference Projections*

Because emissions are of concern over periods ranging from decades to centuries, a fair amount of uncertainty exists in the key determinants of future emissions. To accommodate this reality, a parametric model of global energy use and CO_2 emissions developed for the Department of Energy (DOE) by the Institute for Energy Analysis/Oak Ridge Associated Universities (IEA/ORAU) (Chapter 9) was used to develop a set of reference projections for CO_2 emissions from fossil fuels.

The IEA/ORAU global CO_2 emissions model was used to forecast CO_2 emissions from fossil fuel combustion, in conjunction with three sets of alternative assumptions about the future, to create three future scenarios denoted Cases A, B, and C (Edmonds et al. 1984). This approach permitted disaggregation of resource, economic, and technological factors governing future energy demand so that scenario assumptions could be analytically examined and evaluated. These

Table 10.4
Model Parameter Values Varied to Produce Global Carbon Emissions Cases A, B, and C, 1975 to 2075

Parameter	Case A	Case B	Case C
Global Population in 2075 (10^9 persons) (1975 population equals 4.0)	11.4	8.4	6.3
Base Global GNP (annual percentage increase, 1975-2075)	3.2	2.4	1.9
Solar/Nuclear Electric Power Generation Cost (1975 US$/GJ)[b]	100.0[a]	18.0	9.0
Average Rate of Expansion of Base Coal Production (%/year)	2.9	2.2	1.2
2010 Breakthrough Price of Unconventional Oil Production (1975 US$/GJ)	3.85	7.35	19.25
Rate of Improvement in Energy Efficiency (percent per year)	0.0	0.5	1.0

[a]Effectively prevents development of these technologies.
[b]Units are United States dollars per 10^9 Joules.

Table 10.5
CO_2 Emissions from Fossil Fuel Burning (1975 to 2075) for Three Cases (10^{15} g C)

Scenario	1975	2000	2025	2050	2075
Case A	4.7	8.4	19.0	47.4	91.1
Case B	4.7	7.2	10.3	14.5	18.8
Case C	4.7	6.2	6.7	6.8	6.8

energy production and use scenarios are translated into equivalent CO_2 emissions that can serve as input terms to dynamic simulation models of the global carbon cycle.

The construction of future emissions scenarios focused on those parameters most important in determining global CO_2 emissions and those that have received attention in the literature. A standard case (B) was based on median estimates of model parameters. Scenarios A and C were developed by varying the assumptions concerning population growth, solar and nuclear electric power generation costs, and model factors demonstrated by sensitivity analysis to be important determinants of future CO_2 emissions levels (i.e., energy-efficiency improvement in end use, GNP, and coal and shale oil supplies). Scenarios A and C were obtained when the assumptions were varied jointly to produce extreme scenarios of carbon emissions (cf. Edmonds and Reilly 1983) as opposed to extremes in energy use or energy prices.

A comparison of the sets of assumptions, varied among the three cases, is presented in Table 10.4 (Edmonds et al. 1984). The resultant model outputs based on these three sets of assumptions generate a high carbon emissions case (A), a median emissions case (B), and a low emissions case (C) (Table 10.5). For all three fossil fuel scenarios, the rate of CO_2 emissions is higher in 2075 than in 1975. However, in all three cases the rate of growth of emissions declines toward the end of the period:

	1975–2075 Percent/year	2050–2075 Percent/year
Case A	3.0	2.7
Case B	1.4	1.0
Case C	0.3	0.0

Fossil fuel CO_2 emissions were projected to reach 91×10^{15} g C in Case A by the year 2075. More than half of the carbon emissions from Case A in 2075 are produced from unconventional oil (principally shale oil), with coal and synthetic gas accounting for the bulk of the remainder (Figure 10.3). The role of shale oil in determining CO_2 emissions is minor in all cases except Case A. The heavy use of shale oil in Case A depends heavily on optimistic assumptions about both the resource base and the technology employed to extract the fuels (Edmonds et al. 1984). Carbon emissions in the median case (B) are dominated by coal, with approximately equal contributions from the direct use of coal and coal-derived synthetic oil; projected emissions in 2075 are 18.8×10^{15} g C. Coal is also the principal source of carbon emissions in the low case (C), and the two principal energy sources are coal solids and the solar and nuclear options (Figure 10.4). Emissions in Case C reach 6.8×10^{15} g C in 2050, and then remain constant through 2075.

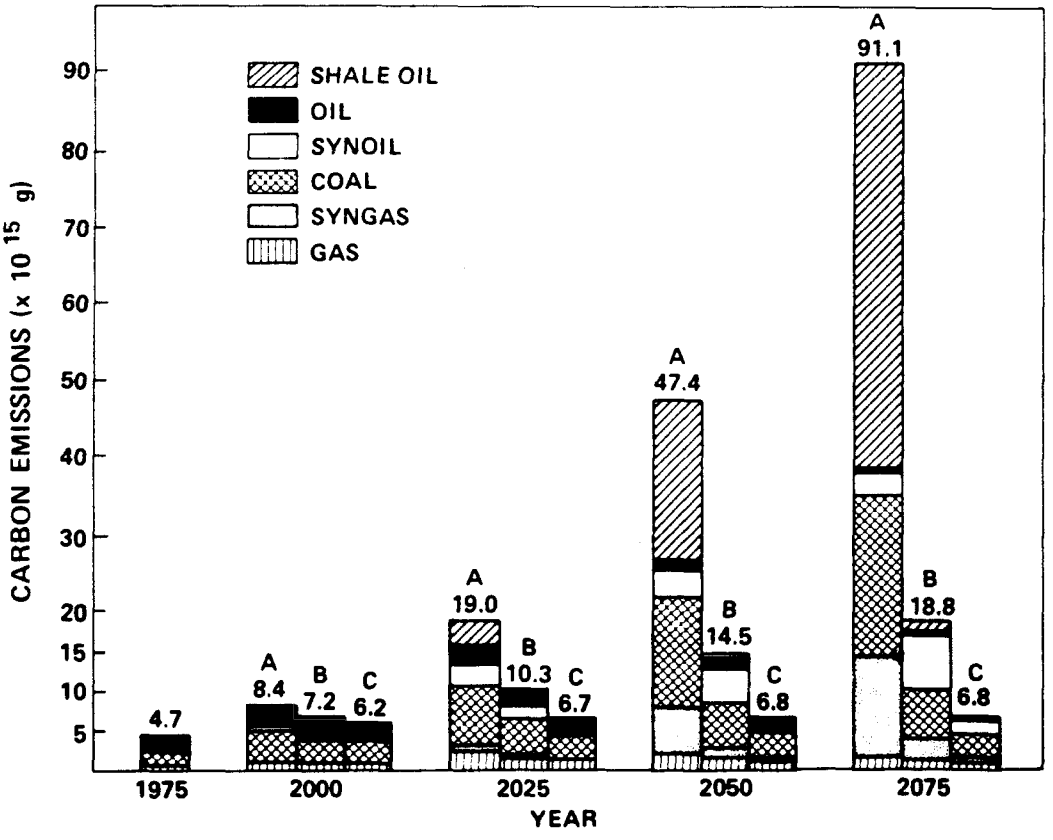


Figure 10.3. Emissions by fuel source for Cases A, B, and C, 1975–2075. (Source: Edmonds et al. 1984.)

10.3.1.2 *Comparisons with Other Estimates of Future CO_2 Emissions*

Figure 10.5 contrasts the set of reference CO_2 projections with other selected energy and CO_2 scenarios. The scenarios presented in the figure are not meant to represent a comprehensive review; in selecting scenarios, fairly complete coverage of extreme scenarios was sought. As a result, there are many other studies that would fall within the broad limits described by Cases A and C (see Table 9.1).

In the figure, one observes that Case C is lower than all but the Niehaus-Williams scenario in their 30-terawatt (TW) solar and nuclear case (NW[L]) and the Lovins et al. (1981) low CO_2 case (L). Case B, on the other hand, is quite similar to the International Institute for Applied Systems Analysis (IIASA) low scenario as well as the Nordhaus and Yohe (1983) 50th percentile or median case (NY[M]). It is also very close to the Reister and Rotty (1983) scenario. Reister and Rotty (1983) and Seidel and Keyes (1983) (neither shown in the figure) developed scenarios independently but used the IEA/ORAU computer model. This model was also used by Rose et al. (1983) to derive a series of 13 scenarios for the 1975–2050 period based on an alternative set of assumptions about end-use efficiencies, solar and nuclear contributions, fuel prices, and availability of Mideast oil. Emissions in the year 2050 ranged from 2.7×10^{15} to 15.2×10^{15} g C/year.

Case A is a high CO_2 emissions scenario with over 90×10^{15} g of carbon emitted in 2075. Although Case A is high relative to most studies, it is lower in all projection years than a simple extrapolation of the pre-1970s experience as shown in the JASON (1979) study.

Given the extreme diversity in CO_2 emissions scenarios, the obvious question arises: Why do these scenarios differ? If that question could be answered in a nontrivial way it would help clarify the agenda for further research (Chapter 9). The problem can be reduced to the identification of components of model structure and input assumptions that are the principal determinants of scenario differences, a task that is not necessarily easy. In many cases the comparison of scenarios involves different assumptions, different model structures, and different study objectives. Thus

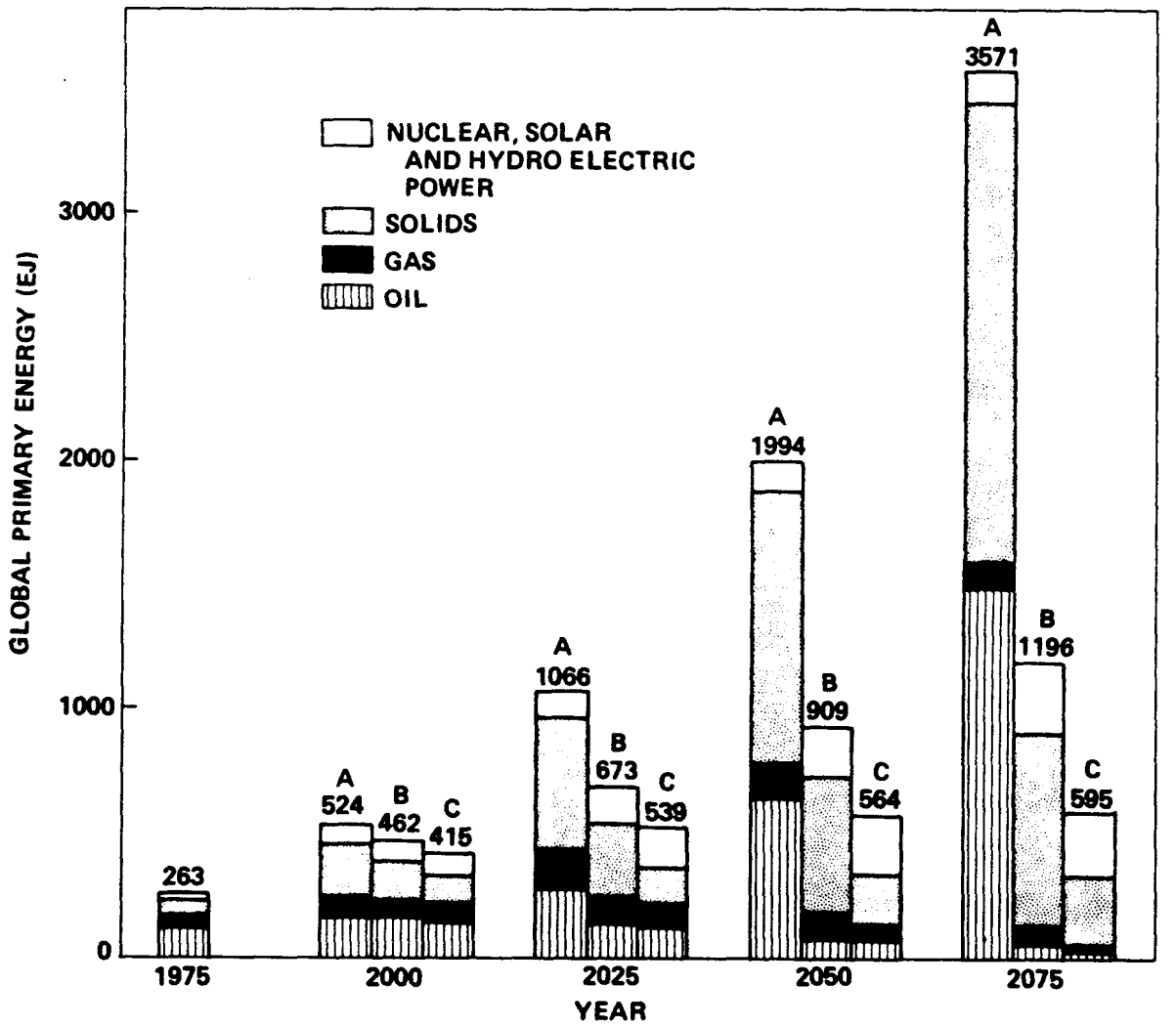


Figure 10.4. Global primary energy by fuel and case, 1975-2075 (in exajoules). (Source: Edmonds et al. 1984.)

comparisons must be made with caution. For example, the JASON projection is a simple extrapolation of historical rates of increase. The Niehaus-Williams and Lovins scenarios both use the IIASA scenarios as a baseline and then make various alterations to demonstrate certain points: Niehaus-Williams to show the potential impact of fuel mix on the dimensions of the CO_2 issue; and Lovins to suggest that energy strategies consistent with the IIASA low assumptions but which yield no CO_2 problem are possible. Nordhaus and Yohe, on the other hand, attempt to make specific predictive statements (e.g., the probability of future emissions falling between two points is X). The IIASA staff develop two scenarios without saying anything about the probability of occurrence other than that they are plausible.

Cases A and C were constructed to explore potential future extremes in carbon emissions rather than in energy use. Case A appears to fit this description well. In addition to the scenario-construction exercise which produced these cases, a Monte Carlo uncertainty analysis of the IEA/ORAU model was conducted (Chapter 9). Comparing the Case A and C scenarios to the Nordhaus and Yohe (1983) and IEA/ORAU Monte Carlo analysis results places Case A at the 99th and 95th percentile levels, respectively [cf. Figure 2.1, Nordhaus and Yohe (1983); Table 9.1 in Chapter 9]. However, Case C falls at the 10th and approximately the 40th percentile level, respectively, of outputs from these same two studies. Thus, it would appear that further exploration of potential extreme scenarios on the low end of the spectrum, in particular, would be

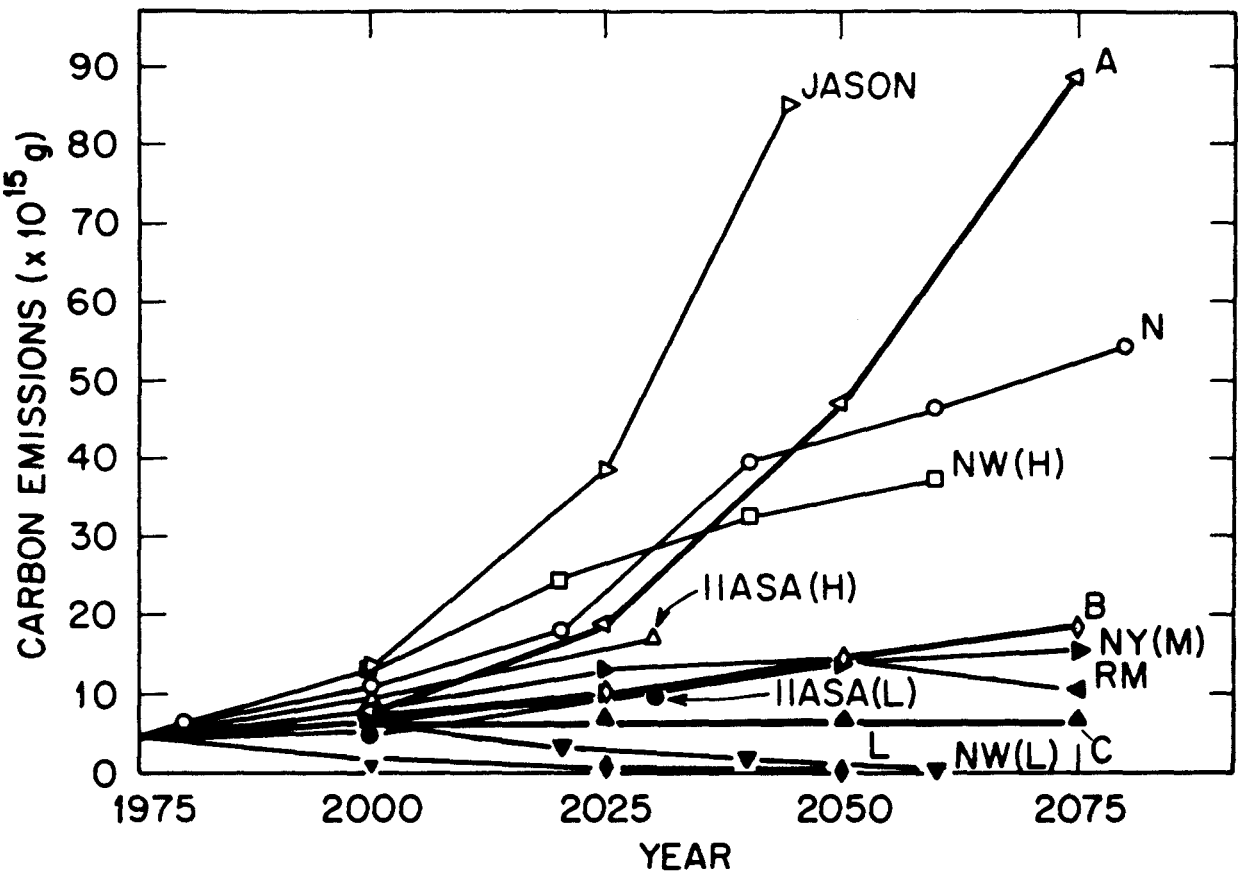


Figure 10.5. Comparison of selected long-term, global CO_2 emissions cases. (Source: Edmonds et al. 1984.)

Key to Figure 10.5:

A	Case A	NY(M)	Nordhaus and Yohe 1983. M = median case
B	Case B	IIASA(H), IIASA(L)	Haefele 1981.
C	Case C	RM	Rotty and Marland 1980.
N	Nordhaus 1979.	JASON	JASON 1979.
NW(H), NW(L),	Niehaus and Williams 1979. H = high case (50-TW—fossil fuel scenario) L = low case (30-TW—solar-nuclear scenario).	L	Lovins et al. 1981

a valuable exercise in the future. The corresponding levels of energy use for Cases A and C in 2075 (Figure 10.4) would lie at the 95th and 15th percentiles of the corresponding Nordhaus and Yohe distributions, respectively. Thus, fuel mix (Figure 10.3) is an important determinant of future CO_2 emissions.

10.3.2 Terrestrial Biospheric CO_2 Releases

Questions arising about the past role of the biosphere and its impact on the contemporary carbon system, including oceanic CO_2 uptake, must be answered if we are to understand the importance of future changes and to predict atmospheric CO_2 (Broecker et al. 1979). The conclusion of the authors, ". . . if the change in forest biomass over the last two decades has been small, the reason must be a chance compensation of opposed effects rather than an indication that this reservoir is neither big enough nor variable enough to play a key role in atmospheric carbon budgets" (pp. 417-418), appears valid today. However, factors other than land-use change that can directly affect terrestrial ecosystems may have to be considered in the context of "opposed effects," the Gaia hypothesis, and the tenuous contemporary global carbon balance.

The energy balance of the Earth and the associated hydrologic cycle are strongly influenced by the Earth's land cover (e.g., Gates 1971; Woodwell 1970; Idso and Brazel 1984). Changes in the land cover as a result of either natural or human disturbance and the succession that follows such disturbance can induce shifts in the energy balance and water cycles (Sagan al. 1979; Salati and Vose 1984). If the perturbation

are sufficiently large, the climate can be altered, with a subsequent feedback effect on the dynamics of terrestrial plant communities.

