ORNL/TM-6651

Simulation of Land-Use Patterns Affecting the Global Carbon Cycle

Y. H. Chan
J. S. Olson
W. R. Emanuel

ENVIRONMENTAL SCIENCES DIVISION
Publication No. 1273

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National Technical Information Service
U.S. Department of Commerce
5285 Port Royal Road, Springfield, Virginia 22161
Price: Printed Copy $9.00; Microfiche $3.00

ORNL/TM-6651

Contract No. W-7405-eng-26

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Submitted as a dissertation by Y. H. Chan to the Graduate Council of the University of Tennessee in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

Date Published: January 1979

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ACKNOWLEDGMENTS

This document is a progress report on part of an assessment of "Global Carbon Cycles and Climatic Risks" for the United States Department of Energy's Office of Technology Impacts. Y. H. Chan wishes to acknowledge additional support by the Graduate Program in Ecology, University of Tennessee. The work reported here has also been provided by Y. H. Chan to the University of Tennessee in dissertation form to fulfill requirements for the Ph.D. degree.

The review and comments on this research by B. L. Dearden, T. W. Kerlin, University of Tennessee and R. I. Van Hook, W. F. Harris, D. E. Reichle, S. I. Auerbach, J. K. Munro and G. G. Killough all with ORNL are gratefully acknowledged. J. R. Jones, ORNL provided assistance with computer graphics. Linda Allison, ORNL edited portions of the manuscript.

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ABSTRACT

CHAN, Y. H., J. S. OLSON, and W. R. EMANUEL. 1978. Simulations of land-use patterns affecting the global carbon cycle. ORNL/TM-6651. Oak Ridge National Laboratory, Oak Ridge, Tennessee. 182 pp.

Past increase of atmospheric CO_2 involves significant contributions from both fossil and nonfossil (biospheric) sources. The latter are controversial, partly because these CO_2 releases may be balanced by accelerated regrowth following clearing of some forests, while others were being converted to agricultural or other nonforest land. A simulation model was used to reconstruct changes since 1860 and project four hypothetical future scenarios of CO_2 injection to 2460. Nineteen compartments and their exchanges of carbon were considered. Areal extent of tropical forests, other wooded ecosystems, and nonforests were incorporated into the model. Rapidly and slowly exchanging pools of carbon per unit area, and net primary production for each pool and ecosystem group, were projected by integrating income-loss differential equations numerically using CSMP programming language.

Estimated cumulative releases of CO_2 from fossil fuels (plus cement) near 120 Gtons of carbon (1 Gton = 10^9 metric tons) from 1860 to 1970 were assumed to equal prompt and delayed releases from forest clearing. Limits of exploitable forest area and biomass were evaluated and found to contribute much less future CO_2 than the usable coal, oil, gas, and oil shale. Ultimate release from the latter ($7500 \pm 2500 \times 10^9$ tons of C) could increase atmospheric CO_2 manyfold: doubling the assumed 1860 levels as early as (1) year 2025 for assumed "nominal"

scenario (expanding releases slightly less rapidly than at present), (2) year 2033 for a "delayed" expansion scenario that would prolong use of fossil reserves (lowering peak carbon release rate from ∿ 43 to ∿ 28 Gtons/year), (3) year 2087 for a "slow burner" scenario (increasing very slowly from present levels), and 94) year 2290 for a "combination" scenario (which assumes low fossil-fuel use, high carbon storage, and high net primary production of forested ecosystems). Depending on the poorly known parameters that were programmed to constrain the organic production rates, cumulative storage, and the response of plants and soils to enhanced atmospheric CO_2, biospheric storage might reach higher levels for all scenarios than the estimates given here. However, maximizing such storage (and helping to minimize atmospheric CO_2 for any given energy policy) in real life would require very much closer understanding and wiser management of ecosystems than history has shown.

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CHAPTER 1

INTRODUCTION

1.1. AN OVERVIEW OF THE GLOBAL CARBON DIOXIDE PROBLEM

People have long recognized that coal, petroleum, and natural gas can readily supply more energy per unit weight or volume than other nonnuclear energy sources such as wood or water power. The industrial revolution initiated an era of unprecedented exploitation of these fossil fuels. Since 1860, human population has increased rapidly. Harvesting of timber has accelerated to meet the demands of industry and homes (as fuelwood). Large areas of forest have been cleared and converted into agricultural lands to produce more food for a growing population. Carbon dioxide released from fossil-fuel combustion and from significantly increased oxidation of organic matter in forests and soils after land clearing became carbon sources to the atmosphere (Baes et al., 1976, 1977; Olson et al., 1978; Woodwell, 1978).

The emissions of carbon dioxide due to human activities are almost certainly responsible for the worldwide 10% increase in atmospheric CO_2 concentration over the past 20 years (e.g., Bolin and Bischof, 1970; Keeling et al., 1976b). An increase in the level of ambient carbon dioxide may possibly enhance the photosynthetic absorption of CO_2 by terrestrial plants (Gaastra, 1959; Lemon, 1977; Loomis, in press). A portion of the excess may enter the oceans (Pytkowicz and Small, 1977; Revelle and Suess, 1957).

Actual measurements of the concentration of atmospheric carbon dioxide show that about 56% of the CO_2 released from estimated fossil-fuel utilization between 1958 and 1974 has remained in the atmosphere (Keeling and Bacastow, 1977). Evidently, the excess carbon dioxide from anthropogenic releases is not fully absorbed by the terrestrial biosphere and the ocean. Presently, it is not clear how to account for the amount of carbon dioxide supposedly released to the atmosphere which has been transferred into the hydrosphere and the terrestrial biosphere (Kerr, 1977; National Academy of Sciences, 1977; Stumm, 1977). Controversies have arisen concerning the role of the biosphere as a net source or sink for excess carbon dioxide in the atmosphere (Andersen and Malahoff, 1977; Woodwell and Houghton, 1977).

The global climatic pattern will be influenced by the increase in atmospheric carbon dioxide. As early as 1861, Tyndall suggested that an increase in the carbon dioxide content of the atmosphere may result in a rise in the air temperature (see also Arrhenius, 1896; Callendar, 1938).

Although water vapor, ozone, and aerosols are important factors influencing the radiative balance of the earth's surface, experts agree that their effects are minor compared with effects of atmospheric CO_2 (Palmer, 1973; Hobbs et al., 1974; Baldwin et al., 1976; Kellogg, 1978). Carbon dioxide, though virtually transparent to shortwave solar radiation, is a good absorber of radiation in the band of the infrared spectrum (12-17 μm), where an appreciable flux of energy is radiated by the earth's surface and the lower atmosphere. This accounts for the

"greenhouse" effect of CO_2. Solar radiation passes through the atmosphere largely unattenuated, but the atmosphere absorbs and retains more longwave radiation from the earth when more CO_2 is present. The result is a net warming of the lower atmosphere, which receives both the solar and returned longwave radiation, and a cooling of the upper atmosphere, which then emits relatively less outgoing flux of radiation to space for any given surface temperature (Plass, 1956, 1972b).

The effect of atmospheric CO_2 on the average temperature of the earth's surface has been studied by many scientists for the last hundred years (e.g., Schneider, 1975). Models of varying complexity have been constructed for predicting climatic changes. Initially, simple models of one-dimensional radiation balance with different feedbacks from other constituents of the atmosphere were developed (e.g., Arrhenius, 1896; Callendar, 1938; Plass, 1956; Möller, 1963). Exemplary of the recent state of the art is the model developed by Manabe and Wetherald (1975). This is a three-dimensional general circulation model (GCM). For a doubling of CO_2 in the atmosphere, their model predicts about a 3 °K rise in the average temperature of the lower atmosphere at middle latitudes and a much higher rise in the polar regions. Most modeling efforts have arrived at the general consensus that the temperature response is approximately logarithmic. Each doubling in the concentration of CO_2 in the atmosphere produces about the same increase in the average temperature of the air at the earth's surface. The amount of the average increase in temperature per doubling of CO_2 given by the various models have been critically reviewed by

Schneider (1975). He estimated that the magnitude of "state-of-the-art estimate" to be in the range of 1.5 to 3 °K, with general agreement that a larger increase can be expected in the polar regions.

Several symposia and workshops have been held in various places in the last few years to facilitate the cooperation and exchange of ideas so important in exploring the many aspects of the problem. These gatherings include SCEP (1970), SMIC (1971), the 24th Brookhaven Symposium in Biology (Woodwell and Pecan, 1973), the Dahlem conference (Stumm, 1977), Office of Naval Research Conference on the Fate of Fossil Fuel CO_2 (Andersen and Malahoff, 1977), the Scientific Workshop on Atmospheric Carbon Dioxide organized by the World Meteorological Organization (1977), and the ERDA Miami Workshop on Significant Environmental Problems (Elliot and Machta, in press). These groups acknowledge, with increasing emphasis and consensus, that the biosphere is important as a resource which will be altered by any major climatic shift, and that the role of the biosphere in carbon cycling had been underestimated.

1.2. SECULAR TREND OF CARBON DIOXIDE CONCENTRATION

Since the detection of CO_2 as a constituent of air by Bergman in 1774, there have been many measurements of the amount of CO_2 in the atmosphere. Callendar (1940, 1958) examined the measurements of atmospheric CO_2 from 1866 to 1956 and argued that the secular increase of CO_2 beginning in about 1900 was due to the burning of fossil fuels by man. Bray's (1959) statistical analysis of the reliability of the

data reached a similar conclusion. He also emphasized the need for further sampling at selected locations. The concentration of atmospheric CO_2 around 1900 was given by him to be about 290 to 293 parts per million by volume (ppmv).

Beginning with the International Geophysical Year in 1958, systematic measurements of the concentration of atmospheric CO_2 became available. These include the nearly continuous record maintained since 1958 at the Mauna Loa Observatory in Hawaii (19° N latitude, 3400 m altitude; Figure 1). The measurements were obtained with a continuously recording infrared gas analyzer, after correction for local effects as described by Machta (1972b), Pales and Keeling (1965), and Keeling et al. (1976b). Scandinavian data derived primarily from flask samples aboard aircraft were discussed by Bischof (1970) and Bolin and Bischof (1970). Shorter records of infrared gas analyzer measurements from Point Barrow, Alaska (73° N latitude) are also available (cf. SCEP, 1970, p. 42).

In the Southern Hemisphere, South Pole data were obtained mostly from flask samples collected since 1957 (Keeling et al., 1976a). New efforts also were initiated to measure the concentration of atmospheric CO_2 around Australia and New Zealand (Pearman and Garratt, 1973; Garratt and Pearman 1973a, 1973b).

Three types of fluctuation are apparent in these records. The first is a seasonal variation with an amplitude of about 6 ppm for Mauna Loa data (Keeling et al., 1976b). The variation has a maximum in spring and a minimum in late summer corresponding to the seasonal changes in net photosynthesis and respiration of the biosphere (Olson, 1970a;

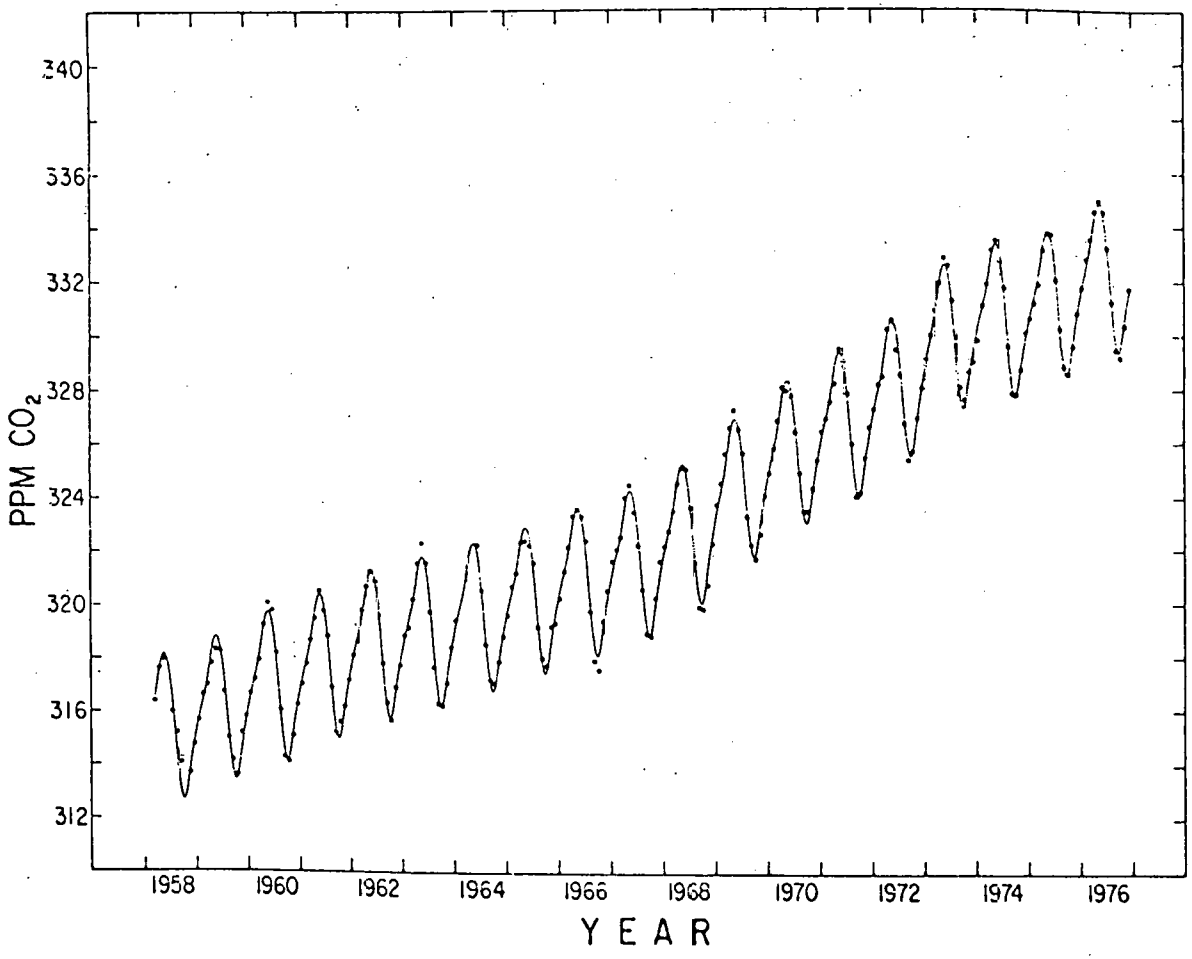

Figure 1. Trend in the concentration of atmospheric CO_2 at Mauna Loa Observatory, Hawaii.

Source: Keeling, C. D., and R. B. Bacastow. 1977. Impact of industrial gases on climate. pp. 72-95. In Energy and Climate. National Academy of Sciences, Geophysics Research Board, Geophysics Study Committee, Washington, D.C.

Lieth, 1963; Junge and Czeplak, 1968). Ground-level seasonal amplitudes of the Scandinavian data are about 15 ppm and damp out with a monthly lag at 10 km altitude, and even more lag at 11 km (average stratosphere altitude, with only 0.7 ppm average seasonal amplitude) (Bischof, 1970; Bolin and Bischof, 1970; Bolin and Keeling, 1963). Data from the Southern Hemisphere also display a seasonal variation (November high vs. March low) but with a smaller amplitude of about 1.6 ppm (Keeling et al., 1976a). The out-of-phase relation with the growing season suggests the presence of a rather strong interhemispheric exchange process with a delay of about six months (Bolin and Keeling, 1963; Junge and Czeplak, 1968).

Dominating these CO_2 time series is a general upward trend of atmospheric CO_2 with time. This averaged about 0.76 to 0.72 ppm per year at Mauna Loa and the South Pole, respectively, from 1959 through 1968, and over 1.0 ppm after 1969. The increase from 1959 to 1971 is 3.4 and 3.1% for Mauna Loa and the South Pole, respectively (Pales and Keeling, 1965; Brown and Keeling, 1965; Keeling et al., 1976a,b). Also noticeable is a change in the annual rate of CO_2 increase. It apparently declined during the early 1960s and increased after 1967. The decline was interpreted by Bainbridge (1971) as a consequence of cooling of surface ocean water by about 1 °K. However, Bacastow (1976) regarded the variation as the result of the Southern Oscillation, a large scale atmospheric and hydrospheric fluctuation with an irregular period of one to five years. The connection, if present, indicates that a principal cause of the variation may be a change in the rate of removal of CO_2 by the oceans (Machta et al., 1977).

1.3. THE TERRESTRIAL BIOSPHERE AS A SOURCE OR SINK OF ATMOSPHERIC CO_2

The concentration of atmospheric CO_2 has been observed to increase steadily since 1958, but the annual increases approximate only 50% of the carbon dioxide released each year by fossil-fuel combustion (Baes et al., 1977; Keeler and Pytkowicz, 1977). Until recently, it had been widely accepted that about 30% of the annual release was dissolved in the ocean and the remainder was assimilated by green plants (e.g., Pearman and Garratt, 1972; Machta, 1972a, 1973; Keeling, 1973a). However, several studies by geophysicists and biologists in the past two or three years implied that the terrestrial biosphere is probably not a net sink for the excess CO_2 and may have functioned, in the past hundred years, as a significant source of atmospheric carbon dioxide (Woodwell and Houghton, 1977; Adams et al., 1977; Bolin, 1977a). The oceanographers, however, maintain that the ocean alone could not accommodate the amount of the excess CO_2 not remaining airborne (Fairhall, 1973; Whitfield, 1974). Increasing importance and controversy have thus arisen concerning the possibilities that trees and humus have been net sources or net sinks for atmospheric CO_2 (e.g., Olson et al., 1978; Loomis, in press; Woodwell, 1978).

Recent estimates of the net release of nonfossil carbon from oxidation of organic matter (living and dead) range from 1 to 8 Gtons of carbon annually (Hutchinson, 1954; Adams et al., 1977; Bolin, 1977a; Woodwell and Houghton, 1977; Woodwell et al., 1978; Wong, 1978; Baes et al., 1976, 1977). The uncertainty of these estimates is generally

high, as the calculations are based on limited data gathered from small geographical areas extrapolated to a global scale. Measurements of the $^{13}C/^{12}C$ ratio in tree rings formed between 1850 and 1950 infer a cumulative release of 120 Gtons of nonfossil carbon during this period (Stuiver, 1978), with considerable release before 1900 (Stuiver, 1978; Wilson, 1978).

Deforestation of the lands is suggested as the main source of nonfossil carbon in the atmosphere. In the boreal and north temperate regions, the total area of forested lands is sometimes assumed to be roughly stable for the last hundred years (Rostlund, 1956). However, in tropical, subtropical, and south temperate forests, clearing definitely prevails over reforestation and natural succession (Persson, 1974; Sommer, 1976). Yet another source of nonfossil carbon is the soil humus which may contain from 1500 to 3000 Gtons of carbon (Bohn, 1976; Schlesinger, 1977). The losses to the atmosphere from soil carbon have been enhanced because of the increased oxidation from swamp drainage, field plowing, and disturbance after logging operations. However, the long-term trend of carbon loss from the soil could have been reversed through humus-building agricultural practices in some well-managed, fertile areas (Loomis, in press). The possibility of net CO_2 loss from some areas and net storage in others leaves the global balance in doubt at this time.

If the biosphere is not providing storage for roughly 20% of the excess CO_2, another sink is implied. Marine chemists dispute the assertion that the excess CO_2 not remaining airborne could be absorbed

by the ocean. Both physical circulation and the buffering action of ocean water are limiting the transfer of atmospheric CO_2 to the deep ocean (Broecker, 1974; Skirrow, 1975).

Constraints of the buffering effects of sea water reduce the uptake of CO_2 in the mixed surface layer (Pytkowicz, 1967). The thermocline acts as a barrier between the mixed layer and the deep water where steep gradients of temperature and salinity slow mixing processes by diffusion and overturn. Furthermore, the deep water circulates very slowly by advection from the North Atlantic to the Pacific, and hence, hinders the penetration of carbonate ions and any shift in the mean concentration. Particulate organic matter settles slowly by gravitation.

If the anthropogenic emissions of fossil plus nonfossil carbon continue unabated at the present rate or faster, chemical buffering of the surface water may cause the ocean to be even less effective in absorbing the excess CO_2 in the future than during the past century (Keeling, 1973a; Revelle and Munk, 1977). The future biosphere does not seem likely to have the capacity to store the large amount of excess CO_2. A finer discrimination of the biospheric submodel in a global carbon model is essential in clarifying the confusion on the role of terrestrial ecosystems as sources or sinks for the excess CO_2 released from human activities.

1.4. SCOPE OF THIS STUDY

The present study extends the previous computer simulation studies of the global carbon cycle. Research activities are concerned mainly

with placing bounds on the pool sizes, exchange rates, and net shifts of terrestrial organic carbon in the global system. This objective is to help answer the questions: Could the biosphere account for a significant fraction of the CO_2 released to the atmosphere, or of the total uptake of excess carbon from the atmosphere? What possible limits of biospheric exchanges and net shifts of atmospheric carbon can be inferred for the past and for the future?

In focusing on the biospheric component of the carbon cycle, it is helpful to divide the terrestrial biota into broad subdivisions. Here the terrestrial ecosystems are grouped into three categories as in Olson et al. (1978). Forests and woodlands north of 30° N latitude are called "Northern Woods"; tropical, subtropical, and other temperate forests south of 30° N, "Southern Woods"; and all other nonforest ecosystems such as grasslands (including open savanna), croplands, deserts, and tundra, the "Nonwoods." Organic carbon in each zone is allocated to a rapidly exchanging pool or a slowly exchanging pool, according to data and judgment on its average turnover time.

The changing land-use pattern is an important factor affecting the global carbon cycle. The present study also explores explicitly the formulation of estimating the quantities of biospheric carbon from preliminary projections of area and mass per unit area for the broad ecosystem groups. Empirical equations also are developed to restrain the net primary production (NPP) and storage capacity of the biota at high level of atmospheric CO_2 and high value of mass per unit area of each ecosystem group.

Current aggregative approaches in modeling the global carbon cycle lack the sensitivity to resolve the precise role of these biospheric reservoirs. Most of the existing carbon models are of lower order systems described by less than ten differential equations for the whole earth. A slightly more complex model, incorporating as much available information as possible, is developed and tested here. This allows effective modification of the model and also precise interpretation of the simulation results, but avoids some of the unmanageability of large models. The biospheric submodel with six pools turns out to be informative at the present stage, even if more or fewer pools are suitable for other purposes.

CHAPTER 2

LITERATURE REVIEW

2.1. THE GLOBAL CARBON CYCLE

The dynamics of the global carbon cycle are dependent on three interconnected natural cycles superimposed with an economic or anthropogenic cycle (Borchert, 1951; Junge, 1963, 1977; Martin, 1970). Two of the natural cycles, the organic matter cycle and the geologic cycle, are regarded as maintaining a steady state over geologic time, while the other biological cycle, termed the "small cycle" by Borchert (1951), has a turnover time of about 40 years (Schlesinger, 1977). The latter is not completely balanced, because a small amount of carbon is continuously transferred to the slower organic matter cycle (Junge, 1977). The economic cycle is only a transient disturbance resulting from human activities (Martin, 1970).

2.1.1. Historical Background

The importance of carbon compounds for life aroused early interest in the cycle of carbon on earth. More than a century ago, the cyclical character of the circulation of carbon between plants and animals through a common atmospheric reservoir was lucidly demonstrated by Dumas (1841). The general concept set forth in that paper has become common knowledge.

One of the earliest efforts to quantify the global carbon cycle was that of Lotka (1924). In a discussion of the various processes related

to the carbon cycle in nature, he partitioned the carbon atoms into several interconnected reservoirs: the atmosphere, lithosphere, ocean, animal, plant, and fossil fuel (coal) pools. Estimates of reservoir size and magnitude of some of the processes such as weathering of rock and combustion of fossil fuels were given in his paper. However, reservoir sizes for the animal and plant pools were not given. Also missing were the fluxes of carbon between the atmosphere and the other three reservoirs (ocean, animals, and plants).

In a review of the global carbon cycle, Bolin (1970) attempted to synthesize the previous findings of carbon circulation in nature, especially in the biosphere. A compartment diagram of the global carbon cycle was presented together with the values for inventories and transfers of the reservoirs. An improvement of this diagram with extensive modification was summarized by the participants of the 24th Brookhaven Symposium on Biology (Reiners et al., 1973).

2.1.2. The Biota and the Carbon Cycle

The contribution of terrestrial ecosystems to modifying or stabilizing the carbon cycle has probably been underestimated in many earlier attempts to understand the kinetics of carbon in nature (cf. Kira et al., 1973). This deficiency was pointed out by Olson (1970a), who also stressed the importance of the forests (the main pool of terrestrial organic carbon). The hierarchical nature of the role of the forest subsystems within the global carbon cycle also was discussed in the same work.

Recently, Botkin (1977) briefly reviewed the effects of CO_2 enrichment on terrestrial and fresh water ecosystems. He considered that CO_2 enrichment may increase the carbon mass in the forests, but the interactions of tree species and the aftermath of increasing deforestation probably will cancel these effects.

2.2. QUANTITATIVE MODELS OF GLOBAL CARBON CYCLE

2.2.1. Introductory Remarks

The quantitative aspects of the global carbon cycle were thoroughly summarized by Bolin (1975) and Baes et al. (1976, 1977). These workers, with different emphases, succinctly reviewed the physical and chemical processes of relevance to the carbon cycle, and also the characteristic features of the ocean circulation. In addition, up-to-date estimates of the reservoir size and magnitudes of the interconnecting fluxes between the reservoirs were also evaluated.

Quantitative models of the global carbon cycle developed so far can be grouped into two categories: the thermodynamic sedimentary model and the kinetic compartmental or box model (Garrels and Lerman, 1977; Mackenzie and Wollast, 1977a). The term "compartment" is used interchangeably with "reservoir" or "box" in most studies of biogeochemical cycles (e.g., Svensson and Söderlund, 1976; Stumm, 1977). In the compartmental model, the elemental carbon is partitioned into a number of physical entities, the compartments or the reservoirs, with arbitrary or distinguishable boundaries interconnected by fluxes of carbon

between adjacent reservoirs. The number of reservoirs chosen for such a model depends both on the previous knowledge of the way carbon is distributed about the earth's surface and on the geographical scale of the transfer under consideration (Mackenzie and Wollast, 1977a; Eriksson and Rosswall, 1976). The concentration and distribution of carbon within a particular reservoir, and to some extent the rates of transfer, are governed by interacting physical, chemical, and biological processes. The resultant behavior of carbon in the cycle is generally described in terms of first-order differential equations (Eriksson, 1971; Mackenzie and Wollast, 1977b).

The mass balance of the sedimentary or rock cycle is based entirely on chemical equilibrium reactions in most of the thermodynamic models (Li, 1972; Garrels and Mackenzie, 1972; Mackenzie, 1975; Mackenzie and Garrels, 1966). Implicitly, these models assume that the distribution of various elements (including carbon) in the earth's crust are the end results of chemical equilibria and that the chemical reactions obey the laws of thermodynamics. Nevertheless, they are inadequate for complete description of the carbon cycle, particularly for the transient behavior, and are usually insufficient as a predictive tool (Mackenzie and Wollast, 1977b; Broecker, 1971).

2.2.2. Compartmental Models and Radiocarbon

A variety of simple compartmental models have been studied in connection with the transfers and steady-state distributions of ^{14}C in the atmosphere and in the oceans. Several workers in the latter part

of the 1950s (e.g., Craig, 1957, 1958, 1963; Revelle and Suess, 1957; Broecker, 1963) took advantage of the naturally occurring ^{14}C as a tracer to obtain the steady-state mean residence times of carbon in the sea and the atmosphere. Most of these earlier models were fairly simple, composed of a two-layer ocean, an atmospheric compartment and, sometimes, an additional biospheric compartment. However, the basic computational procedure for obtaining residence times of carbon in the various compartments later was used for other more complex models suggested by Broecker (1963), and Miyake and Saruhashi (1973).

The detonation of nuclear devices usually produces a significant amount of ^{14}C to the atmosphere. Since 1954, aboveground nuclear tests have been carried out in northern latitudes. The bomb-produced ^{14}C is being redistributed in the carbon cycle. Actual measurements of the excess ^{14}C, which can be considered as individual "spikes," allow the response of the oceans to be revealed (Junge, 1963; Fairhall et al., 1973; Keeling and Bacastow, 1977). Nydal and his co-workers (Nydal, 1967, 1968; Nydal and Lövseth, 1970; Gullikesen et al., 1973) also employed simple compartmental models to characterize the distribution of excess ^{14}C, and to fit statistically the oceanic ^{14}C data compiled from periodic actual measurements from the high-yield nuclear test series of 1957 and 1962.