There is also an indirect effect resulting from changes in the composition of radiatively active atmospheric gases and other pollutants produced by burning of fossil fuels and by changes in the terrestrial landscape (e.g., Crutzen et al. 1979; Logan et al. 1981). The contribution to atmospheric CO_2 concentration because of decreases in terrestrial carbon storage and a potential reverse effect from CO_2 fertilization are indirect influences. A CO_2-induced climate change could influence the land cover and also alter the dynamic responses of terrestrial ecosystems to human disturbance.

The response of this highly interactive system is inextricably linked to understanding the dynamics of the Earth's forests over periods ranging from decades to centuries (Solomon et al. 1984). These dynamics are currently being investigated using simulation models (e.g., Lieth 1984; Shugart 1984). Model simulations allow us to explore hypotheses about the effects on the terrestrial carbon cycle of potentially important phenomena, such as CO_2 fertilization and climate change, in the current absence of needed field data on these effects. Such exercises provided the few estimates of geographic shifts in ecosystem distribution and of alterations in carbon storage using climate model scenarios of CO_2 warming.

Tree-ring records also have the potential for revealing these dynamics. Besides the carbon isotopic history of atmospheric CO_2, tree rings may record past climates (Fritts 1976; Blasing et al. 1981; LaMarche and Hirschboek 1984) and air pollution effects (Baes and McLaughlin 1984), including effects on carbon isotope fractionation (Freyer 1979).

10.3.2.1 *Effects of CO_2 Fertilization*

Tree-ring data are being used to examine the relationship between growth and CO_2 fertilization (e.g., LaMarche et al. 1984). However, a multitude of potential interferences can produce unaccounted-for variability (noise) in any given signal, and this limits the use of tree-ring data (Section 10.2.5). Thus, existing data are often contradictory (Woodwell 1983) or otherwise equivocal. A case in point is the growth enhancement observed in high-altitude tree rings over the period 1850–1859 to 1974–1983 attributed by LaMarche et al. (1984) to a CO_2 fertilization effect. The magnitude of the observed growth enhancement (7 to 11 times the level of a maximal response expected, considering the approximate 20% increase in atmospheric CO_2 over the same interval) indicates that factors other than increased CO_2 must be considered to account for the total effect (e.g., climate change).

The increase in phytomass simulated by seven different forest ecosystem models (temperate deciduous to subtropical rain forest) ranged from 20 to 60% when the growth rate for all trees was uniformly increased by 100% (i.e., less than a 1 to 1 correspondence between phytomass and individual growth stimulation [Shugart 1984; Shugart et al. 1984]). The increase was not due to an elevated maximum phytomass at any point in the landscape, but rather was caused by an increased rate of filling of gaps made in the canopy. For this reason, simulated forests with a high rate of disturbance (e.g., fire, selective timber harvesting) and more growth associated with gap-filling produced a larger landscape-scale fertilization response. This observation might be significant in light of the previous discussion in Section 10.2.9 (i.e., *the growth response is greatest for disturbed systems).* However, as the following discussion suggests, the gulf that separates model simulations and field validation is great indeed.

If changes in growth rates are not uniform across all the species involved (more likely case), the potential expression of CO_2-induced fertilization could involve changes in the community composition and could be very complex, resulting in even smaller increases in forest biomass (Shugart et al. 1984). Another critical uncertainty is whether limitations on growth imposed by current water or nutrient deficiencies, or by changes in plant morphology, will preclude a positive growth response to CO_2 (Kramer 1981). Consideration of the interrelationships between carbon, water, and nutrient pathways in soil-plant systems has led to the opposed hypothesis that stimulation of root and rhizosphere processes by elevated levels of CO_2 could increase both water-use efficiency and nutrient availability, and thus allow an increase in plant growth (Luxmoore 1981; Rosenberg 1981). This hypothesis is supported by the initial results of studies where plants were grown under apparent nutrient limitation in a forest soil (Norby et al. 1984). However, other research on plant-herbivore interactions under conditions of elevated CO_2 (Lincoln et al. 1984) suggests that any increases in carbon storage may be offset by increased consumption.

Increases in carbon-nitrogen ratios and decreases in nitrogen content of plant leaves grown in elevated CO_2 regimes reflect potentially lowered nutritive value to herbivores and, ultimately, to microbes important in leaf litter decomposition, nutrient cycling, and soil formation. In laboratory studies, insect larvae responded by increasingly higher consumption of plants from higher CO_2 atmospheres (Lincoln et al. 1984). Thus, in a field situation, increased herbivory could reduce or eliminate any potential gains in plant carbon

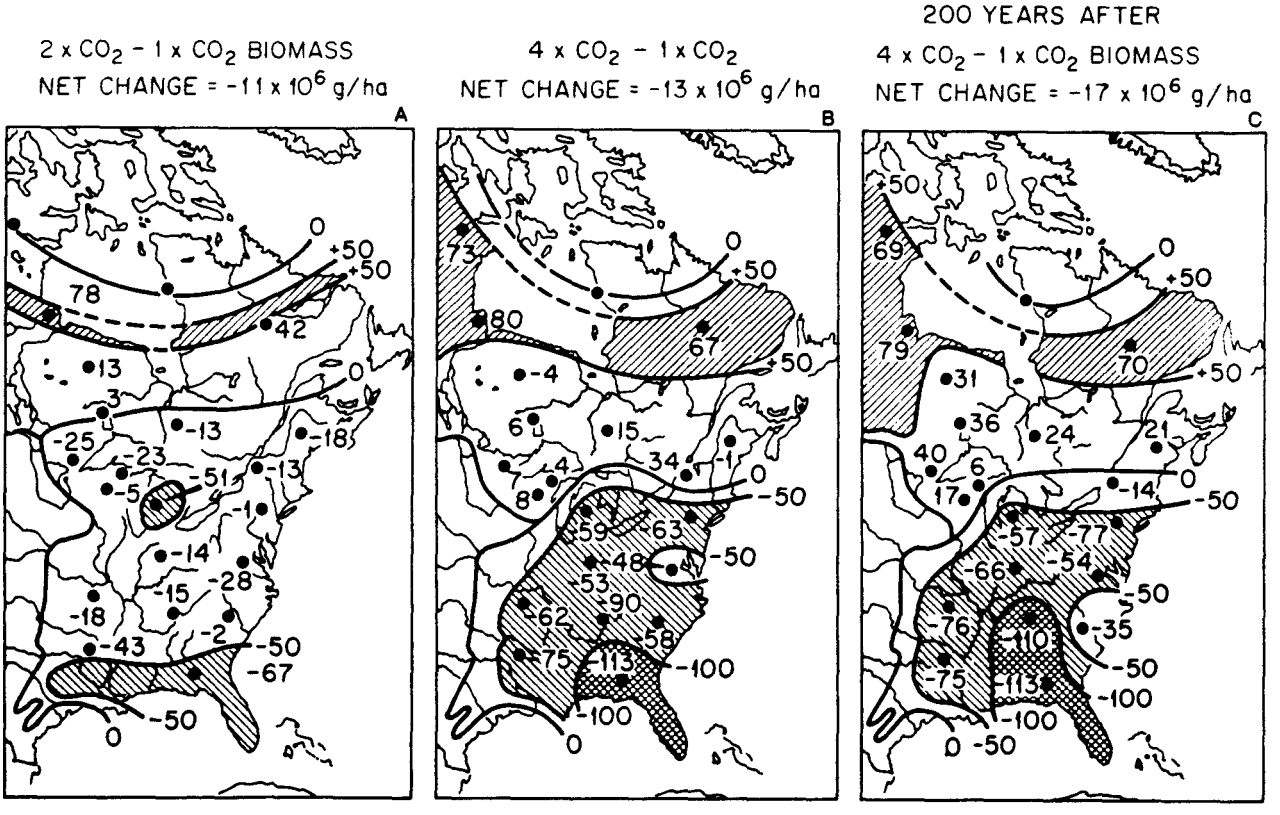


Figure 10.6. Carbon storage dynamics simulated at 21 sites in eastern North America. Maps show carbon storage differences between contemporary climate and (A) 2-x-CO_2 climate, (B) 4-x-CO_2 climate, (C) 200 years after 4-x-CO_2 climate stabilizes. Tree species of Florida, Cuba, and other Gulf islands may be important in the southern United States in the future, but were not simulated (Source: Solomon and Tharp 1985).

storage expected from CO_2 fertilization. Reduced rates of litter decomposition and nutrient cycling caused by lowered litter quality could also terminate an increased growth response to elevated CO_2 without increased herbivory. The role of consumers and decomposers (heterotrophs) in mediating the effects of CO_2 fertilization is potentially very important, but again, very poorly characterized (see Section 8.5.4; Strain 1985). Significant conclusions about the effects of elevated atmospheric CO_2 levels on the global carbon cycle are precluded by the current lack of information (Strain 1985), and thus the parameterization of a CO_2 fertilization response in global carbon cycle models (see Chapter 7) is currently a hypothetical exercise. At this time, there is no direct evidence that net ecosystem production has responded to increasing CO_2, either regionally or globally (Woodwell 1983).

10.3.2.2 *Effects of Climate Change*

Our understanding of potential ecosystem responses to climate change is also rather crude (Sections 8.5.4 and 8.6.1), but some of the factors identified above (ecosystem-specific responses, importance of heterotrophs, sensitivity of nutrient cycles) appear to be common threads binding the interpretation of CO_2 and climatic effects together. The forest stand simulation model approach suggested large decreases in carbon storage in eastern U.S. forests from CO_2-induced climate warming not completely compensated for by increased storage in Canadian boreal forests under generic model-projected climates, which included seasonality (Solomon and Tharp 1985). The projected carbon storage losses (Figure 10.6) amounted to 10 to 20% of the average standing biomass in these systems (Olson et al. 1983). Such losses could offset potentially enhanced carbon storage from CO_2 fertilization. Losses of carbon and changes in forest community structure were still occurring 200 years after a 4-x-CO_2 climate change scenario had been established (Solomon and Tharp 1985). The modifying effects of soil characteristics or heterotrophs are not yet considered in such simulations.

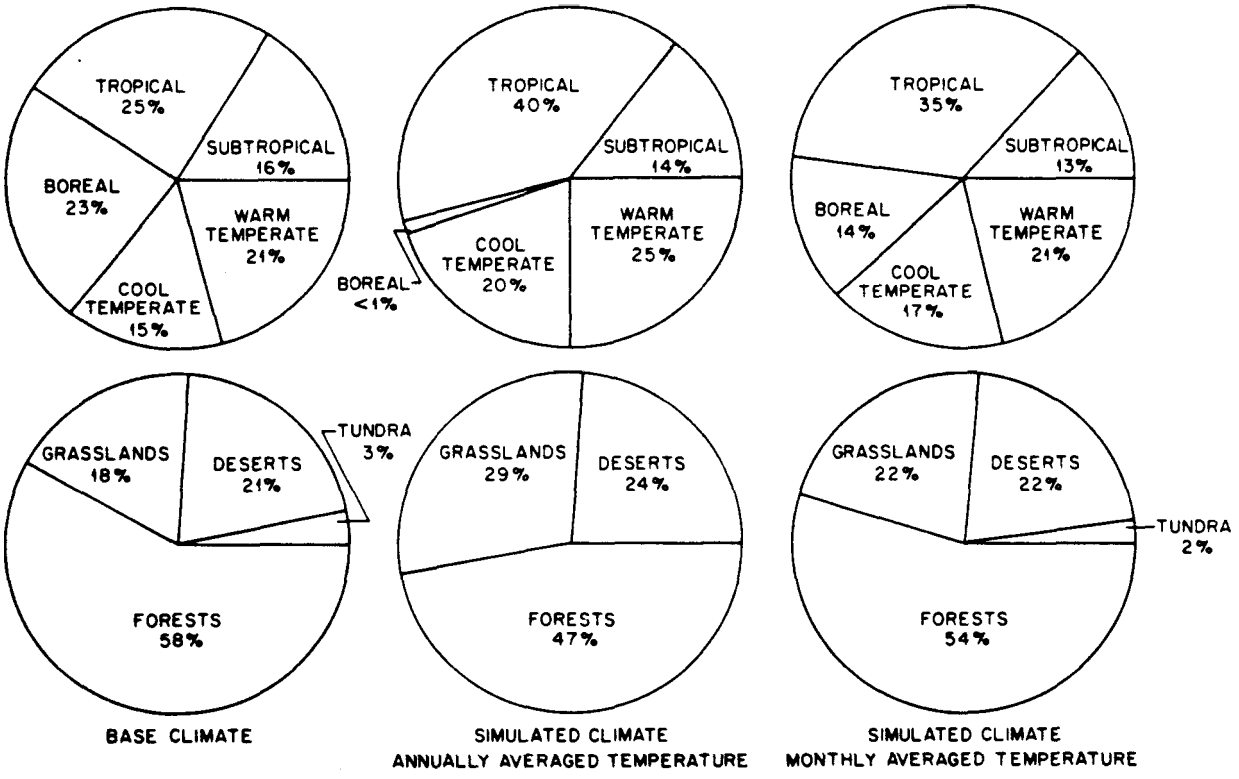


Figure 10.7. Simulated changes in areal extent of major forest types and biomes based on Holdridge Life Zone classification for base case (contemporary climatic conditions) and elevated CO_2 cases (twice base-case CO_2-induced temperature increase, assuming unchanged precipitation) (Sources: Emanuel et al. 1985 and Emanuel et al., in press).

Alterations of the global NPP[5]-climate map of Esser et al. (1982) by model-projected climate data (2-x-CO_2, Mitchell 1983) for Europe and North Africa produced elevated NPP in some northern and middle latitude areas and reduced productivity in some areas south of the Mediterranean region (Lieth 1984). The NPP map is a correlation model of primary productivity, parameterized for annual average temperature and precipitation and adjusted for soil fertility factors, but which does not consider species-specific phenomena or heterotrophic respiration and other losses of fixed plant carbon. Thus, the information of greatest interest, *net* ecosystem production and carbon storage, is not generated using this model. Neither are effects of climatically induced changes in ecosystem structure and function, caused by shifts in geographic distribution of vegetation species, treated in this approach (also see Kauppi and Posch 1985).

Significant changes in the distribution of terrestrial ecosystems were obtained (Emanuel et al. 1985) from a climate-based world vegetation map based on the Holdridge (1947) Life-Zone classification used to simulate the influence of equilibrium temperature change (2-x-CO_2 model results; Manabe and Stouffer 1980) when seasonal temperature changes were not considered. The changes in the distribution of life zones, and therefore of terrestrial ecosystems, were substantial, particularly in boreal regions where the simulated temperature increase is largest and the temperature intervals defining life zones in the Holdridge classification are smallest (Figure 10.7; simulated climate, annually averaged temperature). When seasonality of temperature changes was factored into the analysis of ecosystem distributions (Emanuel et al. in press), boreal regions were again most significantly affected, but changes were much less severe (Figure 10.7; simulated climate, monthly averaged temperature).

Changes in the tropics were less intense and much less affected by seasonality of temperature (Figure 10.7). However, simulated changes in precipitation (or effects of soils) were not considered in this exercise. Because many trees are sensitive to both temperature and

[5]Net Primary Production; see discussion in Sections 6.4 and 8.2.

moisture, the effects of the two are difficult to separate when interpreting ecosystem distributions (Davis 1983). Very large uncertainties are associated with current model predictions of local and regional climatic effects of increased atmospheric CO_2, particularly precipitation and seasonality of climatic variables (Schlesinger 1983). Because climate model simulations (Schlesinger 1983) and paleoenvironmental records[6](e.g., Bradbury et al. 1981; Webb 1985) both indicate that significant areas outside boreal regions, including the subtropics and tropics, may experience reduced precipitation under elevated-CO_2 climates, no conclusions can be drawn at this time about CO_2-climate-induced changes in global carbon storage from this or any other existing simulation exercise.

Model simulations of vegetation for four different scenarios of full-glacial climate have been tested against a pollen-reconstructed temperate forest (mixed boreal and deciduous) in North America. The climate scenarios were based on reconstructions from (1) periglacial indicators, (2) an energy-balance climate model, (3) the most similar present-day pollen analog, and (4) a multivariate analysis of modern pollen and climate data at 135 sites (Solomon and Shugart 1984). Results from scenario (3) were unsatisfactory because simulated winter conditions were too cold for trees in the actual pollen record, and the climate model scenario (2) greatly reduced growth of the same trees caused by cold summers. Thus, neither of these two cases generated a satisfactory representation of the glacial forest. The remaining two scenarios produced simulated forests consistent with the pollen spectra. *Both were characterized by reduced seasonality of climate. However, in the periglacial case (1), winter temperatures were actually 1.4°C higher than at present.* The overall lesson from these results is that the *seasonal distribution of temperature and precipitation, and not just changes in the annual averages* as in most previous examples, is required to produce reasonable predictions of terrestrial ecosystem changes in response to future climate shifts. Achievement of this objective will require significant advances in both climate and terrestrial ecosystem modeling capability.

10.3.2.3 *Air Pollution Stresses*

Finally, a substantial body of circumstantial evidence seems to link major, apparently recent, instances of forest mortality or growth suppression in temperate regions with air pollution stresses related to acid rain, heavy metals, or photochemical oxidants (Johnson and Siccama 1983; Tomlinson 1983). The exact mechanisms contributing to these observed forest declines are still not well known. In some cases, regional changes in climate (e.g., frequency and severity of recent droughts) may play a significant role (Johnson and Siccama 1983). Thus, a potential amplifying effect of changing (o: fluctuating) climate must also be considered in analyzing the causes of these declines.

Taken in the context of the global carbon system, both the areal extent and the loss of biomass from northern forest declines are trivial thus far. However, our concern is for the future. The implications for long-term ecosystem stability, including eventual manifestation of similar effects in tropical regions, given expected population growth and economic development patterns, are troubling, and the potential interactions between climatic and anthropogenic stresses presents yet another complicating factor to be considered in interpreting the effects of CO_2-climate change on terrestrial ecosystems.

10.3.2.4 *Land-Use Conversion Combined with Other Factors*

The past holds at least part of the key to projecting the future. Extensive data bases on fossil emissions, energy use, population and economic growth, and atmospheric CO_2 increases exist and have been used extensively in efforts to project future growth rates of atmospheric CO_2 (Chapters 3, 4, and 9). An exercise as difficult and as fraught with uncertainty as projecting fossil fuel CO_2 emissions may yet be performed within a well-developed (albeit controversial) technical framework (Section 10.3.1; Chapter 9). No comparable sets of reliable data, or models for many key processes described above, yet exist for the terrestrial biosphere, let alone a technical basis for linkage with population growth and socioeconomic patterns for making projections. Projection of future carbon fluxes from terrestrial land-use change based on current information on past fluxes (e.g., Figures 10.1 and 10.2; Chapter 6) would certainly be very difficult.

Even more complications arise when one considers the two opposed suggestions that global large-scale reforestation be used to limit the growth of atmospheric CO_2 and the potential adverse effects of climate change (Brewer 1979; Dyson and Marland 1979), and that large increases in wood production be used to augment energy sources (Armentano 1984). These two options are obviously incompatible because periodically harvested forests contain only a fraction of the carbon they could store if allowed to grow to maximum biomass, and management for maximum sustained yield or financial return will reduce this fraction to one-third or less (Cooper 1983).

We agree with Schelling's (1983) conclusion that large-scale reforestation is not a realistic option for mitigation if one considers the most probable estimates of future CO_2 increase (also see Chan and Olson 1980). This picture could change if the climate response i

[6]Mid-Holocene climate identified as a possible scenario for a CO_2-warmed earth (Kellogg 1978).

more sensitive to CO_2 forcing than currently believed. Biomass energy projections are often incorporated into energy-economic modeling exercises (e.g., Edmonds et al. 1984), and thus specifics will not be discussed here. However, it should be noted that biomass production for energy may have an important impact on the carbon cycle. A recent analysis of fuel-wood production scenarios (Armentano 1984) suggests that U.S. forests will store much less carbon by 2010 than today, thus significantly changing their role in the global cycle.

Contemporary ecosystems outside the tropics may exhibit a small flux of CO_2 with the atmosphere because losses from present-day harvests are approximately balanced by forest regrowth. Hence, the principal focus today is on losses from tropical forests (Chapter 6). However, changing anthropogenic stresses on northern ecosystems could shift atmospheric CO_2 fluxes toward increasingly positive values in the future. This could result in a significant change in the role of northern forests, in particular, in the global cycle.

The key conclusion is that the dynamics of terrestrial landscape responses to combined anthropogenic changes (land use, CO_2 fertilization, CO_2-induced climate change, air pollution, etc.) are both complex and very poorly understood, and that the available data on which an assessment or projection of future response might be based are too fragmentary at present to allow anything more than formulation of hypotheses and recommendations for further research (Chapter 11; Solomon et al. 1984; Strain 1985).

Given this perspective, it is not surprising that few efforts to project future atmospheric CO_2 levels have quantitatively considered the terrestrial biospheric flux. Most often, the biospheric contribution was either ignored or an assumed "status quo" ca. 1980 was simply extrapolated over the projection interval. The effects of variations in these assumptions were examined by sensitivity exercises (e.g., Goudriaan and Ketner 1984), by using several alternative scenarios (see Section 10.3.4), or by application of Monte Carlo techniques (Section 10.3.5).

A "generic" forest-clearing scenario was developed by Trabalka et al. (1986) from historical reconstructions of biospheric fluxes (Figure 10.8; compare with scenarios in Figure 10.1 excluding Freyer and Belacy [1983] data). Although the rate of forest clearing was held constant after 1920, the average flux rate peaked at 1.5×10^{15} g C/year in 1920 and then decayed slowly to 1.2×10^{15} g C/year in 2075 because of partial forest regrowth (Figure 10.8). Deviations from the base forest-clearing scenario were simulated using Monte Carlo techniques and were applied equally to each year from 1800 to 2075 (i.e., varying the elevation of the curve t maintaining the basic shape). The dynamic range net flux from the biosphere to the atmosphere was approximately ±100% (e.g., 0.3×10^{15} to 2.7×10^{15} g C/year in 1980 [cf. Section 10.2.5] and 0×10^{15} to 2.5×10^{15} g C/year in 2075).

Given current uncertainties, and controversy about the role of the biosphere, this crude "generic" scenario and the simulated range of terrestrial fluxes obviously represents a compromise between extreme concepts of the future biosphere: on one hand, as a significant source (tropical forest clearing) and on the other, as a prominent sink (CO_2 fertilization). Experimental evidence does not currently provide compelling support for either extreme suggested above. In fact, analysis of the atmospheric CO_2 record strongly supports the view that the net contemporary biospheric flux with the atmosphere is masked by ignored or unknown sinks (Chapter 3; Section 10.2). Further scenario refinement does not appear to be warranted because a potentially infinite array of scenarios can be generated with currently plausible data, models, and assumptions (e.g., Figure 10.2; also see Enting and Pearman 1986).

The upper range of the generic forest-clearing scenario, which simulates a significant net biospheric flux with the atmosphere ($\sim 2.6 \times 10^{15}$ g C/year) from 1980 to 2075, appears conservative (generous), but may represent a hedge against, for example, the relaxation or reversal of a "chance compensation of opposed effects" (Broecker et al. 1979) or the manifestation of CO_2 climate effects after 1980. Much larger flux rates were proposed, but were probably ruled out by both unrealistic assumptions (e.g., Section 6.5.1) and the increasing incidence of successional forests in the tropics (e.g., Brown and Lugo 1984; but see Fearnside 1985). Concepts of a large net biospheric carbon sink over the next century fly in the face of realities that include a projected minimum doubling of human population in tropical regions (e.g., Edmonds et al. 1984; Myers 1984) and, thus, continued pressure on tropical forest resources (see Fearnside 1985), as well as probable increasing exploitation of Northern Hemisphere resources for fuel and fiber (Armentano 1984). Thus, a pragmatic evaluation suggests, at best, maintenance of an approximately neutral biosphere as the lower extreme scenario for projection purposes (also see Sections 10.3.2 and 10.3.5; Strain 1985).

10.3.3 Methods for Projecting Future Atmospheric CO_2 Levels

The uncertainties in our current understanding of the carbon cycle (Section 10.2) and resultant flaws in existing carbon cycle models have sometimes led to the conclusion that the constant airborne fraction (AF), an empirical estimate of atmospheric CO_2 retention (Chapter 3; Section 10.2.8) and one of the simplest of models, is a more satisfactory and more accurate

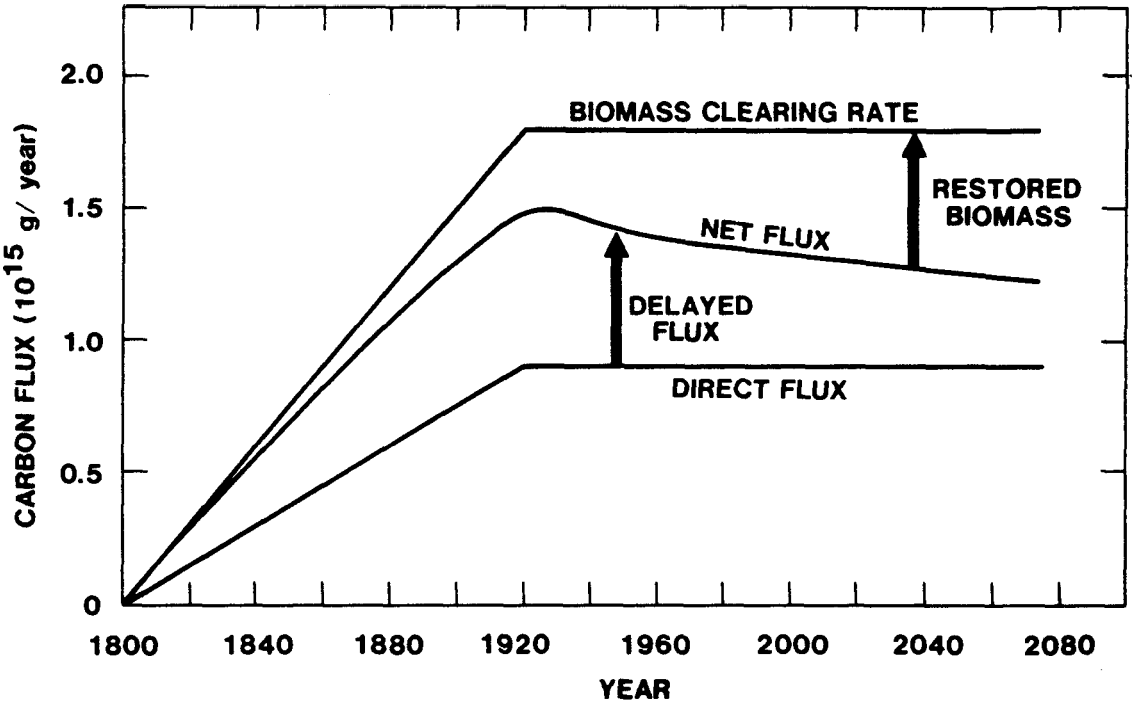


Figure 10.8. Baseline scenario of forest-clearing rate and atmospheric CO_2 fluxes from direct deforestation, delayed decomposition processes and forest biomass regrowth for the period 1800 to 2075 (Source: Trabalka et al. 1986).

alternative for atmospheric CO_2 projections.[7] In the very short term, a few decades or so, this is probably true (see Laurmann and Spreiter 1983). However, the chief advantage of the AF concept, simplicity, is also its major drawback. The global carbon system responds to the total amount of CO_2 in the atmosphere (past, present, future) and not to the small annual increments from fossil fuels and terrestrial land-use changes (Chapters 3, 5, and 7). Despite existing uncertainties, the ability to simulate *changing* conditions in the global carbon balance appears to offer simulation models a clear advantage over the AF in long-term projections. Predictions using the two methods can thus diverge considerably on 50- to 100-year time scales (see Section 10.3.4). Nonetheless, because the AF can take on a variety of forms (e.g., see Section 10.2.8) and has received considerable attention in the literature, it is worthwhile to discuss its various forms and limitations in comparison with carbon cycle models.

Agreement on even the definition of AF has not been possible because of uncertainty about the sign, magnitude, and rate of change of the terrestrial biospheric flux over time (Woodwell et al. 1983) and, t a lesser extent, the choice of atmospheric measuremer data (e.g., time period, location) to be used (Chapter 3 Bacastow and Keeling 1979; Pearman and Beardsmor 1984; Perry 1984). A number of authors (e.g., Nordhau and Yohe 1983; Clark et al., 1982; Laurmann an Rotty 1983; Woodwell et al. 1983), citing evidence for probable terrestrial biospheric flux (Q_b) of 1×10^{15} to $\times 10^{15}$ g C/year during the Mauna Loa record resultin from land-use changes (e.g., Chapter 6), argue that th more commonly used expression of the observed A should include a term for this flux in the denominato of the "fraction" (i.e., a term in addition to ΔQ_f; se Section 10.2.8). This results in estimates of the "total airborne fraction (0.33–0.47 range) that are lower tha the observed fossil fuel AF (0.58). Note that thi concept of a biospheric correction for the AF is quit different from that posed in Section 10.2.8.

The use of a constant total AF is an oversimplifica tion. The basis for its use in making CO_2 projection is a set of rather tenuous assumptions: (1) effect of past releases can be ignored; (2) future biospheri releases will increase exponentially *and at the sam rate as fossil fuel emissions* to maintain a constar AF; and (3) the hypothetical "missing sink" of carb

[7]The simplest "model" would involve an extrapolation of the contemporary atmospheric trend; however, see Figure 3.15 and associated discussion.

needed to "balance" the global cycle by absorbing the excess biospheric carbon will *always* remain equal in agnitude (and opposite in sign) to excess carbon. hese assumptions are challenged by the analyses and discussion in Sections 10.2.7 to 10.2.9, and are particularly questionable if the "missing sink" is caused by "chance compensation of opposed effects," for example, during the contemporary period (Broecker et al. 1979).

In the Nordhaus and Yohe (1983) formulation of the total AF, the effective airborne fraction, AF*, was adjusted to a value of 0.47 in 1975 (range 0.38–0.59) based on the assumption that Q_b had averaged $1(\pm 1) \times 10^{15}$ g C during the Mauna Loa record (and would continue to do so to the year 2100). Incorporation of an empirical ocean seepage term to accommodate the excess carbon from Q_b during the period of the Mauna Loa record allowed AF* to vary by approximately ±20% with variations in rates of future CO_2 emissions. However, the potential uncertainty in AF* appears greater than that suggested by a 0.38–0.59 range (Section 10.2.8), and because only a fraction of the ocean carbon sink (associated with the seepage factor) is permitted to vary with atmospheric CO_2 content (cf. Section 5.2.1), the Nordhaus and Yohe (1983) variant does not produce the dynamic range of responses attainable with carbon cycle models.

Is any definition of the airborne fraction useful in making projections? If the global carbon cycle can be represented by a linear mathematical model with constant coefficients, an exponentially rising input of CO_2 to the atmosphere would imply a constant AF after times that are long compared to the time constant of the exponential input (Bacastow and Keeling 1979). This does not appear to be a reasonable set of assumptions because several have already been or will soon be violated, including:

- Past emissions rate changes from 1914 to 1945 and since 1973 (Chapters 3 and 4);
- Emissions scenario uncertainties in the future (Chapter 9; Siegenthaler and Oeschger 1978; Perry 1984; Section 10.3.1);
- Potentially nonexponential historic releases of CO_2 to the atmosphere due to land-use changes (Figures 10.1 and 10.2; Chapter 6);
- Nonlinearity in the ocean buffer factor with increasing atmosphere CO_2 content (Chapters 5 and 7);
- Potential nonlinear response of land biota in the future caused by combined land-use change, climatic, CO_2 fertilization, and air pollution effects (Section 10.3.2);
- Unknown behavior of missing sinks; and
- Potential nonlinear response of ocean uptake of CO_2 because of climatic effects on biotic response, CO_2 exchange, and circulation patterns (Chapters 5 and 7).

Further, the rate of growth in atmospheric CO_2 content varies significantly on an annual basis (e.g., Mauna Loa data in Figures 2.1 and 3.16). It is apparent that the magnitude of the variability in the associated carbon cycle processes that produce these effects cannot be ignored (Section 10.2.8; Keeling 1983; Chapter 3). Calculating the AF for each individual year during the period of the Mauna Loa record produces a wide range of values from 0.2 to 1.1. Thus, the empirical evidence supports the argument made on theoretical grounds that the magnitude of the AF is a time-dependent variable, not a constant, that will respond to a variety of changes in the global carbon cycle, including those induced by climatic variations (Section 3.4.3). Nonetheless, the suggestion that a constant AF may be used to represent the complex, global carbon system is a pervasive concept.

The carbon cycle model-derived AF, on the other hand, is dependent both on the model and on the input scenarios for *past and future* CO_2 emissions; AF *varies continuously* with time for a given model/scenario combination. At very high future fossil fuel CO_2 release rates, there are limits on annual fractional mass transport from the atmosphere to the mixed layer because of the buffer effect, resulting in significant increases in AF (Chapters 5 and 7). Potential decreases in AF can also occur if emissions rates decrease or even remain relatively constant in the future (Perry 1984). The dynamic range in AF obtained with carbon cycle models (same model, different scenarios; Table 10.6; also see Perry 1984) is significantly greater than that attainable with any formulation of the AF. In this respect, global carbon cycle models, even with current limitations, have a distinct advantage in long-term projections of future atmospheric CO_2 levels.

The airborne fraction should not be represented exclusively as a constant because it is a system-dependent and, therefore, a time-dependent variable. The effective AFs calculated in Section 10.2.8 represent only "starting values" for CO_2 projections. These values are *expected* to change over time, for the reasons cited above. Of greatest concern is the fact that the magnitude could increase significantly in the future as a result of the combined effects of the buffer factor and potential CO_2-climate feedbacks on the carbon cycle. Thus, constant AFs appear to be useful only for relatively short-term projections (i.e., before major changes occur in the global carbon system, ~25 years into the future [Laurmann and Spreiter 1983; Section 10.3.4]). Although we may be forced to accept constant

Table 10.6
Model Projections of Airborne Fraction from 1860 to 2230, to 2280, and to 2130

Airborne Fraction 1860 to			References[a]		
2030	2080	2130	Source of Projection	Model Used	Scenario Used
0.73	0.80	0.77	Sie 78	Sie 78-1	Sie 78-H
0.58	0.54	0.55	Sie 78	Sie 78-1	Sie 78-L
0.65	0.68	0.60	Sie 78	Sie 78-2	Sie 78-H
0.51	0.46	0.43	Sie 78	Sie 78-2	Sie 78-L
0.72	0.84	0.81	Per 84	Bjo 79	Sie 78-H
0.61	0.73	0.72	Per 84	Bjo 79	Per 84-A2
0.58	0.63	0.57	Per 84	Bjo 79	Per 84-A3
0.54	0.53	0.44	Per 84	Bjo 79	Per 84-A4
0.74	0.82	0.80	Per 84	KE3B	Sie 78-H
0.69	0.74	0.73	Per 84	KE3B	Per 84-A2
0.67	0.67	0.62	Per 84	KE3B	Per 84-A3
0.66	0.61	0.55	Per 84	KE3B	Per 84-A4
0.61	0.58	0.55	Per 84	KE3B	Per 84-A5

Note: The atmospheric CO_2 concentration in 1860 was assumed to be 290– 295 ppm.

[a]See key to Table 10.6 references:

Bjo 79	Björkström (1979); model with biota growth (β) factor modified by Perry (1984).
Per 84	Perry (1984).
Per 84-(A2-A5)	From Perry (1982); CO_2 release scenarios A2 (highest), A3, A4, and A5 (lowest).
Sie 78-1	Siegenthaler and Oeschger (1978); model without biota.
Sie 78-2	Siegenthaler and Oeschger (1978); model with biota growth (β) factor.
Sie 78-H	Siegenthaler and Oeschger (1978); high CO_2 release scenario, also used by Perry (1982) as his extreme high release scenario.
Sie 78-L	Siegenthaler and Oeschger (1978); low release scenario.
KE3B	Killough and Emanuel (1981); model 3b.

AFs to simulate lower bounds in longer term projections (i.e., because models cannot now generate effective AF values as low as 0.5 [Section 10.2.8]), we must also recognize that such results probably represent conservative (low) estimates. The axiom "simplest is best" is a concept that seems inappropriate when applied to a system as complex as the global carbon cycle.

10.3.4 Reference Projections with Globally Averaged Carbon Cycle Models

The three reference fossil fuel emissions scenarios (Cases A, B, and C) described in Section 10.3.1.1 were used as inputs to two globally averaged ocean carbon cycle models (Edmonds et al. 1984; Trabalka et al. 1986). The two carbon cycle models used the three reference projections of fossil fuel CO_2 emissions and three potential terrestrial biospheric release scenarios (zero net flux, MBL, and Freyer and Belacy [p] Figure 10.1) to construct a spectrum of 18 atmospher CO_2 projections to the year 2075 (Figure 10.9). The restrial biospheric release scenarios were hypothetica extended into the future by making the assumption th the annual release rates in 1980 remained constant un 2075. Given the current uncertainties about the role the biosphere, the assumption of a constant annual fl after 1980, but representing a range of 0.0×10^{15} 2.7×10^{15} g C, was deemed a suitable compromi (Edmonds et al. 1984; also see Sections 10.3.2.4 a 10.3.5).

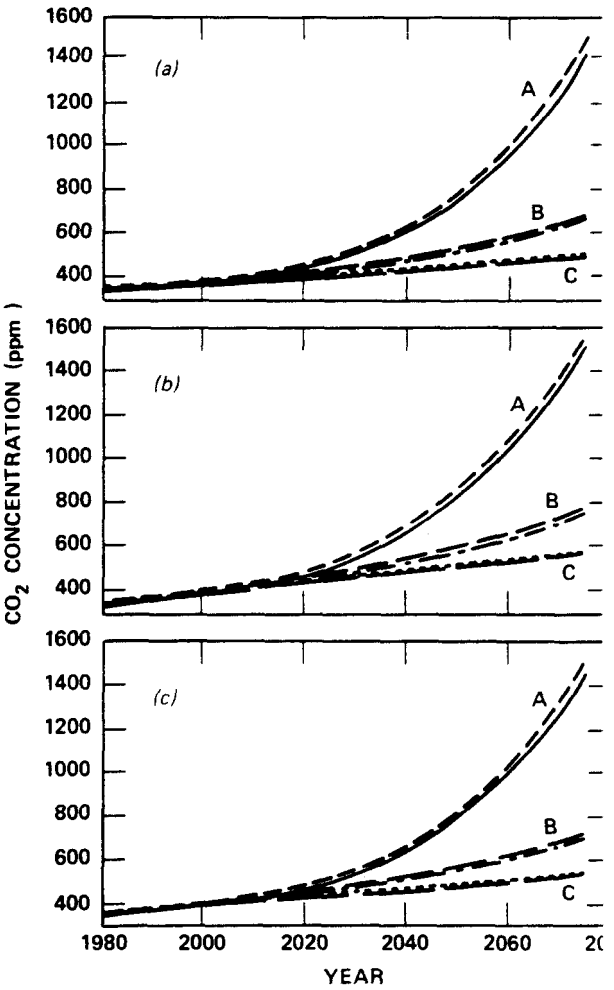


Figure 10.9. Simulated atmospheric CO_2 concentration. (a) terrestrial source (neutral case); (b) terrestrial source based on [13] data; (c) terrestrial source based on land-use and demographic dat A = high fossil fuel CO_2 emissions case; B = standard fossil fuel C emissions case; C = low fossil fuel CO_2 emissions case Upper cur represents the response of the modified Björkström (1979) model a the lower curve the response of the Emanuel et al. (1984b) model each paired comparison for a given emissions case (Source: Edmo et al. 1984).