The dilution of ^{14}C activity by the addition of fossil-fuel carbon, the Suess effect (Revelle and Suess, 1957) has generally been disregarded in previous models. In view of the increasing production of fossil fuel carbon, this dilution effect will become increasingly significant

(Baxter and Walton, 1970). Walton et al. (1970) took account of the magnitude of this effect in their estimates of the mean residence times of carbon reservoirs.

2.2.3. Dynamic Models of the Carbon Cycle

The carbon dioxide/climate problem has stimulated many studies in the past, especially during the last ten years, to obtain a better understanding of the circulation of carbon atoms in our environment. Results from these studies have provided mathematical models of the world carbon cycle that offer an initial framework for interfacing with climatic models. Most of these time-dependent models, with varying degrees of refinement, are able to integrate the improved information concerning the various components of the carbon cycle from previous static models. The latter assumed steady-state conditions to simplify estimation of the transfer rates and exchange time of the many fluxes. In a comprehensive study of a carbon cycling model, Keeling (1973a) has summarized and reviewed some of the early work. Several later models also have been reviewed by Bolin (1975).

2.2.3.1. Model Structures

Topologically, many of the carbon models are compartmental or box models. The carbon in the compartments is treated as if well-mixed or at least randomly mixed; the transfer processes are assumed to obey first-order rate equations. Nonlinearities in the model equations are usually associated with primary production of living plants and the

chemical dissociation equilibrium of carbonates in the ocean (cf. Keeling, 1973a).

Depending on the objective of the studies, the earlier models either neglected the biosphere altogether (Revelle and Suess, 1957), or assumed the ocean to be one well-mixed reservoir (Eriksson and Welander, 1956). A few workers (Cramer and Meyer, 1972; Gowdy et al., 1975), whose carbon models were based on Bolin's (1970) inventory, also put in compartments for the marine plants (phytoplankton) and animals (zooplankton). While Oeschger et al. (1975) and Siegenthaler and Oeschger (1978) modeled with only one biosphere compartment, others have expediently lumped the whole biosphere with the atmosphere (Plasset and Latter, 1960; Plasset and Dugas, 1967; Dugas, 1968) or with the mixed layer (Zimen and Altenhein, 1973b).

The various carbon compounds found in the biosphere decompose at different rates (Swift, 1977; Minderman, 1968). Considerable knowledge would be gained by subdividing the biospheric carbon pool. Eriksson and Welander (1956) divided the land biosphere into an "assimilating plant" and a much larger "dead organic matter" pool. This practice was followed by many workers, such as Bolin and Eriksson (1959), Bolin (1970), Young et al. (1972), Gowdy et al. (1975), Smil and Milton (1974), and Niehaus (1975, 1976). However, the residence time of a carbon atom in the terrestrial biosphere can be very different before the atom is returned to the atmosphere, depending on which pool it is in. This is also true for the dead organic matter or humus as it is oxidized to CO_2 at different rates by microorganisms (Minderman, 1968). It is reasonable

and advantageous to differentiate carbon pools of the land biosphere into various categories according to the residence time of the carbon fractions (cf. SCEP, 1970; Wagener and Förstell, 1972). An appropriate division is to have a rapidly exchanging carbon pool with a residence time of the order of a few years, and a slowly exchanging pool with a residence time of the order of decades to centuries. Models adopting these criteria of division were developed by Keeling (1973a), Bacastow and Keeling (1973), Ekdahl and Keeling (1973), Machta (1972a, 1973), Killough (1977), Olson and Killough (1977), Olson et al. (1978), Revelle and Munk (1977), and Zimen et al. (1977).

Bolin and Eriksson (1959) noted that the top 50 to 100 m of sea water is agitated by the wind such that it is relatively well mixed and rapidly equilibrated with atmospheric carbon dioxide. They accordingly divided the ocean compartment into a surface layer of 75 m depth exchanging with a much larger deep ocean layer beneath. Most of the later workers followed their formulation. Oeschger et al. (1975) made a significant improvement in representing the carbon movement in the ocean, replacing the 2- or 3-reservoir oceanic model with a "box-diffusion model" that allows carbon flux across a gradient beneath the mixed surface layer. This approach showed many promising results and has been readily adopted by many other workers (e.g., Jørgensen and Mejer, 1976; Killough, 1977; Olson and Killough, 1977; and Siegenthaler and Oeschger, 1978). A comparison of the box-diffusion model with the reservoir or compartmental model was analyzed thoroughly in an Appendix at the end of the report by Keeling and Bacastow (1977).

Other variables also can be included in the system of equations to characterize the initial states of the various compartments besides concentration and level of carbon in the model. These variables may include human population (Young et al., 1972), ambient temperature or the earth's heat balance (Jørgensen and Mejer, 1976; Niehaus, 1975, 1976; Mulholland et al., 1977), concentrations of nitrogen and phosphorus in the biota (Jørgensen and Mejer, 1976), or societal activities such as fossil-fuel production and consumption (Smil and Milton, 1974). Hoffert (1974) has made a refinement in the atmospheric subsystem by including latitudinal distribution and mixing of atmospheric carbon in his model.

2.2.3.2. Source Functions of Fossil Carbon

Reliable data on fossil-fuel production from 1860 A.D. to the present have been compiled and documented by the United Nations (Keeling, 1973b; Rotty, 1973, 1975, 1977). Based on these figures, together with the appropriate carbon content of each type of fuel, the annual amount of carbon released to the atmosphere from combustion of fossil fuels alone can be calculated (Revelle and Suess, 1957; Baxter and Walton, 1970; Keeling, 1973b; Rotty, 1977; Zimen et al., 1977; Perry and Landsberg, 1977). Similarly, CO_2 from kilning of limestone for cement also can be compiled. For at least the past 110 years, with the exception of the period between the two World Wars, the world's consumption of fossil fuels and cements has increased exponentially at about 4.3% per year. It is, therefore, conceivable to approximate

within this time period the input function of fossil-fuel carbon (with about 2% of the total CO_2 from production of cement) as an exponential function (e.g., Baxter and Walton, 1970; Keeling, 1973a; Ekdahl and Keeling, 1973; Bacastow and Keeling, 1973; Machta, 1972a, 1973). It is improbable that the exponential growth of fuel consumption will continue unabated until fuel supplies are completely exhausted. In order to represent more realistically the exhaustion of finite fossil-fuel resources, Dugas (1968) computed the consumption rate with a piecewise continuous function. Fuel consumption was allowed to continue exponentially until around 2000 A.D. and then to level off following a hyperbolic tangent function when fossil fuels become scarce.

The utilization rate of a nonrenewable resource is more appropriately illustrated by a sigmoid curve (Hubbert, 1971). Two different types of functions that can generate a sigmoid curve, a Gaussian normal curve and the family of logistic curves, have been used by various workers to model the annual injection of fossil-fuel carbon to the atmosphere. The normal curve (cf. Hoffert, 1974; Gowdy et al., 1975; Mulholland et al., 1977) is generally expressed as

$$\dot{P}(t) = P_\infty / \sqrt{2\pi \cdot \sigma \exp[-(t-t_m)^2/2\sigma^2]} \quad , \qquad (2.1)$$

where $\dot{P}(t)$ is the amount of fossil carbon released between time t and $t+\delta t$, and P_∞ is the total inventory of recoverable fossil fuels to be consumed.

The logistic function first introduced by Verhulst (cf. Lotka, 1924), can be generalized by the equation (Goel et al., 1971),

$$\dot{P}(t) = rP[1 - (P/P_\infty)^n], \quad n > 0 \quad . \tag{2.2}$$

The simple logistic function, with the exponential constant $n = 1$, was first adopted independently by Young et al. (1972) and Zimen and Altenhein (1973a,b) to describe the production of CO_2 from industrial activities. Similar functions, but with different values of the exponent (for $0 < n \leq 1$), were used as input functions in later modeling efforts (Killough, 1977; Olson and Killough, 1977; Olson et al., 1978; Keeling and Bacastow, 1977; Revelle and Munk, 1977; Zimen et al., 1977; Siegenthaler and Oeschger, 1978).

An input function which differs slightly from the logistic function was developed by Rotty (1976, 1977). The function

$$\dot{P}(t) = rP[1 - P/P_\infty]^\alpha, \quad \alpha \geq 1 \quad , \tag{2.3}$$

shows a different behavior from the other two sigmoid curves by being more asymmetrical. It, too, is equivalent to the simple logistic function if $\alpha = n = 1$.

2.2.3.3. Biospheric Fluxes

The assimilation rate (or primary production) of plants is dependent both on the concentration of the atmospheric carbon and on the biomass of the plants themselves. This is suggested in Eriksson and

Welander (1956) discussion of the behavior of the global carbon cycle in which the land biota is coupled to the atmosphere and the ocean, with the possibility of self-sustained oscillation of atmospheric carbon dioxide. However, plant respiration was considered to be a function of biomass only. This concept of donor and recipient controls in primary production of plants was also found in the models of later studies (e.g., Keeling, 1973a; Killough, 1977; Olson and Killough, 1977; Revelle and Munk, 1977; Olson et al., 1978).

The rate of transfer between living biomass and dead organic matter is probably not a function of the present level of biomass, but is rather a function of the biomass level some years earlier (Eriksson and Welander, 1956). Hence, time-delays were introduced in some models to characterize the responses of the biospheric compartments (Eriksson and Welander, 1956; Young et al., 1972).

The problem of modeling the behavior of land plants responding to an increase in atmospheric carbon dioxide was studied briefly by Bolin and Eriksson (1959). However, they still assumed that the primary production follows first-order rate equation depending only on the donor (i.e., the atmospheric) compartment. Moreover, time-delay was not incorporated in the system. The donor-controlled linear compartmental model of Gowdy et al. (1975) was another example in this regard.

Niehaus (1975, 1976) took a totally different approach to model the fertilization effect of increasing CO_2 on the net primary production (NPP) of the biosphere. He used an empirical Mitscherlich-type function to compute the relative growth rate as a function of normalized partial

pressure of atmospheric carbon dioxide. Parameters in this equation were derived from field experiments on several tree species as reported in the literature. This equation, as given by Niehaus (1976) is

$$y = y_{\infty}[1 - \exp(-k/2.466)] \quad , \tag{2.4}$$

where $y_{\infty} = 3$ is the specific asymptotic growth rate and k is the partial pressure of atmospheric CO_2 normalized at 320 ppmv. A 10% increase in atmospheric carbon up to five times the current concentration would cause approximately an 8% rise in the net production rate.

In many laboratory experiments, the organic carbon increases logarithmically with an increase in ambient carbon dioxide (Keeling, 1973a). Furthermore, a 10% increase in CO_2 results in up to about 5% more net production in plants. This potential growth rate would further be reduced, if limiting nutrients were taken into account. Consequently, in Keeling (1973a), the increase in net primary production (NPP) was described by a logarithmic function adjusted by a biotic growth factor, β. The latter factor ranged from 0 to 0.4. This approach to describing biospheric responses to additional CO_2 was used widely in subsequent models of the carbon cycle (e.g., Oeschger et al., 1975; Killough, 1977; Revelle and Munk, 1977; Zimen et al., 1977; Siegenthaler and Oeschger, 1978; Olson et al., 1978).

The biosphere may not serve as a major sink for the extra CO_2 injected into the atmosphere. Broecker et al. (1971) and Hoffert (1974) argued that even though the increasing CO_2 would enhance the photosynthetic rate to a certain extent in greenhouse experiments, the

availability of water and other necessary nutrients would eventually become limiting under field conditions (cf. Attiwill, 1971; Lemon, 1977). All of them preferred to keep the mass of the biosphere constant in their modeling studies.

Biospheric carbon can be assumed to increase initially and could remain constant or decrease after reaching a certain maximum level (Bacastow and Keeling, 1973; Killough, 1977). This has been accomplished by setting the biotic growth factor $\beta = 0$ at some specific time during the simulation (Bacastow and Keeling, 1973), or by multiplying the growth function by a limiting factor (e.g., Killough, 1977).

Several recent appraisals and reviews on the contribution of biospheric carbon to the atmosphere have provided a wide range of tentative values which underscore the necessity of further research. In their review, Baes et al. (1976) assumed that there was little net change of biospheric carbon in the Northern Woods and that the annual rate of cutting the Southern Woods might approach 1%/year. At this rate, their preliminary estimates of biospheric carbon releases from the Southern Woods were 1.2 Gtons/year from prompt releases and about 2 Gtons/year from delayed releases.

Adams et al. (1977) estimated the probable maximum cutting rate of forests to be 1 ton per capita per year (or about 4 Gtons/year currently). No allowance for the accelerated decay of soil humus and organic matter was included. Using a slightly different approach, Bolin (1977a) estimated that in the early 1970s, about 1 Gton of carbon per year was released directly to the atmosphere from the tropical forests assuming

also little net change in the temperate forests. Based on the rates of harvests reported in literature, Woodwell et al. (1978) judged that the most probable range for the total world release from the biota annually is 4 to 8 Gtons/year of carbon. The maximum value might reach 18 Gtons/year. Another estimate was given in the recent comprehensive review of Wong (1978). He provided gross and net estimates of carbon releases from the various sources such as forest clearings, wood burning, desertification, etc. His estimate for the net input of nonfossil carbon was about 1.6 Gtons/year of which 1.5 Gtons/year was assumed to come from the clearing of new tropical forests. The latter figure was taken as an intermediate value from the estimates of Baes et al. (1976) given above.

An early estimate of carbon released from biomass burning was given as 0.1 Gtons/year by Robinson and Robbins (1972). Much of their estimate of biomass burning as well as the corresponding estimates of Adams et al. (1977), and Wong (1978) were probably allowed for in the 7 Gtons/year of forest burning in Baes et al. (1976, 1977). Except for the estimates given by Baes et al. (1976), most of the other estimates did not give explicitly the proportion of carbon released directly to the atmosphere or the amounts of carbon shifted to the rapidly and slowly exchanging carbon pools. The delayed releases of live organic matter and soil humus would significantly increase the total release in the near future.

2.2.3.4. Deforestation Rates

It is necessary to incorporate the increase of deforestation explicitly in a new carbon model in order to visualize the effects of diminishing biospheric storage on the global carbon cycle. In an analysis of man's impact on the global carbon cycle, Young et al. (1972) first introduced a variable reflecting the carbon stored in forest cuttings. They defined its rate of change as the per capita consumption rate of wood times the world population level.

The effects of deforestation on the carbon cycle have been modeled by shifting a fixed portion (1% or less annually) of the carbon stored in tropical forests to the nonforest compartments (Olson et al., 1978). These initial findings suggested that an annual rate of deforestation of tropical forests of 1% since 1860 would drastically reduce the carbon stored in the world's forests in the near future. Concomitantly, an unrealistically high amount of carbon would remain in the atmosphere.

Deforestation rates are probably varying with other conditions. Revelle and Munk (1977) conceived the present remaining forests as nonrenewable resources (cf. Gómez-Pompa et al., 1972) and described the release of carbon from large-scale deforestation by another logistic equation. An alternative approach was taken by Zimen et al. (1977) in their recent paper. They viewed the forest as another energy source supplementing the usual fossil fuels, and so the release of carbon from forest was considered to vary proportionally with the carbon input from combustion of fossil fuels.

2.2.3.5. Sea Water Buffering Factor

The buffering action of carbonate ions in sea water plays a significant role in the ocean's uptake of atmospheric CO_2. Carbon dioxide is in a dissociative equilibrium with the inorganic carbon species HCO_3^- and CO_3^{--} in sea water (Pytkowicz, 1972; Broecker, 1974; Skirrow, 1975). If the atmospheric carbon dioxide concentration increases by X%, the resulting relative increase of oceanic carbon dioxide and inorganic carbon ions in equilibrium will be approximately X/ζ% (Keeling, 1973a). Here the buffering factor, ζ, has been used as in previous studies to relate the evasion potential of an infinitesimal change in sea-surface partial pressure owing to an infinitesimal transfer of CO_2 across the air-sea interface. Bolin and Eriksson (1959), in explaining the chemistry of the dissociative equilibria of the carbonate-bicarbonate system in sea water, suggested a buffering factor of 12.5. This value is slightly overestimated because the influence of borate in sea water is neglected (cf. Keeling, 1973a).

A comprehensive exposition of carbon chemistry in the sea was given by Keeling (1973a) and summarized by Baes et al. (1976, 1977). Keeling (1973a) and Plass (1972a) also dealt with the variation of the buffering factor under various conditions of pH and CO_2 partial pressure. These relations were expressed by equations that are cumbersome for numerical computation. Algorithms for obtaining the incremental form of the buffering factor (or evasion effect) from pH and CO_2 concentration were coded independently by Bacastow and Keeling (1973), Jørgenson and

Mejer (1976), and Killough (1977). Their calculation gave the valuc of preindustrial buffering factor at around 9 to 10. A constant value of 10 was assumed by Oeschger et al. (1975) in their box-diffusion model.

To simplify further the computation of the buffering factor, separate empirical formulas were derived from the relative change of atmospheric or sea surface partial pressure of CO_2 to obtain the corresponding value of buffering factor:

(i) Revelle and Munk (1977),

$$\zeta = 9 + 4(X - X_0)/X_0 \quad , \tag{2.5}$$

where X is the concentration of atmospheric CO_2 and X_0 is the preindustrial concentration of atmospheric CO_2.

(ii) Zimen et al. (1977),

$$\zeta = -0.122p^2 + 5.36p + 3.6 \quad , \tag{2.6}$$

where p is the partial pressure of CO_2 in the surface water relative to the preindustrial value.

(iii) Siegenthaler and Oeschger (1978),

$$\zeta = 9 + 4.9p - 0.1p^2 \quad , \tag{2.7}$$

where p is the same as above.

Eq. (2.5) is used in Chapter 3 with X and X_0 replaced by c_a and c_a^*, respectively. Theoretically, ζ increases with increasing temperature and increasing inorganic carbon in the ocean (Keeling, 1973a; Jørgensen and Mejer, 1976). This means that more carbon dioxide will

be released to the atmosphere as the oceanic carbon increases. This temperature refinement is not included in the present model.

CHAPTER 3

MODEL DEVELOPMENT

The global carbon model developed in the present study is basically a deterministic compartmental model. As defined by Eriksson (1971), a compartment is any homogeneous part of nature with clearly defined but arbitrarily chosen boundaries. The compartmental model is designed to describe the time development of average properties of the compartments, in particular where a detailed description of the processes in the compartments is too complex to formulate (Eriksson, 1971; Bolin and Rodhe, 1973).

The basis of the present compartmental model rests on the hypothesis of mass balance which provides a dynamic description of carbon movement in the environment. The calculation of steady-state mass balance involves equating the input of carbon to a particular compartment with the fluxes out of that compartment. For linear systems, first-order rate equations are assumed in all fluxes. However, nonlinearities can be incorporated in the model if the kinetics of the transfer processes are better described by nonlinear functions, i.e., by flows which are not directly proportional to the amounts in the source compartment.

3.1. QUANTITIES OF CARBON

Table 1 lists the symbols and initial values of the 25 state variables used in modeling the global carbon cycle. Inventories of

Table 1. Estimated Value of State Variables at Steady State (1860 A.D.)

Symbol	State Variable	Value
c_1	troposphere	508 Gtons
c_2	stratosphere	90
c_3	aquatic live organisms	1.5
c_4	dead organisms and organic matter in mixed layer	29
c_5	dead organisms and organic matter in thermocline and deep ocean	1,620
c_6	dissolved and miscellaneous carbon on land	592
c_7	mixed surface layer	650
c_8	thermocline	6,600
c_9	deep ocean	31,820
c_{10}	Nonwoods (rapidly exchanging carbon)	34
c_{11}	Nonwoods (slowly exchanging carbon)	490
c_{12}	Northern Woods (rapidly exchanging carbon)	65
c_{13}	Northern Woods (slowly exchanging carbon)	560
c_{14}	Southern Woods (rapidly exchanging carbon)	48
c_{15}	Southern Woods (slowly exchanging carbon)	580
c_{16}	detritus carbonates and calcareous tests	1,290
c_{17}	sedimentary carbonates	30,000,000
c_{18}	deep humus and reactive organic sediment	1,000
c_{19}	lithified organic sediment	6,600,000
q_{10}	mass/area of compartment c_{10}	0.46 kg/m^2
q_{11}	mass/area of compartment c_{11}	6.62
q_{12}	mass/area of compartment c_{12}	2.24
q_{13}	mass/area of compartment c_{13}	19.31
q_{14}	mass/area of compartment c_{14}	1.72
q_{15}	mass/area of compartment c_{15}	20.0

carbon in the 19 compartments are partially based on the reviews of Baes et al. (1976, 1977) and Pytkowicz (1973). The values represent the conditions which existed before the large-scale anthropogenic emissions of fossil carbon dioxide. A block diagram of the carbon model is presented in Figure 2. The number of significant figures is necessary for mass-balance purposes but does not reflect the accuracy of the data and numerous assumptions which have been reviewed.

3.1.1. Atmospheric Compartments

In most previous carbon models, the preindustrial concentration of atmospheric carbon dioxide was assumed to be 290 ppmv (615 Gtons C). This value seemed compatible with 293 ppmv as found by Bray (1956) for the mean of reported observations somewhat later—between 1857 to 1906. The increase of about 20 ppmv of atmospheric carbon dioxide due to deforestation and industrialization during this period (Stuiver, 1978; Wilson, 1978) has not been taken into account until very recently.

An intermediate value of 280 ppmv, equivalent to 598 Gtons of carbon, was taken as the 1860 concentration of atmospheric carbon dioxide. According to Machta (1973), the troposphere (c_1) contains about 85% of the atmospheric carbon and the remaining 15% is in the stratosphere (c_2). This ratio also was adopted in this study in partitioning the atmospheric carbon pool into the troposphere and the stratosphere compartments.

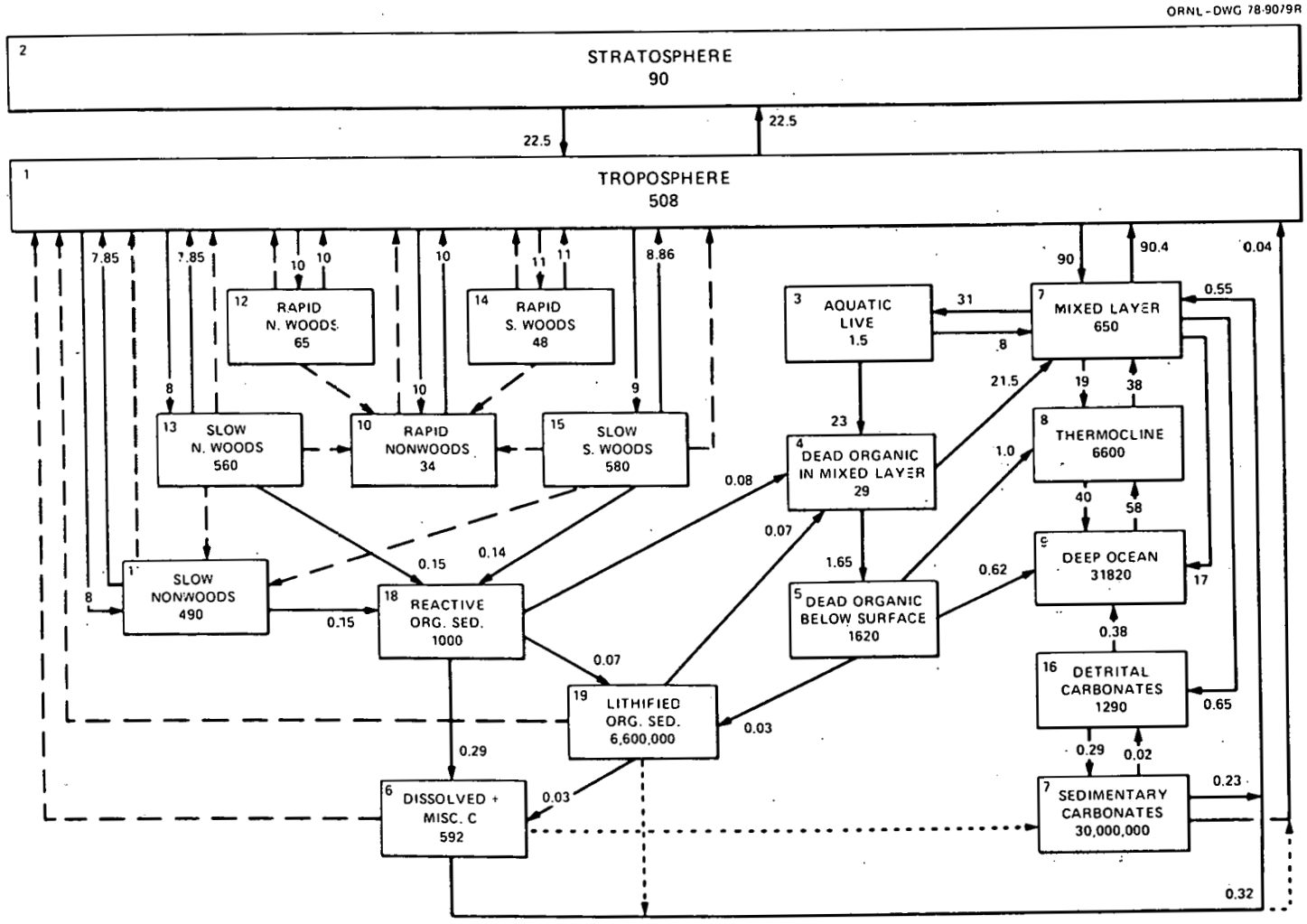

Figure 2. The 1860 A.D. global carbon cycle. Numbers in boxes are pool sizes in unit of Gton ($=10^{15}$g). Solid arrows are natural fluxes in unit of Gton/year. Broken arrows are anthropogenic fluxes. Dotted arrows are natural fluxes whose values are assumed to be negligible.

3.1.2. Marine Compartments

For modeling purposes, the ocean submodel was divided into three parts—the mixed surface layer (c_7), the intermediate waters (including the main thermocline where it exists, c_8), and the deep ocean (c_9) (cf. Baes et al., 1976, 1977; Broecker et al., 1971; Miyake and Saruhasi, 1973). The preindustrial partial pressure of inorganic carbon in the mixed layer is nearly in equilibrium with atmospheric carbon dioxide. Following the procedure of Killough (1977), the initial amount of inorganic carbon in the mixed layer was recalculated to cover the whole ocean area of 361×10^6 km^2. An average pH of 8.275 (Keeling, 1973a) was used in the calculation. A value of 650 Gtons of carbon was obtained (Table 1). Refinements, e.g., separating warm and cold surface waters, are possible but are merely reflected in some flux estimates in this report.

Inorganic carbon pools in the thermocline and the deep ocean followed the estimates given by Baes et al. (1976) taking account of the smaller projected area of the deeper ocean layers.

Another pool of inorganic carbon scattering in the ocean water is the calcareous test (shells) and detrital carbonates (c_{16}). This material is formed by sea plankton in the surface waters, and some of it survives dissolution in cold deep waters to be deposited on part of the sea floor of less than 4 km depth (Broecker, 1974). A rough estimate of 1290 Gtons was given by Pytkowicz (1973) for that part of the settled detrital carbonates which are in exchange with sea water. This amount is about three times the unpublished estimate of W. S. Broecker (see

Baes et al., 1976; Olson et al., 1978). At least part of the submarine and lithospheric sediment (c_{17}), which contains an amount of 30×10^6 Gtons, is eventually returned to the weathering environments on the continents.

The total living biomass (c_3) in the sea is estimated to contain about 1.5 Gtons of carbon. This is close to the estimate of 1.8 Gtons by Whittaker (1975). The other pool of dead organic matter is about three orders of magnitude larger than the living biomass. Both dissolved organic matter and particulate organic matter are present in sea water. Baes et al. (1976) partitioned the material into two compartments—29 Gtons in the mixed layer (c_4) and 1620 Gtons in the lower waters (c_5).

3.1.3. Terrestrial Ecosystems

Pool sizes of live and detrital organic carbon in the terrestrial ecosystems were reviewed by Olson et al. (1978), Baes et al. (1976, 1977) and Schlesinger (1977). Recent estimates of live organic carbon for the biosphere range from 557 to 1080 Gtons. The highest figure could well approximate the pattern of global carbon distribution at the preagricultural or prelogging state (Rodin et al., 1975). A cluster of intermediate values of about 815 ± 20 Gtons of carbon were obtained by Reiners et al. (1973), Whittaker and Likens (1973, 1975) and Ryabchikov (1975). The lowest estimate given by Olson et al. (1978) is used in this model for the year 1970.