Table 10.7
Model-Calculated Future Atmospheric CO_2 Concentrations Based on Global Carbon Emissions Cases A, B, and C for Fossil Fuels Combined with Various Biospheric Inputs for the Years 2000, 2025, 2050, and 2075

	Atmospheric CO_2 Concentration (ppm) (Year)			
Case	2000	2025	2050	2075
A	380–400	480–530	760–840	1400–1550
B	380–400	440–480	540–600	670–760
C	370–400	420–460	470–520	510–580

Note: See text for a full description of Cases A, B, and C (Section 10.3.1).

The two ocean models are radically different in structure (Chapter 7); the modified Björkström formulation is an advective/diffusive model, whereas the Emanuel et al. (1984b) model is diffusive in character, with diffusivity effectively varied with depth (see Edmonds et al. 1984). The Emanuel et al. model is able to provide a satisfactory fit to the Mauna Loa trajectory, assuming past inputs of fossil fuels alone (Edmonds et al. 1984). The CO_2 uptake range spanned by simulations with these two models is highly representative of state-of-the-art formulations (fossil fuel AF* for 1958–1980, 0.67 and 0.61, respectively; cf. Table 5.4), but was not intended to cover the entire range in the published literature. The range in effective airborne fraction (1980–2075) was expanded in part by forcing these models to accept both fossil fuel and biospheric fluxes (AF* = 0.58 to 0.80 with the combinations described above; cf. Section 10.2.8).

The results from the carbon cycle modeling exercise (Edmonds et al. 1984) are shown both in Figure 10.9 and in Table 10.7 for reference in the discussion that follows. The data in Table 10.7 are presented as a range of values for each date and each fossil fuel emissions case, because all six combinations of the two globally averaged carbon cycle models and the three potential terrestrial biospheric release scenarios were run for each emissions case. The range of values serves to provide the reader with some sense of the uncertainty contributed by carbon cycle modeling to the overall projection exercise (also see later discussion in Section 10.3.5). Table 10.7 shows that, given the carbon cycle models and the fossil fuel/terrestrial scenario combinations used, variations in the fossil fuel emissions cases clearly dominate the atmospheric CO_2 response by year 2050. A relatively tight range of projected atmospheric CO_2 concentrations is produced for the year 2000 (370- to 400-ppm range will accommodate results from all cases), but significant divergence occurs after the year 2025. Median atmospheric CO_2 concentrations are 540, 710, and 1480 ppm in 2075 using emissions Cases C, B, and A, respectively.

The results of projections based on calculations using assumed constant AFs are presented in Table 10.8 (Edmonds et al. 1984). Atmospheric concentrations calculated with constant AFs do not deviate radically from carbon cycle model projections until after 2025. Deviations become quite large thereafter, particularly for calculations with total AFs using emissions Case A; projected concentrations are as much as 40% lower in 2075. The possible reasons for these deviations, including anomalies associated particularly with total AFs, were discussed previously. It appears, however, that projections based on AF concepts may not differ markedly from those based on carbon cycle models on time scales of 25 to 50 years (also see Laurmann and Spreiter 1983). The analysis in Section 10.2.8 suggested that the low estimate of the effective airborne fraction could be 0.5. Currently, carbon cycle models cannot simulate this condition without resorting to unrealistic assumptions on ocean uptake processes or biospheric CO_2 fertilization (cf. lowest values in Tables 10.2, 10.3, and 10.6). However, the assumption of a constant AF of 0.5 to provide the lower bound estimate for all atmospheric CO_2 projections is also questionable because:

1. The value of 0.5 is already a conservative (low) estimate (Section 10.2.8);
2. Changes in ocean chemistry at high atmospheric CO_2 levels (i.e., emissions Case A) should increase the AF significantly over the projection interval (~300-ppm difference in atmospheric CO_2 levels in 2075; cf. Table 10.8 with Table 10.7) unless ocean uptake is drastically misrepresented by current simulations (but see Siegenthaler 1983);
3. Missing sinks would have to absorb this excess CO_2 to maintain a constant airborne fraction;
4. The living terrestrial biomass would either have to double in size ($\sim 600 \times 10^{15}$ g) or an equivalent amount of carbon would have to be stored in soils, sediments, or peat to accommodate the latter condition; and
5. This set of assumptions cannot be supported on the basis of current information, for example, average uptake rate of the "missing sink" would be $\sim 6 \times 10^{15}$ g C/year over the 1975–2075 time interval.

Table 10.8
Airborne Fraction-Calculated Future Atmospheric CO_2 Concentrations Based on Global Carbon Emissions Cases A, B, and C for Combined Fossil Fuel and Terrestrial Biospheric Emissions for the Years 2000, 2025, 2050, and 2075

		CO_2 Concentration (ppm)			
		Terrestrial History (airborne fraction)			
		Neutral Biosphere[b]			
Case[a]	Year	(0.58)[c]	(0.50)[c]	$\delta^{13}C$ (0.43)[d]	MBL (0.34)[e]
C	2000	370	360	370	370
	2025	410	400	410	400
	2050	460	440	450	440
	2075	510	490	490	480
B	2000	370	370	370	370
	2025	430	420	420	410
	2050	510	490	480	470
	2075	630	590	580	550
A	2000	370	370	370	370
	2025	460	440	440	430
	2050	670	630	600	570
	2075	1100	1000	950	850

Note: See text for a full description of Cases A, B, and C (Section 10.3.1).
[a]Biospheric releases of 1.2×10^{15} and 2.7×10^{15} g C year, respectively, were added to fossil emissions scenarios from 1980 to 2075 in calculations based on the $\delta^{13}C$ and Marine Biological Laboratory (MBL) terrestrial histories.
[b]Assumes zero net atmosphere-biosphere flux from 1800 to 2075 (fossil fuel airborne fraction).
[c]Values based on calculations for 1959-1982 described in Section 10.2.8.
[d]Airborne fraction reduced from 0.58 to 0.43 to account for addition of 1958–1980 biospheric contribution (1.29×10^{15} g C/year) estimated from ^{13}C deconvolution based on Freyer and Belacy (1983) tree-ring data.
[e]Airborne fraction reduced from 0.58 to 0.34 to account for addition of 1958–1980 biospheric contribution (2.75×10^{15} g C/year) estimated from MBL terrestrial reconstruction (Houghton et al. 1983).

We believe, therefore, that calculations based even on the observed AF (0.58; Table 10.3) provide a *conservative lower* bound on future atmospheric CO_2 levels if high levels of CO_2 emissions are assumed. A compromise, but still conservative, solution could include calculations with the lower value of 0.5 for emissions Case C in combination with results using the observed value (0.58) for emissions Cases A and B (Table 10.8).

Given the existing uncertainties about the historic level of atmospheric CO_2 for climatic response projections (260-300 ppm), the estimated date for a doubled atmospheric concentration falls between 2025 and 2045 in results from Case A, between 2035 and 2070 using Case B, and from 2050 to well beyond 2075 for Case C. Thus, this projection exercise indicates that doubling of atmospheric CO_2 is unlikely to occur before the year 2025 and may not occur within the next century. This conclusion is very similar to that reached by Nordhaus and Yohe (1983) using a totally different methodology for energy-CO_2 emissions and atmospheric CO_2 projections.

Long-term atmospheric projections obviously are not made without some trepidation, given the current state of the art. However, some of our uncertaint may be irreducible; thus we cannot remain immobilized ostrich-like, by excessive caution. We do, however need to analyze potential contributions to uncertaintie as quantitatively as current understanding permits.

10.3.5 Carbon Cycle Uncertainty Analyses

10.3.5.1 *Background*

A variety of techniques have been developed t determine the dependence of predictions of simulatio models on inputs, initial conditions, and parameters. It is important to select from these techniques the one mo appropriate for the particular model and question(s being examined.

Sensitivity studies are often employed to examin the impact of variations in individual parameters o model output, and are generally an integral part o model development (e.g., Broecker et al. 1979; Killoug and Emanuel 1981; Siegenthaler 1983). Parameter have been varied individually (e.g., Laurmann an Spreiter 1983; Machta 1983; Goudriaan and Ketne 1984), in systematic combinations (Bacastow a Björkström 1981), or simultaneously in their entire

for efficiency (Enting and Pearman 1986; Gardner and Trabalka 1985). Such analyses provide guidance to those aspects of carbon cycle models likely to be significant in predicting future atmospheric CO_2 concentrations, but reveal nothing about the ability of models to represent nature except perhaps by indirect inference (Machta 1983). For example, virtually all sensitivity studies conducted on the current generation of ocean models indicate that the patterns and rates of vertical mixing beneath the surface mixed layer are the most critical parameters (Chapter 7), yet current representations of this process in models are not physically realistic (Chapters 5, 7, and 11; Brewer 1983).

Likewise, most sensitivity studies on models containing a biospheric component suggest that varying the so-called β-factor, the surrogate parameter for a hypothetical carbon sink attributed to a CO_2 fertilization effect, could have a major effect on predictions of atmospheric CO_2. However, it is not realistic to adopt a model parameterization for CO_2 fertilization based on current information (Section 10.3.2). Thus, the β-factor is most often used currently as a "garbage can" (e.g., see Goudriaan and Ketner 1984) to absorb excess carbon not taken up by the ocean models, and thus bring recalcitrant models into agreement with contemporary atmospheric measurement data. When β is set equal to zero (no fertilization effect; our preferred case at the present time), sensitivity studies indicate that parameters that control the rate and timing of terrestrial releases (e.g., forest-clearing rates, biomass densities, and succession) are the most significant biospheric model contributors to prediction uncertainty (Houghton et al. 1983; Chapter 6). However, the current treatment of carbon succession accompanying human disturbance is highly qualitative because of the paucity of field measurements (Chapters 7 and 11).

Sensitivity studies do not provide estimates of overall (combined) parameter uncertainty contributions to model predictions. One approach to simulate carbon cycle uncertainties was described in Section 10.2.8 (variations in effective airborne fraction). Another, designed to incorporate uncertainties in future fossil fuel CO_2 emissions scenarios, was discussed in Section 10.3.4 (permutations and combinations of fossil fuel emissions cases, terrestrial flux reconstructions, models, and airborne fractions) (also see Bacastow and Björkström 1981). Monte Carlo methods are also appropriate for questions involving carbon cycle simulations (e.g., Nordhaus and Yohe 1983).

10.3.5.2 *Monte Carlo Studies*

Recent developments have drastically improved both the efficiency and the precision of Monte Carlo techniques (Gardner in press). However, there are limitations in using this analytical tool with carbon cycle models because of the interdependence of model parameters and contemporary patterns of atmospheric CO_2 increase.

Each Monte Carlo iteration requires three steps: (1) selection of random, independent sets of values for parameters, (2) initialization of the carbon cycle model, and (3) simulation of global carbon cycle dynamics from year 1800 to year 2075 (see Gardner and Trabalka [1985] for details). Initialization of the carbon cycle model involves a calibration step in which parameter values and initial conditions are calculated, consistent with an assumed steady-state atmospheric CO_2 content and radiotracer depth-distribution in the ocean. Similarly, the terrestrial biota are assumed to be at equilibrium. Carbon quantities change during a simulation as a function of inputs resulting from the combustion of fossil fuel and releases of carbon from the terrestrial biota as a consequence of forest clearing. The initial conditions and the time-dependent releases of carbon from fossil fuel emissions and forest clearing from 1800 to 2075 constitute a particular model scenario.

A problem arises because, although model simulations may approach the trajectory of the modern atmospheric CO_2 record under baseline conditions (parameters at midrange or selected best-estimate values), Monte Carlo variations in parameters tend to deflect the iterative solutions away from the baseline trajectory. In practical terms this means that one must either (1) relax the condition that the model output satisfactorily approaches the trajectory of the modern atmospheric CO_2 record on its path from an assumed 19th century steady state toward the future, or (2) reject simulations that do not reasonably satisfy this constraint. The decision is based on the information desired: condition (1) is used for statistical analysis of relative parameter contributions to model output errors, whereas condition (2) is better for setting bounds on atmospheric futures because more use is made of all existing data. However, the latter can require a very large number of model runs to obtain a sample of outputs large enough for a meaningful analysis.

Nordhaus and Yohe (1983) overcame this difficulty by starting all simulations in the year 1975 and by using an AF model of the carbon cycle (see Section 10.3.3) that was subjected to variation in making predictions of future CO_2 levels. A separate experiment was conducted to rank the relative contributions of individual model parameters, including the AF, to variability in predicted concentrations of atmospheric CO_2 in the year 2100 (end point of their projection exercise). It was not possible to estimate the absolute contribution of each parameter to the variance in the atmospheric CO_2 level based on the information provided, but

Table 10.9
Percentage Contributions From Carbon Cycle Model Parameter Variations to Variance in Atmospheric CO_2 Concentrations With Future Fossil Emissions Case B

Parameter (Range of allowable variation)	Year 1959–1980[a] Mean	1959–1980[a] Trajectory[b]	2025	2075
Forest-Clearing Rate (±80%)	48.3	22.8	35.1	21.6
Ocean Tracer Profile[c]	16.5	46.1	41.9	61.1
Initial Atmospheric CO_2 Concentration in 1800 (240–280 ppm)	27.2	0.9	12.8	6.7
Pre-1980 Fossil Emissions (±12%)	1.6	22.8	0.9	0.3
2-x-CO_2 Temperature Increase in Ocean Surface Layer (1.5°–4.5°C)	0.7	1.8	1.6	2.1
Others (±20%)[d]	4.4[e]	3.6[e]	6.0[e]	5.9[e]
Total Variance Explained	98.7	98.0	98.3	97.7

Note: Analysis of 600 Monte Carlo iterations (Trabalka et al. 1986; discussion of parameter ranges in Gardner and Trabalka 1985).

[a]Mauna Loa record (Keeling et al. 1982).
[b]Deviation of model-calculated atmospheric CO_2 trajectory from observed trajectory represented by annual mean values.
[c]Range of equivalent vertical diffusion coefficient calculated by the method of Oeschger et al. (1975) was 0.95–5.2 cm^2/s.
[d]25 parameters varied by ±20% to approximate uncertainty; however, initial ocean surface temperature in 1800 was allowed to vary only ±2°C (Gardner and Trabalka 1985).
[e]No individual contribution greater than 2.3% at any time.

uncertainty in five energy-economic parameters that determine future fossil fuel use produced the greatest impact on atmospheric CO_2 levels in the year 2100 (see Chapter 9). The AF contribution was next in level of importance. Potential uncertainties in long-term fossil CO_2 emissions scenarios also have been suggested to be predominant over uncertainties in carbon cycle models from earlier research (Laurmann and Spreiter 1983; Oeschger and Heimann 1983; Goudriaan and Ketner 1984; Perry 1984) and the projections in Section 10.3.4.

Monte Carlo simulations with a globally averaged carbon cycle model (described in Emanuel et al. 1984b; also see Killough and Emanuel 1981 and Emanuel et al. 1984a) also were applied to quantitatively estimate contributions from carbon cycle model uncertainty to atmospheric CO_2 predictions (Gardner and Trabalka 1985; Trabalka et al. 1986). A sensitivity analysis was used to identify those parameters requiring detailed specification of uncertainties. Ranges and uncertainties were approximated for 25 parameters by setting the probability distribution function to a uniform distribution with a range equal to ±20% of the nominal value. In most cases, this appears to be a satisfactory treatment of actual uncertainty (Gardner and Trabalka 1985), but because the results are insensitive to the ranges for these 25 parameters, the approximations do not affect prediction uncertainties.

The ocean model used in the uncertainty analysis was the Emanuel et al. (1984b) formulation used in reference projections described in Section 10.3.4. Atmospheric emissions were generated by the generic forest-clearing scenario described in Section 10.3.2 (Figure 10.8) in combination with the "standard" or median fossil fuel emissions scenario (Case B) from Section 10.3.1. Fossil emissions scenarios were not subjected to Monte Carlo variation because the focus was on comparative estimates of errors introduced by carbon cycle and fossil CO_2 emissions modeling. Thus, an analysis based on the median estimate of fossil fuel emissions provides useful comparative information (also see Laurmann and Spreiter 1983).

The principal contributions to variation in predicted atmospheric CO_2 from 1959 to 1980 were from uncertainties in the forest-clearing rate and in the tracer data that control the critical rate of vertical mixing in the ocean model (Table 10.9). Although forest clearing, past fossil emissions, and the estimate of the historic atmospheric CO_2 content were initially significant sources of uncertainty in atmospheric CO_2 projections, their importance declined with time. (The contribution of the initial atmospheric CO_2 parameter to prediction uncertainty could be lower than indicated in the table because of recent data improvements [Chapter 3; Section 10.2.1], but the parameter rankings should be unaffected.) By the year 2075, uncertainty in the ocean model was most important because fossil emissions (19×10^{15} g C in Case B) constitute over 90% c the atmospheric CO_2 inputs. Thus, atmosphere-ocea

exchange dominates the global carbon system by 2075, 'd the ocean tracer data, the surrogate for uptake by .e deep ocean CO_2 sink, contribute most to uncertainty in predictions of atmospheric CO_2. (This is hardly a surprising result and could have been predicted a priori based on past sensitivity studies.)

Contributions from only 3 of the 31 model parameters subjected to variation provided a satisfactory explanation of uncertainties in future atmospheric CO_2 levels (Table 10.9). The other 28 parameters combined contributed less than 9% of the variation, and additional efforts to remeasure these parameters or other exercises designed to "fine-tune" these values will fail to appreciably reduce prediction uncertainties unless significant changes are made in the structure of the model and/or in the emissions scenarios. The effects of such changes have yet to be quantitatively analyzed in a systematic treatment.

This limitation makes it difficult to generalize the results from an uncertainty analysis (e.g., to establish priorities for future research). The structural limitations of globally averaged models provide yet another obstacle to interpretation. For example, the forest-clearing rate in Table 10.9 is a carbon loss rate (Figure 10.8) from globally averaged pools of carbon stored in trees, which are then redistributed among the atmosphere and other terrestrial pools (Emanuel et al. 1984a). Forest area is not a model parameter. Thus, critical geographical and ecosystem-specific variations in terrestrial parameters identified in Chapters 6 and 7 (e.g., tropical forest areal changes, stand carbon densities, and carbon succession patterns) are not treated in this globally averaged representation of the carbon cycle. Thus, it would be unwise to infer that the parameter rankings in Table 10.9 represent much more than reinforcement of the need for model improvements (i.e., more physically realistic ocean models and geographically resolved terrestrial models) (Chapters 5 to 7).

Higher rates of fossil emissions would produce an even stronger dependence of atmospheric CO_2 on variability in ocean model parameters for reasons discussed above, whereas significantly lower fossil fuel release rates (e.g., emissions Case C) would increase the relative importance of terrestrial parameters—exactly how much depends on the scenarios. However, some idea of the latter may be obtained by comparing the results from simulations for 2025 with those for 2075 in Table 10.9; annual fossil emissions in the 1980–2025 period in Case B (5.3×10^{15} to 10×10^{15} g C) are reasonably comparable with the post-1980 annual releases in Case C (6.2×10^{15} to 6.8×10^{15} g C). Future improvements in model structure are likely, but their pact currently is very difficult to predict. Thus, it ..ould be clearly understood that the results in Table 10.9 are unique to the structural characteristics of this carbon cycle model, to fossil emissions Case B, and to the generic terrestrial release scenario.

How robust are the conclusions derived from data in Table 10.9 with respect to variations in the biospheric flux time series? Without answers to questions posed in Sections 10.2.9 and 10.3.2, the answer is bound to be incomplete and the response is unlikely to satisfy all critics. The bulk of the available evidence, including the most recent ice-core data, supports the concept of the peak nonfossil (biospheric?) flux to the atmosphere having been reached before 1940 (Section 10.3.2). Thus, if the annual fluxes over the 1980–2075 period fall within the range of uncertainty in 1970–1980 values (0.0×10^{15} to 2.6×10^{15} g C), then the results in Table 10.9 appear to represent a reasonable statement. Those who believe either that net tropical forest-clearing fluxes have increased recently at a rapid rate and could increase significantly in the future (Section 6.5.1), or that a significant net biological sink from CO_2 fertilization exists and could also increase over the next century (discussion in Sections 10.3.2 and 10.3.5), are likely to view the results as conservative, underplaying the significance of the biosphere, albeit for totally different reasons. Those who feel that the role of the biosphere is currently negligible and will be relatively insignificant for the foreseeable future will likely conclude that the interpretation of the importance of the biosphere suggested by Table 10.9 is too liberal.

The key point is that there exists significant unresolved uncertainty, and some controversy, because current data are quite contradictory. What is needed are new data and analyses from careful research designed to resolve the outstanding issues about the role of the biosphere in the carbon cycle (Chapter 11). Otherwise, we may be continually engaging in speculation. If it becomes possible to develop an accurate time series for atmospheric CO_2 concentrations from ice-core data (e.g., Neftel et al. 1985), more detailed scrutiny of existing terrestrial time series may be warranted (also see Enting and Pearman 1986).

Only 47 of the 600 model solutions (Trabalka et al. 1986) were able to meet a constraint that they pass through a ±2-ppm "window" around the average measured atmospheric CO_2 concentration (Mauna Loa record, 1959–1980, Keeling et al. 1982) on their path into the future. The effect of the window on model output was to reduce the variance in atmospheric CO_2 predictions and to strengthen the dependence of future atmospheric CO_2 levels on ocean model parameters, akin to the effect produced by increasing fossil emissions rates. This occurred primarily because very high rates of forest clearing were then incompatible with the ranges adopted for other parameters, particularly for initial atmospheric CO_2 content.

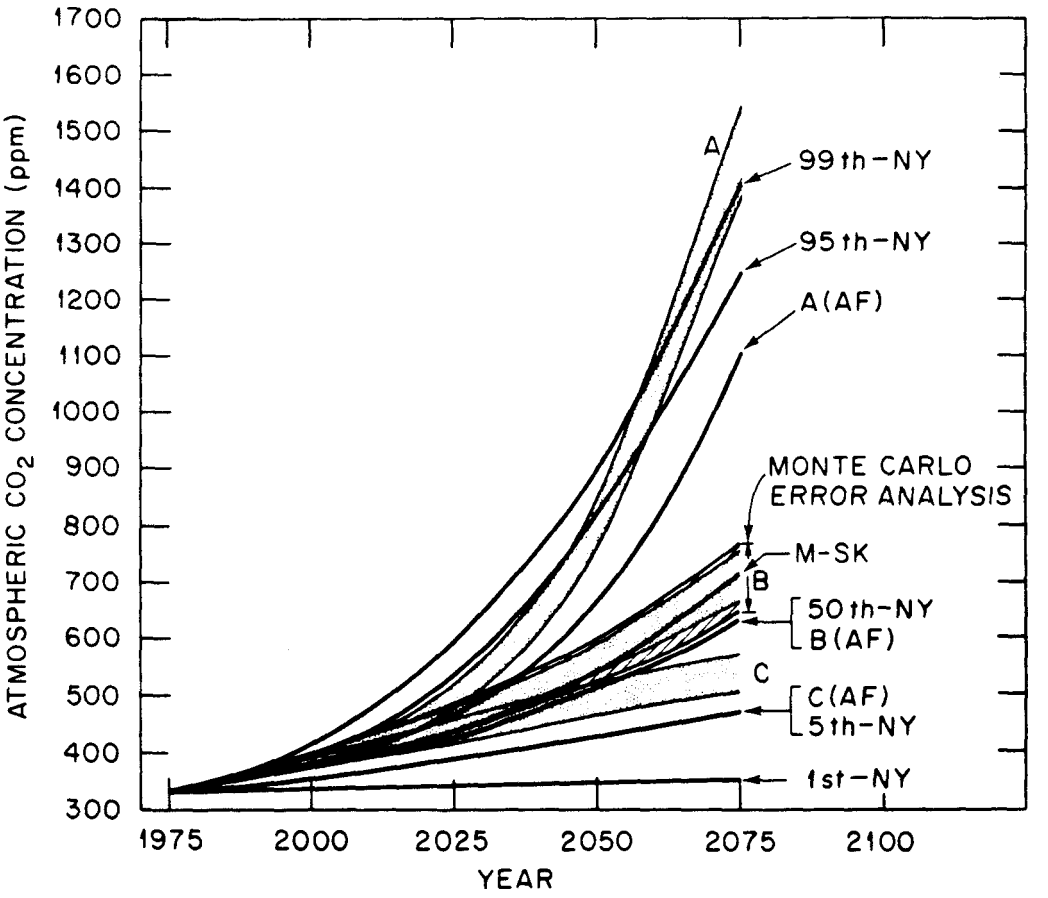


Figure 10.10. Projected atmospheric CO_2 trajectories (and ranges of uncertainty) for carbon cycle simulations with emissions Cases A, B, and C, along with selected examples from other recent studies. (See key to Figure 10.10 below.)

Results of six simulations representing two carbon cycle models and three potential histories of terrestrial biospheric flux for each emissions case (shaded areas); constant AF calculations (A[AF], B[AF], and C[AF]). AF of 0.58 used for Cases A and B, and 0.5 for Case C. The 95% confidence band of uncertainty associated with Case B from a Monte Carlo analysis of parameter uncertainties (shaded area plus cross-hatching) for one of the two models (see text).

Individual CO_2 Projection Cases:

99th-NY, 95th-NY, 50th-NY, 5th-NY, 1st-NY	High (99th percentile), 95th percentile, median (50th percentile), 5th percentile, and low (1st percentile) atmospheric CO_2 trajectories from Nordhaus and Yohe (1983).
M-SK	Midcase atmospheric CO_2 trajectory from Seidel and Keyes (1983).

Why did relatively few of the model runs pass through the Mauna Loa window to the future? This result is simply a reminder that our current understanding of the carbon cycle is unsatisfactory (see Figure 10.2); no deeper meaning should be sought. Although this application of Monte Carlo techniques may seem rather inefficient, the results obtained with the window serve to reinforce several key points: the importance of the contemporary atmospheric CO_2 measurement data as constraints on models; the high potential variability in some carbon cycle model parameters, particularly those used to simulate the past role of the biosphere; and many other limitations of current models (Section 10.2; also see Chapter 11).

10.3.6 Uncertainties in Atmospheric CO_2 Projections

10.3.6.1 *Reference Projections*

Results from the restricted set of 47 Monte Carlo iterations of the carbon cycle model using future fossil fuel emissions Case B (Trabalka et al. 1986) are shown in Figure 10.10 along with results from Section 10.3.4

for comparison with other recent CO_2 projections. Within-model variation (i.e., as estimated using one model with a Monte Carlo analysis of parameter uncertainties) was somewhat greater than, but still quite comparable to, variation introduced by using two carbon cycle models and by varying terrestrial history (cf. Case B uncertainties in Figure 10.10). The estimated contribution from the carbon cycle model component to *overall* uncertainty in projected atmospheric CO_2 concentrations declines with time from 1975 to 2075, even though the absolute magnitude of the uncertainty (in ppm CO_2) increases. Case A projections are distinguishable from the carbon cycle "noise" by 2025, and Case B and Case C projections are distinctly separable by 2060. The apparent carbon cycle contribution to overall uncertainty in atmospheric CO_2 in the year 2075 is relatively small (<14% of the total variation if airborne fraction calculations are excluded). This is so, even though some parameter uncertainty estimates were used that would be considered conservative (high) by some authorities. However, our basic conclusions about relative parameter-related contributions to projection uncertainties are similar to those estimated by others using different carbon cycle models (Laurmann and Spreiter 1983; Oeschger and Heimann 1983; Goudriaan and Ketner 1984).

The potential impact of future improvements in carbon cycle model structure, however, is likely to be significant and obviously cannot be simulated totally at the present time (see Section 5.8.6 and Chapter 7). One cannot determine the importance of processes that are poorly parameterized (or nonexistent) in any model without having another model or a more refined parameterization with which to make comparisons (e.g., Section 5.8.6; Killough and Emanuel 1981; Siegenthaler 1983). Any globally averaged model contains many oversimplifications (see Chapter 7), and the Emanuel et al. (1984b) model is no exception. The treatment of ocean vertical mixing in a one-dimensional formulation is as refined as current understanding permits, but parameterizations for high-latitude deep water injection or for the marine biosphere are not incorporated. Neither are there formal links between the land, freshwaters, and the ocean—a generic characteristic of current models. The globally averaged terrestrial biosphere in the Emanuel et al. (1984b) model does not permit a geographically detailed treatment of critical issues such as region-specific, forest-clearing rates and stand carbon densities (Chapter 6) or ecosystem- and perturbation-specific carbon succession patterns (Chapter 7), and the treatment of soil carbon responses to human perturbations is rudimentary (Emanuel et al. 1984a). The latter problem is also currently generic to all carbon cycle models (Chapter 11) and has generated significant errors in biospheric carbon fluxes (Schlesinger 1986; Pastor and Post, in press) as well as other unrealistic results (e.g., a soil carbon pool which is over 50% charcoal [Goudriaan and Ketner 1984])! One cannot criticize past modeling efforts for having stayed within the limited confines of the available data. Rather, these observations illustrate why one is currently forced to accept conservatively broad estimates of parameter ranges as a hedge against potential uncertainties contributed by model structure. Reduced uncertainties will require significantly improved models with, more importantly, better data bases to support their development and application.

Potential variation in that portion of the model input function related to future fossil fuel emissions has the greatest impact on model output uncertainty. Variations in future terrestrial sources in simulations with emissions Case C (Section 10.3.4) produce the greatest percentage variation in atmospheric CO_2 loadings because of the relatively low level of fossil emissions. However, the maximum variation in projected atmospheric CO_2 concentration at any point in time through 2075 is still only 14%. At the high fossil fuel loadings in Case A, the maximum percent variation among any of the permutations and combinations of models and terrestrial history cases is 8%, even though the calculated atmospheric CO_2 reservoir has more than quadrupled by the year 2075 (Figure 10.10). Scenarios of lower emissions rates of carbon to the atmosphere exhibit stronger dependencies between model parameters and model outputs (cf. Laurmann and Spreiter 1983).

These conclusions about the relative importance of emissions and carbon cycle uncertainties still apply qualitatively, but not *quantitatively,* if AF calculations are included in the comparisons. Inclusion of AF data (Figure 10.10) produces a significant widening of the band of uncertainty by extending the low end of the spectra, particularly significant in Case A. Another result is a significant increase in the proportion of the overall uncertainty in projected atmospheric CO_2 concentrations attributable to the carbon cycle.

The impact of future fossil fuel emissions uncertainties on an analysis of atmospheric CO_2 prediction uncertainties is so dramatic that it invites erroneous conclusions. These results do not imply, for example, that we can dismiss research on the past or future role of the terrestrial biosphere in favor of ocean model development. Nor do they mean that we can ignore the carbon cycle and, instead, concentrate our efforts on improved assessments of future fossil CO_2 releases. The results are, in fact, a direct reflection of the difficulty of making accurate projections on century time scales for a phenomenon that is tied to our highly uncertain global geopolitical, social, and

economic systems. The aftermath of the 1973 OPEC oil embargo on the global energy-economic structure (e.g., Marlay 1984) illustrates that projections over periods as short as a decade are problematic enough. What the results from uncertainty analyses do imply very strongly is that, if the consequences of CO_2 climate change are deemed unacceptable, our principal leverage for mitigation or control of future levels of CO_2 probably lies in manipulation of energy use and carbon emissions patterns, and not in alterations of the natural carbon cycle. However, we must be cautious in drawing even this conclusion, because society undoubtedly will not be free to choose any potential energy-economic path within the total range of uncertainty.

Thus, it appears worthwhile to elaborate further. Assume that improved understanding of the magnitude of climate response to atmospheric CO_2 and the consequences of CO_2-climate change leads to the conclusion that the atmospheric concentration must not exceed 500 ppm or, alternatively, must be reduced to <350 ppm as quickly as possible to minimize deleterious environmental and human impacts. Our view of the uncertainty analysis is then quite different. All major systems in the global carbon cycle now contribute significantly to uncertainty in atmospheric CO_2 content (see results for year 2000 or 2025 in Tables 10.7, 10.8, 10.9, and Figure 10.10). We would then require much more accurate projections of atmospheric levels, and uncertainty in all systems of the global geochemical cycle as well as fossil fuel CO_2 emissions would have to be reduced. This might force the adoption of a limited range of energy-use-CO_2 emissions strategies (see Chapter 9) or technological mitigation measures (e.g., Steinberg et al. 1984). Further, mitigation strategies related to management of the terrestrial biosphere and the effects of lesser CO_2-climate feedbacks involving the natural carbon cycle, which would not be considered at very high levels of atmospheric CO_2 (e.g., in the context of limits to atmospheric content), now could play an important role in controlling atmospheric concentrations.

It should also be recognized that immediate implementation of mitigation strategies might not be reflected in stabilized or reduced atmospheric CO_2 levels for 50 to 100 years because of the inertia inherent in responses of human political, economic, and technological systems and of the global carbon cycle (Chapters 4, 5, and 9; Perry et al. 1982; Bach 1983; Laurmann 1983). Further, the time required for a reduction of elevated levels to contemporary atmospheric levels ranges from *centuries to millennia,* depending mainly on the pattern of CO_2 releases (Section 5.6.2; also see Bach 1983). Thus, the temporal factor provides yet another set of constraints on potential responses (see discussion in Perry et al. 1982).

These exercises demonstrate that the level of acceptable uncertainty in understanding of the global carbon cycle is critically linked to our perceptions of uncertainties in other major aspects of the CO_2-climate problem, including the magnitude of CO_2-climate changes and the indirect effects of such changes. Will increased atmospheric CO_2 levels result in an unacceptable climatic warming or will they simply counteract the cooling effect of a projected minimum in solar activity over the next 100 years (Rampino 1985)? How would impacts on society be manifested in these two alternatives? What other possibilities exist? Uncertainties in these other areas appear to be so large at present (see Chapter 1) that we must press forward to develop a much more accurate understanding of the global carbon cycle. This must be done as a hedge against the time when a rapid and reasoned response may become necessary to avert or mitigate the deleterious effects of CO_2-climate change. Prudence dictates that this include the timely development of plans to provide alternate, environmentally benign, energy sources and strategies for their implementation with a minimum of political, economic, and other human impacts.

10.3.6.2 *Comparisons with Other Atmospheric CO_2 Projection Exercises*

Comparison of other recent atmospheric CO_2 projections with those resulting from emissions Cases A, B, and C provides yet another view of potential uncertainties. The atmospheric CO_2 levels from Seidel and Keyes' (1983) midrange case fall within the envelope of estimated carbon cycle uncertainty about Case B (Figure 10.5). Their projections were made with the IEA/ORAU energy-economic CO_2 emissions model coupled to the same carbon cycle model used in the Monte Carlo error analysis. They chose a set of midcase energy-economic assumptions that resulted in lower rates of CO_2 emissions early in the 1975–2075 period, but also increased to produce a 50% higher rate of release than in Case B by 2075. Nine alternative CO_2 emissions scenarios were developed by varying the assumptions on primary energy sources, economic growth, and bans on certain fuels, but the CO_2 results were not reported. The terrestrial biosphere was assumed to make no contribution to carbon emissions (Seidel and Keyes 1983).

The Massachusetts Institute of Technology (MIT) projection exercise (Rose et al. 1983) produced a spectrum of 11 global carbon emissions cases through the year 2050, but was not extended to the atmospheric CO_2 projection step. However, based on comparison

with the fossil emissions rates from the other studies graphed in Figure 10.10, we estimate that the MIT cases would result in atmospheric CO_2 levels in 2050 ranging from slightly higher than those generated by Case B to very nearly equal to those obtained from the lowest Nordhaus and Yohe (1983) scenario (first percentile). It also appears that the nine alternative scenarios of Seidel and Keyes (1983) would fall within this same range.

The IIASA High and Low scenarios (Haefele 1981) exhibit characteristics very similar to Cases A and B, respectively, up to their termination in 2030 (Figure 10.5). Atmospheric CO_2 projections reported by Bach (1983) and by Goudriaan and Ketner (1984), using carbon cycle models containing a biotic growth response to increasing levels of CO_2, fall very close to the lower extremes of the corresponding uncertainty envelopes (Cases A and B) in Figure 10.10. The concentrations reported in these studies for the year 2030 (not shown in the figure for purposes of clarity) fall between 480 and 500 ppm and 430 and 450 ppm, respectively, for the High and Low scenarios (cf. Table 10.7). Thus, these results are not very different from those obtained with emissions Cases A and B, based on the assumption of a neutral biosphere (Section 10.3.4). The reason for this similarity is that a biotic growth (CO_2 fertilization) factor was used in the two models, and this factor was "tuned" (adjusted) to remove excess carbon from contemporary fossil or biospheric releases not absorbed by the ocean models. Because the status of the biosphere was not altered during the period of these projections, the results were similar to those from state-of-the-art ocean models representing a more efficient CO_2 sink, but assuming a neutral biosphere.

A subset of the atmospheric CO_2 projections developed by Nordhaus and Yohe (1983) is also provided for comparison with the others shown in Figure 10.10. It should be remembered that their AF model produces lower estimates of atmospheric retention than the carbon cycle models. For example, the high emissions case (99th percentile) of Nordhaus and Yohe releases significantly more carbon to the atmosphere by 2075 than Case A because the initial rate of increase in fossil CO_2 emissions is much more rapid (cf. Figure 2.1 in Nordhaus and Yohe 1983). The Nordhaus and Yohe high case releases over 600×10^{15} g more carbon by 2075 than the lowest simulations based on Case A (e.g., with a neutral biosphere) yet the predicted atmospheric CO_2 concentrations in 2075 are essentially the same (Figure 10.10). Based on model and scenario comparisons made by Perry (1984), and in this work, we conclude that the atmospheric retention estimated by Nordhaus and Yohe (1983) for their high emissions cases is 10 to 20% less than that of the carbon cycle models. This could amount to a difference of a decade or more in the estimated date for doubling or quadrupling the current CO_2 level in the atmosphere. (However, it also must be pointed out that other uncertainties in such an exercise currently exceed those contributed by differences in carbon cycle simulations.) Further, the Nordhaus and Yohe 95th percentile case falls well above the atmospheric CO_2 concentration curve based on our Case A and a constant AF. Thus, our conclusions about the earliest date for a "CO_2 doubling" are quite similar to theirs.

Differences between projections based on the Nordhaus and Yohe AF methodology and the carbon cycle models appear to be somewhat less for cases with lower absolute levels of carbon emissions. Note that the projected atmospheric CO_2 concentrations from the Nordhaus and Yohe (1983) 50th and 5th percentile cases are quite similar to Case B and Case C projections, respectively, with constant AFs. Using either of our carbon cycle models to project CO_2 levels from the Nordhaus and Yohe midcase (emissions rates are comparable to our Case B simulation with a neutral biosphere) would result in a 1975–2075 concentration trajectory that would fall within the central Case B projection envelope (Figure 10.10). However, the CO_2 concentration trajectories predicted with the lowest case (1st percentile) from Nordhaus and Yohe (1983) would still fall well below the lower bound on our set of Case C projections (including the AF calculation).

We chose to include comparisons with the two extreme cases presented by Nordhaus and Yohe (1983), rather than with the more often cited 5th and 95th percentiles, because our primary concern was to document and compare potential future extremes in carbon emissions and atmospheric CO_2 levels. The subjective or "judgmental" nature of projection exercises makes it difficult to determine an exact probability assignment, particularly when a given case falls on the tails of a statistical distribution. The lower estimated atmospheric retention of the airborne fraction formulations, when compared to the models, serves to highlight how difficult the latter exercise can become. However, the uncertainty analysis involving the IEA/ORAU model (Chapter 9) indicates that Cases B and C from Edmonds et al. (1984) essentially bracket a median estimate of CO_2 emissions and, therefore, Case C may not be a very suitable lower bound for projections. The Nordhaus and Yohe 1st percentile case provides a reasonable match to the 5th percentile case resulting from the uncertainty analysis performed on the IEA/ORAU model. Thus, for the present, we consider the Nordhaus and Yohe lowest case to represent a more appropriate lower bound on atmospheric CO_2 levels than do Case C simulations.