The SCEP (1970) report gave an estimate of 1580 Gtons as the amount of organic carbon circulating actively within the terrestrial ecosystems.

This value is slightly larger than the value of 1533 Gtons summarized by Reiners et al. (1973). Baes et al. (1976, 1977) and Olson et al. (1978) readjusted these values to 1760 Gtons by recognizing a larger contribution from the dead organic matter (about 1200 Gtons). Their estimate represents the distribution of active organic carbon around the year 1970. It includes approximately 1600 Gtons of carbon with relatively rapid turnover rate and about 160 Gtons of carbon with slower turnover rate. Each of these carbon pools was further apportioned to the three biotic zones as given in Table 1 (page 33). Their pool sizes at 1970 also served as target values in this study to adjust the initial carbon pools of the six terrestrial biotic compartments, c_{10} to c_{15} (Table 1).

The initial slow carbon pools add up to 1630 Gtons comprising about 1×10^3 Gtons of the partially decomposed litter and soil humus from the upper part of the soil horizon. There is another pool of approximately 1×10^3 Gtons of carbon from buried peat and deep humus (c_{18}) which oxidizes slowly and has a still slower turnover rate (residence time about 1×10^3 years or longer) than the three pools of slowly exchanging carbon. Altogether, the amount of humus and dead organic material approaches 2×10^3 Gtons, and is within the range of 1.5×10^3 to 3×10^3 Gtons estimated recently by Schlesinger (1977) and Bohn (1977), respectively. The carbon dioxide produced from the slow oxidization of this reactive deep humus is likely to dissolve in ground waters. Pytkowicz (1973) gave an estimate of 592 Gtons for this pool. This estimate seems very high, unless it includes a store of other carbon sources such as the carbonates or caliche found in many of the

alkaline soils. This pool (c_6) may also include a relatively small amount of inorganic carbon which is dissolved in inland waters. Further clarification of this estimate is still desirable. Estimates of lithified carbon in the sedimentary rocks (c_{19}) vary widely. A value of 6.6×10^6 Gtons C was used in this model (cf. Baes et al., 1976, 1977). This source of uncertainty has essentially no effect on the results studied in the present report.

3.2. CIRCULATION OF CARBON

The symbol and definition of the fluxes between adjacent compartments together with their initial values at steady state (1860 A.D.) are given in Table 2. Many of the values are taken from Baes et al. (1976), Garrels et al. (1975, 1976), and Pytkowicz (1973). Some of the fluxes are estimated from other related sources; a few are computed by mass-balancing the total influxes and outfluxes of the adjoining compartments in the model.

3.2.1. Atmospheric Fluxes

The residence time of carbon dioxide in the stratosphere is about four years (Walton et al., 1970). This value is used in the present study although a shorter time of about two years is used in the models developed by Machta (1973) and Keeling and Bacastow (1977). Assuming that the amount of carbon transferred from the stratosphere to the troposphere balances the flux in the opposite direction, then $F_{1,2} = F_{2,1} = 22.5$ Gtons/year.

Table 2. Assumed Value of Carbon Fluxes at Steady State (1860 A.D.)

Flux	Value (Gton/year)	Process
$F_{1,2}$	22.5	Transfer of inorganic C from troposphere to stratosphere
$F_{2,1}$	22.5	Transfer of inorganic C from stratosphere to troposphere
$F_{1,7}$	90	Absorption of C from troposphere to mixed layer
$F_{7,1}$	90.4	Evasion of C from mixed layer to troposphere
$F_{1,10}$	10	NPP of Nonwoods (rapidly exchanging C)
$F_{1,11}$	8	NPP of Nonwoods (slowly exchanging C)
$F_{1,12}$	10	NPP of Northern Woods (rapidly exchanging C)
$F_{1,13}$	8	NPP of Northern Woods (slowly exchanging C)
$F_{1,14}$	11	NPP of Southern Woods (rapidly exchanging C)
$F_{1,15}$	9	NPP of Southern Woods (slowly exchanging C)
$F_{3,4}$	23	Death of aquatic organisms
$F_{3,7}$	8	Respiration of aquatic animals
$F_{4,5}$	1.65	Gravitational sinking of organic detritus
$F_{4,7}$	21.5	Oxidative decay in mixed surface layer
$F_{5,8}$	1	Oxidative decay in thermocline
$F_{5,9}$	0.62	Oxidative decay in deep ocean

Table 2 (continued)

Flux	Value (Gton/year)	Process
$F_{5,19}$	0.03	Deposition of organic sediments from marine organic matter
$F_{6,7}$	0.32	Dissolved inorganic C transported by streams
$F_{7,3}$	31	NPP of aquatic plants and phytoplanktons
$F_{7,8}$	19	Transfer of inorganic C from mixed layer to thermocline
$F_{7,9}$	17	Transfer of inorganic C from mixed layer of polar seas to deep ocean
$F_{7,16}$	0.65	Formation of calcareous tests
$F_{8,7}$	38	Transfer of inorganic C from thermocline to mixed layer
$F_{8,9}$	40	Transfer of inorganic C from thermocline to deep ocean
$F_{9,8}$	58	Transfer of inorganic C from deep ocean to thermocline
$F_{10,1}$	10	Oxidative decay of rapidly exchanging C in Nonwoods
$F_{11,1}$	7.85	Oxidative decay of slowly exchanging C in Nonwoods
$F_{12,1}$	10	Oxidative decay of rapidly exchanging C in Northern Woods
$F_{13,1}$	7.85	Oxidative decay of slowly exchanging C in Northern Woods
$F_{14,1}$	11	Oxidative decay of rapidly exchanging C in Southern Woods
$F_{15,1}$	8.86	Oxidative decay of slowly exchanging C in Southern Woods
$F_{11,18}$	0.15	Deposition of organic matter from Nonwoods

Table 2 (continued)

Flux	Value (Gton/year)	Process
$F_{13,18}$	0.15	Deposition of organic matter from Northern Woods
$F_{15,18}$	0.14	Deposition of organic matter from Southern Woods
$F_{16,9}$	0.38	Dissolution of calcareous tests and detrital carbonates
$F_{16,17}$	0.29	Sedimentation of inorganic C
$F_{17,1}$	0.04	Metamorphism of sedimentary rocks and volcanism
$F_{17,7}$	0.23	Weathering of carbonates
$F_{17,16}$	0.02	Physical weathering of sedimentary rocks
$F_{18,4}$	0.08	Organic C transported by streams
$F_{18,6}$	0.29	Oxidation of deep humus
$F_{18,19}$	0.07	Lithification of organic matter
$F_{19,4}$	0.07	Organic C transported by streams
$F_{19,6}$	0.03	Oxidation of old organics

About 60 to 100 Gtons of carbon enters the ocean annually (Pytkowicz, 1973; Woodwell and Pecan, 1973). The figure given by Baes et al. (1976), i.e., $F_{1,7}$ = 90 Gtons/year, is retained here as the flux of carbon from the troposphere to the surface ocean. This implies a residence time of carbon transferring from the atmosphere to the sea of about 6.6 years. On land, the vegetation assimilates about 56 Gtons of carbon as annual net primary production (SCEP, 1970; Lieth, 1975). This amount is partitioned among the six biotic compartments as shown in Figure 2 (page 35). Gross primary production is about twice this rate of income, on average; respiration of the autotrophic plants is not treated here, because it is so rapid and has little effect on storage (Olson et al., 1978).

3.2.2. Marine Fluxes

Recent estimates of marine primary production approach 31 Gtons/year (Platt and Subba Rao, 1975; Fogg, 1977). This value is adopted here as the net primary production of marine ecosystems. Although many of the observations from short-term ^{14}C feeding experiments lead to some intermediate value between net production and gross production (Bunt, 1975; Fogg, 1975), these overestimations are probably offset by not including the benthic production, loss of dissolved organic matter from cells and also photosynthesis at low light intensity (Bunt, 1975).

About 36% (perhaps more) of the marine primary production is consumed by marine animals (Whittaker, 1975). Of this amount, about 70 to 80% returns to the water from animal respiration and egestion

(Tait, 1971). Thus, the flux $F_{3,7}$ = 31 Gtons/year × 0.36 × 0.75 = 8 Gtons/year. On balance, the living aquatic organisms contribute 23 Gtons of organic carbon annually to the dissolved and particulate organic carbon pools in the mixed layer. Most of it oxidized there. A small fraction of the degraded organic matter (about 1.65 Gtons/year) is being oxidized as it settles slowly to the deep ocean (cf. Broecker, 1974).

Pytkowicz (1973) suggested that nearly 98% of the annual organic influx is oxidized in the near surface waters. A value of 21.5 Gtons/year is assigned to oxidization in the mixed layer, leaving the remainder of 1 Gton/year to the thermocline. Moreover, about 0.03 Gtons of the organic carbon is deposited annually in the sediment (Garrels et al., 1975). This leaves the balance of 0.62 Gtons/year as the flux of carbon representing the oxidization of organic matter in the deep ocean.

Taking the residence time of the subsurface layers as 1000 years (Broecker, 1963; Stuiver, 1973), the upwelling rate of inorganic carbon to the mixed layer is calculated as equal to 38 Gtons/year. This is balanced by a slightly smaller downward advection of inorganic carbon, i.e., $F_{7,8} + F_{7,9}$ = 36 Gtons/year. The flux $F_{7,9}$ is a surrogate for the sinking of cold polar water to the deep layers of the warm oceans (i.e., the deep Atlantic, Pacific, and Indian Oceans). In estimating the magnitude of this particular flux, it is assumed that the transfer of carbon from the mixed layer depends on the partial pressure of CO_2 in the water and the ratio of surface areas between the polar waters

and the warm oceans. Though the polar waters cover approximately one-sixth of the ocean surface, they absorb up to five times as much CO_2 per unit area as the warm surface waters (Broecker, 1963). Somewhat arbitrarily here, the polar waters are allowed to carry about 17 Gtons of carbon annually to the deep oceans, and the remaining flux of 19 Gtons/year is assumed to pass through the thermocline. Bolin (1977b) outlines other approaches which are beyond the scope of the present work. The transfer of inorganic carbon from the thermocline to the deep oceans is assumed to be 40 Gtons/year which is slightly greater than the upwelling rate from the thermocline (cf. Miyake and Saruhashi, 1973).

Pytkowicz (1973) estimated that 1.3 Gtons of the bicarbonate carbon is being utilized annually by marine organisms for test formation. In the course of forming carbonates from bicarbonates, one-half of the carbon content is released as CO_2 to the surrounding water so that only 0.65 Gtons ($F_{7,16}$) of carbon is consumed annually in the formation of calcareous tests (cf. Pytkowicz, 1973). As the solid tests and other detrital carbonates pass slowly through the unsaturated deep ocean, about 0.38 Gtons of carbonate carbon per year ($F_{16,9}$) react with an equal amount of CO_2 produced from the oxidization of organic matter and redissolve as bicarbonates (Pytkowicz, 1973). The deposition of carbonates to the sediment, $F_{16,17}$ is estimated by Pytkowicz (1973) as 0.29 Gtons/year. Concurrently, 0.02 Gtons of carbon ($F_{17,16}$) is returned annually to the ocean from physical weathering of sedimentary rocks (Pytkowicz, 1973).

Chemical weathering of sedimentary rocks by CO_2 dissolved in ground waters probably consumes about 0.32 Gtons of carbon annually ($F_{6,7}$) from the CO_2 alone, and at the same time, releases about 0.23 Gtons of carbon from sedimentary carbonates, ($F_{17,7}$). A total of 0.55 Gtons of inorganic carbon is transported annually to the ocean from runoff and river discharge (Pytkowicz, 1973).

It is assumed that an average of 0.04 Gtons of carbon ($F_{17,1}$) is released yearly to the atmosphere from metamorphism of sedimentary rocks and from volcanism originated in the deep lithosphere (Baes et al., 1976; Pytkowicz, 1973).

3.2.3. Terrestrial Fluxes

The net production of organic carbon for the six biospheric compartments follow the estimates given by Baes et al. (1976, 1977) and Olson et al. (1978). The total terrestrial NPP is about 56 Gtons/year. For the rapid carbon pools, an equal amount of organic carbon from oxidative decay is hypothesized to return to the atmosphere. However, a small fraction of the net production from the slow carbon pools is incorporated in the soil as slow-decaying humus.

The annual deposition of organic carbon to the sediment ($F_{18,19}$) is estimated to be about 0.07 Gtons/year (Garrels et al., 1976). At steady state, only 0.03 Gtons of carbon from the organic sediment is oxidized and dissolved in the ground waters annually ($F_{19,6}$) and thus, 0.29 Gtons of carbon per year ($F_{18,6}$) is required to balance the loss from weathering of sedimentary rocks (Garrels et al., 1976).

Another 0.15 Gtons of carbon as dissolved and particulate organic matter is transported annually to the ocean by rivers and streams (Garrels and Mackenzie, 1972; Mackenzie, 1975). It is assumed that the organic matter is derived from both the deep humus pool and the old organic sediment pool. From mass-balance calculations, $F_{19,4} = 0.07$ Gtons/year and so $F_{18,4} = 0.08$ Gtons/year. Since a total of 0.44 Gtons ($F_{18,4} + F_{18,6} + F_{18,19} = 0.08 + 0.29 + 0.07$) of organic carbon per year is lost from the reactive organic sediment pool (c_{18}), an influx of an equal amount of carbon from the slow carbon pools is required to balance the fluxes at steady state. For lack of appropriate data, an approximately equal contribution from each of the three slow carbon pools is assumed in this model. Thus, $F_{11,18} = 0.15$ Gtons/year, $F_{13,18} = 0.15$ Gtons/year and $F_{15,18} = 0.14$ Gtons/year.

Besides the anthropogenic fluxes discussed in Section 3.3.4, some of the carbon in the terrestrial compartments also may transfer to the atmosphere because of human activities. Such flows may be the result of accelerated oxidization of organic matter from the plowing of croplands or from the drainage of swamps and bogs. Not enough information is available to determine the magnitude of these fluxes but they are supposed to be relatively small. These fluxes are also drawn as broken arrows on the left of the block diagram (Figure 2, page 35) in addition to the broken arrows which show the shifting of carbon after land-clearing activities.

Several other natural fluxes are shown as dotted arrows in the block diagram (Figure 2). These fluxes represent the direct deposition

of carbonate caliche in alkaline soils and its reverse process releasing CO_2 to the atmosphere directly. The values of these fluxes are also assumed to be negligible.

3.3. MAJOR WORKING ASSUMPTIONS AND THEIR BACKGROUND

The working hypothesis extending throughout the present model is that the biospheric organic carbon could be both a source and sink for atmospheric CO_2—in different places at the same time. Both rapidly (τ < 10 years) and slowly ($\tau \simeq n \times 10$ years) exchanging pools from "Southern Woods" (mainly tropical and subtropical forests) and perhaps "Northern Woods" (forest occupied areas north of 30° North latitude) have been shifted partially to various Nonwoods ecosystems (including agricultural lands, grasslands, deserts, and fringe areas in urban and rural communities). Some extra CO_2 will be released to the atmosphere promptly (e.g., by burning, much more in recent years than in former centuries). More CO_2 will return to the atmosphere gradually as organic remains decay.

3.3.1. Assumption of Quasi-Steady State Before Large-Scale Industrialization

Over the last half-billion years, the ratios of carbon isotopes in marine carbonates have varied within a relatively narrow range. This strongly supports the postulate that the global carbon cycle has maintained a dynamic steady state for a long time (Broecker, 1970; Mackenzie, 1975; Schidlowski et al., 1975, Junge et al., 1975; Garrels et al., 1976). The steady state probably exhibited a damped oscillation

out of phase with the advances and retreats of the continental glaciers during the Pleistocene epoch. However, the emergence of human society has inadvertently perturbed this steady state by hastening the release of carbon dioxide more rapidly than the normal fluctuations through agriculture, forest clearing, and the combustion of fossil fuels.

Since early Neolithic time, forests have been cleared to provide land for the growing of crops (Olson et al., 1974). About 90% of the original forests in China and other European countries are gone (Sears, 1956; Darby, 1956; Young, 1976). The vast tropical forests are diminishing rapidly (Meijer, 1973; Goodland and Irwin, 1975; Hamilton, 1976; Brünig, 1977). This has created some unforeseen consequences. Soil carbon is diminishing from the constant plowing and accelerated oxidation. Wood is used as fuel to provide heat. From around the 10th century in China and possibly four or five hundred years later in the other European countries, forest areas were regressing rapidly. Coal was used increasingly to replace fuelwood and charcoal (Tuan, 1968, 1970; Nef, 1977). However, fossil fuels were not used extensively worldwide until a much later date—probably after the last half of the 19th century (Putnam, 1953). The increasing anthropogenic emission of CO_2 annually is now two to three magnitudes larger than the average annual degassing of CO_2 from the earth's mantle (cf. Baes et al., 1976). In actuality, it has already perturbed the global carbon cycle from its original steady state.

At present, there are no reliable data giving the amounts of carbon dioxide released before 1860 A.D. from anthropogenic production.

Estimates of annual production of fossil fuel and cement from 1860 onward are being compiled and updated by the United Nations; whereas, data on deforestation are still scarce and mostly unreliable. In view of this fact and the later adoption of part of Keeling's (1973b) compilation of annual production of industrial CO_2, the year 1860 is arbitrarily taken as the onset of the large-scale perturbation. Hence, all simulation runs start with t_0 = 1860. Much of the colonial clearing of tropical forests to be simulated in the present study also accelerated around that time (Sioli, 1973; Goodland and Irwin, 1975). Similar treatment of earlier land clearing will require longer simulations and better estimates than are now available.

3.3.2. Source Functions of Carbon Dioxide from Fossil Fuel and Cement

Keeling (1973b), Rotty (1973, 1975, 1977) and Zimen et al. (1977) have compiled similar estimates of the annual production of carbon dioxide from the combustion of fossil fuels and the kilning of cement covering the period from 1860 to 1974. These tabulated figures show a growth coefficient of about 4.35%/year for the periods 1860 to 1910 and 1945 to 1974 but with a disrupted growth in between. Hence, it is impractical to describe the source function for this period with a smooth curve. Instead, the actual historical data have been used as inputs in the model. From 1860 to 1949, the input rates of carbon dioxide to the atmosphere were interpolated linearly from the list compiled by Keeling (1973b). The annual inputs for the remaining years up to 1974 were interpolated from the values compiled by Rotty (1977).

No correction for oilhead flaring was applied to Keeling's (1973b) original figures, since the practice of flaring was almost negligible before 1950 (Rotty, 1974).

The extrapolation of the exponential increase of carbon dioxide production is not appropriate when modeling long-term projections. The future input of carbon dioxide to the atmosphere depends very much on the potential demand and production of fossil fuels, which in term are dictated by the ultimate supply of recoverable fossil-fuel resources (Hubbert, 1971, 1973). Thus, a logistic function or one of its modifications, is an adequate equation for projecting the future release of fossil-fuel carbon.

A modified logistic equation similar to the one introduced by Keeling and Bacastow (1977) was employed in this study to project the input rate of carbon dioxide from fossil fuel and cement beyond the year 1974. This equation is expressed as,

$$u(t) = \dot{P}(t) = rP[1 - (P/P_{\infty})^{n}] \quad , \tag{3.1}$$

where P_{∞} is the ultimate release of carbon in recoverable fossil fuels and cement,

P is a function of time and represents the cumulative amount of carbon released up to that time,

n is a variable quantity to place more or less emphasis on the release rate of the remaining fuels, and

r is the initial fractional rate of increase.

By integration (cf. Turner et al., 1976), the above equation gives the cumulative amount of carbon released at time t, i.e.,

$$P(t) = P_\infty/\{1 + [(P_\infty/P_0)^n - 1] \exp[-nr(t-t')\}^{1/n} \quad , \tag{3.2}$$

where P_0 is the cumulative production of carbon from 1860 to $t' = 1974$.

Resources of fossil fuels and shale oil are poorly known but may be assumed to contain 10^4 Gtons of carbon (Olson et al., 1978). However, only a fraction of these may be recoverable as "reserves" even taking into account the improved economic and technical conditions in the near future. Keeling and Bacastow (1977), Revelle and Munk (1977) and Zimen et al. (1977) used an estimate of about 5000 Gtons (cf. Perry and Landsberg, 1977) in each of their models. The value of 7500 Gtons used in the present simulation is essentially the same as the 7300 Gtons of Baes et al. (1976, p. 32). It may prove "high" but that is intentional for contrast with other scenarios discussed below.

The exponent n in Eq. (3.1) can be decreased to project a slower release rate and to allow the ultimate supply to taper off more slowly. To assure a relatively smooth fit to the historical data, the value of n was calculated with a predetermined growth rate (r = 4.35%/year) and the 1974 annual production $\dot{P}$ and cumulative production P_0. From the differential equation, n was obtained as

$$n = \log (1 - \dot{P}/P_0 r)/\log(P_0/P_\infty) \quad . \tag{3.3}$$

The percentage growth rate parameter, r, may be treated as a constant which can be calculated also from the differential equation

with a predetermined exponent n and the 1974 production estimates ($\dot{P}$ and P_0; equivalent to Killough, 1977, page 42), i.e.,

$$r = \dot{P} P_\infty^n / P_0 (P_\infty^n - P^n) \quad . \tag{3.4}$$

Alternatively the parameter r may be treated as nonconstant and may depend on the interactions of many social, economic, and political factors. Hence, r may be regarded as a slowly decreasing function depending on the cumulative production of carbon from fossil fuel. The rate might stabilize at a lower level, probably in the middle of the next century, if the human population begins to stabilize (Echols, 1976).

In one of the scenarios simulated in this study (the "slow burner" case), r was assumed to decrease to an asymptotic value of 1.35%/year according to the expression,

$$r(P) = \{74[1 - \exp(-0.00271P)]\}^{-1} \quad . \tag{3.5}$$

Combining Eqs. (3.4) and (3.5) results in a source function that accounts for both the resource limitation and the constraints from human actions, i.e.,

$$u(t) = \{74[1 - \exp(-0.00271P)]\}^{-1} P[1 - (P/P_\infty)^n] \quad . \tag{3.6}$$

Because this equation cannot be integrated easily for any arbitrarily chosen values of the constants, it was computed by updating r of Eq. (3.4) with the current value of p(t), and subsequently, the new

value of r was used in the following step to obtain the values of $\dot{P}(t)$ and $P(t)$.

3.3.3. Change in Land Areas of the Terrestrial Ecosystems

Very preliminary estimates of the area and trends of different terrestrial ecosystems are given in Table 3 (cf. Olson et al., 1978, Table 2.2). The preagricultural extents of major ecosystems are regrouped from the works of Rodin et al. (1975) and their map modified by Olson (1970b). The areal extents of each ecosystem for 1860 and 1970 are tentative approximations after parts of the areas had been cleared for the development of agricultural lands (cf. Wilson, 1978) and urban growth. These estimates have incorporated proper adjustments for unvegetated lands and other fringe areas (Ryabchikov, 1975) and in this sense are preferred over the land-area table compiled by Whittaker and Likens (1973). The latter's figures are generally higher than the estimates given in Table 3 for 1970, but may in some cases approximate conditions earlier in the historic clearing of the tropics (Sommer, 1976; Brünig, 1977), and other temperate regions (Rostlund, 1956; Darby, 1956).

The future growth of human population is used in the projection of area trends of the three land categories. According to some optimistic projections by Frejka (1973) and Echols (1976), human population may stabilize at about 6 to 8 billion people around 2075 A.D. Most of the forested lands that are suitable for conversion to intensive agriculture will probably be cleared during this period of growth. A

Table 3. Estimated Partitioning of Terrestrial Ecosystem Area Trends, and Organic Carbon Pools[a]

Reservoir	Estimated Area (10^6 km^2)			Estimated Carbon Pool							
				Live (Gton)							
	Preagricultural[b]	1860[c]	1970[d]	Preagricultural[b]	1860[c]	1970[d]	Density[e] kg/m^2	Dead[f] Gton	Total Gton	Rapid[g] Gton	Slow[h] Gton
1. WOODS COMPLEXES											
A. Boreal + Temperate											
Boreal (taiga)	10.10	9.5	9	101	88	81	9				
Semiboreal	6.91	5.6	5	64	48	40	8				
Cordilleran	3.77	3.5	3	68	61	45	15				
Other cool temperate	3.76	3	2	68	33	20	10				
Warm temperate	5.76	4.5	3.8	108	57	38	10				
Semiarid	3.83	2.5	2	25	15	10	5				
Arid moistland	1.07	0.4	0.2	13	2	1	5				
TOTAL	35.2	29.0	25.0	447	304	235	(9.4)				
NORTHERN WOODS SUBTOTAL (excluding S of 30° N: add below)		27.9	24		284	226		384	610	70	540
		1.1	~1		20	9					
B. Tropical + Subtropical											
Wet site, rainforest	4.56	4.3	3.3	84	68	59	18				
Other tropical moist	8.83	7	5.3	216	126	90	17				
Montane, seasonal	1.18	1	0.7	38	16	8	12				
Montane, humid	2.42	2.2	2	60	26	20	10				
Arid moistland	0.32	0.2	0.1	3	2	1	10				
Woody savanna, scrub	14.99	13.2	11.5	139	91	80	7				
TOTAL	32.3	27.9	22.9	540	329	258	(11.4)				
SOUTHERN WOODS SUBTOTAL		29.0	23.9		349	267		303	570	50	520
WOODS TOTAL		56.9	47.9	987	633	493		687	1180	120	1060

Table 3 (continued)

Reservoir	Estimated Area (10^6 km^2) Preagricultural[b]	1860[c]	1970[d]	Estimated Carbon Pool: Live (Gton) Preagricultural[b]	1860[c]	1970[d]	Density[e] kg/m^2	Dead[f] Gton	Total Gton	Rapid[g] Gton	Slow[h] Gton
2. NONWOODS COMPLEXES											
Agro-urban[i]											
Crops	0	5	12		4	12	1				
Fringe area[j]	1	4	7	2	8	14	2				
Buildings, etc.[k]	0	1	3	—	0	0	0				
		10	22	2	12	26	(1.2)				
Other Land											
Tundra-like, bogs	13.53	13	12	21	17	12	1.0				
Grasslands	22.96	21	20	31	21	14	0.7				
Desert, semidesert	29.35	30	29	9	15	17	0.6				
	65.84	64	61	61	53	43	(0.7)				
NONWOODS TOTAL	67	74	83	63	65	69		511	580	40	540
EARTH minus water, ice		130.9	130.9		693	562		1198	1760	160	1600
3. LAKES, RIVERS (including reservoirs)	3		3.1			0.03		60?	60?		60
4. GLACIERS	15+		~15			0.0					
EARTH minus oceans			149								
5. OCEANS	350+		361			1		1650	1651	2	1649
						558		2908	3471	162	3309
EARTH TOTAL			510								

Table 3 (continued)

[a]Source: Olson, J. S., H. A. Pfuderer, and Y.-H. Chan. 1978. Changes in the Global Carbon Cycle and the Biosphere. ORNL/EIS-109. Oak Ridge National Laboratory, Oak Ridge, Tennessee.

[b]After Rodin, Bazilevich and Rozov (1975), and Olson (1970b).

[c]Very preliminary judgement of forest clearing before and after 1860 (subject to revision).

[d]After Olson (1970a, and new estimates).

[e]1970 estimate only; parenthetical averages are weighted.

[f]1970 estimate of relatively active dead pool, probably excluding significant amounts of peat (histosols) and other resistant humus (estimated as having residence time near or greater than 1000 years).

[g]1970 estimate that includes materials with a probability distribution of fairly short residence times (such as living and dead stages of leaves, flowers, fruits, small roots, most animals, and their unstable residues.

[h]1970 estimate of most woody parts of live plants and dead residues having residence times averaging many years.

[i]After Ryabchikoff (1975).

[j]Fringe areas include decorative and wild vegetation, abandoned fields, roadsides, and other more or less vegetated areas around towns or other settlements.

[k]Relatively unvegetated areas in towns or industrial areas, mines, quarries, highways, and other disturbed areas besides agricultural fields.

relatively large portion of the remaining forests on tropical latosols and in mountainous regions would be harvested for timber and firewood, or parts would be cleared to provide lands for highways, industrial construction, urban development, etc.