Based on the preceding discussion, it would appear that, although current estimates of the projected atmos-

pheric increase by 2075 range from small increments to over 1000 ppm above the 1983 level (342 ppm), increases of 150 to 300 ppm by 2075 appear to be much more probable.

10.4 SUMMARY AND CONCLUSIONS

10.4.1 Simulating the Global Carbon Cycle

Prediction of future atmospheric CO_2 concentration must be based on carbon cycle models that accurately depict quantitative rates of carbon exchange among the major reservoirs. As a first approximation in the validation of models, we should be able to compute a balanced global carbon budget for the contemporary period; we currently cannot do this and we are uncertain why not. This obviously has some impact on our ability to project future atmospheric CO_2 content with existing models. Uncertainties about the historical role of the terrestrial biosphere in response to anthropogenic changes in land use and, to a lesser extent, the rate of carbon uptake in the open ocean and in freshwater and coastal marine sediments, appear to be the prime contributors to the global imbalance. These uncertainties can be treated to some extent, albeit crudely, in a projection exercise. However, the influence of other important phenomena, such as effects of natural subfossil sinks, climate change, increased CO_2, anthropogenic pollutants, use of biomass for energy, and potential mitigation measures on the global carbon cycle cannot be treated in a substantive manner at present. In longer term projections, on the order of a century, these phenomena could have a significant impact. However, research on most of these factors is still in its infancy.

Ultimate resolution of the imbalance in the global carbon cycle for the contemporary period may not be possible without more basic understanding of fundamental ecosystem patterns and processes (e.g., steady-state concepts and nature of succession) or ocean circulation (Chapter 11). Thus, although there is a tendency to view carbon cycle research primarily as an applied effort (e.g., adding up carbon budgets, measuring changes in ecosystem areas, monitoring atmospheric change, recording fossil fuel use, or measuring oceanic tracer data profiles), we must not lose sight of the fact that our inability to balance the carbon cycle, both now and in the future, may hinge largely on resolution of more basic research questions (Chapter 11). It may well be that there always will exist significant errors in a balance as complex as that of global carbon; and the risks associated with this situation need to be addressed.

Analysis of the atmospheric CO_2 measurement records with three-dimensional tracer models, coupled with creative application of developing satellite image technologies, may provide an effective means of determining the actual sources of imbalance. The sine qua non in global carbon cycle research is the continuation of the existing global network of CO_2 measurements at remote clean-air locations. Expansion of this effort to key locales in major land biomes may provide the additional data needed to identify whether the greatest errors in CO_2 fluxes occur in the representations for the terrestrial biosphere or the oceans, or are due to some currently unaccounted-for phenomena (Chapter 11).

10.4.2 Atmospheric CO_2 Projections

Although a number of well-posed recent efforts were made to project the concentration of atmospheric CO_2 into the future (Bach 1983; Goudriaan and Ketner 1984; Keeling and Bacastow 1977; Laurmann and Rotty 1983; Laurmann and Spreiter 1983; Markley and Hurley 1983; Niehaus and Williams 1979; Nordhaus and Yohe 1983; Oeschger and Heimann 1983; Pearman 1980; Perry 1984; Revelle and Munk 1977; Seidel and Keyes 1983; Siegenthaler and Oeschger 1978), it is difficult to make exacting comparisons between the results of this effort and those of others, because of major differences in assumptions (1) in models to represent energy/economic forecasts (including exponential growth as a simple model), (2) in models to represent the carbon cycle (including airborne fraction as a simple model), (3) in initial atmospheric CO_2 concentration, and (4) in terrestrial biospheric CO_2 release history.

The effects of potential variations in the historical time series of CO_2 concentration caused by anthropogenic changes in the terrestrial biosphere generally have not been rigorously considered as part of a CO_2 projection effort. In only four of these previous efforts (Bach 1983; Goudriaan and Ketner 1984; Nordhaus and Yohe 1983; Seidel and Keyes 1983) were results from an energy-economic model of CO_2 release from fossil fuel emissions incorporated into the projections. In only one instance (Nordhaus and Yohe 1983) was an attempt made to estimate the statistical confidence level of the projections of future CO_2 emissions. In the latter exercise, a modified airborne fraction concept rather than a global carbon cycle model was used to make projections of atmospheric CO_2 levels. The assumptions used in this airborne fraction model result in lower predicted retention of atmospheric CO_2 than state-of-the-art carbon cycle models and make it somewhat more difficult to compare with other recent studies. It should be noted that our more recent resul

to fall well within the envelope of results reported in these other exercises, and despite differences in methodologies, our basic conclusions about the future growth of atmospheric CO_2 are virtually identical to those of Nordhaus and Yohe (1983).

Because of the obvious difficulty associated with forecasts of future energy-economic CO_2 emissions, our ability to accurately project future atmospheric CO_2 levels is very limited. Given the global emissions projections cases presented, we are reasonably confident that the atmospheric CO_2 concentration in the year 2000 will fall within the 360- to 400-ppm range. We agree with the conclusions of others that simple models based on the projection of existing atmospheric trends (e.g., airborne fraction) are not likely to result in substantial prediction errors over the next 25 years.

Our analysis also indicates that doubling of atmospheric CO_2 levels is unlikely to occur before the year 2025 and may not occur within the next century. Thus, our results agree reasonably quantitatively with the conclusions of Nordhaus and Yohe (1983), even though totally different energy-economic-CO_2 emissions and carbon cycle modeling approaches were used in the two studies. Although current estimates of the projected atmospheric increase by 2075 range from small increments to over 1000 ppm above the 1983 level, more likely estimates of this increase are on the order of 150 to 300 ppm.

One must be cautious about making atmospheric projections, particularly at high rates of fossil fuel emissions, because of the potential for future expression of feedbacks or nonlinear responses in the global carbon system. Existing carbon cycle models or approximations based on the airborne fraction concept are probably not valid at the highest atmospheric CO_2 concentrations, or at highest release rates, or at the later dates discussed here. However, these concerns do not mean that we cannot or should not attempt predictions with existing models, but rather, that we are aware of their limitations. Potential carbon cycle feedbacks could increase the rates of growth of atmospheric CO_2. However, these concerns are currently overshadowed by the magnitude of the potential uncertainty in future fossil fuel emissions.

10.4.3 Future Fossil Fuel CO_2 Emissions

Although projections of global fossil fuel emissions are apparently the greatest source of errors in predicting future atmospheric CO_2, it may not be reasonable to assume that a significant reduction in this uncertainty is possible if projections on century-long time scales are needed. There are obvious limits to our ability to predict economic/energy-use futures, as very recent history including the aftermath of the 1973 oil embargo has taught us. Because all aspects of the complex "human equation" cannot be foreseen, including (1) the changing spectrum of available technologies and fuels, (2) a host of difficult-to-predict demographic, economic, social, and geopolitical factors, and (3) the global response to a climate change only dimly perceived at present, much of this uncertainty may be irreducible. Uncertainties in emissions sources may always remain dominant in atmospheric CO_2 projections. The real challenge may be to make policy decisions intelligently in the face of this uncertainty.

Yet the study of CO_2 emissions from fossil fuel use has resulted in significant changes in society's thinking about the problem of CO_2 emissions on both the qualitative and quantitative levels. The mode of analysis shifted dramatically from simple time trend analysis to energy-economic analysis. The vehicles used for projecting future CO_2 emissions changed from simple log-linear extrapolation models to behavioral energy-economic models designed specifically to investigate long-term global energy-economic relationships. Thus, the focus of research shifted from CO_2 emissions to long-term global energy systems and to the demographic, socioeconomic, and technological factors that will define and limit the energy choices available to society over the next century (Chapter 11). The analytical perspective broadened. The single "best estimate" scenario was supplanted by an explicit recognition of uncertainty (e.g., Nordhaus and Yohe [1983] constructed CO_2 emissions projections using a Monte Carlo simulation and Edmonds et al. [1984] developed subjective scenario bounds) and by alternative scenario development (e.g., Rose et al. 1983; Haefele 1981; Seidel and Keyes 1983). As a consequence, our understanding of likely future CO_2 emissions rates changed dramatically as well, and despite the broad uncertainty that remains, the current consensus view of CO_2 emissions from fossil fuels finds a continuation of the postwar time trend extremely unlikely. The CO_2 emissions from fossil fuel use are likely to increase, but at a lower rate than in the past, and the "best estimate" scenarios show that that rate is not only lower but significantly lower than the postwar time trend.

10.4.4 Interpretation of Uncertainty Analyses

Variability in fossil fuel emissions has such a dramatic impact on atmospheric CO_2 prediction uncertainty that one can be misled by concluding that further research on the global carbon cycle has a very low priority or, alternatively, that carbon cycle research ought to be focused exclusively on the oceanic component. In fact, what these results imply is quite complicated, and the level of

allowable uncertainty in our understanding of the carbon cycle is irrevocably connected to uncertainties in CO_2-climate forcing and the indirect effects of CO_2 climate change. If atmospheric CO_2 must be held to <500 ppm to minimize deleterious environmental and human impacts, then all major components in the global carbon balance, including the carbon cycle, make significant contributions to prediction uncertainties. Under these conditions we would require more accurate projections of atmospheric content, and uncertainties in all systems, including future fossil fuel emissions, would have to be reduced. The latter implies that mitigation strategies or controls on energy technologies and fossil fuel use would have to be implemented on a timely and on an international basis.

Projections of future atmospheric CO_2 levels are based on current understanding of the operations of the carbon cycle and associated geochemical and climatic systems over the past few centuries. It is important to recognize that human activities or even natural phenomena have the potential for altering the dynamics of these systems enough to invalidate some current CO_2 projections. Changes in atmospheric CO_2 and climate could be expressed in the form of feedbacks—both unbalancing and stabilizing. These feedbacks could affect the carbon cycle, energy systems, and human ecology. The magnitudes of these feedbacks could be very important. Yet basic understanding is lacking, even in the short term, of critical processes. These limitations mean that particular outcomes cannot be defined with confidence. Thus, contingency planning for technological and social adjustments will continue to be argued over a range of possibilities that only can be narrowed or elaborated on the basis of new research or insight.

At present, uncertainties in the principal aspects of the CO_2-climate issue other than the carbon cycle are so large (see Chapter 1) that we must develop a much better understanding of the global biogeochemical cycle of carbon (linked to climate and other critical element cycles) to be prepared to respond in a timely and reasoned manner to avoid or lessen the deleterious impacts of CO_2 climate change. Planning for potential development and implementation of alternative environmentally benign energy sources should be undertaken well in advance of the time when a response may be needed to lessen the political, economic, and other human impacts of these changes. The need for international cooperation in framing a response presents serious problems (Schelling 1983; Goldemberg 1984), and may only be possible if a truly international research and planning effort is achieved.

10.4.5 Other Important Considerations

The assumption that CO_2-climate change will be gradual and predictable is not unequivocally supported by evidence in the geologic record (Chapters 3 and 8). Past relationships between systems of the global carbon cycle and climate changes (Chapters 3 and 8) may be inferred from proxy data. Understanding such relationships may be critical to understanding future responses of the carbon cycle to anthropogenic perturbations (Chapter 11). Proxy records for atmospheric CO_2 variations in ocean sediments suggest that concentrations have remained within a relatively narrow range ($\sim$270 $\pm$80 ppm) over the past 340,000 years (e.g., through several glacial-interglacial cycles) (Chapter 3). Proxy records of CO_2 variations and major climate changes appear to be coupled, and time-series analysis of the data suggests that atmospheric CO_2 *leads* the climate change. The ice-core CO_2 record (30,000–40,000 B.P.) contains evidence for more rapid, concurrent changes in both CO_2 and climate on time scales on the order of a century (Oeschger and Stauffer 1986; Chapter 3). These past atmospheric CO_2 shifts are comparable to anthropogenic changes brought about since the early 19th century ($\sim$70 ppm). Thus, within the next decade, atmospheric CO_2 content should increase to a level outside the range indicated by geologic records for the Quaternary period, and the magnitude of the overall anthropogenic increase will be comparable to that associated with significant changes in past climates.

Although there still exists considerable uncertainty about the interpretation of the geologic records for CO_2 and climate, these data provide us with a perspective of the CO_2-climate issue that we cannot ignore. It suggests that our current models, which emphasize human perturbations, may be missing critical natural feedback components in both terrestrial and marine systems, perhaps even "mode switches" (Broecker et al. 1985) that involve changes in ocean circulation, which could be very important to our understanding of both climate and the carbon cycle over the next century (Chapter 11). Thus, it seems prudent to expand the effort devoted to understanding atmospheric CO_2 fluctuations in the geologic past (Chapter 11). However, it should also be recognized that CO_2 is only one, albeit the most important (other than water vapor!), of the radiatively active trace constituents of the atmosphere that can produce a climate change.

The presence of other atmospheric trace constituents (e.g., methane, nitrous oxide, chlorofluorocarbons) could significantly augment the predicted climatic response caused by CO_2. This might be of particular

significance in the analysis of policy options if trace gas concentrations continue to increase at rates observed over the past several decades and if fossil fuel CO_2 releases continue at the reduced rates observed over the past several years (Wuebbles et al. 1984; Marland and Rotty 1983; Ramanathan et al. 1985; low-emissions cases reviewed in this chapter). Readers should realize, therefore, that we have presented only a portion of the atmospheric gas concentration data needed to attempt modeling of future climatic responses, and that the carbon cycle is only one of several element cycles that must be understood to address climate-related issues (Chapter 11).

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11. RESEARCH REQUIREMENTS

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11.1 RESEARCH NEEDS

The large, geologically active pools of living and dead organic matter in vegetation and soils were largely neglected in early carbon cycle research. These pools were often treated as conduits irrelevant to carbon storage rather than as state variables and parameters that might change and modify physical rates. Recently acquired information dictates that we develop improved data and models for cycles of carbon and of climate that contain explicit formulations for more active participation by these organic agents, including marine biota (Chapters 3, 5, and 8; henceforth all chapter and section references are to this volume).

Omitting these components, or considering them as simple extensions of the atmospheric or surface ocean carbon pools, might have been justified for certain exercises dealing with extremely brief perturbations around a steady-state condition or geologically slow balances. However, much of our current uncertainty about relationships between fossil fuel use and atmospheric composition was produced by a grossly oversimplified treatment of the carbon cycle in global models, particularly those carbon exchanges mediated by the Earth's biota.

To recapitulate our most pressing difficulty once more, ocean carbon dioxide (CO_2) models are marginally able to absorb enough carbon from the atmosphere to account for past fossil fuel burning, but they cannot accept other contemporary atmospheric sources (i.e., forest clearing). This incompatibility is exacerbated by a detailed history of atmospheric CO_2 in ice cores (Chapter 3), which suggests either that nonfossil CO_2 sources were nil after 1940 or that these sources were masked in the atmospheric record by ignored or unknown sinks. We have strong evidence that forest clearing after 1920 was at least as significant as that before, even though the focus of clearing shifted from temperate to tropical regions (Chapter 6). Thus, there are several possible explanations: errors in the land-use time series, additional sinks not represented in current models, and masking of sinks by oversimplified models. As one simple answer, all three could be valid (Chapters 8 and 10), each obscured by the others.

Oversimplification included the omission of such subfossil carbon pools as peat or buried soils. These were omitted partly because we had difficulty characterizing and differentiating the proper pools, and difficulty estimating interpool transfer rates. This omission was rationalized by the assumption that responses or net change would be too small to affect the "fast" atmosphere/surface-ocean/live-organic-carbon exchanges or the exchanges in very slow systems (e.g., stratigraphic sediments and rocks). Because carbon pools in forests, in soils, and in other forms of detritus may be responsive to the very climatic changes motivating the research on CO_2, methane (CH_4), and other "greenhouse gases," their neglect in modeling can no longer be justified.

Another explanation often raised, but one that we think is illogical (to one who has read the previous chapters), is that a better model plus the existing data will suffice. This is another straw man, and was properly dispatched earlier (Section 5.8.6):

> There is a tendency, especially within that part of the community not directly involved in the technical and scientific details, to believe that modeling alone will (suffice) . . . . This is definitely not the case. Nor is it true that measurements alone will suffice. Rather, it will be the hand-in-hand cooperation between experimentalists and modelers that will advance our capability (p. 107).

Many other uncertainties remain, either about the nature of individual subsystems or about the ways in which they interact in the global carbon cycle. Examples of specific suggestions for future research to resolve these questions are specified throughout Chapters 3 to 10. Many of these will not be reiterated here. The reader is referred to the concluding sections of these chapters for authors' detailed recommendations for research associated with specific areas of inquiry.

Our purpose below is to build on the synthesis of information in Chapter 10 and to focus attention on the critical areas identified therein (i.e., the key research components needed to draw and expand current efforts into a more coherent whole). The reasoning for this as well as a pertinent introduction to the following discussion was provided in Section 8.6.2 by Olson et al.:

> Emphatic statements about gaps in existing knowledge and requirements for extending present knowledge have been rendered at dozens of places above, through either reasoning or literature sources to back up the authors' conclusions. Naturally we vary in sensitivity to disciplinary specialties. Neither this review nor the many outside opinions received need to be smoothed out into one bland endorsement. Many worthwhile things must be done sooner or later. In the following sections, we highlight just a few thrusts made toward understanding our geosphere and biosphere that will require strenuous persistence in work already begun or shifts in emphasis (p. 203).

Despite the seeming incongruity, however, we need to begin by examining ways to better understand that part of the global carbon system that now seems to contribute most to future uncertainties, but where the

prospect of reduction in uncertainty is least encouraging (i.e., human ecology and future CO_2 emissions).

11.1.1 Future CO_2 Emissions

The fossil fuel CO_2 emissions analyses of the past 10 years have gradually changed our understanding of the CO_2 issue. The 4.4%/year rate of growth—a benchmark of early analysis—is now thought to be highly unlikely. This has reduced the urgency of expensive investments to respond to the effects of climate change and given time for thoughtful analysis. There remains a recognizably broad range of uncertainty associated with CO_2 emissions projections, however, and much of this may be irreducible; some may be directly related to the options open to society.

Future work must recognize this irreducibility and seek approaches for assisting decision-makers. Some actions must be taken with limited information, even if the actions relate only to contingency plans and decisions on how much effort should be devoted to research on CO_2.

We see several directions for future research in the area of emissions, which can help decision-making and scientific investigation to interact (see Chapter 9; also see Reilly and Edmonds [in press] for a more detailed treatment). Emissions modeling provides a necessary tool to facilitate and complement these processes. The following are specific research efforts needed.

11.1.1.1 *Historical Fossil Fuel Emissions Data*

The fossil fuel emissions time-series data base, revised and updated annually, is a key reference source for both emissions and carbon cycle modeling. This data base not only should be maintained, but also expanded to provide the seasonal and geographic distribution of CO_2 emissions on a routine basis (Section 4.3).

11.1.1.2 *Exploration of Extreme CO_2 Emissions Scenarios*

The Monte Carlo analysis using the IEA/ORAU model (Chapter 9) generated a range of future emissions rates by random selection of parameter values. Such an analysis provides only very summary statistics. A useful exercise would be an analysis of specific scenarios at the extremes of the distribution. Such an examination may suggest structural change in the model and its supporting data base. Together these changes might contribute to reducing uncertainty, by isolating critical areas for further study. It would at least provide a detailed set of reference scenarios for researchers and decision-makers. The real future presumably may not approach either of the extremes, but we will better understand why a family of outcomes below a midcase is most likely to approximate the future.

11.1.1.3 *Further Examination of Input Assumptions*

Sensitivity and scenario analyses conducted to date suggest several "key" parameters (end-use energy efficiency, income elasticity of energy demand, labor productivity, rate of interfuel substitution, and rate of electrification [Chapters 4 and 9]) that need to be explored in greater detail if we are to reduce uncertainties. A better understanding of technological potentials and the process by which new technologies penetrate markets is also essential. Possible climatic feedback effects on energy use (e.g., Rust and Kirk 1982) also need to be explored. Uncertainty analysis is helpful in ranking the importance of different model assumptions and parameter values and in aiding in setting priorities for future research. However, a closer examination of the way we think about the determinants of global energy and CO_2 emissions, aided by state-of-the-art techniques for defining parameter values (e.g., expert elicitation), will result in better definitions of data base development and energy modeling priorities.