Revelle (1974, 1976) carefully assessed the extent of the potential arable land available in various regions. Including only lands which can produce at least one crop with or without irrigation, he estimated that another 11×10^6 km^2 of the noncroplands would be developed further for intensive agriculture. This figure does not include about 0.5×10^6 km^2 of the humid tropical forests growing on low-fertility lateritic soils. Of the total potential cropland, only 3×10^6 km^2 are in the Northern Woods region and the other 8×10^6 km^2 are in the Southern Woods. At present, human settlements may occupy approximately 3×10^6 km^2 of the land surface containing considerable vegetation. A doubling of the human population in the next century would require an equal area for the same functions. A net area of only about 1×10^6 km^2 might be needed in the Northern Woods region, because nearly 20% of the population increase would occur in the developed countries of the North temperate latitude (Frejka, 1973). Assuming further that the entire potential arable land would be used for agriculture, an estimated 4×10^6 km^2 net of the Northern Woods area would be reduced gradually in the coming years. In the Southern Woods region, even assuming 1×10^6 km^2 of the semideserts and deserts could be utilized in urban expansion, this still would require 9×10^6 km^2 net of urban and agricultural lands to be converted from forested lands.

In the present model, a logistic equation was used to project the decrease of the Northern Woods land area, i.e.,

$$\dot{A}^N = A^N R^N (1 - A^N/A_\infty^N) \quad , \tag{3.7}$$

where $A_\infty^N = 20 \times 10^6$ km^2 is the asymptotic areal extent of the Northern Woods, and $R^N = 4.82 \times 10^{-3}$ yr^{-1} is the rate of clearing (Figure 3).

A simple logistic equation does not fit the area change scenario conjected for the Southern Woods region. Several facts have to be taken into consideration. Beginning from the early years of the 19th century, there was lavish exploitation of timber resources from the tropical forests (cf. Goodland and Irwin, 1975), and also a rapid expansion of new croplands in New Zealand, Australia, and South Africa (Wilson, 1978). The clearing rate, which was increasing exponentially due to these activities, was magnified by the synergistic effects of a 3% annual population growth (Revelle and Munk, 1977) and the destruction of forests during wars and the subsequent rebuilding of cities afterwards. It was assumed in this model that the clearing rate would stay constant after 1980 when the net loss of forested lands would be slowed down by reforestation and also afforestation in arid lands. By 2000 A.D., the clearing of forests would be limited by the forest resource itself and would be approximated by a logistic function. Hence, the change of Southern Woods land area was expressed by the following equations:

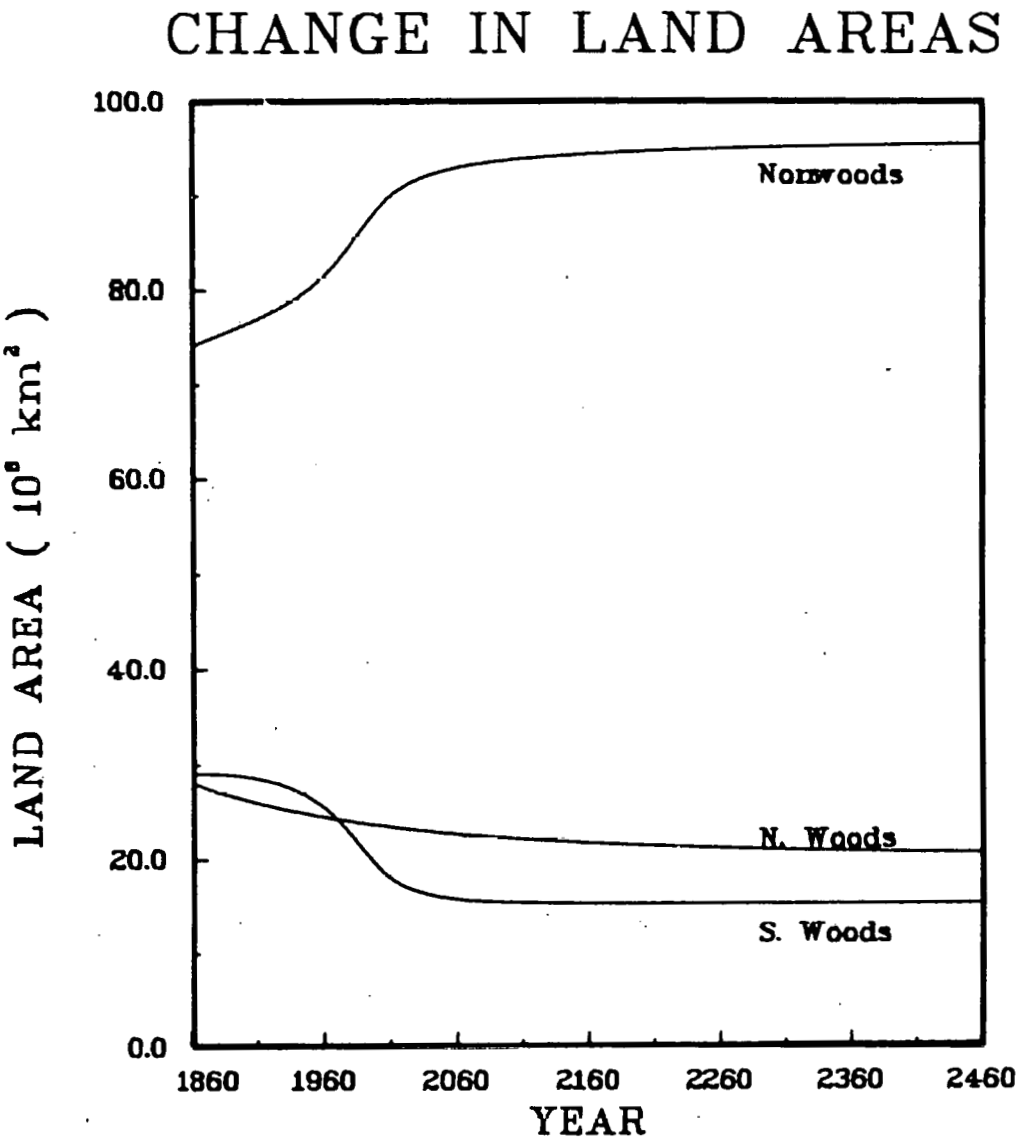

Figure 3. Projected area trends of Nonwoods, Northern Woods, and Southern Woods.

$$\dot{A}^S = A^S R^S(t) \quad , \tag{3.8a}$$

$$\text{where } R^S(t) = \begin{cases} -k_1 e^{k_0(t-1980)}, & 1860 \leq t \leq 1980 \\ -k_2, & 1980 < t \leq 2000 \\ k_3(1-A^S/A_\infty{}^S), & 2000 < t \leq 2400 \end{cases} , \tag{3.8b}$$

$k_0 = 0.03\ yr^{-1}$, $k_1 = 2.22 \times 10^{-4}\ yr^{-1}$, $k_2 = 8.13 \times 10^{-3}\ yr^{-1}$, $k_3 = 3.1 \times 10^{-2}\ yr^{-1}$, and $A_\infty{}^S = 15 \times 10^6\ km^2$ is the asymptotic areal extent of Southern Woods.

The constant k_0 was taken to be the same as the recent annual population growth rate of 3%. The other constants were chosen accordingly to obtain a relatively smooth curve (Figure 3) which would pass through both the estimated values of Southern Woods area for 1860 and 1970.

3.3.4. Shifts of Terrestrial Carbon After Deforestation

As the land is cleared, some of the carbon would be released promptly to the environment either from burning of fuelwood and slash or from enhanced respiration of heterotrophic soil organisms. A slightly bigger release of carbon would be delayed for a few decades or longer as a result of the slow decomposition processes of refractory carbon compounds (Swift, 1977). Presently, not enough systematic data have been published giving the relative amount of carbon released promptly to the environment after logging operations. As first approximations, 40% of the total carbon removed from the Southern Woods'

rapidly exchanging compartment was assumed to return promptly to the atmosphere, and the remaining portion to the Nonwoods' rapid carbon pool. The equivalent shifts from the rapidly exchanging carbon of Northern Woods were set at 30 and 70%, respectively. For the slowly exchanging carbon in the Southern Woods region, 20% of the carbon removed was considered to transfer promptly to the atmosphere, 35% to the Nonwoods' rapid carbon pool, and the other 45% to the Nonwoods' slow carbon pool. A slightly smaller fraction (20%) from the Northern Woods' slow carbon pool was modeled to shift promptly to the atmosphere, whereas 30% was shifted to the rapid carbon pool in the Nonwoods, and the remaining 50% to the Nonwoods' slow carbon pool.

3.3.5. Primary Production of Marine Organisms

Although carbon is found in abundance in sea water, the majority of it is not available for assimilation by plants in the sea. Most of the marine plants depend on dissolved CO_2 for photosynthesis. Only a small fraction of the plant species can make use of bicarbonate ions (Steemann Nielsen, 1975). Except under certain circumstances (clear, mid-day in summer) in some eutrophic lakes (cf. Schindler and Fee, 1975), the amount of CO_2 supplied by invasion from the atmosphere is rapid enough so that it is usually not limiting to the photosynthetic production of marine plants. Increase of the partial pressure of atmospheric carbon dioxide might induce a corresponding increase in the NPP of marine plants, but field measurements obtained so far are still not conclusive (Small et al., 1977).

The growth of marine microphytes (algae and diatoms) is known to be limited by the availabilities of nonchelated nitrogen, phosphorus, and silicon in the ocean (Stengel and Soeder, 1975). Thus, these microphytes may not be able to utilize fully the predicted addition of carbon dioxide to the ocean. Seaweeds and other aquatic macrophytes growing along the coastal regions probably get sufficient nutrients from river runoff; however, the contributions of these organisms to the total budget of marine organic carbon are probably small (Botkin, 1977). It is probably valid to assume that the change in NPP of aquatic plants would remain small within the time span of this simulation, or at least in the early decades before drastic change of ocean circulation and climate might occur. Hence, for this study, NPP of marine plants is kept at the present rate of 31 Gtons/year (cf. Platt and Subba Rao, 1975; Fogg, 1977).

3.3.6. Primary Production of Terrestrial Biota

As in Eriksson and Welander (1956), net primary production (NPP) of terrestrial biota is assumed to vary positively with the amount of atmospheric carbon and the mass of the biota. Net primary production can be modeled as the product of a logarithmic function of the relative increase of atmospheric carbon adjusted by a biotic growth factor, β, and the relative increase of carbon in the biota (Keeling, 1973a; Bacastow and Keeling, 1973; Keeling and Bacastow, 1977; Olson et al., 1978). The latter relation, especially the early rising phase, can be approximated roughly by a power function of the relative change in the carbon mass per unit area of each ecosystem. The employment

of the mass per unit area ("density") takes into account the recession in forest areas caused by various human activities.

A slightly modified form of the flux of carbon from the atmosphere to the land biota at a high concentration of atmospheric carbon dioxide is developed in this study. This is an adaptation of Keeling's (1973a) formulation. The function is expressed by the formula

$$F_{ab} = F_{ab}^* \left[1 + \beta \ln(c_a/c_a^*)\right] (\varepsilon_1) (q_b/q_b^*)^{2/3} (\varepsilon_2) \quad , \qquad (3.9)$$

where ca = mass of carbon in the atmosphere at time t,

c_a^* = mass of atmospheric carbon at steady state,

q_b = carbon mass per unit area at time t,

q_b^* = carbon mass per unit area at steady state,

ε_1 = environmental coefficient of CO_2 enrichment, and

ε_2 = limiting effect of increasing density of carbon mass.

The biotic growth factor, β, accounts for the availability of limiting resources for photosynthesis under natural conditions and for the genetic variation of the individual tree species in utilizing carbon dioxide. Values for β may range from 0 to 0.4 (Keeling, 1973a), but ought to reflect the wide spectrum of potential production responses expected for various ecosystems. In this model, β for the Northern Woods is set equal to 0.15; for both the Southern Woods and the Nonwoods, $\beta = 0.2$.

Since gas exchanges in photosynthesis and respiration are carried out mainly through the boundary layers of leaf and bark surfaces, it might be appropriate to relate the fluxes of carbon dioxide with change

in surface area than with change in biomass. The exponent 2/3 (i.e., the area:volume ratio) in the equation changes the dimension relation from a function of mass (or volume) to a function of area.

In general, the plant's response to a particular environmental factor can be defined by an idealized growth response curve (Russo and Knapp, 1976). This curve usually has a "tolerance" range, a narrower "growth" range and within which, an "optimum" range of values. Physiological activities of the plant can only occur between the upper and lower limits of the growth range. Growth rate is maximum within the optimum range. For each value (v) of any growth process, there is a related environmental coefficient (ε) which is considered as a measure of the plant's physiological response to this process. A hyperbolic decrease in any process is assumed outside the optimum range. Thus, the coefficient ε would assume one of the following values,

$$\varepsilon = \begin{cases} 1, & v \leq m \\ \tanh\ [\pi(v-g)/(m-g)], & m < v \leq g \\ 0, & v > g \end{cases} , \qquad (3.10)$$

where g is the upper value of the growth range, and

m is the upper value of the optimum range.

Measurements of NPP of tree species at very high ambient CO_2 level are few and the results are inconclusive. Furthermore, almost all of these measurements are done in leaf or growth chambers where the conditions may be more favorable for photosynthesis than those in the field. Some of the recent studies are reviewed by Lemon (1977), and an

annotated bibliography on the effects of global CO_2 enrichment on plants have been compiled by Strain (1977).

The increase of total NPP of a plant community under an enriched CO_2 atmosphere may be hindered by other limiting factors (cf. Attiwill, 1971; Botkin, 1977; Delwiche and Likens, 1977; Zinke, 1977). However, there are also measurements of photosynthesis in stressed or polluted environments that show substantial increases in NPP (e.g., Wright, 1974; Hurt and Wright, 1976; Bryan and Wright, 1976; Green and Wright, 1977). Even under very high ambient CO_2 concentration with favorable light and nutrient supplies, NPP also may be slowed down by the closure of leaf stomata or from plant tissue injury (Gaastra, 1959; Zelitch, 1971), or even by the high starch content in the leaves (Hofstra and Hesketh, 1975).

Several experimental results suggest that CO_2 enrichment effect may approach the upper ceiling at atmospheric concentration of from 650 ppm for Eucalyptus (Brittain and Cameron, 1973) to about 2400 ppm for the cucumber (Aoki and Yabuki, 1977). In general, plants following the C_4 pathway of photosynthesis saturate at a CO_2 level of about 1000 ppm, whereas most other plants with C_3 type metabolism only show inhibition at about 2500 ppm (Leopold and Kriedemann, 1975; Noggle and Fritz, 1976). These figures would be lowered considerably in natural conditions (cf. Totsuka, 1966; Lister and Lemon, 1976) because of mutual shading and competition for nutrients. In the present study, the optimal atmospheric CO_2 concentration for NPP [variable m in Eq. (3.10)] was set at 1000 ppm. The average NPP would be allowed to decrease with

the further increase of atmospheric carbon dioxide. At higher than 5000 ppm atmospheric CO_2 level [variable g in Eq. (3.10)], NPP is assumed to cease completely (cf. Moore, 1976; Bolhår-Nordenkampf, 1976).

Only scant field measurements are available for relating plant biomass to NPP (cf. Kira, 1975). Japanese studies have shown that leaf biomass and gross production rate tend to reach an asymptotic value within a relatively short period after the closure of forest canopy, whereas total plant biomass and community respiration continue to increase with age slowly but steadily. As a result, net production rate is expected to reach a maximum after the complete closure of canopy, and thereafter to decline gradually as the mass and plant respiration per unit area increase and the irradiation below the canopy decreases (cf. Hellmers, 1964; Kira, 1975; Reichle, in press).

Except for some open woodlands and savannas, the mature natural forests are usually closed forest. Thus, any increase in the carbon "density" will probably retard the rate of net production, if other conditions are held constant. The 1860 values of carbon "density" in Northern Woods, Southern Woods, and Nonwoods were assumed to be the optimal values [m's in Eq. (3.10)] in this study (see below). However, thinning may have proceeded so far by that time that the optimum is really higher.

In preagricultural time when the destruction of forest by man was still insignificant, the growth of forest would approach or even overshoot an equilibrium with the particular site and climate of the region. The "density" of carbon in these forests would indicate the

probable average maximal value supported under local environmental constraints—allowing for natural fluctuations. For live biomass, these values were taken as 1.35 and 1.4 times the 1970 values for the Northern Woods and Southern Woods, respectively. Similar values for dead organic matter might be larger due to the prolonged accumulation of humus and peats. Hence, the maximal values [g in Eq. (3.10)] for carbon "density" in the terrestrial biosphere were set at 1.5 times the 1970 values for both the rapidly exchanging carbon and the slowly exchanging carbon. Higher values are possible over limited areas, and possibly could apply over wide areas if forest management allowed it.

3.4. SYSTEMS REPRESENTATION AND MATHEMATICAL FORMULATION

In the present model, the total carbon in nature is considered to cycle within a closed system. Nineteen compartments are described by state variables for their carbon mass and ancillary treatment covers changes of area, and of amounts per unit area. Mathematically, the dynamics of the change in carbon mass in these compartments can be represented by a system of nonlinear ordinary differential equations. Let c_i^* be the mass of carbon at steady state in compartment i as given in Table 1 (page 33), and $F_{i,j}^*$ be the steady-state carbon flux from compartment i to compartment j as given in Table 2 (page 40). The rate constant a_{ij} in the differential equations is given by

$$a_{ij} = F_{ij}^*/c_{ij}^* \quad . \tag{3.11}$$

The dynamic equations for changes of elemental carbon in the troposphere (c_1) and in the stratosphere (c_2) are:

$$\dot{c}_1 = u(t) + a_{7,1}[\zeta(c_7 - c_7^*] + a_{2,1}c_2 + \sum_{i=10}^{15} a_{i,1}c_i + F_{17,1}^* - (a_{1,2} + a_{1,7})c_1 - \sum_{i=10}^{15} f_{1,i} \quad , \tag{3.12}$$

$$\dot{c}_2 = a_{1,2}c_1 - a_{2,1}c_2 \quad , \tag{3.13}$$

where u(t) is the input due to fossil fuel combustion; and $q_i(t)$, i=10, ..., 15 is the density of carbon in compartment i.

Only one equation is used to describe the time course of carbon in the living aquatic organism (c_3). However, the dead organisms and other organic matter in the mixed layer (c_4) and in the subsurface waters (c_5) are treated separately:

$$\dot{c}_3 = F_{7,3}^* - (a_{3,4} + a_{3,7})\, c_3 \quad , \tag{3.14}$$

$$\dot{c}_4 = a_{3,4}c_3 + a_{18,4}c_{18} + a_{19,4}c_{19} - (a_{4,5} + a_{4,7})\, c_4 \quad , \tag{3.15}$$

$$\dot{c}_5 = a_{4,5}c_4 - (a_{5,8} + a_{5,9} + a_{5,19})\, c_5 \quad . \tag{3.16}$$

The differential equation describing the change of inorganic carbon dissolved in ground waters is as follows:

$$\dot{c}_6 = a_{18,6}c_{18} + a_{19,6}c_{19} - a_{6,7}c_6 \quad . \tag{3.17}$$

The dynamics of inorganic carbon in the mixed layer (c_7), the thermocline (c_8) and the deep ocean (c_9) are modeled by the following equations:

$$\dot{c}_7 = a_{1,7}c_1 + a_{3,7}c_3 + a_{4,7}c_4 + a_{6,7}c_6 + a_{8,7}c_8 + a_{17,7}c_{17} - a_{7,1}[\zeta(c_7 - c_7^*) + c_7^*] - F_{7,3}^* - (a_{7,8} + a_{7,9} + a_{7,16})\, c_7 \quad , \tag{3.18}$$

$$\dot{c}_8 = a_{5,8}c_5 + a_{7,8}c_9 - (a_{8,7} + a_{8,9})\, c_8 \quad , \tag{3.19}$$

$$\dot{c}_9 = a_{5,9}c_5 + a_{7,9}c_7 + a_{8,9}c_8 + a_{16,9}c_{16} - a_{9,8}c_9 \quad . \tag{3.20}$$

For the six terrestrial biotic compartments, the mass balance equations are:

$$\dot{c}_{10} = f_{1,10} - a_{10,1}c_{10} + 0.7\dot{A}^N q_{12} + 0.3\dot{A}^N q_{13} + 0.6\dot{A}^S q_{14} + 0.35\dot{A}^S q_{15} \quad , \tag{3.21}$$

$$\dot{c}_{11} = f_{1,11} - (a_{11,1} + a_{11,18})\, c_{11} + 0.5\dot{A}^N q_{13} + 0.45\dot{A}^S q_{15} \quad , \tag{3.22}$$

$$\dot{c}_{12} = f_{1,12} - a_{12,1}c_{12} - \dot{A}^N q_{12} \quad , \tag{3.23}$$

$$\dot{c}_{13} = f_{1,13} - (a_{13,1} + a_{13,18})\, c_{13} - \dot{A}^N q_{13} \quad , \tag{3.24}$$

$$\dot{c}_{14} = f_{1,14} - a_{14,1}c_{14} - \dot{A}^S q_{14} \quad , \tag{3.25}$$

$$\dot{c}_{15} = f_{1,15} - (a_{15,1} + a_{15,18})\, c_{15} - \dot{A}^S q_{15} \quad , \tag{3.26}$$

where $f_{1,i} = F_{1,i}^* \{1 + \beta_i \ln[(c_1 + c_2)/(c_1^* + c_2^*)]\}$

$$\epsilon_1 (q_i/q_i^*)^{2/3} \epsilon_{2,i} \quad , \quad i = 10, 11, \ldots, 15 \quad . \tag{3.27}$$

The densities q_i associated with the terrestrial compartments are given by

$$q_i = \frac{C_i}{A_i} \quad , \quad i = 10, 11, \ldots, 15 \quad , \tag{3.28}$$

where $A_{10}, A_{11} = A^O$; $A_{12}, A_{13} = A^N$; and $A_{14}, A_{15} = A^S$.

Two of the remaining four equations describe the time behavior of inorganic carbon pools in calcareous tests and detrital carbonates (c_{16}) and sedimentary carbonates (c_{17}). The other two equations define the rate of change of organic carbon in the reactive sediment pool (c_{18}) and the lithified organic sediment (c_{19}). These equations are:

$$\dot{c}_{16} = a_{7,16} c_7 + a_{17,16} c_{17} - (a_{16,9} + a_{16,17}) c_{16} \quad , \tag{3.29}$$

$$\dot{c}_{17} = a_{16,17} c_{16} - F_{17,1}^* - (a_{17,7} + a_{17,16}) c_{17} \quad , \tag{3.30}$$

$$\dot{c}_{18} = a_{11,18} c_{11} + a_{13,18} c_{13} + a_{15,18} c_{15} - (a_{18,4} + a_{18,6} + a_{18,19}) c_{18} \quad , \tag{3.31}$$

$$\dot{c}_{19} = a_{5,19} c_5 + a_{18,19} c_{18} - (a_{19,4} + a_{19,6}) c_{19} - u(t) \quad . \tag{3.32}$$

In the computer program, the calculation is organized slightly differently to allow for future modifications of the carbon densities,

q_i. The rates of change of the six biospheric compartments are computed directly as the derivative of the product of (area) × (density), i.e.,

$$c_i = q_i \dot{A}_i + A_i \dot{q}_i \quad , \quad i = 10, \ldots, 15 \quad , \tag{3.33}$$

The values of mass/area (density) are estimated at present from the mass-balance equations (i.e., Eqs. 3.20 to 26; with c_i replaced by $\hat{c}_i$, i = 10, ..., 15, in the computer program) and the appropriate area of the ecosystem groups.

CHAPTER 4

COMPUTER SIMULATIONS AND RESULTS

4.1. IMPLEMENTATION OF THE MODEL

The nonlinear system of differential equations which defines the present model has been solved by numerical integration on the computer. The computer program in the Appendix is coded in IBM's Continuous Simulation Modeling Program-III (CSMP-III) simulation language (IBM, 1975). This special language offers a standard framework for representing the model structure, provides many functional units which are easily incorporated into a model, and requires only a few control statements to obtain line-plots and tabulation of the outputs. This language is a proprietary software package and depends on the use of IBM machines and usually occupies a substantial core requirement in the computer. Yet the relative ease and speed of achieving a workable program makes it a helpful tool in the developmental stage of many simulation models. All simulation runs have utilized the IBM 360/75 or 360/91 machine at the Oak Ridge National Laboratory. The plotting programs are written in DISSPLA (Integrated Software System Corporation, 1970) language which is a special software package for preparing graphic displays on the various output devices.

In the present model, the time constants (or fractional turnover rates) associated with each compartment have a large range of values (from > 1000 years to ~ 1 year or less). Systems of differential

equations with this characteristic are usually numerically stiff (Gear, 1971). The CSMP-III package provides an integration method called STIFF to handle this type of system. However, the subroutine is available only in single precision operation, and is likely to accumulate errors during the many integration steps when the system is solved for a long time period. It is useful to note that the present model is a closed system in the sense that carbon neither enters nor leaves the system. To keep track of these errors, differences in final and initial values of the total mass of carbon were used to check the extent of deficit from round-off.

4.2. SIMULATION SCENARIOS

Four separate cases differing mainly by the different forcing functions (Figure 4) were simulated. The time period for each run was the same for all cases—from 1860 to 2460. These scenarios can be contemplated as conceivable strategies or management policies of fossil-fuel consumption and land clearing in the future. We do not know the exact future scenario, but can examine cases that hopefully bracket the real future. The 19 initial pool sizes, 44 fluxes, and about 20 other parameters were specified as discussed in Sections 3.1 and 3.2. System responses from alternative choices of these parameters will be discussed in Chapter 5.

Case 1 is the "nominal" or reference run. It is chosen as the standard against which the other simulation experiments are compared. The annual input of fossil carbon used in this simulation run is assumed

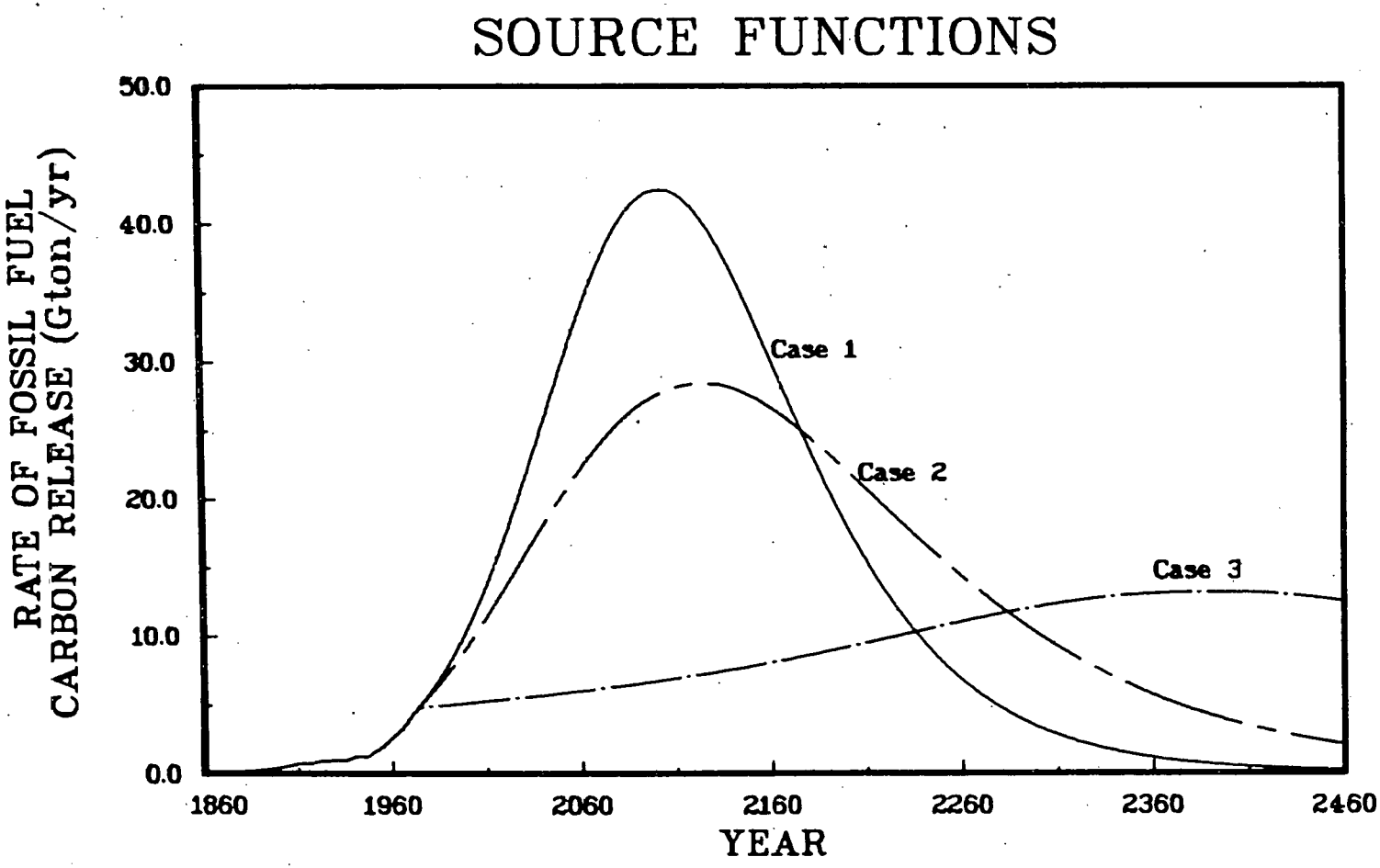

Figure 4. Different scenarios of annual carbon input from fossil fuel and cement. Case 1: "nominal" scenario. Case 2: "delayed consumption" scenario. Case 3: "slow burner" scenario.

to represent a slight moderation of the present trend (Figure 4). Projection of the annual release of CO_2 from fossil-fuel combustion and cement manufacturing beyond 1974 was calculated from Eq. (3.1) with the initial rate parameter, $r = 0.0435\ yr^{-1}$, and the exponential constant, $n = 0.427$ derived from Eq. (3.3).

Case 2 represents a "delayed consumption" scenario. A similar modified logistic equation was employed to project the future release of fossil carbon. To obtain a low-release curve, the exponential constant was reduced to $n = 0.1$. The initial rate parameter was computed as in Eq. (3.4) from which a value of 0.11 was obtained.

Case 3 typifies the "slow burner" scenario with drastic (probably very unrealistic) and sudden decrease in the expansion of fossil-fuel burning. The source function of fossil carbon was given by Eq. (3.6) with a decreasing rate parameter. It was computed according to the procedure discussed in Section 3.3.2. This scenario can be considered as one representing possible reduction of the future rate of fossil-fuel burning. The exponential constant in Eq. (3.6) had the same value as in Case 1, i.e., $n = 0.427$.

Case 4 combined the reduced input of fossil carbon as in Case 3 but with an additional damping action from an improved biomass management. It may be called the "combination" scenario. Besides a decreasing rate of fossil-fuel consumption, it was assumed that a larger portion of the slowly exchanging carbon in the wooded compartments would be stored in the reactive sediment-deep humus compartment after 2000 A.D. This might be the result of better soil management practices in agriculture

and forestry, increases of carbon in wetlands, and in deep soil horizons or the deliberate utilization of timber as materials in construction and durable commodities industry, with deliberate delays in recycling. In the model, these varied possibilities for delayed oxidation after 2000 A.D., were simulated simply as if 1% per year of the slowly exchanging carbon in the wooded compartments were transferred continuously to the sediment-deep humus compartment (compartment 18). At the same time, the biotic growth factors for the Northern Woods and the Southern Woods compartments were raised to 0.2 and 0.25, respectively.

The following sections will first discuss the behavior of the "nominal" run in considerable detail. Responses of the model in the "slow burner" scenario will serve for comparison to illustrate the differences. Results from Cases 2 and 4 will be discussed in Chapter 6 where they are most relevant to alternative policies of fossil-fuel consumption and biospheric management.

4.3. TRIAL RESULTS AND DISCUSSIONS

4.3.1. Anthropogenic Releases of Fossil Carbon

Estimates of the annual input of fossil carbon used in the different scenarios are given in Table 4. The projections associated with the "nominal" case are slightly less than the predicted levels of fossil-fuel consumption suggested by Hamilton (1977). The emission rate at 2000 A.D. was near the 11 Gtons/year assumed by Niehaus (1976) as the optimistic equilibrium strategy in his study. This value also came within the range of 8.5 to 15 Gtons/year given by Rotty (1977) who based his estimates

Table 4. Annual and Cumulative Release of Carbon from Fossil Fuel and Cement

Scenario	1975	1980	1985	2000	2020	2025	Peak (Year)
			Annual Release (Gton/year)				
Nominal	5.05	5.92	6.90	10.62	17.42	19.41	42.48 (2101)
Delayed	5.03	5.77	6.57	9.32	13.62	14.75	28.41 (2121)
Slow burner or combination	4.90	4.96	5.03	5.23	5.50	5.57	13.20 (2382)
			Cumulative Release (Gton)				
Nominal	142.2	169.6	201.6	331.5	608.5	700.5	
Delayed	142.2	169.2	200.0	318.6	547.2	618.1	
Slow burner or combination	142.1	164.7	185.8	244.2	317.9	336.2	

on fuel-use trends and the different regional patterns of energy demand. Hamilton's (1977) projection of fossil-fuel consumption for the year 2000 is about 14 Gtons/year.

The peak rate of release for the "nominal" scenario would occur in 2101 A.D. when the fossil carbon would release at the rate of 42.5 Gtons/year. In the "delayed consumption" scenario, the peak would be delayed for another 20 years, i.e., at 2121 A.D., and the maximum rate of release would be about 28.4 Gtons/year (Figure 4, page 75).

The other source function used by both the "slow burner" and the "combination" scenarios projected a very slow increase in the fossil-fuel consumption and cement manufacturing over an extended period of time. The peak would be delayed further until 2382 A.D. when only 13.2 Gtons/year of fossil carbon would be released (Figure 4).

This source function represents a very modest role for fossil-fuel consumption in the future. It can be argued that in the next 25 or 50 years, the supply of fossil fuel would be a determinant in formulating any national energy policy. The demand for energy in the developed countries would be met increasingly by nonfossil sources, most likely nuclear or solar (Weinberg and Rotty, 1977). Most of the developing countries have neither the resources (geological or economic) nor the social infrastructure to adopt such an energy-intensive lifestyle. Hence, the demand for fossil fuel in the near future might possibly be lower than the current estimates proposed by Rotty (1977) and Niehaus (1976). The higher peak releases projected in the "nominal" and the high but "delay consumption" scenarios might not be realized as the

unacceptable ecological impacts of a high level of atmospheric CO_2 gradually induce all of human society to shift to a strategy of moderate release (Olson et al., 1978, Fig. 6-3).

By 1970, an estimated 120 Gtons of carbon had been released from cement and fossil fuel. According to the "nominal" case, this value would increase to 332 Gtons by the end of this century. In the "slow burner" scenario, the cumulative release would be about 244 Gtons in 2000 A.D., and less than 5000 Gtons of carbon would be released to the atmosphere between 1860 and 2460 (Table 4, page 75).

4.3.2. Anthropogenic Releases of Biospheric Carbon

Table 5 summarizes one hypothesis of credible annual and cumulative releases of biospheric carbon from land-clearing activities whose consequences are examined in the simulations that follow. As mentioned in Section 3.3, only those parts of the organic carbon that are released directly to the atmosphere or are shifted from slowly to rapidly exchanging pools will have any appreciable influence on the near-term behavior of the carbon cycle. Note that the turnover time of delayed releases of biospheric carbon require several years, so the actual amounts reaching the atmosphere as the model operates are lower than the estimates given in Figure 5 for years of rising release, and higher in years of declining release.

About 1.33 Gtons/year of organic carbon are estimated to have been cleared from terrestrial forests in 1860 A.D. Of this amount, only 0.28 Gtons were released promptly to the atmosphere in the same year;

Table 5. Annual and Cumulative Amounts of Biospheric Carbon Assumed to be Removed and Released

	1900	1950	1970	1975	1980	2000	2050	2100
				Annual Transfer (Gton/year)				
Total removed*	1.340	2.632	4.040	4.520	5.060	4.731	1.081	0.443
Prompt release	0.285	0.568	0.876	0.980	1.097	1.022	0.231	0.094
Delayed release	0.474	0.962	1.488	1.667	1.867	1.740	0.388	0.155
Relocated	0.581	1.102	1.676	1.873	2.096	1.969	0.462	0.194
				Cumulative Transfer (Gton)				
Total removed*	52.91	143.99	210.49	232.23	256.56	354.01	468.56	502.47
Prompt release	11.20	30.74	45.14	49.85	55.13	76.22	100.82	108.04
Delayed release	18.43	51.28	75.70	83.71	92.69	128.59	170.35	182.39
Relocated	23.27	61.98	89.66	98.67	108.75	149.20	197.39	212.04

*Difference between releases and total removed is assumed to persist as humus with slow turnovers in nonwooded compartments or sediments relocated by soil erosion. Numbers do not add up because of rounding.

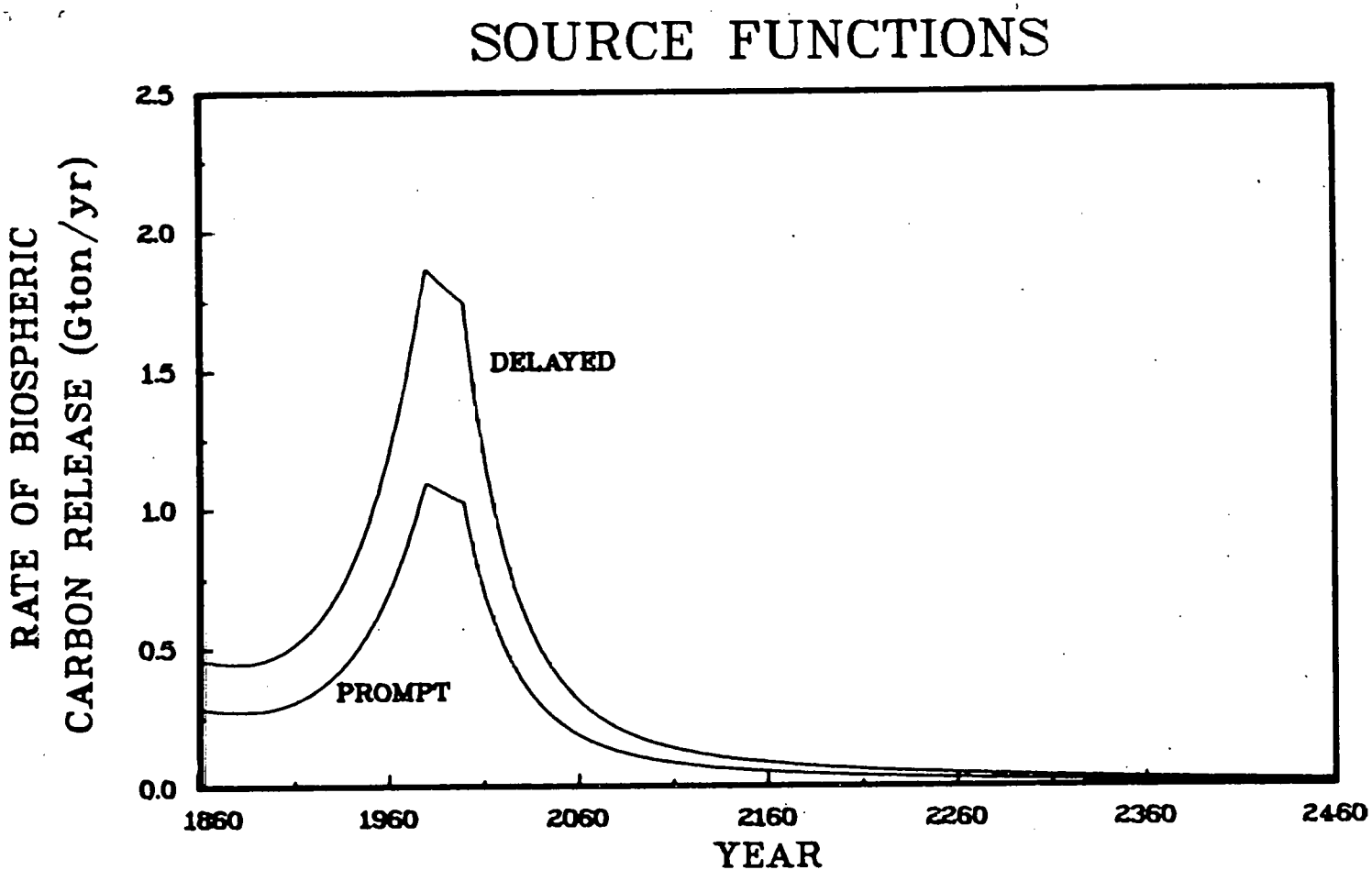

Figure 5. Annual releases of biospheric carbon from deforestation ("nominal" scenario).

0.46 Gtons are assumed to constitute the delayed release (Figure 5). The remainder was assumed to persist as humus with "slow" turnover rate nearly unchanged, even though soil had changed status from wooded to nonwooded ecosystems. Figure 5 also shows that the release of both categories of carbon increase abruptly after 1890 and reach a peak value of 2.96 Gtons/year in 1980. If the deforestation rate in the Southern Woods region is considered to stabilize and decrease after 1980, the total biospheric release would also decrease slowly to 2.71 Gtons/year in the year 2000 and then would drop precipitously in subsequent years.

The steep increase of biospheric carbon releases since 1890 also is shown by the cumulative releases (Figure 6). From 1860 to 1950, the cumulative amount of biospheric carbon from deforestation activities, transferred from both the Northern Woods and the Southern Woods regions, is estimated to be 144 Gtons. Only about 82 Gtons of the carbon is released to the atmosphere either promptly or delayed for a few years. That which is not released as CO_2 is at least redistributed to the slow carbon pool of the Nonwoods compartment. The total cumulative release was estimated to increase to 121 Gtons in 1970 and to reach 205 Gtons in 2000 A.D. (Table 5). Thereafter, the excess carbon coming from nonfossil sources increase very slowly.

The amounts of biospheric carbon released to the atmosphere are very much the same in the "nominal," the "delayed," and the "slow burner" scenarios. In the "combination" scenario, however, the contribution of nonfossil carbon is less than the other cases, especially after the peak

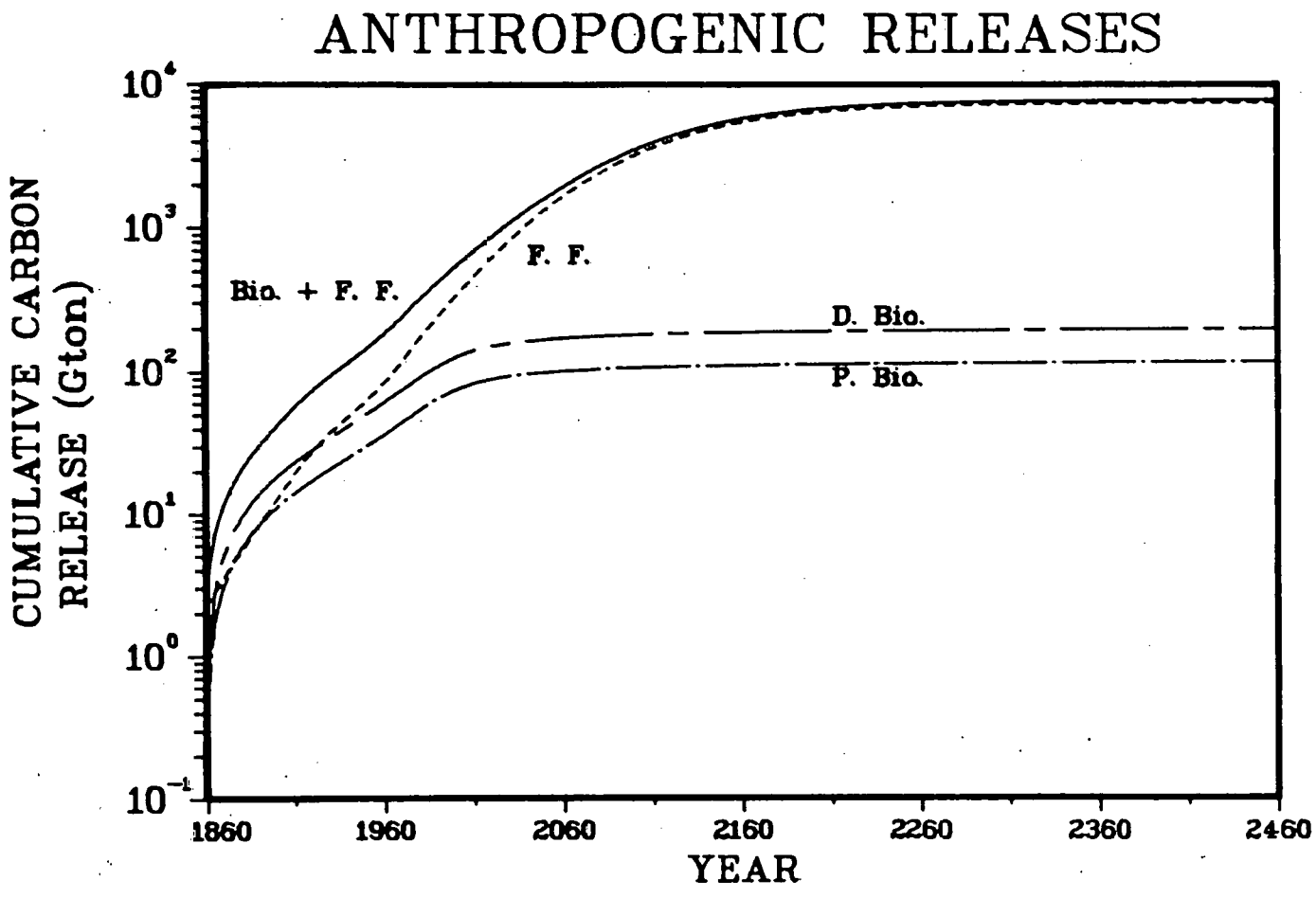

Figure 6. Cumulative releases of carbon from deforestation and combustion of fossil fuel ("nominal" scenario). P. Bio: prompt release of biospheric carbon. D. Bio: delayed release of biospheric carbon. F. F.: fossil carbon.

release in 1980. This is the result of additional fluxes of carbon going to the deep humus compartment and the effect of higher biotic growth factors on storage of carbon.

Comparing the cumulative releases of fossil and nonfossil carbon, there is a greater contribution initially from the biota than from the fossil sources as was suggested by Stuiver (1978). However, the share from fossil fuel increases rapidly and by 1970, both sources have already contributed approximately an equal amount, ∿120 Gtons, to the atmosphere over the period since 1860.

Figure 7 shows the cumulative releases of biospheric and fossil carbon in the "slow burner" scenario. In this case, the total production of fossil carbon, and hence the total anthropogenic release, will not increase as rapidly as those of the "nominal" case. However, the cumulative amount of biospheric carbon released from both active categories is about the same as the amount estimated in the "nominal" case.

The annual release of biospheric carbon estimated by the present study was about the same as given in Baes et al. (1976). The projections for the prompt release of biospheric carbon from the wooded compartments for the next 25 years were about 0.98 to 1.01 Gtons/year. Corresponding projections for the delayed release were 1.67 to 1.87 Gtons/year (Table 5, page 81). The overall release was higher than the estimates of Bolin (1977a) and Wong (1978) but was near the lower limit of the range of values suggested by Woodwell et al. (1978).

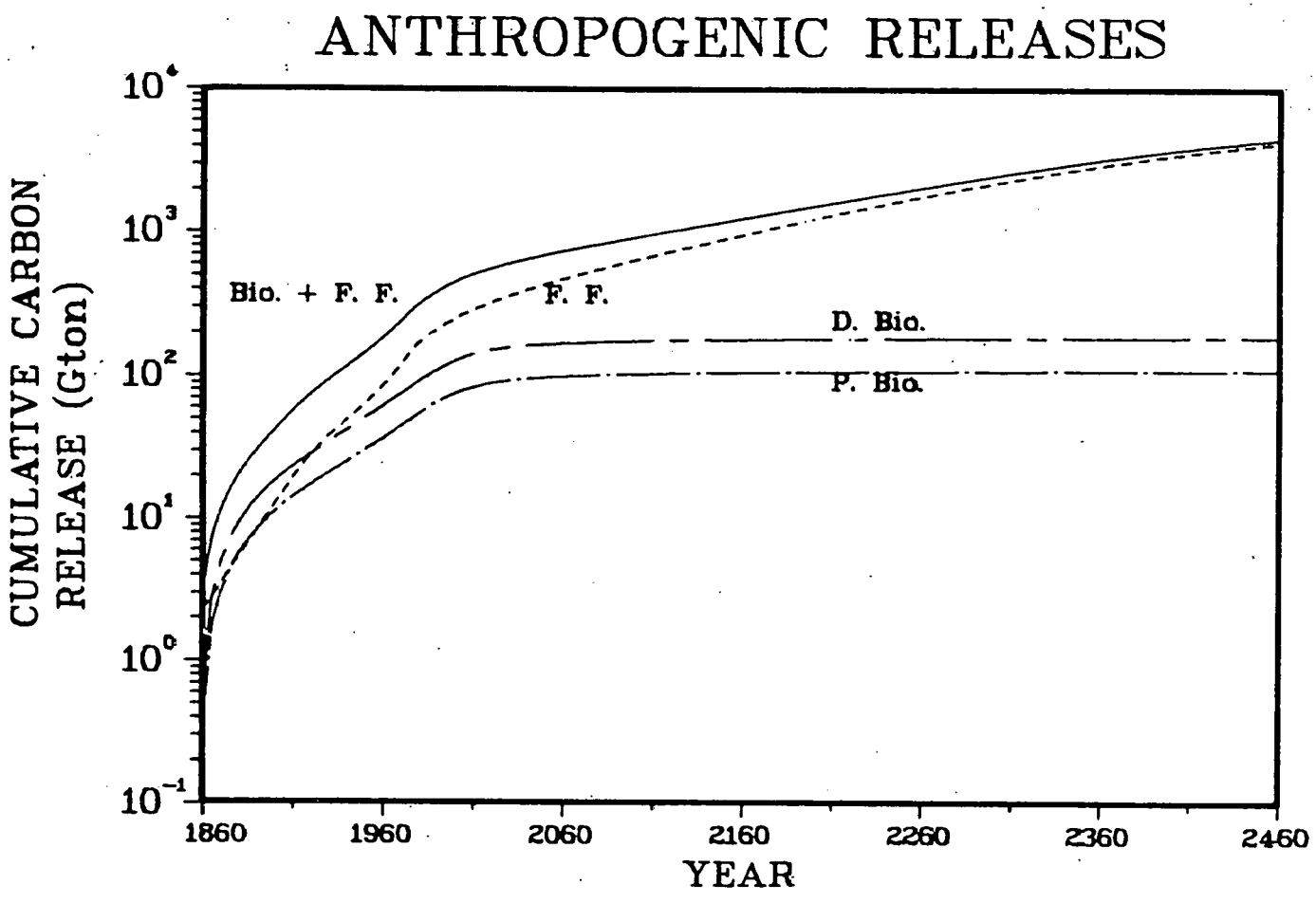

Figure 7. Cumulative release of anthropogenic carbon ("slow burner" scenario). P. Bio: prompt release of biospheric carbon. D. Bio: delayed release of biospheric carbon. F. F.: fossil carbon.

Only a few estimates of the cumulative input of biospheric carbon are available. Except for the rough estimate given by Hutchinson (1954), the other values are based on the measurements of radiocarbon ratios from a limited sample of dated tree rings. Hutchinson's (1954) estimate for the period 1900 to 1935 was about 20 Gtons of carbon. This value might be too low, as Farmer and Baxter (1974) gave an estimate of about 30 Gtons over the 20-year period after 1900. The former value compared favorably with the result of about 17 Gtons obtained in this study (Figure 6, page 84). The calculation given by Stuiver (1978) for the hundred-year period beginning from 1850 was about 120 Gtons. This value is higher than the result obtained in this study, even when the possible contribution from the 10-year period before 1860 was added to the estimate from the simulation study. However, Wilson (1978) suggested that the total release of nonfossil carbon for the brief period between 1860 and 1890 would possibly be 110 Gtons. This was 5 times the estimate of 22.2 Gtons obtained for the same period in this report. Until more estimates from radiocarbon measurements are available, it is difficult to judge the accuracy of these estimates. Refinement of the model would be required if further measurements confirm that a substantially larger amount of the biospheric carbon was released before the end of the last century.

4.3.3. Projections of Atmospheric Carbon

Time courses of carbon level in the atmospheric compartment obtained from all four scenarios are shown together in Figure 8. In the

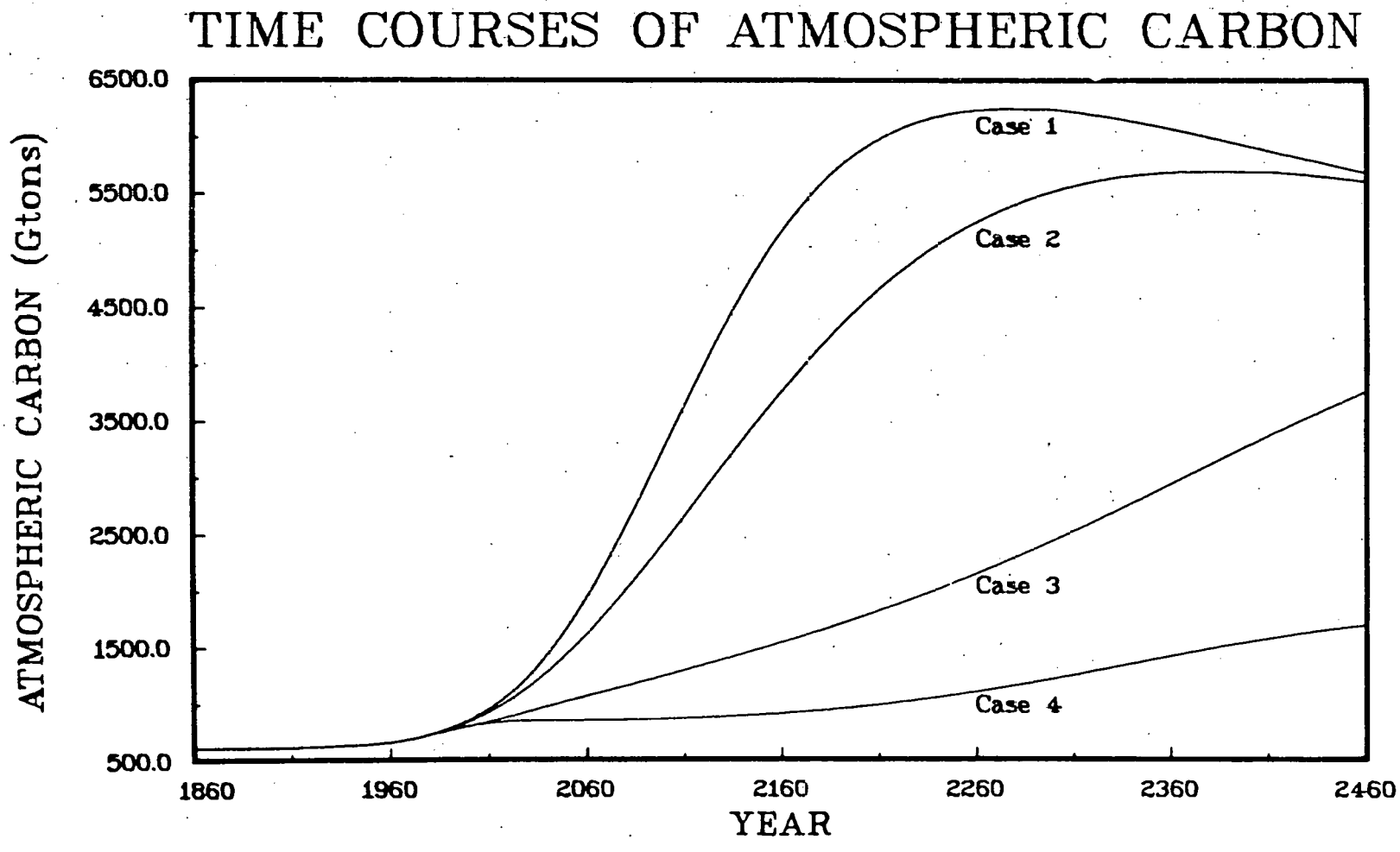

Figure 8. Simulated changes of carbon level in the atmosphere (troposphere + stratosphere). Case 1: "nominal" scenario. Case 2: "delayed consumption" scenario. Case 3: "slow burner" scenario. Case 4: "combination" scenario.