11.1.1.4 *Regional Disaggregation*

One of the key results from the scenario and uncertainty analyses reviewed in Chapter 9 was the importance of regional detail, especially with regard to the developing world. Areas that need special attention are (1) the role of development in shifting aggregate energy-use patterns and (2) determination of market versus nonmarket fuel choices (e.g., conservation, passive solar, biomass). Nonfossil fuel sources of CO_2 (e.g., biomass) should be an integral part of such an analysis.

11.1.1.5 *Near-Term Analysis*

The CO_2 emissions question often is explored in terms of very long time horizons—100 years or more. Whereas many aspects of the problem demand this perspective, certain elements (e.g., role of capital stocks and investment decisions in moderating energy production and consumption choices) require greater resolution in earlier years. This is particularly crucial to understanding the role near-term events play in narrowing the long-term possibilities, and could be more critical if other trace gases exacerbate climatic effects, thus moving dates of significant climate change nearer to the present.

11.1.1.6 *Role of Model Structure*

The impact of model structures on projections is not accounted for formally as are the random errors in parameter values (Chapter 9). An examination of assumptions and results of different models would

enable researchers to estimate the extent to which onfidence intervals need to be broadened to account for the uncertainty in model structure, even while parameter errors are being narrowed.

11.1.1.7 *Trace Gas Emissions*

CO_2 is not the only greenhouse gas. A partial list of other greenhouse gases would include methane (CH_4), nitrous oxide (N_2O), chlorofluorocarbons, ozone (O_3), methyl chloride (CH_3Cl), ethylene (C_2H_4), sulfur dioxide (SO_2), and ammonia (NH_3). Emissions and atmospheric concentrations of many of these gases are growing. Research in the area of emissions has been based either on pure speculation (e.g., What if concentrations doubled?) or time-trend analysis. Emissions projections for trace gases thus have neither been consistent nor grounded on a strong analytical base (see Ramanathan et al. 1985). Emissions of some of these gases are directly related to energy use and land-use changes and provide a natural extension for current efforts on energy-related CO_2 emissions.

11.1.2 Improvements in Carbon Cycle Models

Carbon cycle modeling was long pervaded by the incorrect assumption that modern human influences were preeminent and that the imprint of humankind was felt only since the dawn of the so-called Industrial Era (circa 1860). Some such working assumption and starting date appears to have been necessary to make reasonable progress in global modeling with the limited data bases available a decade ago, even though it was recognized that the pace of population growth and agricultural and industrial expansion was quickening much earlier than 1860. These simplifying assumptions led to formulation of models of perturbation around previous assumed equilibria for the carbon cycle. It appears to have led to an oversimplified treatment of a very complex system responding to earlier modern and ancient historic influences (Chapters 7 and 8). The gears and wheels of this complex machinery are controlled to a large extent by a set of feedback loops involving climate and biogeochemical cycles of elements other than carbon (see Chapters 5 to 8).

Thus, current models are calibrated on the assumption that a steady-state or equilibrium condition existed for centuries before being upset by burgeoning human population growth and associated rapid agricultural expansion and industrial development. In this assumed condition, the sum of all natural and human-induced fluxes between the atmosphere and other important rbon reservoirs must equal zero (averaged over riods on the order of a few decades). Recent attempts to reconstruct the history of atmospheric CO_2, using a variety of geologic records and modeling techniques (see particularly Sections 3.2, 8.4.2, and 9.3.5; also Broecker et al. 1985a), exposed the potential brittleness of the pervasive steady-state hypothesis. It is an approximation, at best, and requires scrutiny and alternatives.

Thus, it appears necessary to frame a very different modeling program than that portrayed in most literature from the past decade. Prerequisites for restoring confidence in the underlying science and working assumptions about the nature of human impacts are better perspectives of the natural geochemical cycles of many elements besides carbon (nitrogen [N], oxygen [O], phosphorus [P], sulfur [S], and silicon [Si]) and of their relationships with naturally varying climate (Chapter 8; also see Ramanathan et al. 1985). *Because of the interactions between the geochemical and climatic cycles, research cannot remain isolated by discipline as it typically has been up to now.*

The sections that follow present a listing of poorly exploited data resources or phenomena unaccounted for; properly interpreted, some of these can lead to changes in our understanding of atmospheric sources and sinks needed to balance the carbon cycle with models. Specific research efforts could improve understanding and reduce uncertainties. The sections are presented in estimated order of relative potential for explaining the contemporary imbalance in the global cycle. This ordering should not be misconstrued, however, to be a ranking of priorities for future research on the carbon cycle. Understanding the history of relationships between the carbon cycle and climate also is an important undertaking, although not necessarily the most critical factor in resolving the contemporary imbalance in the carbon system.

11.1.2.1 *Historical Records of Changes in Terrestrial Land Use*

A high priority should be assigned to completing temporally detailed reconstructions (by decade as far back as possible) of land use on a highly resolved geographical scale (counties, districts) on the basis of primary sources, both published documents and archival records, and where necessary, correlates such as demographic patterns, topography, transportation corridor development, and so forth. To the extent that a satellite map of contemporary land use can be constructed to identify areas impacted by humans, the task of historical research can focus on those lands actually transformed completely by new uses (Chapter 6). However, it is expected that the use of satellite imagery will confirm suspected degradation and, hence, lowered carbon storage of many forests

and grazing lands that were not simply replaced by crops. In many regions, particularly in the Northern Hemisphere, some of these lands experienced significant regrowth of vegetation from past cycles of disturbance; history of land use would thus be masked by current status. Thus, historical research must be a primary new focus of reconstruction efforts. *This research will also require a fairly fine geodetic grid, geographically referenced framework for modeling land-use changes on a geophysical and geochemical basis* (i.e., more highly resolved than the one-degree or several-degree grids that are usually used for global climate modeling) (Chapter 7).

11.1.2.2 *Carbon Succession in Terrestrial Ecosystems After Changes in Land Use*

It seems quite conceivable that researchers might develop a more accurate and comprehensive data base on land-use change patterns, including totally revised carbon release estimates, and still be unable to "balance the carbon cycle." This is because the existing treatment of steady state and succession for terrestrial systems "is a theoretical one derived from a logical assemblage of information on natural history and biochemistry; it is not based on actual measurements of extant steady state ecosystems" (Bormann and Likens 1979a, p. 665). The alternative paradigms of "overshoot and crash" or of frequently depleted landscapes require much wider testing (Olson 1981, 1985).

One paramount issue is whether ecosystems that have been changed most by humans during the past 250 years (e.g., clearcutting for wood products or agriculture), and are now in successional stages (e.g., returning to forest), were pristine, diverse systems at the time of first presumed disturbance. Or, were they already lower in diversity, species, or age-distributions, and hence in carbon content, because of still-earlier human or natural damage (e.g., Olson 1981; Richards et al. 1983; Brown and Lugo 1984; Sanford et al. 1985)? Misclassification of a recovering (successional) system as an undisturbed, natural ecosystem has produced serious errors in inventory estimates and modeling, leading to high estimates of carbon fluxes over the entire period of interest (Chapter 6; Olson et al. 1978, 1983; Houghton et al. 1983).

The sequence of disturbance is also critical for several other reasons. Second-growth forests can produce an *overshoot* response, higher accumulation of total biomass, than a "climax" or "virgin" forest, during some period in the 100 to 200 years following disturbance (Bormann and Likens 1979a, 1979b; Shugart 1984; Pastor and Post in press). Similar responses are expected from forests naturally low in species diversity (Shugart 1984), such as northern coniferous forests of the U.S.S.R., Canada, northern United States, and northern Europe. Forests in these regions ar now undergoing significant regrowth or succession from several cycles of past disturbances (Chapter 6). This overshoot phenomenon could also be occurring in many regions formerly covered by temperate deciduous forests and in subtropical and tropical regions as well. If the timing of past disturbances has been such that overshoot has been the rule rather than the exception during the latter part of the 20th century, appreciable terrestrial sinks could have resulted (i.e., "chance compensation of opposed effects", Broecker et al. 1979, p. 417).

Even more critical, however, may be the recent realization that changes in forest soil organic matter following disturbance (i.e., clearcutting and regrowth) are not generalizable to a simple loss-recovery relationship. Patterns of soil carbon succession are related to differences in successional species and initial organic matter levels; increased soil carbon content after disturbance may be commonplace in some forests of the Northern Hemisphere (Pastor and Post in press).

It follows that if the species composition (e.g., ecosystem structure) changes appreciably during succession (i.e., in ways favoring species that enhance soil carbon content), the equilibrium values of storage in the large soil reservoir could increase significantly (100% in areas where soil carbon density was low initially; Pastor and Post in press). Changes in forest species composition and other effects on ecosystem structure have been *imposed* by humans in cases too numerous to document here. The causes range from selective harvest and other silvicultural practices to deliberate importation of exotic species, and to accidental spread of disease vectors (e.g., chestnut blight, dutch elm disease). The concept that enhanced equilibrium carbon storage may follow human disturbance of forest ecosystems represents a critically important departure from the existing treatment in carbon cycle models. (Yet even this optimistic view may have to be leavened with the reality that some landscapes have been degraded to potentially very low status for storing carbon and recycling essential elements. Many of these cases, especially where humus has been degraded, may not be fixable simply by changes in ecosystem structure or by imports of fertilizer and water.)

Some current models of forest response to land-use change are based on the concept that the carbon content is rapidly reduced by a fixed percentage and then proceeds on a straight line to an asymptotic steady state without an aggradation or overshoot phase (Chapter 7). These differences in concepts of natural ecosystems and responses to small or large disturbances could have significant implications for a modern-day carbon balance affected by an evolving set of ecosystem responses

to both past disturbance and more recent conditions. *The temporal carbon response en route back to steady state might be quite different than current carbon cycle simulations, might vary among existing climates, and hence might be sensitive to shifts in climate, burning, and altered human pressures (or protection).*

Thus, it would appear that re-examination of both the "steady-state" paradigm and theories of ecosystem succession is as critical as the development of better empirical data bases on land-use change and soil and vegetation carbon succession patterns. Answers to these questions will not be possible without painstaking field measurements in a variety of ecosystems over time (Section 8.6.1). *Measurements and new models alone, however, will not suffice.*

11.1.2.3 *Deciphering Biospheric Contributions from Atmospheric Records*

Just as it appears unlikely that we can resolve our questions about the historical role of terrestrial ecosystems without basic research on patterns and processes (steady state and succession), it also seems unreasonable to assume that a satisfactory estimate of the global carbon balance can be achieved solely with better models incorporating improved representations for these paradigms and for records of land use. The diversity of ecosystems and potential disturbances, and the temporal and geographic variability associated with each, produces a heterogeneous mixture of carbon fluxes, the sum of which may always be uncertain in detail. The logistical problems involved in detailed coverage of all places would be formidable (Section 8.6.2), so we seek a complementary approach that takes advantage of the atmosphere's integration of exchanges with all other carbon reservoirs.

A synthesis of the atmospheric CO_2 measurements from clean-air sites around the world represents the most precise and integrated record available of the workings of the global carbon cycle, of the fluctuating CO_2 exchanges of the atmosphere with both the ocean and the biosphere, and of the response of this complex system to strong perturbations of both natural (El Niño-Southern Oscillation, ENSO) and human (fossil fuel combustion) origins (Chapter 3). *We must draw increasingly on the information contained in the contemporary records through the application of three-dimensional atmospheric tracer models and other techniques for analysis of the global atmospheric CO network data* (Chapters 3 and 7), *and on still newer developments associated with satellite imagery* (e.g., greenness index; Chapters 3 and 6). The latter will require an extensive period of development nd testing before rigorous application to the global arbon balance problem, but early results are very promising (cf. Tucker et al. 1985; Tucker et al. 1986; Tucker et al. in press; Woodwell et al. 1986).

Systematic determinations of $^{13}CO_2$ at a representative subset of the clean air sites of the global CO_2 program are also needed to augment $^{12}CO_2$ data, sufficient to define the $^{13}CO_2$ latitude dependence and regional seasonal variations. Such isotopic time series will be invaluable for separating the relative importance of land and ocean CO_2 contributions to important sources and sinks.

We must also develop more dependable records of past atmospheric CO_2 variations, reaching back in time as far as possible. The tree-ring ^{13}C record of atmospheric CO_2 appears to be compromised by physiological variability, and is limited principally to the late Holocene in temporal coverage. A satisfactory physiological model for isotope fractionation by plants exists (Farquhar et al. 1982), and the potential contributors to ^{13}C variations were reviewed by Francey and Farquhar (1982). Uncertainties may be attributable to sampling, site, growth rate, climate, air pollution, and even CO_2 fertilization. Attempts to reduce variability in ^{13}C records from tree-ring data by careful selection of records and by averaging large numbers of records (Freyer 1986), by corrections for growth-rate effects (Stuiver et al. 1984), or by corrections for both climatic variability and growth-rate effects (Leavitt and Long 1983a,1983b; Leavitt and Long in press) have so far only been marginally successful (Stuiver et al. 1984). More research on methods for identifying and quantifying the effects of physiological factors on isotope fractionation under field conditions will be necessary if the tree-ring record is to become a useful proxy for the history of CO_2 in the atmosphere or as an indicator of CO_2-climate modulation of atmosphere-biosphere exchange. At present, it seems difficult to judge the probability of success.

Currently, the ice-core records appear to provide the best source of information on past atmospheric CO_2 (and CH_4, CO, N_2O) fluctuations prior to the Mauna Loa record. These data, including the carbon isotopic records contained therein, must be exploited to the fullest extent possible to assist not only in defining the past role of the biosphere, but also in understanding the climatic significance of atmospheric CO_2 fluctuations in the geologic past (Chapters 3 and 10).

11.1.3 Changing Ocean Circulation and Atmosphere-Ocean CO_2 Exchange

Both the internal distribution of carbon species and the net uptake of CO_2 from the atmosphere by the ocean reservoir are the result of the interplay between biological processes and ocean circulation

and chemistry (Chapters 5, 7, and 8; Viecelli 1984). Marine photosynthesis, decomposition, and the tightly coupled nutrient cycles are controlled to a large extent by physical factors: temperature, irradiance, vertical mixing (e.g., effect on irradiance; Lewis et al. 1984), and the patterns of water circulation (Chapter 5). Climate change could affect the latter, producing significant changes in ocean CO_2 uptake mediated by the marine biota (Chapter 8). Explicit parameterizations for the active participation of biota in ocean models, including the quantitative effects of qualitative differences (e.g., species composition of the biota), have not been treated previously in sufficient detail. These can no longer be ignored because knowledge of biologically mediated fluxes is critical to accurate interpretation of numerous tracer distributions used in model calibration (Chapters 5 and 7) (see Betzer et al. 1984).

The relative simplicity of current ocean CO_2 models may introduce unnecessary error into the results (Section 10.2.8). To develop an explicit treatment of water circulation and the potential feedback relationships with climate change, including biotic phenomena, the surface ocean component of the model must be parameterized for seasonally and regionally varying heat and freshwater (continental and atmospheric) inputs, sea ice, and wind stress (Broecker et al. 1985a). *Thus, development of a general circulation model of the living ocean is needed to provide the most rigorous estimates of ocean CO_2 uptake, and, more importantly, to delve into the critical relationship between evolving CO_2-climate change and the atmosphere-ocean carbon system.* Such a model may be a critical need for development of more realistic climate models (see Chapter 1 and MacCracken and Luther 1985).

A set of outstanding tools consisting of oceanic chemical and radioisotope tracers is available for studying natural CO_2-climate relationships (see also Chapters 5 and 8). Because the climate problems involve long-term and large-scale phenomena, tracer studies apply particularly well. The information provided by tracers integrates on the time scales of decades, centuries, or longer. This is needed to provide an understanding of the past history of the carbon cycle and the potential perturbations by climatic phenomena. The only comprehensive tracer studies on the oceans with relevance to CO_2-climate research are the GEOSECS (Geochemical Ocean Section Study) program (1972–1978) and the TTO (Transient Tracers in the Ocean) project started in 1981. The GEOSECS program provided excellent but limited data, essentially along two longitudes in the Atlantic, Pacific, and Indian oceans. The purpose of the TTO program was to repeat and expand tracer studies to obtain a fully three-dimensional picture of tracer distributions in all oceans. It now appears, however, that after the first successful experiment (the North Atlantic Study) and a severely reduced Tropical Atlantic Study, the completion of the program is in jeopardy.

A mechanism for determining the rate of ocean uptake independently of ocean model calculations (e.g., in situ) would obviously be very important in validating ocean carbon cycle models. A number of options have been proposed (e.g., monitoring dissolved carbon chemical species in the water column at a number of fixed geographic locations; Chapter 5). However, significant logistical and theoretical difficulties are currently associated with these options, and it will be some time before their efficacy can be demonstrated (Chapter 5). Long-term monitoring of the oceanic carbon cycle (e.g., Scientific Committee on Ocean Research 1984) independent of synoptic tracer measurement programs is clearly important, but, taken alone, it is an unsatisfactory alternative because ocean circulation and climate effects will not be adequately described from such data. *Thus, long-term measurements of the oceanic carbon system should be a component in a coordinated global tracer research program.*

11.1.4 Carbon Storage or Release of Subfossil Pools (Peat, Sediments, Paleosols)

The direct links between terrestrial, freshwater, and coastal marine systems via the hydrosphere, rather than through the atmosphere, have essentially been ignored in the current generation of carbon cycle models. This is another manifestation of the assumptions of current models; if *humans* have not measurably altered a carbon flux, it need not be considered. However, data obtained under the auspices of the Transport of Carbon and Minerals in Major World Rivers Program sponsored by SCOPE/UNEP (Chapter 5) now indicate that *both natural and human-induced carbon storage in sedimentary deposits of rivers and coastal marine environments may be important in the context of a contemporary global balance sheet.*

The *enhanced* annual carbon sink resulting from human intervention in fluvial systems may approach 1×10^{15} g C (Chapter 5; Degens et al. 1984; Kempe 1984). Combining estimated *natural* sinks from peat formation and from riverine, lacustrine, and coastal sediment deposition yields another sink of about the same magnitude (Section 8.5.2). Thus, it appears that the land-river-ocean interface can no longer be treated as superficially in modeling as it has been up to now.

An interesting dilemma arises because long-term geochemical models require that the average annual flu-

rate into subfossil carbon pools be $\leq 0.1 \times 10^{15}$ g C (Chapter 8). This leads to several important questions: If the former estimate of a natural freshwater and coastal carbon sink of about 1×10^{15} g/year is correct, what was the natural return flux to the atmosphere that counterbalanced the subfossil sink? Could this have entailed significant releases of methane as well as CO_2 from the subfossil pools, even in the natural condition? How have human manipulations of rivers and their watersheds, such as channelization, dredging, impoundment, eutrophication, erosion, and storage of eroded materials in local alluvia (floodplains, low-order channels), upset the natural order? How transient are the estimated contemporary sinks induced by human interference?

These are very difficult questions. Our existing data bases and perspectives are very poor, particularly for phenomena linking inland catchments and river floodplains with estuaries and coastal deltas. *This appears to be an area of research where the judicious gathering and interpretation of isotopic records is needed in combination with better inventories and flux measurements of peat, sedimentary deposits, and terrestrial and floodplain soils. Rethinking of the atmospheric CO_2 reconstructions is also indicated* (Sections 10.2.7, 11.1.2.3, and 11.1.5).