"nominal" case, the trajectory of atmospheric carbon rises rather rapidly to a peak around 2275 A.D. and, after passing this peak, declines slowly. The estimated value at 1975 was 712 Gtons; it increases to 861 Gtons in the year 2000. At the beginning of the 22nd century, the concentration increases to more than five times the original value in 1860. A doubling of the preindustrial level to 1196 Gtons is reached at the year 2027. The projected peak value reaches about 6253 Gtons, or 10.5 times the preindustrial value. As compared to the studies of Revelle and Munk (1977) and Keeling and Bacastow (1977), the results obtained in this report show a faster increase of atmospheric carbon and a doubling time which both are achieved earlier in the time history of the simulation.

In the "slow burner" scenario, the level of atmospheric carbon increases at a much slower pace than in the "nominal" case and still shows an increasing trend at the last time step of the simulation. Compared to the "nominal" case, the estimated value in this simulation run for 1970 was about the same as in the former case. Atmospheric CO_2 then increases to 809 Gtons in 2000 A.D., but only 1257 Gtons in 2100 A.D. Furthermore, it delays for another 60 years (i.e., 2087 A.D.), the time required to reach twice the amount of the preindustrial level. At the final time step of the simulation, the projected carbon level reaches 3376 Gtons, or six times the 1860 concentration.

The peak values of atmospheric carbon concentration predicted by many workers (e.g., Revelle and Munk, 1977; Oeschger et al., 1975) are about two to five times the concentration before 1860 A.D.

Exceptions are the higher estimates given by Bacastow and Keeling (1973) and Keeling and Bacastow (1977), reaching five to eight times the preindustrial value. Options in their models to limit the mass of the biosphere seem to retain a higher fraction of the injected carbon in the atmosphere. By putting several restraints on net primary production (see Section 3.3.6), the present model predicted a still higher retention of the atmospheric carbon, reaching slightly more than 10 times the initial concentration.

In retrospect, the low peak values of atmospheric carbon concentration obtained from some of the carbon models might be explained by: (1) the unlimited growth of biomass absorbing most of the additional carbon dioxide (see example Fig. 10.1 and Fig. 10.2 of the first model of Revelle and Munk, 1977); (2) the somewhat lower estimate of the fossil carbon resource (3 to 4 $\times 10^3$ Gtons) ultimately released (Dugas, 1968; Killough, 1977; Young et al., 1972); or (3) the deliberate constraints imposed on the rate of anthropogenic production of carbon dioxide in their simulations (Smil and Milton, 1974; Zimen et al., 1977; Siegenthaler and Oeschger, 1978).

4.3.4. Marine Compartments

The increases of carbon levels in the marine compartments are slow and steady initially, but accelerate in the next two or three centuries as atmospheric carbon increases (Figure 9). In the "nominal" case, the amount of carbon in the mixed layer reaches a maximum of 780 Gtons, peaking at about the same year (i.e., 2275 A.D.) as in the atmospheric

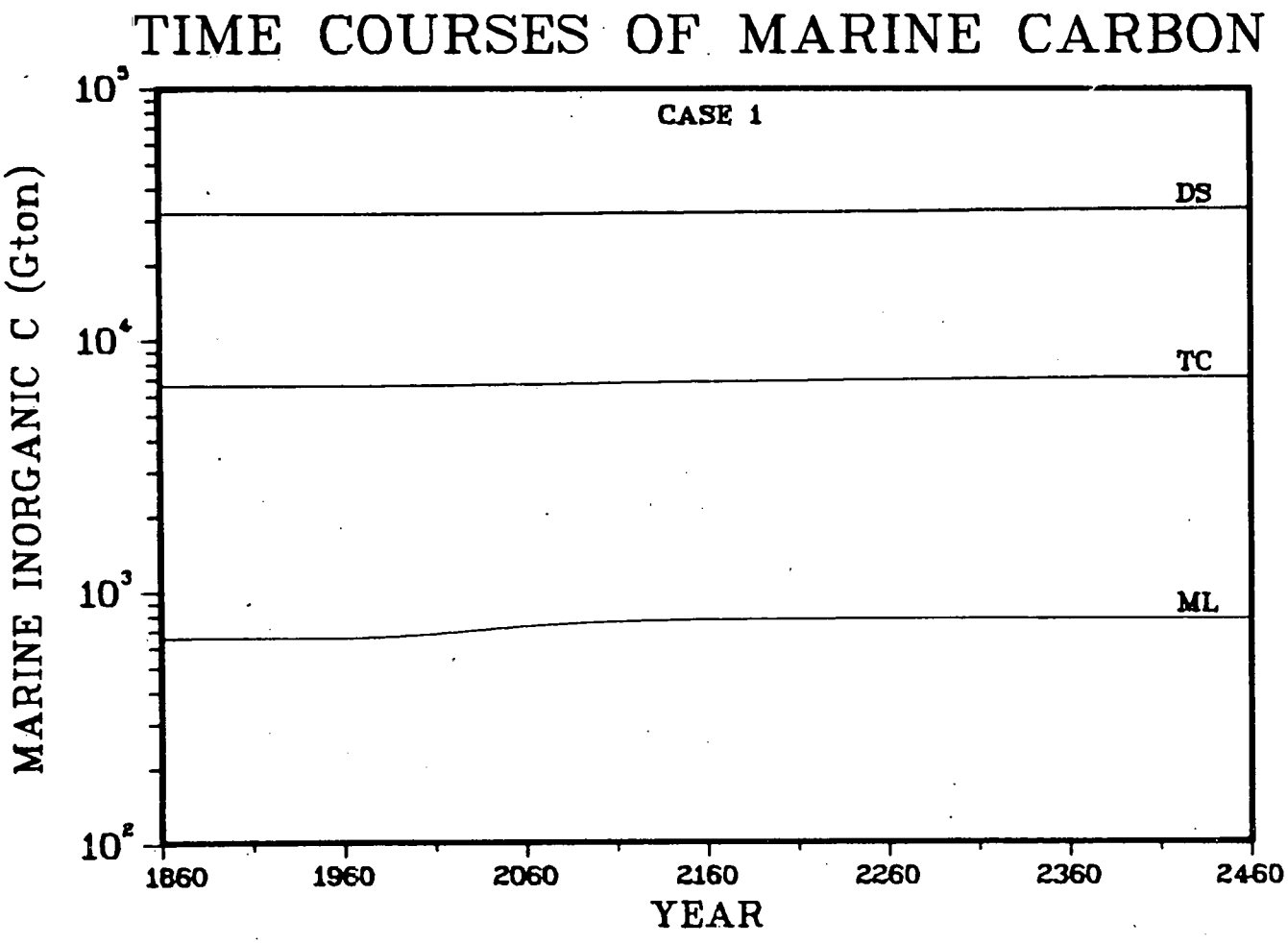

Figure 9. Projections of carbon level in the mixed layer (ML), the thermocline (TC) and the deep sea (DS) ("nominal" scenario).

compartment. This finding of a synchronous peaking time suggests that the chemical equilibrium of sea water in the mixed layer responds instantly to an increase in atmospheric carbon dioxide with very short time delay. This observation was in line with a similar time-delay of about two years calculated by Broecker (1977). By 1975, about 33 Gtons of carbon are added to the marine compartments during the previous 115 years of fossil-fuel combustion and land-clearing.

Generally, carbon levels in the marine compartments obtained with the "slow burner" scenario are slightly less than the corresponding values obtained from the first case. There is no peaking of mixed layer carbon observed in the "slow burner" scenario. This is also true for the time courses of carbon levels in the thermocline and the deep ocean compartments for both scenarios.

4.3.5. Net Primary Production

The net primary production of wood compartments displays a similar behavior in both the "nominal" and the "slow burner" scenarios. Since the increase of NPP is proportional to the increases in the mass of atmospheric carbon and the biospheric "density," the NPP is expected to show an increase during the early time-step of the simulation. Figure 10 shows the responses of the NPP of both rapidly and slowly exchanging carbon obtained with the "nominal" case.

In the wood compartments, the NPP decreases a little for the first few years, probably because of lower mass densities resulting from extensive deforestation. As the atmospheric carbon is increasing, the

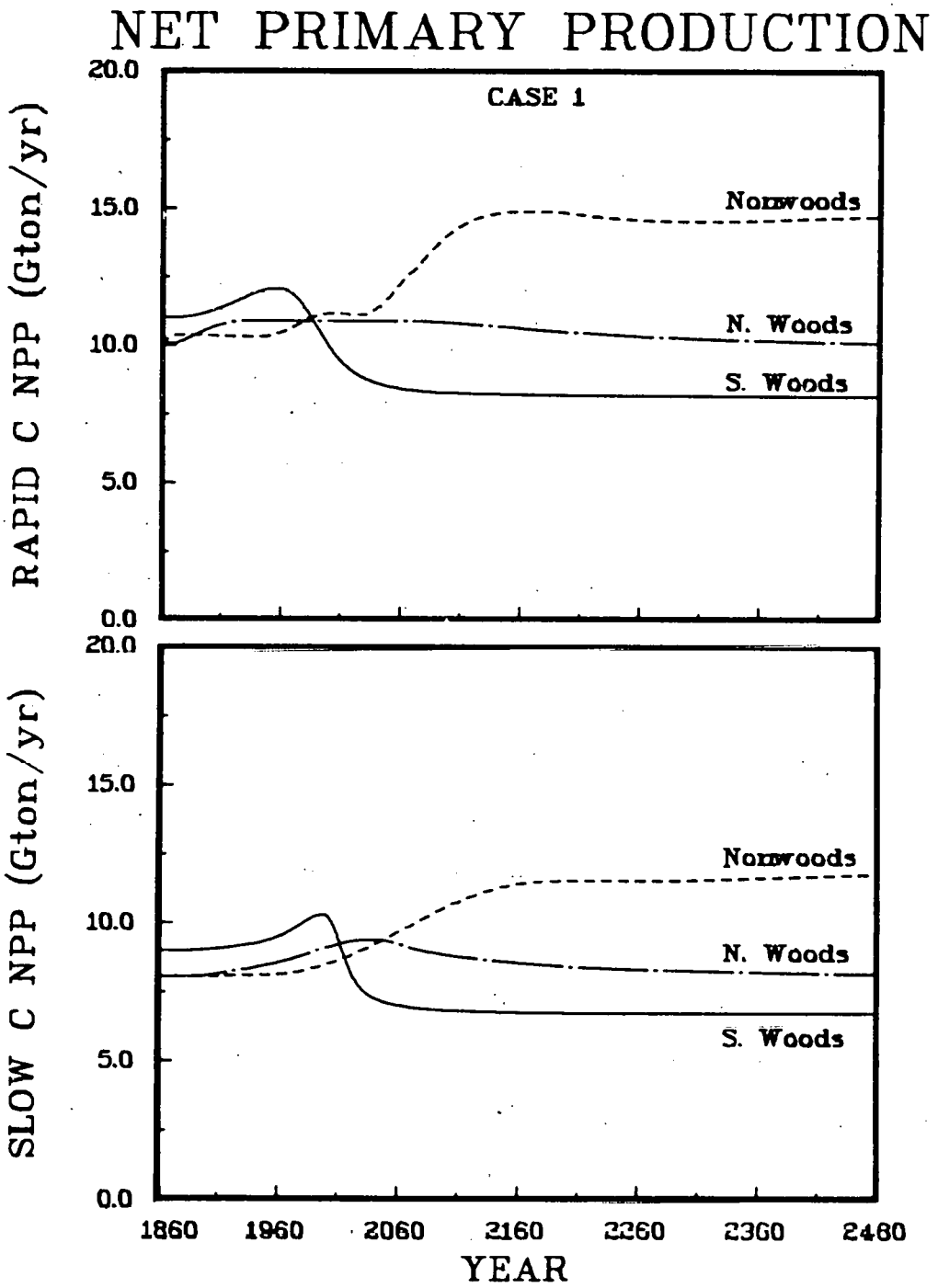

Figure 10. Simulated net primary production of terrestrial ecosystems ("nominal" scenario).

NPP of the rapidly exchanging carbon in both compartments also increase above the preindustrial values (10.88 Gtons/year in the Northern Woods and 12.07 Gtons/year in the Southern Woods). The slowly exchanging carbon in the wood compartments also shows a similar trend, but the period of increase continues further to around 1995 for the Southern Woods (10.29 Gtons/year) and 2030 for the Northern Woods (9.37 Gtons/year). The subsequent decline of these compartments is likely a reflection of the limiting conditions of the high atmospheric carbon and the high mass "density" imposed in the structure of the model.

The NPP obtained for the Nonwoods is strongly influenced by the addition of carbon from land-clearing activities. The later period is again limited by the high level of atmospheric carbon and a large carbon mass per unit area. In this simulation, the NPP shows an increasing trend in the beginning and then levels off at about 2160 A.D., reaching 14.8 Gtons/year in the rapid carbon pool and 11.7 Gtons/year in the slow carbon pool.

Results obtained from the "slow burner" scenario show that the changes of NPP in the wood compartments are almost identical with those from the "nominal" case, except that the values of NPP from the former scenario are slightly slower between the period from 2000 to 2150. The Nonwoods compartments give different responses in their rates of organic production. Generally, the magnitudes of NPP are lower than similar items in the "nominal" case, and the NPP also depicts a much slower increase over the entire period of the simulation (Figure 11). This behavior is most likely caused by the low level of atmospheric

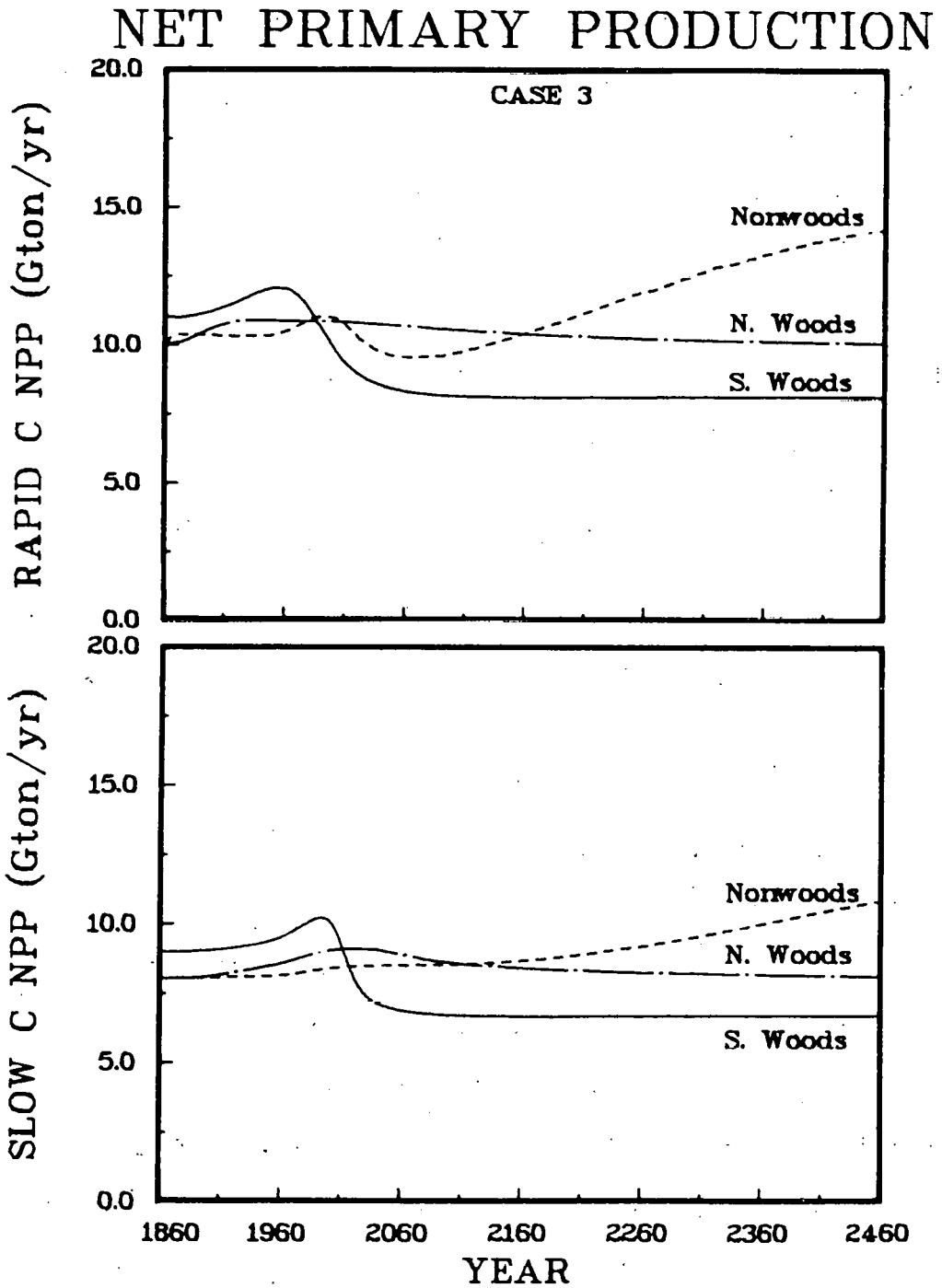

Figure 11. Simulated net primary production of terrestrial ecosystems ("slow burner" scenario).

carbon projected in this scenario. The "benefits" of CO_2 fertilization are not fully reflected in the biosphere growth, and the NPP might actually be higher if the actual constraints on photosynthetic rate and on ultimate storage are less severe than are assumed.

4.3.6. Mass Per Unit Area of Biospheric Carbon

Since the value of mass/area ("density") of biospheric carbon in this model is formulated as a mass-balance equation depending on the changes of organic carbon, its behavior is strongly linked to the net primary production and the atmospheric level of carbon. The upper frame of Figure 12 depicts the changes of mass density of the rapidly exchanging carbon (e.g., foliage and other short-lived plant parts and residues) in the "nominal" case. The two wood compartments have curves of the same shape, but the Southern Woods shows a time-delay of several decades related to clearing and regrowth history.

In the first few years, the density values of the Southern Woods remain almost constant on the average. After this period, the values of carbon/area increase for about a hundred years as regrowth dominates those cut areas which remain in forest. The values of biomass/area and NPP level off for the remaining period of the simulation. The period of increase is probably the result of the temporarily accelerated NPP. As the biota limits its own foliage and NPP at some upper value in the future, the density also remains relatively unchanged. In reality, it may decrease if turnover times are shortened by intense harvesting—e.g., for biomass fuels. At the assumed equilibrium stage, the

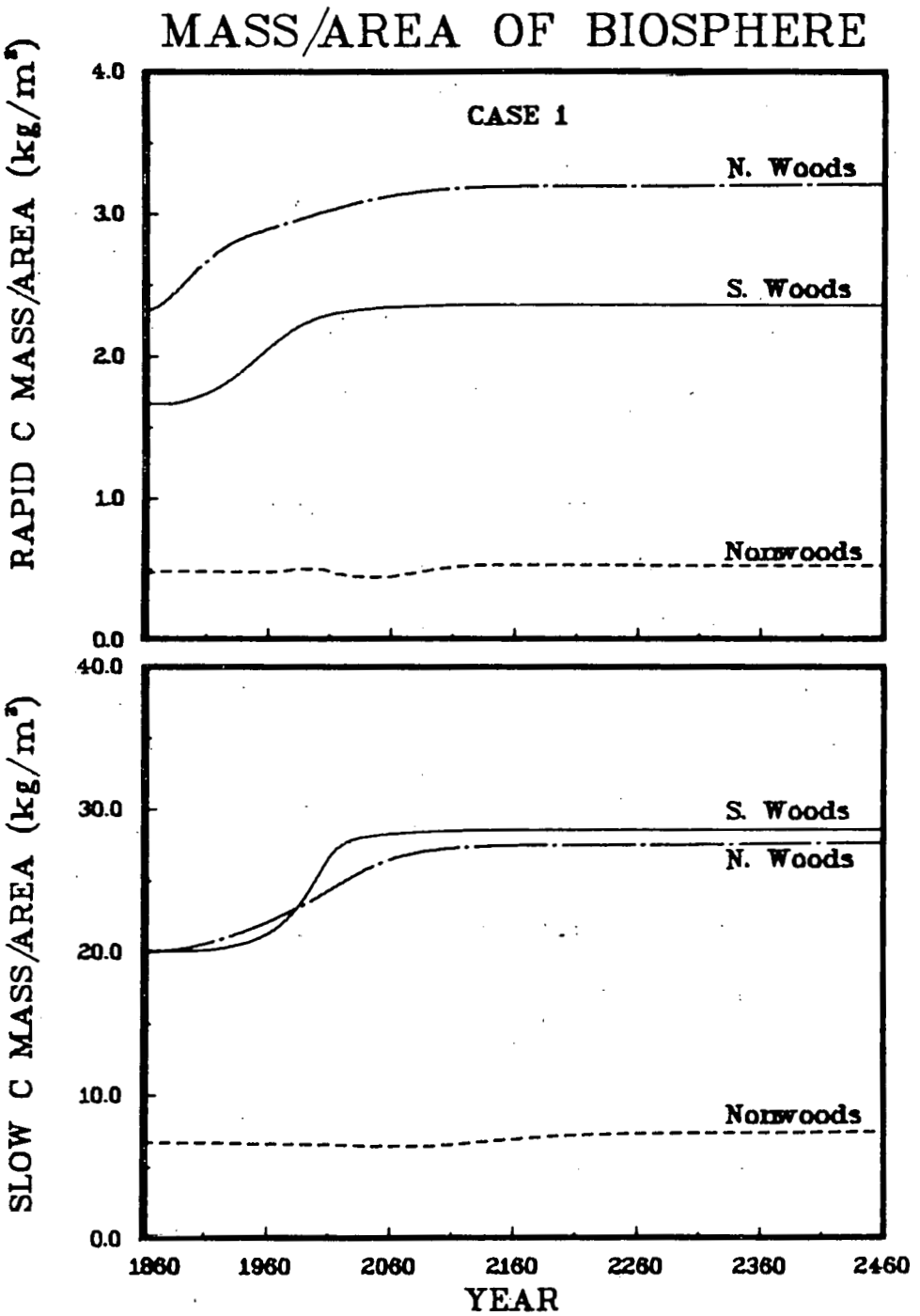

Figure 12. Predicted mass per unit area of biospheric carbon ("nominal" scenario).

densities of the Northern Woods and the Southern Woods rise to 3.3 kg/m^2 and 2.4 kg/m^2, respectively, in the present model.

The behavior of the densities of slowly exchanging carbon pools (lower frame of Figure 12) is quite similar as those of the rapid carbon pools. In the Northern Woods the density increases slowly until around 2600 A.D. and then levels off at a value of 27.7 kg/m^2. The density of the Southern Woods remains at the same level for the first 100 years, follows by a sharp increase to around 2000 A.D. and remains level at about 28.6 kg/m^2 for the remainder of the simulation. Again, this may reflect assumed regrowth and storage in those forest areas which are allowed to regrow instead of being converted to nonforest.

There are only slight increases in the densities for the rapid and slow carbon pools of the Nonwoods (Figure 12). The rapid carbon pool only stores a maximum of 0.75 kg/m^2 of carbon.

In the "slow burner" scenario (Figure 13), the carbon densities of both rapid and slow carbon pools from the wood compartments are almost the same as those obtained from the previous case. However, values of mass/area of both carbon pools from the Nonwoods are slightly below the results reached in the "nominal" case. These observations are anticipated as the rates of NPP for the two wood compartments have almost the same values from both scenarios. Moreover, in the "slow burner" case, both rapid and slow carbon pools of the Nonwoods have a slightly lower rate of NPP as those from the "nominal" case, and the modest benefit of CO_2 fertilization is even less effective.

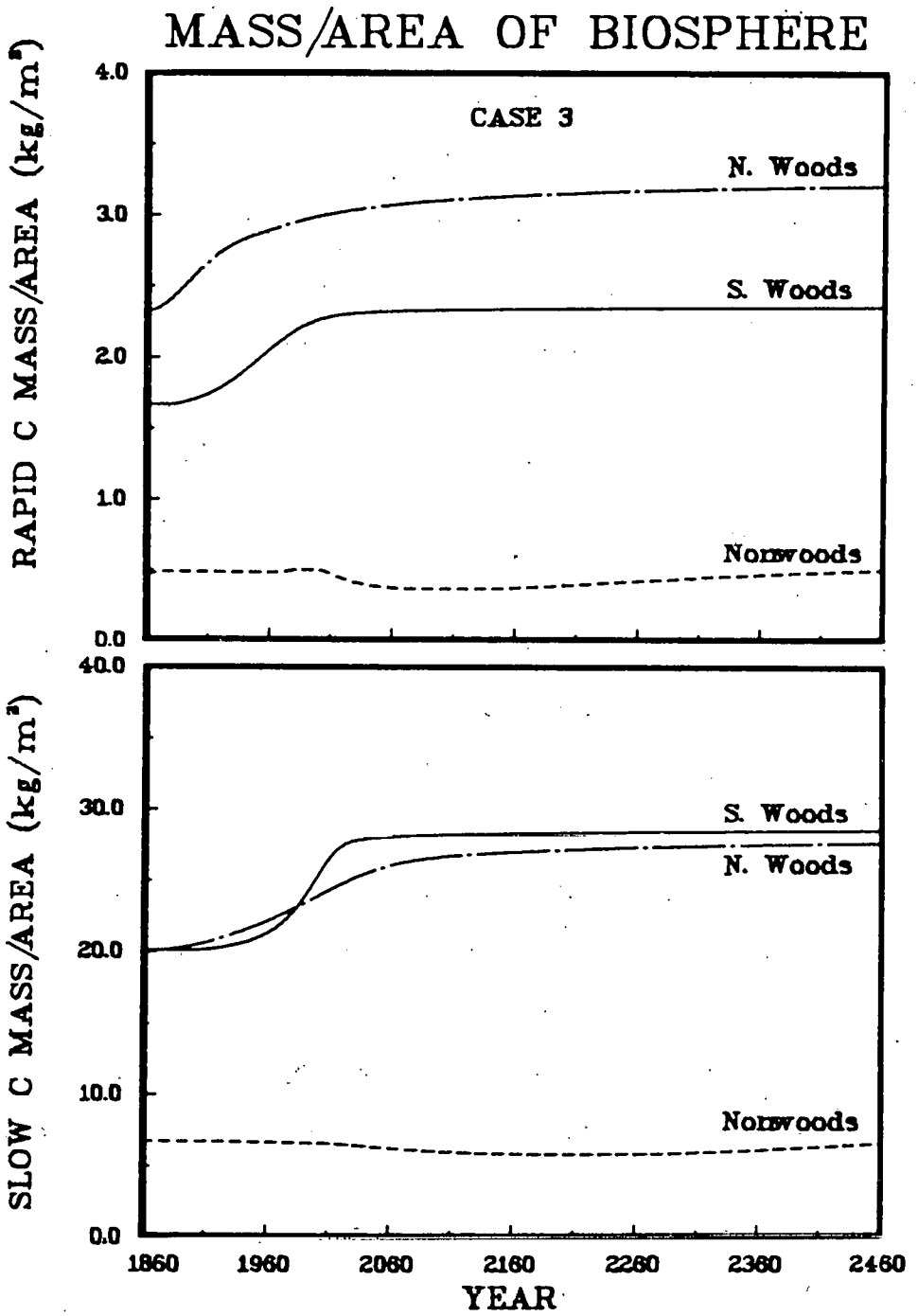

Figure 13. Predicted mass per unit area of biospheric carbon ("slow burner" scenario).

4.3.7. Projections of Biospheric Carbon

The level of biospheric carbon in a terrestrial compartment depends both on the area and the density of carbon in the mixture of ecosystems. As mentioned in the previous section, the carbon densities of the wood compartments obtained from the "nominal" and the "slow burner" scenarios are nearly identical. The carbon levels of similar compartments obtained in these two scenarios also follow nearly the same courses. The following discussion on the projections of carbon level in the wood compartments are therefore pertinent to both scenarios (Figures 14 and 15).

In the rapid carbon pool, the level of carbon in the Northern Woods shows a slight decrease initially for about 10 years and then a rapid increase to 70.2 Gtons in 1947. After this maximum, a long and slow decrease occurs which stabilizes at about 65.3 Gtons.

The amplitude of carbon estimates for the Southern Woods is larger than that of the Northern Woods. It also decreases slightly for the first 10 years similar to the mass/area curves. The carbon level then increases slowly to 51.8 Gtons in 1954, follows by a rapid drop to 43.8 Gtons around 2000 A.D. Thereafter, the level stabilizes at about 35.5 Gtons.

The general trend of the rapid carbon pool in the Nonwoods is about the same as the wood compartments except that the level shows an increase from clearing since the beginning of the simulation. In the "nominal" run (Figure 14), there is a peak of about 43.8 Gtons around 2000. The

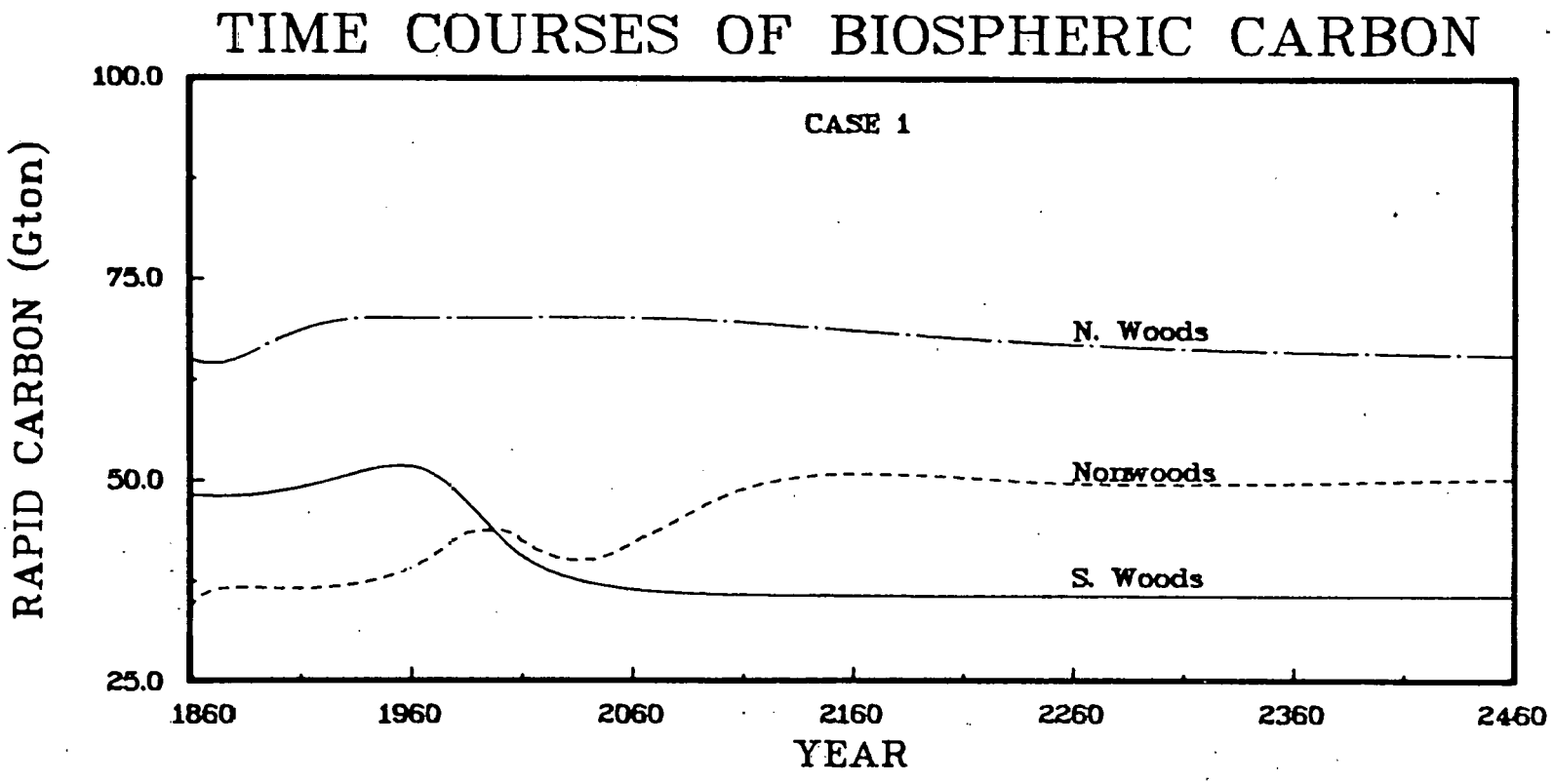

Figure 14. Projections of rapidly exchanging carbon in terrestrial ecosystems ("nominal" scenario).

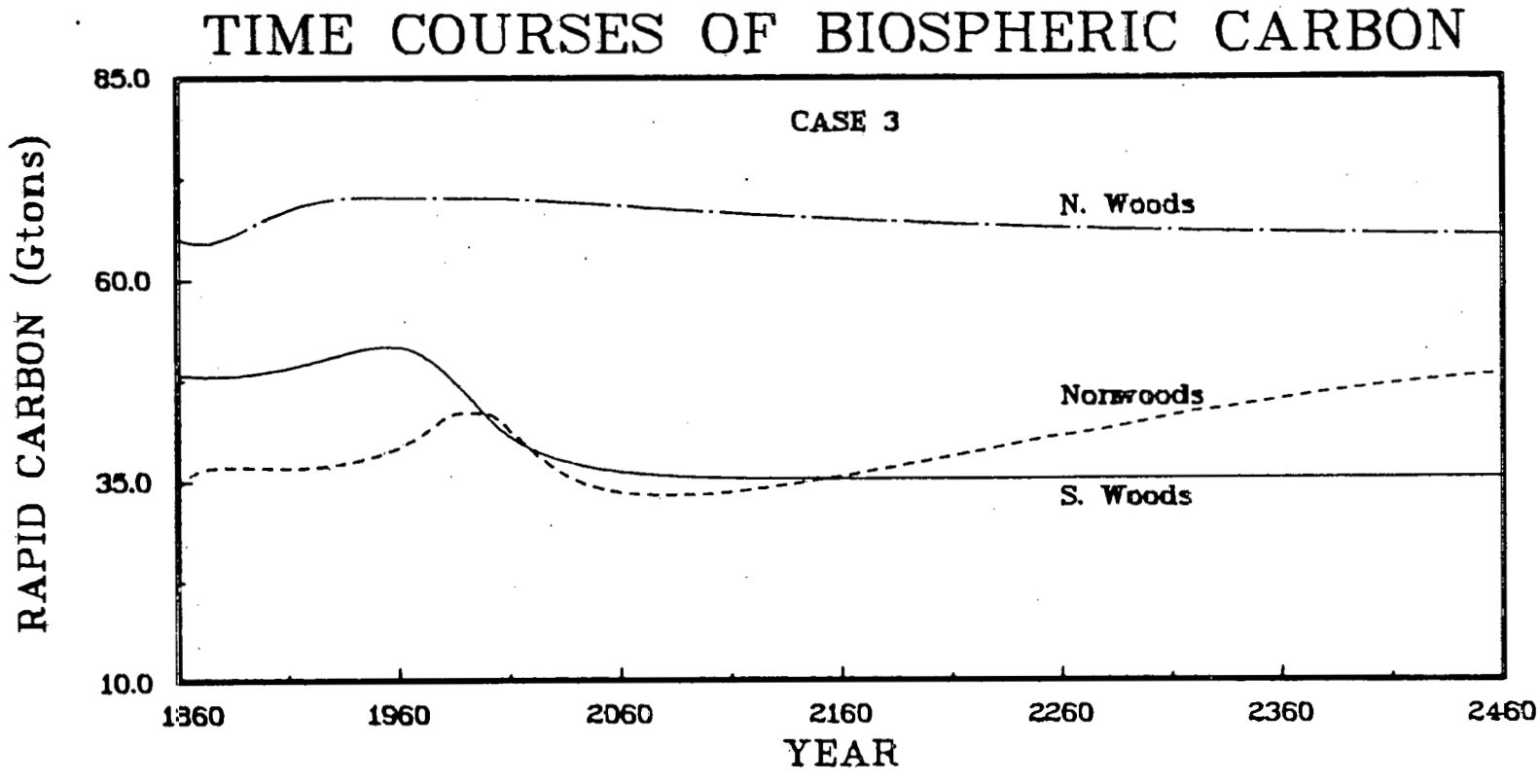

Figure 15. Projections of rapidly exchanging carbon in terrestrial ecosystems ("slow burner" scenario).

curve then decreases to a low value of 35 Gtons in 2035, because the low input of carbon from deforestation no longer exceeds the rapid losses of the carbon residues which are assumed converted from slow to rapid pools. As the high concentration of atmospheric carbon begins to dominate after this time, the carbon content in the Nonwoods rises again and stays at a level of about 50.9 Gtons for the subsequent years.

For the "slow burner" scenario (Figure 15), the projection of carbon in the Nonwoods shows a minimum of about 33.1 Gtons around 2080. It increases again slowly to about 48.3 Gtons in the later time history of the simulation.

There is a similar situation in the slow carbon pools of the wood compartments where results produced from both scenarios are almost identical (Figures 16 and 17). The carbon level in the Northern Woods decreases to 530 Gtons gradually from the start to around 1925 because of deforestation. As the NPP rises slowly, the carbon in this compartment also increases to 598 Gtons around 2085. The level declines in subsequent years to about 563 Gtons.

A different time behavior is observed in the level of slow carbon pool of the Southern Woods (Figures 16 and 17). The level decreases gradually in the beginning and then accelerates after 1900 due to the increase clearing of forests. After about 2060 A.D., the projected level stays at about 429 Gtons of carbon.

Since some of the carbon in both wood compartments is shifted to both the rapid and slow carbon pools of the Nonwoods, a general increase in the levels of rapid and slow carbon pools in the Nonwoods are

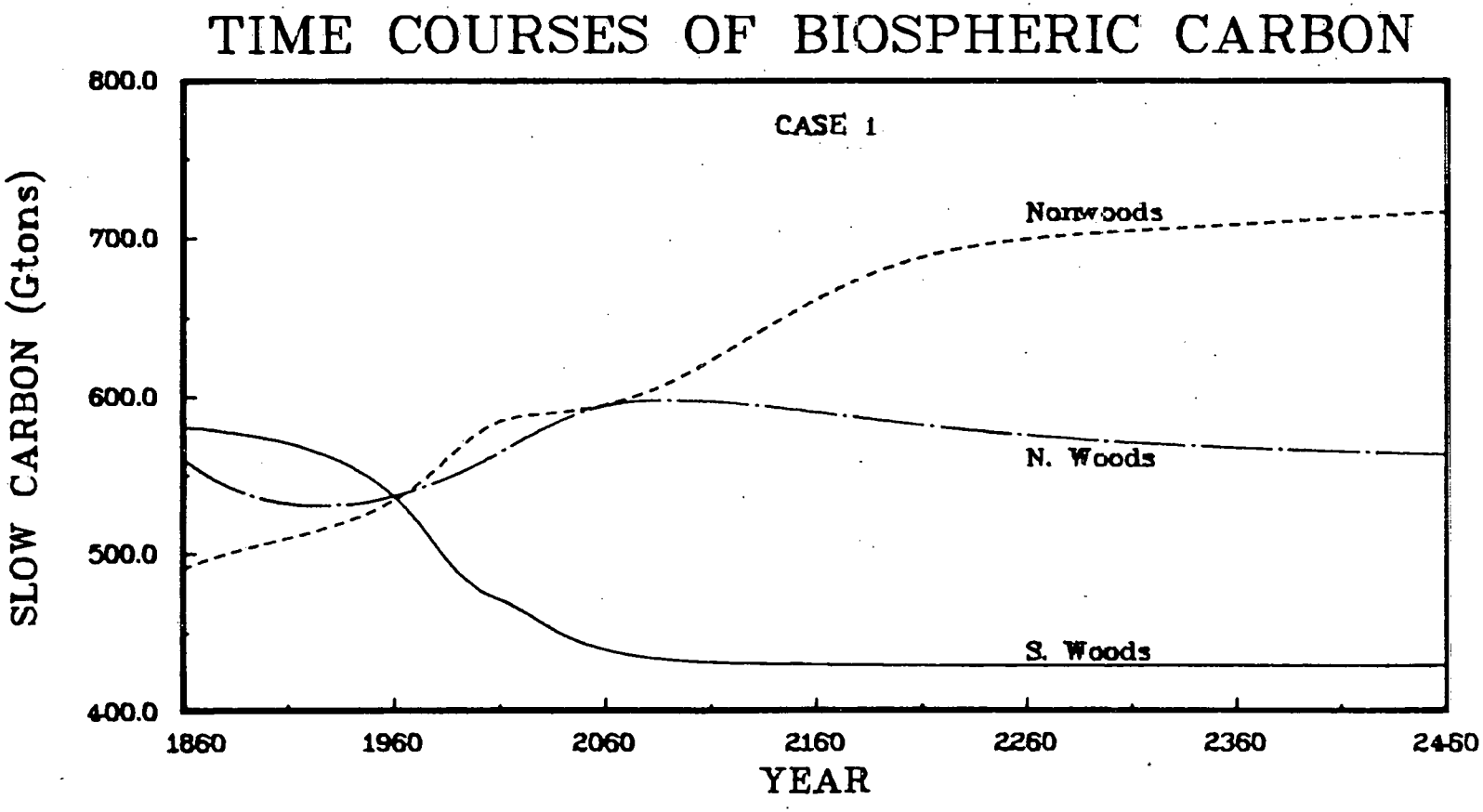

Figure 16. Time courses of slowly exchanging carbon in terrestrial ecosystems ("nominal" scenario).

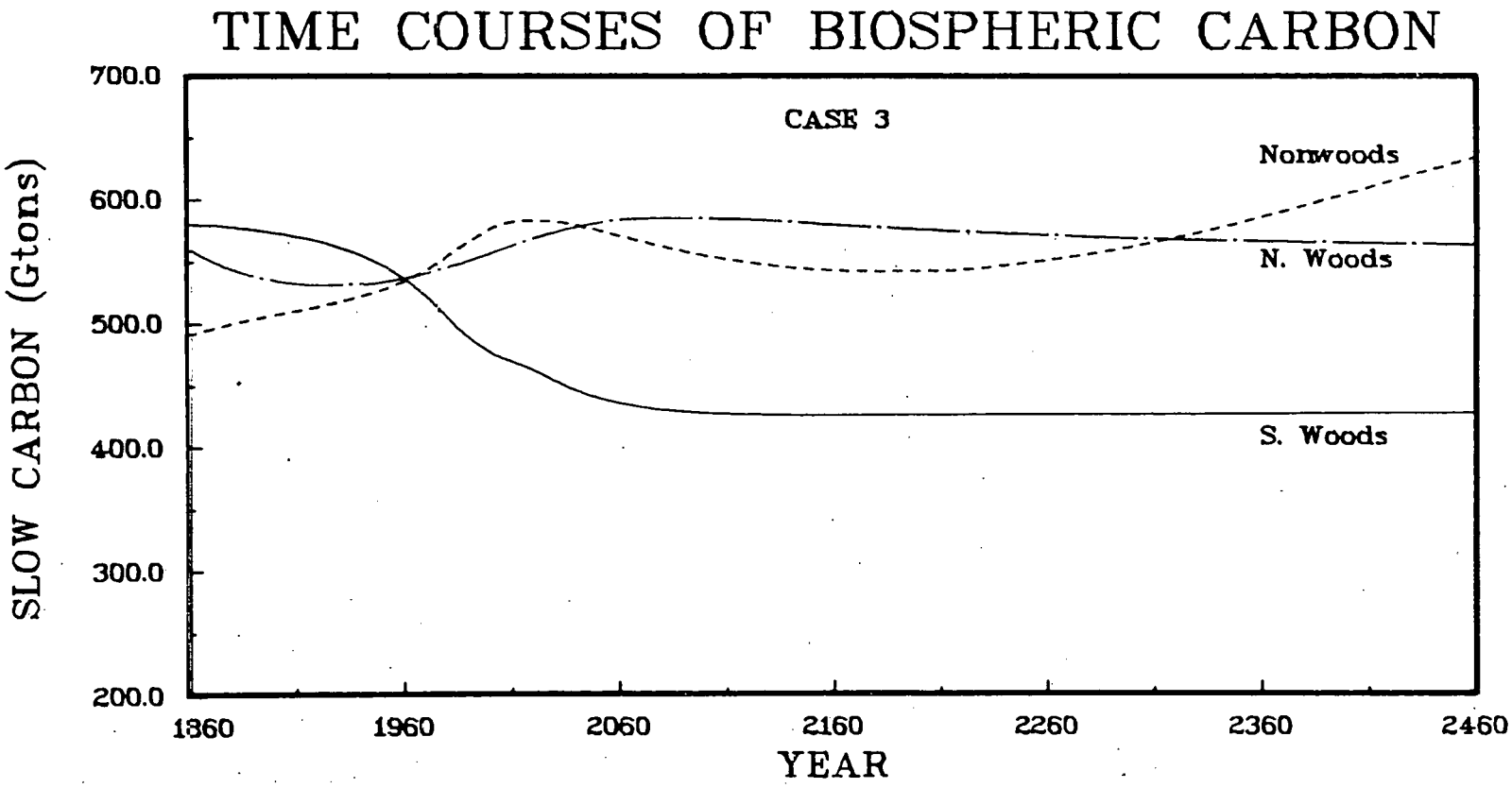

Figure 17. Time courses of slowly exchanging carbon in terrestrial ecosystems ("slow burner" scenario).

observed. In the "nominal" run (Figure 16), the increase is completed in several steps. The first step, between 1860 and 2010, shows a slow increase. The second step is observed between 2010 and 2060, when the carbon level remains about the same. This period corresponds to the possible declining rate of deforestation at that time. Subsequently, the rate of NPP is affected more strongly by the increasing concentration of atmospheric carbon, and the projected carbon level in this compartment shows a slow rise to a value of 717 Gtons.

In the "slow burner" case (Figure 17), the slow increasing of atmospheric carbon projected in the future is sufficient to slow down the increase of carbon level in the Nonwoods. The curve shows a slow decrease from around 2020 to 2180; after that the projection increases again to 634 Gtons in the year 2460.

4.3.8. Partitioning of Excess Carbon

Table 6 gives the estimated fractions of the excess carbon including carbon released promptly or slowly from fossil fuel and the biosphere as it is distributed among compartments of the atmosphere, the marine, and the terrestrial biosphere. Since the biospheric carbon from delayed release also was used in the calculation, these numbers are likely underestimates. Partitioning of excess carbon to the sediment compartments and other minor compartments is not included, and these presumably cover the remainder of the excess carbon.

In 1950, about 35% of the annual release of anthropogenic carbon remained in the atmosphere. The percentage is projected to increase

Table 6. Partitioning of Excess Carbon (as fraction of the total excesses)

Compartment	1950	1960	1970	1975	2000
			Annual Fractions		
Atmosphere	0.35	0.45	0.51	0.53	0.61
Marine	0.12	0.13	0.13	0.13	0.15
Biosphere	0.009	-0.007	-0.015	-0.016	0.03
			Cumulative Fractions		
Atmosphere	0.36	0.37	0.40	0.41	0.49
Marine	0.11	0.12	0.12	0.12	0.13
Biosphere	-0.066	-0.052	-0.043	-0.042	-0.016
Biosphere + releases	0.50	0.49	0.46	0.44	0.37

gradually to 60% in the year 2000. The marine compartments are estimated to absorb a total of about 12% of the annual excess between 1860 and 1970. The percentage was about half that estimated by Revelle and Munk (1977). Currently, the terrestrial biosphere is responsible for a net release of carbon to the atmosphere. As the NPP of the biota increases, the terrestrial ecosystems absorb more carbon dioxide and retain a total of about 3% of the excess CO_2 released in 2000 A.D.

Table 6 also shows the partitioning of the excess CO_2 as fractions of the total cumulative releases. About half of the cumulative releases of carbon would be retained in the atmosphere by the year 2000. At the same time, the marine compartments absorb a total of about 13% of the total excess. In the biosphere, the pools of slowly exchanging carbon continue to release carbon. It follows that the terrestrial ecosystems would contribute continuously a total of 1.6% of the cumulative excesses to 2000 A.D. Total partitioning of excess carbon to the terrestrial biosphere plus the releases, prompt and delayed, of biospheric carbon (cf. Revelle and Munk, 1977), also are given in Table 6.

CHAPTER 5

MODEL EVALUATION

5.1. SENSITIVITY ANALYSES

Many details and interactions of components of the global carbon cycle are not yet thoroughly known. Furthermore, wide ranges of values have been reported for the pool sizes and fluxes of the compartments. This condition is not unexpected, if one considers the limitations of the measuring techniques and the large amount of variability that are characteristic of natural systems.

One way to assess the usefulness of the model is to find out how sensitive the responses of the components are under parameter perturbations. However, sensitivity analysis (Tomović, 1964) of complex nonlinear system with many discontinuities, such as in the present model, is cumbersome. The practical and direct method used in this study was to observe the response of atmospheric carbon to systematic variations in the coefficients, initial conditions, or forcing functions of the model. A tremendous number of computer runs would have to be performed if all the parameters and their combinations were analyzed. Of these runs, some may not yield significant information. Hence, only a few of the parameters whose values are considered especially critical were tested in the present sensitivity analysis.

5.1.1. Initial Value of Atmospheric Carbon

The first parameters to be investigated was the initial value of the atmospheric compartment. Two other sets of values were used in subsequent simulation runs to compare the responses of the atmospheric carbon with that from the "nominal" case. One of these values was 616 Gtons used by Keeling (1973a). The model was simulated with the following changes: c_1^* = 523.6 Gtons, c_2^* = 92.4 Gtons, and $F_{1,2}^*$ = $F_{2,1}^*$ = 23 Gtons/year. Stuiver (1978) and Wilson (1978), supported by evidence from the $^{13}C/^{12}C$ ratios of dated tree rings, suggested that the preindustrial level of atmospheric carbon might be near 270 ppmv, or 576.7 Gtons. Consequently, another simulation run was carried out with c_1^* = 490.2 Gtons, c_2^* = 86.5 Gtons, and $F_{1,2}^*$ = $F_{2,1}^*$ = 21.6 Gtons/year.

Figure 18 shows the changes of atmospheric carbon with different starting values. These changes do not alter the timing or the characteristics of the time trajectories of atmospheric carbon. Only the magnitudes of these curves are altered; insignificantly in the early period but with increasing differences at the end. The trajectory with the lowest starting value increases to 837 Gtons in 2000 A.D., and reaches a peak value of 6210 Gtons around 2275 A.D. If the initial level of atmospheric carbon were taken to be 616 Gtons, it rises to 882 Gtons in 2000 A.D. and peaks at the same time as the others with a value of 6289 Gtons.

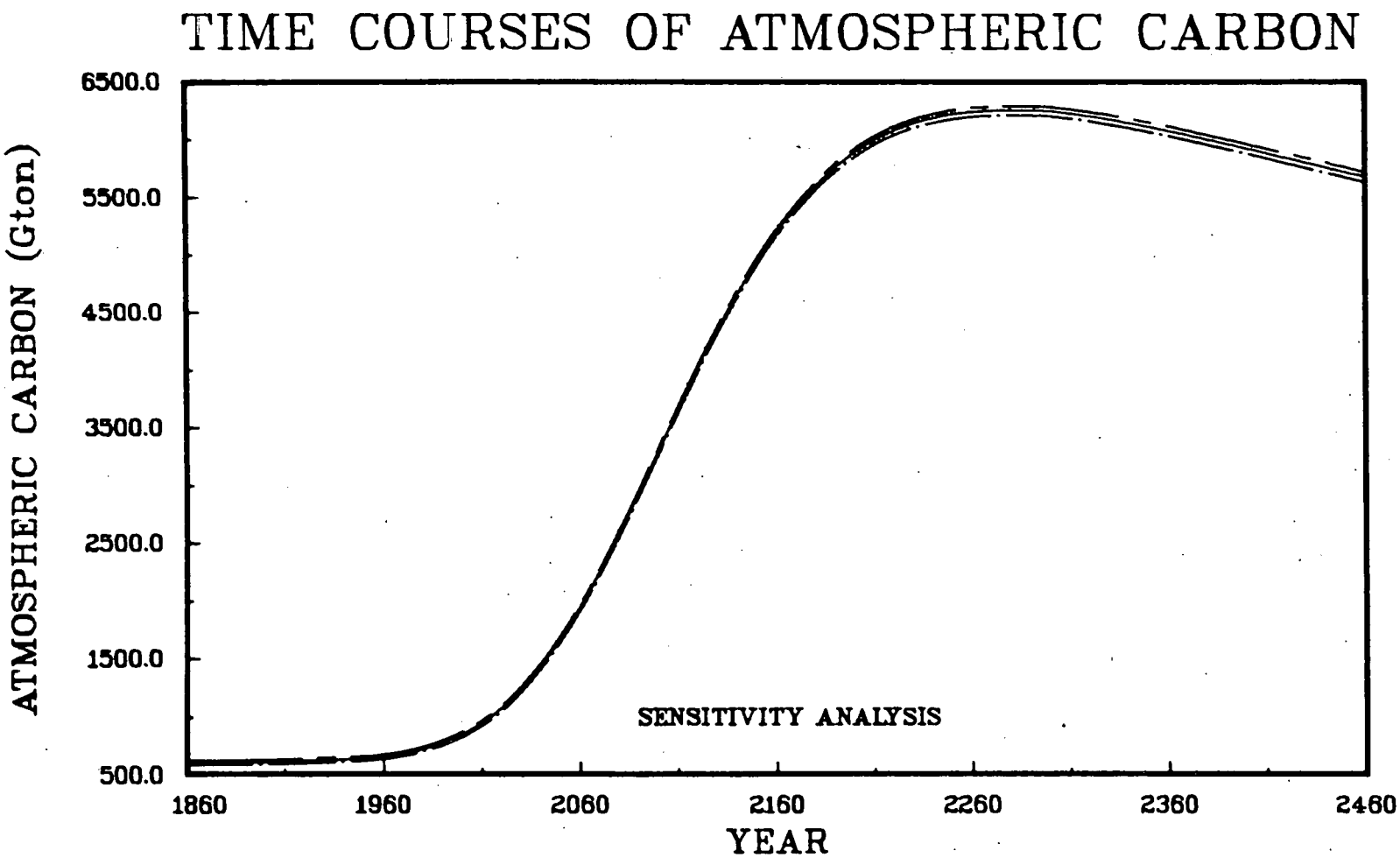

Figure 18. Responses of atmospheric carbon with different initial pool sizes in the atmosphere. Solid curve: nominal values. Upper curve: 616 Gtons C. Bottom curve: 576.7 Gtons of carbon.

5.1.2. Biotic Growth Factor

Some of the constants employed in the equations to estimate the net primary production of biospheric compartments are subjected to large variation. The biotic growth factor is one of the coefficients whose impact has been studied by several workers. This factor is usually assumed to range from 0.0 to 0.4 (Keeling, 1973a; Oeschger et al., 1975) for large areas.

In the present sensitivity analyses, the three biotic growth factors for the Northern Woods, Southern Woods, and Nonwoods compartments are increased or decreased by 10% at the same time. As found by Olson et al. (1978), the level of atmospheric carbon was relatively insensitive to small changes in the growth factors, but the concentration of atmospheric carbon does decrease with a higher value of β, and vice versa (Figure 19). When the growth factors are decreased by 10%, the projection of atmospheric carbon at 2000 A.D. is 865 Gtons, and peaks in 2275 A.D. at a value of 6297 Gtons. If the growth factors are increased by 10%, the levels of atmospheric carbon in 2000 and 2275 A.D. are 858 Gtons and 6215 Gtons, respectively.

5.1.3. Maximum Mass/Area

The other parameter in the NPP equations to be analyzed is the maximum density of carbon that can be attained in each of the biospheric compartments. In the simulations, the probable maximum density of each biospheric compartments is not supported by actual observation.