11.1.5 Effects of Climate and CO_2 Fertilization on Global Carbon Storage

Climate change should alter measurably the seasonal and geographical patterns of CO_2 concentrations recorded by the international network of CO_2 monitoring sites. The details of the CO_2 response to changing climate will depend on many factors (e.g., rate of fossil fuel use, ocean thermal lag, role of other greenhouse gases). Atmospheric CO_2 records now give the most precise and spatially integrated signal of the regional imbalance of respiration and photosynthesis. But similar records of CH_4 and CO (and eventually $^{13}CH_4$, ^{13}CO) (i.e., the full suite of all three principal gases of the carbon cycle) are also needed to clarify oceanic versus biospheric exchanges with the atmosphere and to specify regionally important biospheric and soil processes (e.g., CO_2 fertilization, losses from tropical deforestation, soil respiration). Of particular interest is the latitude dependence of the increasing atmospheric content of CH_4 and CO (affected by heterotrophic responses to deforestation or retreat of the boreal permafrost) (Section 3.5). An evolving biospheric-atmospheric exchange, as seems to be occurring now in the growing CO_2 seasonal amplitude in the Northern Hemisphere, may be the early signature of climate change.

The implications of climate change, particularly in boreal regions (Section 10.3.2), *focus attention on the need to resolve current disagreements about the potential climatic responses of northern ecosystems (Section 6.6.1) and to suggest greater emphasis in research on climatic effects in northern areas.* Furthermore, the implied changes in land cover indicate the potential for significant changes in the Earth's albedo and in regional surface hydrology that should be considered in simulating the sensitivity of climate to changes in methane and CO_2 concentrations. Hastened releases of CO_2 as well as methane from better-drained bogs and other sites in the warming boreal or taiga zone could provide a positive feedback, accentuating a further excursion.

For the global biosphere of the future in the potentially warmer world of higher greenhouse gases, the relative importance of enhanced photosynthetic uptake (CO_2 fertilization, β-factor response), enhanced respiration, wildfires (see Section 8.6.1), *and human disturbance during the next century cannot yet be predicted with confidence, and may only be determined from the future CO_2 record itself* (Chapters 3 and 10).

11.2 RECOMMENDATIONS: A GLOBAL VIEW

The preceding examples of research needed on future CO_2 emissions and the whole biosphere, lithosphere, and hydrosphere in relation to the atmosphere continually remind us of the need for worldwide inference, even where data from certain economic regions, ecosystems, geologic strata, ship cruises, or aircraft tracks are merely increments to collective global knowledge, extending from a century of prior synthesis.

For the next century, researchers will no doubt continue to be opportunistic in rising to the challenges just outlined, and to others perceived dimly, if at all, right now. Efforts and funding will shift directions, partly as new means are found to reach these ends. *But there is also an overriding responsibility to reconnect the isolated thrusts into a view that is more global—not only worldwide, but transcending disciplinary or mission specialties* (e.g., Ramanathan et al. 1985).

11.2.1 Future CO_2 Emissions

In no aspect of the carbon system is such global inference more difficult than in modeling fossil fuel CO_2 emissions: Enter the complex human equation, complete with the threat of global nuclear disaster and unstable political and socioeconomic structures, but also with the latent capability for adaptation to major inventions and social and environmental changes.

Thus, our future is unknowable, and much of the uncertainty in global estimates of energy-related CO_2 emissions is doubtlessly irreducible. *Modeling emissions nevertheless provides a necessary tool for treating these uncertainties quantitatively, even if the potential for foreseeing all analytical aspects of the human equation is slim.* Differences in structure and basic assumptions among current emissions models are such that we cannot now rank the key determinants of CO_2 releases, even on a semiquantitative basis. Achieving some degree of consensus on the key issues in emissions modeling is timely, but should not be enforced in a way that encourages bandwagons or penalizes valid dissent. When the uncertainty of model structure, as well as parameter values, is more fully addressed, it will again be profitable to extend this analytical tool to assess emissions of other important greenhouse gases.

11.2.2 The Biosphere

Ecologists have a motivation for and a tradition of integrating findings from diverse systems such as forests, tundra, and grassland into a global picture. But there are problems in relating global synthesis to the scale at which investigators examine local carbon cycle issues (e.g., food webs and habitat variations). *A significant challenge remains in using principles developed from local studies to improve on current regional estimates, and then to apply judgment in weighting regional averages for scaling up to the global biosphere.* Problems presented by the rapid mixing of CO_2 from local sources into the rapidly mixed atmosphere represent a classic challenge for testing whether seasonal trends can be resolved into sensible combinations of income and loss rates and the correct landscape mosaic.

Tracing waves of change in land conversions gives explicit attention to human history (Richards et al. 1983), reaching back earlier than the 1860 milestone date generally estimated as the start of significant fossil fuel releases. *Yet we can only treat the changes by studying age sequences and carbon time series for recovering vegetation and soil. We need to resolve income and loss sides of the major ecosystems' budgets for carbon; also for nutrients or toxic materials that can affect either side of Equations (8.1) and (8.2). This cannot be accomplished without developing sound regional models for the carbon cycle in both soils and vegetation.*

The role of heterotrophs (that is, consumers and decomposers of organic food) in modifying the annually recycled flux of over 50×10^{15} g/year of carbon is one of the basic problems at the ecosystem level of biology. Clearly, this flux, and that of mineral nutrients and compounds of nitrogen that regulate this flux, could be altered by farming and famines, as when climatic averages and spatial patterns became altered by CC CH_4, or other atmospheric gases. Whether primary production could be much altered by direct or indirect feedbacks from animals and decomposers, as well as by CO_2 itself, is a thorny problem, requiring that global ecological and physiological approaches become combined in order to progress.

11.2.3 The Atmosphere

Field surveys and landscape experiments will always be unevenly pursued over the wide horizons of land and sea. Logistical problems will always provide some limitations. Even static inventories cannot be spread over all the Earth's gradients to gain really close interpolations needed to explain changes in the well-mixed atmosphere.

At present, the atmospheric signal of CO_2 is not strictly explainable by coupling with a hydrosphere represented only by the ocean or an idealized spatially homogeneous biosphere. *The differences in spatial scales at which atmospheric and biospheric processes operate must be considered more fully in future research, and the direct coupling of the biosphere and hydrosphere requires explicit treatment of the land-river-ocean link.*

Important clues are certainly imbedded in the complex response of the global carbon cycle to extreme natural fluctuations like the 1982/83 ENSO event. *Global atmospheric CO_2 monitoring networks should be extended to carefully selected sites in the major land biomes in both tropical and higher latitudes in close coordination with intensified research on the use of satellite remote sensing and related field measurement studies* (see Section 8.6.1) *to quantify the photosynthetic activity and carbon exchanges of these biomes* (Sections 3.5, 6.6.1, and 8.6). *Accomplishing the foregoing objectives will also require more realistic atmospheric circulation models, better parameterized for atmospheric convection and for biospheric exchanges, based on knowledge of the biota rather than derived from simple models* (Section 7.5.1).

Evidence from ice cores and other geologic records of past CO_2-climate fluctuations indicates that the global climate change predicted for the coming generations will doubtless impress its signature on this atmospheric CO_2 record, which indeed may offer some of the most sensitive possibilities for its early detection. *A synthesis of all such data must be applied to the outstanding questions of the global carbon cycle: Is the biosphere currently a net source or sink? What will be the future source and sink roles of the biosphere and ocean when the carbon cycle is more strongly perturbed by great rates of fossil fuel input?*

11.2.4 The Hydrosphere

An ocean research program devoted to understanding climate change and carbon cycle relationships is a massive undertaking, much too large to be supported either by a single U.S. government agency (Webster 1984a) *or by a single nation.* Massive new efforts, depending critically on satellite remote sensing of ocean current structure, surface wind stress, and meteorological variables, have been proposed, but little in the way of new research has actually begun (Webster 1984b). It has been argued that a better description of the ocean's structure and circulation is needed before we can attempt to understand its role in climate (Webster 1984b). *Although we support proposed new efforts in global ocean climate research, including an expanded role for satellite imagery, we must point out that a major source of information needed for better understanding of ocean circulation exists in oceanographic tracer data.* Support for continued tracer measurements is now being included in planning for some major programs such as the World Ocean Circulation Experiment (WOCE)(Webster 1984a).

The value of oceanographic tracer studies in establishing constraints on ocean circulation models was amply demonstrated (e.g., Broecker et al. 1961; Sarmiento 1983; Wunsch 1984; Broecker et al. 1985b). Previous tracer studies also proved their worth by uncovering a number of surprises about the inner workings of the oceans (Chapter 5). This demonstrates the necessity to obtain the largest, most diverse, tracer data base possible to determine the nature and driving forces of ocean circulation.

Given the importance of the oceanic carbon cycle/climate link, tracer research ought to receive a higher priority than currently assigned. At the present time, it appears that completion of the TTO program, appropriately integrated with physical oceanography, provides a major opportunity to obtain the data required to achieve carbon cycle and climate research objectives. The required integration can be achieved through more efficient sampling strategies and multiship coordination of small-volume (^{3}H, ^{3}He, chlorofluorocarbons, chemical tracers) and large-volume (^{14}C, ^{39}Ar, ^{85}Kr, and ^{228}Ra) sampling. Over the longer term, development of tandem accelerator mass spectrometry capabilities for radiocarbon measurements may allow use of small-volume (1 liter) samples and greatly reduce logistical problems. *The projected time scale for ocean circulation model development and the apparent inadequacies in current data and understanding of ocean circulation indicate that this undertaking should be regarded as a basic research effort and likely to remain so for decades. A worldwide tracer program like the TTO concept is of utmost importance for the long-range success of the effort of modeling the carbon cycle, along with climate and ocean interactions. Experimental worldwide data are needed to test the models and for credible predictions. A TTO or another similar program will have to be carried out for a few decades starting as soon as possible, particularly while the tracer transients initiated by nuclear weapons testing are still evolving* (see Chapter 5).

11.2.5 The Lithosphere

We can and must wring far more information from the geological records of atmospheric CO fluctuations and paleoclimate, CO_2-climate relationships (i.e., from ice and deep-sea sediment cores), and critical links between cycles of carbon and other elements in the recent and distant past (Chapters 3 and 8). *Because we cannot perform truly global experiments, we need to decipher natural records of these phenomena* (Broecker et al. 1985b).

Our current knowledge of the concentrations of carbon and sulfur isotopes in rocks as a function of time is rudimentary; there are only a few thousand carbon isotope determinations of rocks spanning the last 700 million years—averaging one determination for each 350,000 years! The carbon, oxygen, and sulfur isotope concentrations in rocks and sediments reveal changes in the metabolism of the earth: the movement of materials between sedimentary reservoirs, the rates of production and storage of oxygen and carbon, and the imprint of climate change on these geochemical cycles. It is important that we be able to understand what drives this slow-breathing system on which shorter term human perturbations have been imposed, this flywheel that has kept the Earth habitable for these billions of years. We should welcome the opportunity to provide more evidence of its self-regulation (i.e., through natural feedbacks), especially if that were to moderate the human cost of regulation or adaptation to change (Section 8.6.2).

The coupling of C with S, and with N and P, and the associated uncertainties discussed in Chapter 8 provide a few important examples out of many that will require additional attention before the natural carbon cycle itself can be fully understood. Also emphasized is the linkage of silicate weathering and even "reverse weathering," relating the rock cycles of calcium and other metallic nutrients to soil formation, erosion, and marine processes (see Chapter 8).

It should be feasible to explore how the atmospheric changes between glacial and interglacial times, over tens or hundreds of thousands of years and in diverse global climates, might compare with the shifts over millions of

years of slowly cooling average climate since the warm Cretaceous and Eocene times. Doing this, however, will require attention to transient conditions and time constants of several kinds, because the conditions of interest may be changing before equilibrium (i.e., balancing around the whole cycle) can even be attained. *We reiterate that the large organic pools of vegetation and soils will require treatment not only as conduits irrelevant for carbon storage, but as state variables and parameters that modify physical rates such as rock weathering* (Berner and Barron 1984).

11.2.6 Modeling and Evaluation

Besides the problems inherent in modeling each of the different spatial and temporal scales, future research on how to link models to cover the widely different space/time intervals relevant to global cycling will require more attention. This volume has often emphasized a globally averaged scale. Global modeling provisionally accepts simplification to get the essence of the answers to important questions. Yet far more effort has been applied to smaller scale problems, mostly with a single geochemical species. Thus, there still remains a critical need (Olson 1971) for more modeling across a hierarchy of space-time scales.

One approach to the desired coupling is to start from intermediate levels of land and ocean complexes. These can be modeled now with fuller insight than earlier, drawing on local data or modeling of subsystems. Obviously, much remains to be accomplished! *Our principal objective now must be to develop sound models of the critical subsystems: soils, vegetation, oceans, freshwater interfaces, and atmosphere.* Yet, after such research has proceeded to the point where sound models of subsystems are available, modelers must turn to the problem of global coupling: of atmosphere with all other subsystems; of land with oceans, via rivers, lakes, and estuaries; and of cycling of carbon with other elements, particularly the biologically limiting nutrients, phosphorus and nitrogen. Which rate-limiting factors interact? How changeable are the ranges of response times and the quasi-periodic forces on such cycling (see Figure 8.4)?

From regional modeling, we can reach both higher and lower levels of scale in time and space and of complexity. Longer time scales will include not only geologic eras but also subfossil deposits that feed partly to the stratigraphic column and partly back to the episodic releases of extra CO_2 and CH_4. Climatic effects will be explicit, as they are shown to condition the daily and yearly cycles for local pulses of life. A hierarchic strategy of modeling can evolve as required to reach across all time scales that have meaning when coupled together. Then, insight on the natural cycling of elements can even better help us put human disturbances in global perspective. After learning more about the past, from thousands to millions of years ago, we can better renew the emphasis on centuries and decades, future as well as past.

11.3 CONCLUDING COMMENTS

The system comprising the Earth's surface, atmosphere, and biota might be viewed as if in homeostasis so that the chemical composition of the atmosphere and the climate are both regulated by biota to *its preference*. No intricate planning, foresight, or cooperation appears to be needed. Regulation can arise as a simple colligative property of life and its environment. One such view, after Vernadsky (1926) and Hutchinson (1954), was recently restated as the Gaia hypothesis (Hitchcock and Lovelock 1967; Lovelock 1979, 1983). It offers a challenging perspective on the natural cycles of the elements.

Such a view suggests, for example, that both oxygen and CO_2 are regulated by biota, even though their steady states shifted greatly over geologic time. The maintenance of both of these gases within the narrow range of acceptable levels for higher forms of life over the last hundreds of millions of years might otherwise seem like a noteworthy coincidence. The atmospheric content of oxygen cannot have varied outside 15 to 25% because of the steep function relating flammability to the oxygen-mixing ratio. Carbon dioxide may have changed severalfold, but seems constrained by the requirements of photosynthesis and of the planetary radiation balance (Sections 8.3, 8.5, 8.6, and 11.2.5). One mechanism for long-term biotic regulation of CO_2 involves the effect of biological activity on CO_2 uptake from the atmosphere via rock weathering (Berner and Barron 1984).

This planetary life-support system has been remarkably effective, yet is not perfect, judging from the evidence of periodic mass extinctions, including the most-discussed one occurring at the Cretaceous-Tertiary transition. We still have much to learn from (or about) long-term biospheric disturbances, to understand the human-modified zone of life, the "noosphere" (e.g., Vernadsky 1926). Human effects on atmospheric composition and the size and operations of the terrestrial ecosystems represent major excursions that may yet overwhelm the life-support system crafted in nature over billions of years. Or analysis might reinforce our wish or faith that negative feedbacks must counteract any wide departure, leaving life and human cultures to adjust in many small ways instead of large ones.

In closed systems, disturbances of the system environment lead to adjustments toward a new equi

librium that partly compensate for the perturbation. ow much analogy can be found in the complex, open ystems of the real, living world? To answer, it is necessary to develop models for cycles of elements and of climate that contain explicit formulations for more active participation by the organic agents. We need to compare such models' predictions with those of the geochemical cycles, which allow more of a spectator role for biota, in order to determine whether tentative simplifications made early (e.g., steady state), to make entree into such a complex inquiry, can persist and be robust enough to be shared by the different models designed for quite contrasting domains of time or space.

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GLOSSARY

2B	Two-box model
AD	Advection-diffusion model
AF	Airborne fraction
AF*	Effective airborne fraction
AIT	Action initiation time
AMS	Amsterdam Island
API	American Petroleum Institute
ASC	Ascension Island
ASTM	American Society for Testing Materials
AVHRR	Advanced very high resolution radiometry
AVI	American Virgin Islands
AZR	Terceira Island, Azores
B-A	Box-advection model
B-A-D	Box-advection-diffusion model
B-B	Two-box model
B-D	Box-diffusion model
BD	Box-diffusion model
BJMO	Björkström ocean model
BRW	Barrow, Alaska
CAC	Climate Analysis Center
CAM	Crassulacean acid metabolism
CBA	Cold Bay, Alaska
CDIC	Carbon Dioxide Information Center
CEC	Commission of the European Communities
CFI	Continuous forest inventory
CMO	Cape Meares, Oregon
COS	Cosmos, Peru
CSIRO	Commonwealth Scientific and Industrial Research Organisation
DOC	Dissolved organic carbon
DOE	U.S. Department of Energy
EIA	U.S. Energy Information Administration
ENSO	El Niño/Southern Oscillation
EOF	Empirical orthogonal function
EOR	Enhanced oil recovery
EPRI	Electric Power Research Institute
ER	Edmonds and Reilly
ERGB	Edmonds, Reilly, Gardner, and Brenkert
FAO	U.N. Food and Agricultural Organization
FGGE	First Global Geophysical Experiment

FLK	Falkland Islands
FY	Fiscal year
GCM	General circulation model
GEOSECS	Geochemical Ocean Sections
GMCC	Geophysical Monitoring for Climatic Change
GMI	Guam, Marshall Islands
GNP	Gross national product
GPP	Gross primary production
HBA	Halley Bay, Alaska
IAEA	International Atomic Energy Agency
ICSU	International Council of Scientific Unions
IEA	International Energy Agency
IEA/ORAU	Institute for Energy Analysis, ORAU
IGY	International Geophysical Year
IIASA	International Institute for Applied Systems Analysis
ITCZ	Intertropical convergence zone
KEY	Key Biscayne, Florida
KPA	Kitt Peak, Arizona
KUM	Point Kumukahi, Hawaii
LP	Linear programming
MB	Multibox model
MBC	Mould Bay, Canada
MBL	Marine Biological Laboratory
MIT	Massachusetts Institute of Technology
MKO	Mauna Kea, Hawaii
MLO	Mauna Loa, Hawaii
M-SK	Midcase trajectory from Seidel and Keyes
NAS	National Academy of Sciences
NBS	National Bureau of Standards
NEP	Net ecosystem productivity
NOAA	National Oceanographic and Atmospheric Administration
NORPAX	North Pacific Experiment
NPP	Net primary production
NRC	National Research Council
NW	Niehaus and Williams
NWR	Niwot Ridge, Colorado
NY	Nordhaus and Yohe
NZL	Kiatorete Spit, New Zealand
OECD	Organisation for Economic Cooperation and Development
OPEC	Organization of Oil Exporting Countries
ORAU	Oak Ridge Associated Universities
ORNL	Oak Ridge National Laboratory
pCO_2	Partial pressure of carbon dioxide
PPTN	Precipitation
PSA	Palmer Station, Antartica
R/V	Research vessel
SCOPE	Scientific Committee on Problems of the Environment
SCOR	Scientific Committee on Ocean Research
SEY	Seychelles Islands
SIO	Scripps Institution of Oceanography
SK	Seidel and Keyes
SMO	American Samoa
SOA	State of the Art [Report]
SOI	Southern oscillation index
SPO	South Pole

SST	Sea surface temperature; Station surface temperature
STC	Weathership Charlie
STM	Weathership M
TA	Total dissolved alkalinity
TC	Total dissolved inorganic carbon
TTO	Transient Tracers in the Ocean
T.U.	Tritium unit
UNEP	United Nations Environment Programme
UNESCO	United Nations Educational, Scientific, and Cultural Organisation
WEC	World Energy Conference
WMO	World Meteorological Organization
WOCE	World Ocean Circulation Experiment

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