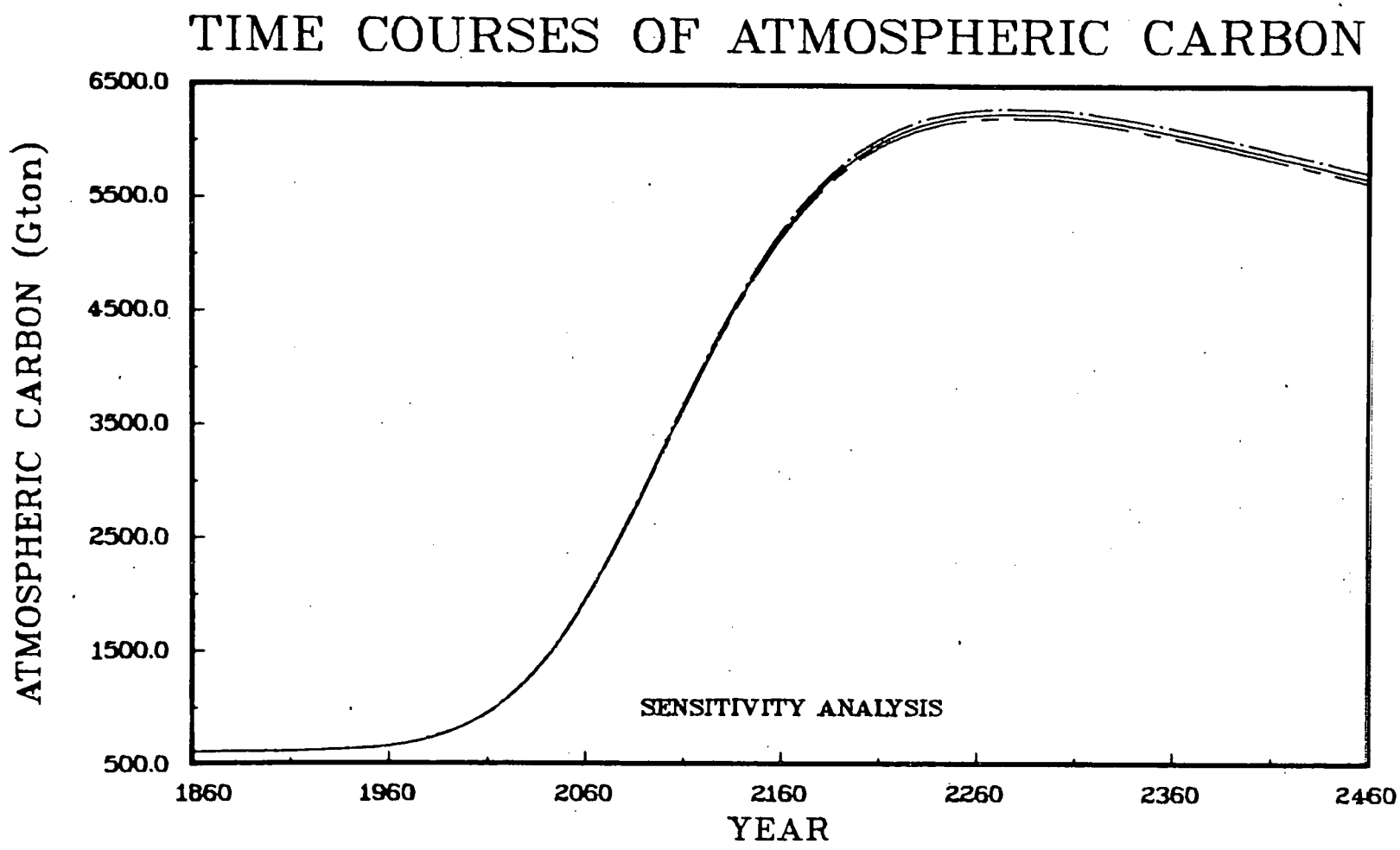

Figure 19. Sensitivity of atmospheric carbon level to different values of the biotic growth factors. Solid curve: nominal values. Upper curve: 90% of nominal values. Bottom curve: 110% of nominal values.

Tentative values assumed in the "nominal" run are 1.5 times the preindustrial estimates. This factor is increased or decreased by 10% (i.e., 1.65 or 1.35 times the preindustrial estimates) in the sensitivity analyses.

As in the previous test, the basic pattern of the atmospheric responses is similar but is slightly more sensitive (Figure 20). A slightly higher level of atmospheric carbon is observed when the model specified lower values of the maximum of mass density. The projected level of atmospheric carbon in 2000 A.D. is 883 Gtons, rising to a peak of 6345 Gtons around 2275 A.D. When the maximum densities are set at 1.65 times the initial estimates, atmospheric carbon decreases to 851 Gtons in 2000 A.D. and peaks at a value of only 6164 Gtons in 2275 A.D.

5.1.4. Flux of Polar Waters

One component of the total flux that transports about 36 Gtons/year of carbon from the mixed surface layer to the deeper waters passes through the thermocline. The other component occurs in the polar regions where cold polar waters sink directly to the deep waters of the warm oceans. Estimates of the latter flux used in the "nominal" run may be near an upper bound. Consequently, only the response to a smaller down flux of polar waters was tested. The magnitudes of fluxes from the mixed layer were assumed proportional to the ratio of the sea surface areas where the two fluxes occur. Hence, the changes were: $F_{7,8}^{*}$ = 30 Gtons/year, $F_{7,9}^{*}$ = 6 Gtons/year and $F_{9,8}^{*}$ = 47 Gtons/year. The last change was necessary for mass-balancing the system.

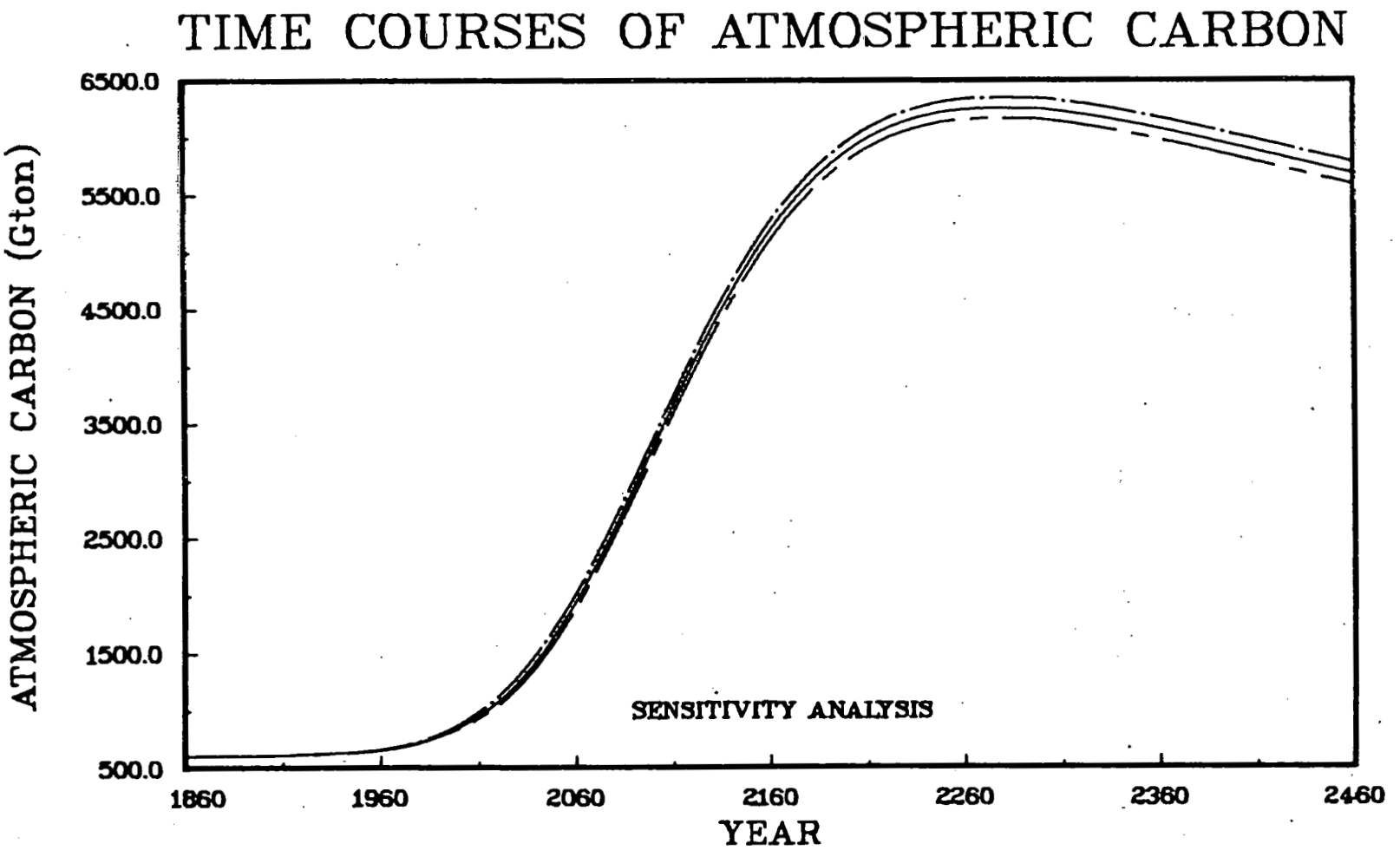

Figure 20. Effect of increasing and decreasing the maximum values of mass/area on the simulated level of atmospheric carbon. Solid curve: 1.5 times preindustrial values. Upper curve: 1.35 times preindustrial values. Bottom curve: 1.65 times preindustrial values.

The results (Figure 21) show that the carbon level in the atmosphere is relatively insensitive to variations in these fluxes in the early time period of the simulation. The sensitivity increases at the end, when the atmospheric carbon increases from 863 Gtons in 2000 A.D. to a maximum of 6390 Gtons in 2285 A.D. The peaking time of the atmospheric carbon is delayed for about 10 years as compared to the "nominal" run.

Differing viewpoints were expressed by Dugas (1968) and Broecker et al. (1970) regarding this direct transfer of atmospheric carbon to the deep ocean. Dugas' (1968) results showed that the concentration of atmospheric carbon was not critically affected by the inclusion of this transfer in her model. Yet Broecker et al. (1970) maintained that this particular transfer played an important role in moving extra carbon dioxide from the atmosphere to the deeper ocean. The sensitivity analysis given here suggests that the addition of such a flux in the model is very significant when the atmospheric carbon reaches a high level.

5.2. VERIFICATION WITH HISTORICAL DATA

Measurements of carbon level in the atmosphere at selected localities can be used to validate the responses of the model. The observed concentrations of atmospheric carbon at Mauna Loa since 1958 (Keeling et al., 1976b), together with the estimates obtained from the present model, are listed in Table 7. Although the data cover only 17 years, they can still be used as a yardstick for some future refinements in certain aspects of the model.

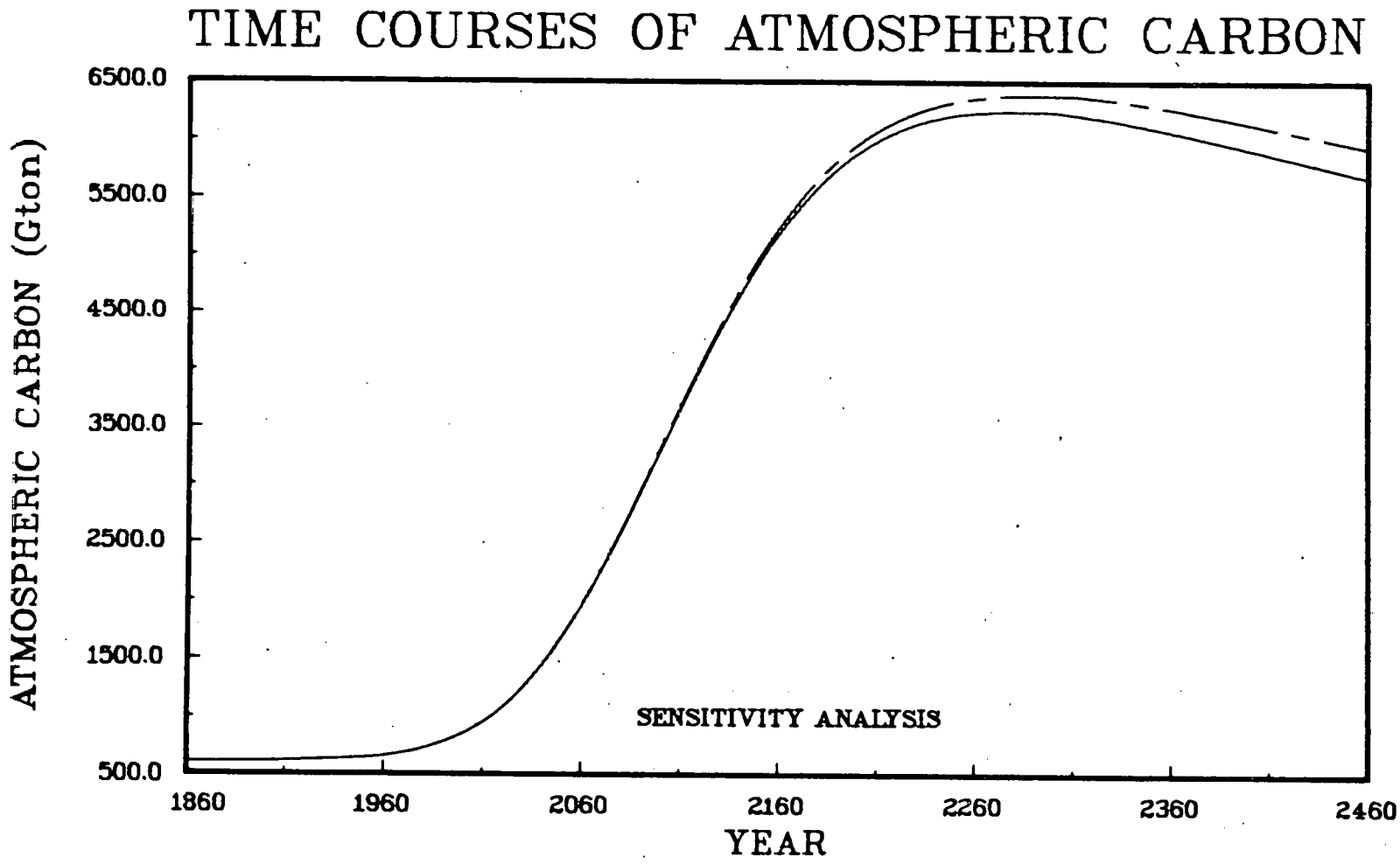

Figure 21. Effect on simulated level of atmospheric carbon of lowering the flux of polar waters. Solid curve: nominal values. Upper curve: $F_{7,8}^{*}$ = 30 Gtons/year, $F_{7,9}^{*}$ = 6 Gtons/year, $F_{9,8}^{*}$ = 47 Gtons/year.

Table 7. Concentration of Carbon in Air Observed at Mauna Loa, Estimated and Simulated for Atmosphere

	Keeling et al. (1976)		This Report
Year	ppm (mol)[a]	Gton C	Gton C
1958	(315.49)	669.72	662.25
1959	316.14	671.09	664.19
1960	317.03	672.98	666.27
1961	317.71	674.42	668.39
1962	318.57	676.25	670.54
1963	319.03	677.22	672.83
1964	319.63	678.50	675.28
1965	320.25	679.81	677.86
1966	320.92	681.24	680.57
1967	321.70	682.89	683.38
1968	322.58	684.76	686.31
1969	324.47	688.77	689.43
1970	325.79	691.57	692.77
1971	326.83	693.78	696.28
1972	(328.01)	696.29	699.93
1973	(330.15)	700.84	703.75
1974	(330.76)	702.13	707.72

[a]Estimates in brackets are taken from Zimen et al. (1977).

Estimates obtained from the present model show a steeper slope during 1958 to 1974 than the observed data. The largest error, which occurred in 1958, is about 7.47 Gtons, or about 1.12% less than the Mauna Loa measurement. The difference may be explained by the net release of a substantial amount of organic carbon from deforestation during or immediately before this period. According to Wilson (1978) and Stuiver (1978), the net contribution of biospheric carbon to the atmosphere may already have tapered off by 1900 A.D., whereas the present model still assumes a large biospheric release during this period and which reaches a maximum release between 1980 and 2000.

According to Keeling and Bacastow (1977), the average airborne fraction of fossil CO_2 between 1959 and 1973 inclusive was about 56%. The estimate for the similar fraction in the present study was about 75% for the same period. This higher value evidently reflects the steeper gradient of the time history mentioned above. The possibilities of relatively more deforestation or other nonfossil release early in the last century, and less during these critical years of empirical monitoring, therefore deserve closer study in the continuation of the present work.

CHAPTER 6

IMPLICATIONS OF ALTERNATIVE SCENARIOS

It is expected that changes in the global carbon cycle and CO_2 concentrations will become a dominant ecological and societal problem in the near future if mankind continues the upward trend of fossil-fuel utilization. Unless recent models of carbon cycles and climatic change are both missing key points, the climate will change significantly; this in turn may affect human life support systems. The urgency of the carbon dioxide/climate problem is directly linked to the combustion of fossil fuel and therefore calls for caution regarding future policies of using the fossil carbon resources in the developed nations as well as in the less developed countries (cf. Rotty and Weinberg, 1977; Niehaus, 1976).

Generally, decision makers are reluctant to formulate any vital policy which is based on analyses with so many uncertainties (Spofford, 1971). However, the choice has to be made soon in view of the occurrence of long-lasting effects associated with many of the anticipated impacts. Technical bases underlying several alternative energy systems have to be made before the end of this century if society is to have time to anticipate and cope with the main implications of these choices.

6.1. Alternative Scenarios of Ultimate Fuel Use

The sensitivity analyses discussed in Chapter 5 are concerned mainly with the coefficients, initial conditions, etc., within the model for a relatively high release, namely 7500 Gtons of carbon over

∿600 years (peak rate 42.5 Gtons/year in 2101 A.D.). It is interesting to see the response of the atmospheric carbon to different values of the ultimate amount of fossil carbon that would be released to the environment. As noted in Section 3.3.2, a medium-level estimate used by Keeling and Bacastow (1977) and Zimen et al. (1977) was about 5×10^3 Gtons of carbon. However, if most of the presently known sources of fossil carbon were consumed as fuel, then ultimate release could be about 10×10^3 Gtons (cf. Olson et al., 1978). These two estimates were substituted in the modified logistic equation for additional runs. The possibility of releasing lower fractions of the potential carbon resources as fuel, generating CO_2, is discussed below.

Figure 22 shows the responses of atmospheric carbon to this parameter. Similar results have already been discussed by Olson et al. (1978, Fig. 2.2, Table 4.6) and Zimen et al. (1977). With a 5000 Gtons total release of fossil carbon, the future level of atmospheric carbon is tentatively projected to 860 Gtons in A.D. 2000, and peaks at 4159 Gtons in 2225 A.D., half a century earlier than the "nominal" run. The peak value is about seven times the 1860 estimate. In the opposite direction, the very high estimate of the ultimate release also sustains a corresponding higher projection of atmospheric carbon. It reaches about 882 Gtons by the end of this century. The peak level, reached in 2340 A.D., is 8795 Gtons, or about 15 times the preindustrial value.

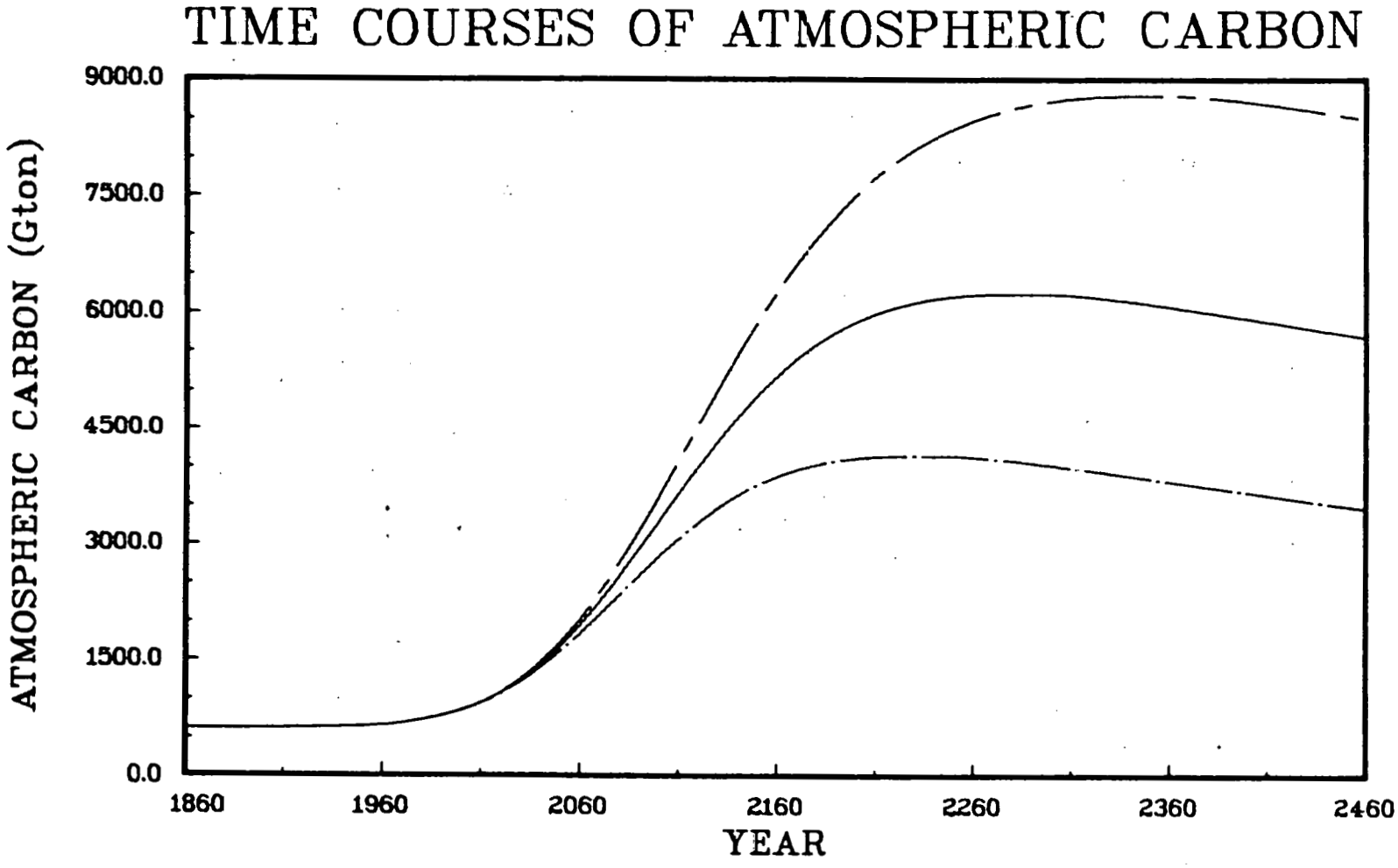

Figure 22. Range of simulated atmospheric carbon using different values of ultimate releases of carbon from fossil fuel. Solid curve: nominal value, 7,500 Gtons. Upper curve: P_∞ = 10,000 Gtons. Bottom curve: P_∞ = 5,000 Gtons.

6.2. RESULTS OF ALTERNATIVE SCENARIOS

As pointed out by Brooks (1977), a continuing, sustained effort, supported by steady public attention and visibility, is required to deal with any long-term environmental problem. Simulation of different scenarios centering on the many possibilities of fossil-fuel consumption and the future management of the ecological systems are essential in supplying the needed information for assessment of societal and ecological impacts. Management policies can be formulated to optimize some chosen objective functions under several important constraints such as maintaining a balanced and diversified pattern of ecosystems (Olson et al., 1978). Alternatively, minimizing the predicted future value of CO_2 and maximizing the gross and net primary production and carbon storage of the biosphere and its uses by man are both important. Besides the two scenarios which have been discussed more fully in Chapter 4, the other scenarios (the "delayed consumption" scenario and the "combination" scenario) are discussed below in a search for possible alternative strategies related to the objective functions just mentioned. Even though actual data may be revised as models improve, the perspective on large or small changes seems likely to be helpful.

If the 7500 Gtons consumption of fossil fuel is retained, but is delayed according to the conditions suggested in the "delayed consumption" scenario, the emission of carbon from fossil fuel would be reduced to 9.3 Gtons/year in the year 2000. The rate would increase to about 65% of the maximum release obtained in the "nominal" case a hundred years

later. This percentage would increase further to 68% sometime around 2121 when the emission rate of the "delayed consumption" scenario would reach its own maximum. After 2175, the annual release of fossil carbon from the "nominal" case would stay below the level released from the "delayed" scenario (Figure 4, page 75).

Values for the NPP and the density of the biota obtained in the "delayed" scenario are slightly behind the corresponding values from the "nominal" case for the first 100 to 150 years, because the "delayed" scenario is assumed to have a lower excess of carbon dioxide in the atmosphere. Later, as more and more fossil carbon is released, values for NPP and mass/area show slight increases over the values obtained in the "nominal" case. These trends are also shown in the carbon pools of the biospheric compartments (Figures 23 and 24).

The response of the atmospheric compartment (Figure 8, page 88) in the "delayed" scenario shows that doubling would occur by 2033, about six years later than the time reached in the "nominal" case. The peak value occurs in 2380, reaching a value of about 5700 Gtons, or slightly more than nine times the preindustrial concentration.

In the "combination" scenario, release of fossil carbon is assumed to follow the same course as that in the "slow burner" scenario (Figure 25). In both cases, a total of about 4270 Gtons are released by the year 2460. Moreover, the "combination" scenario also assumes higher values in the biotic growth factors after 2000, and at the same time, an additional 1% of the slowly exchanging carbon from the wooded ecosystems is assumed to be diverted into long-term storage. The live

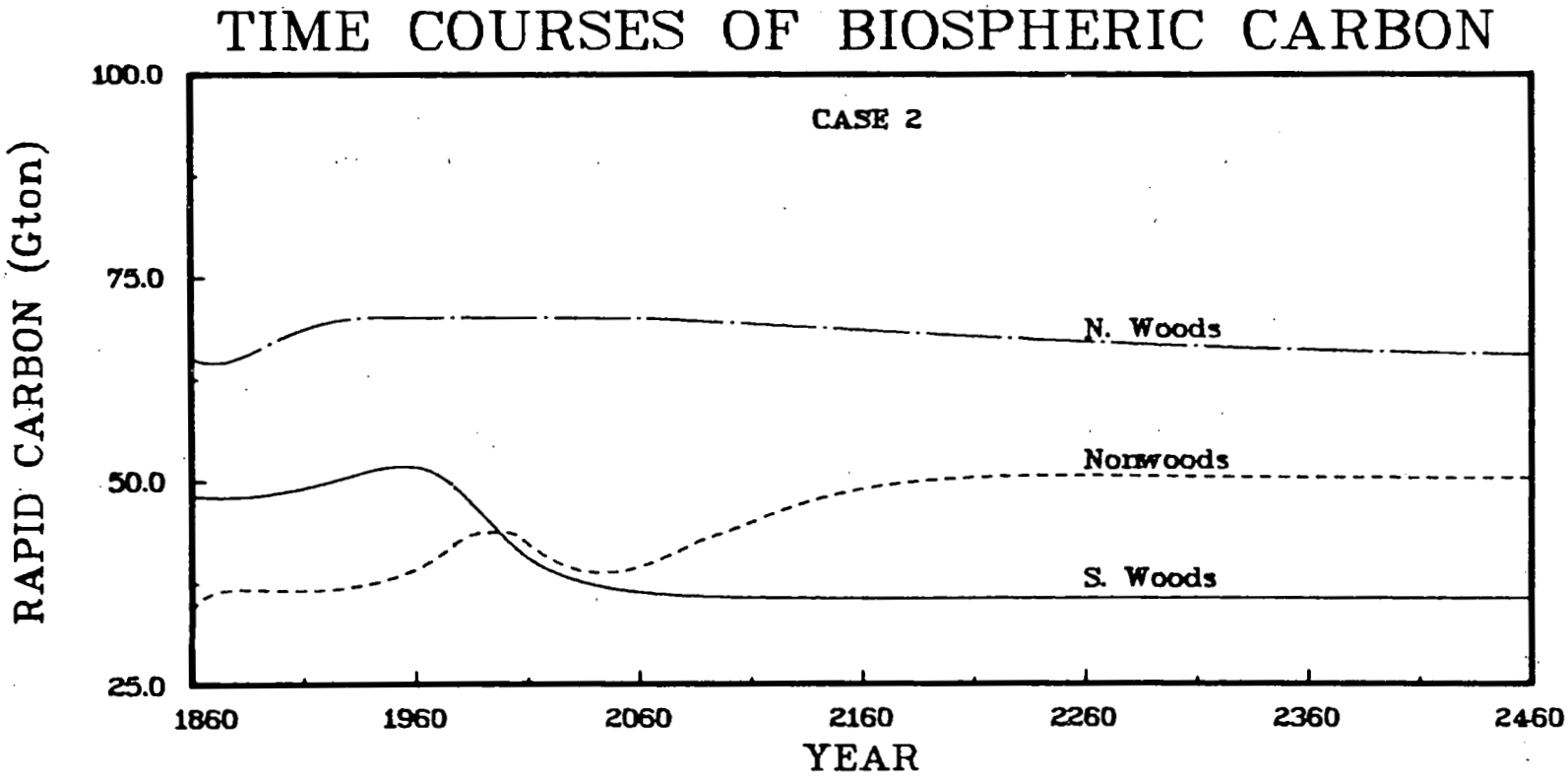

Figure 23. Projections of rapidly exchanging carbon in terrestrial ecosystems ("delayed consumption" scenario).

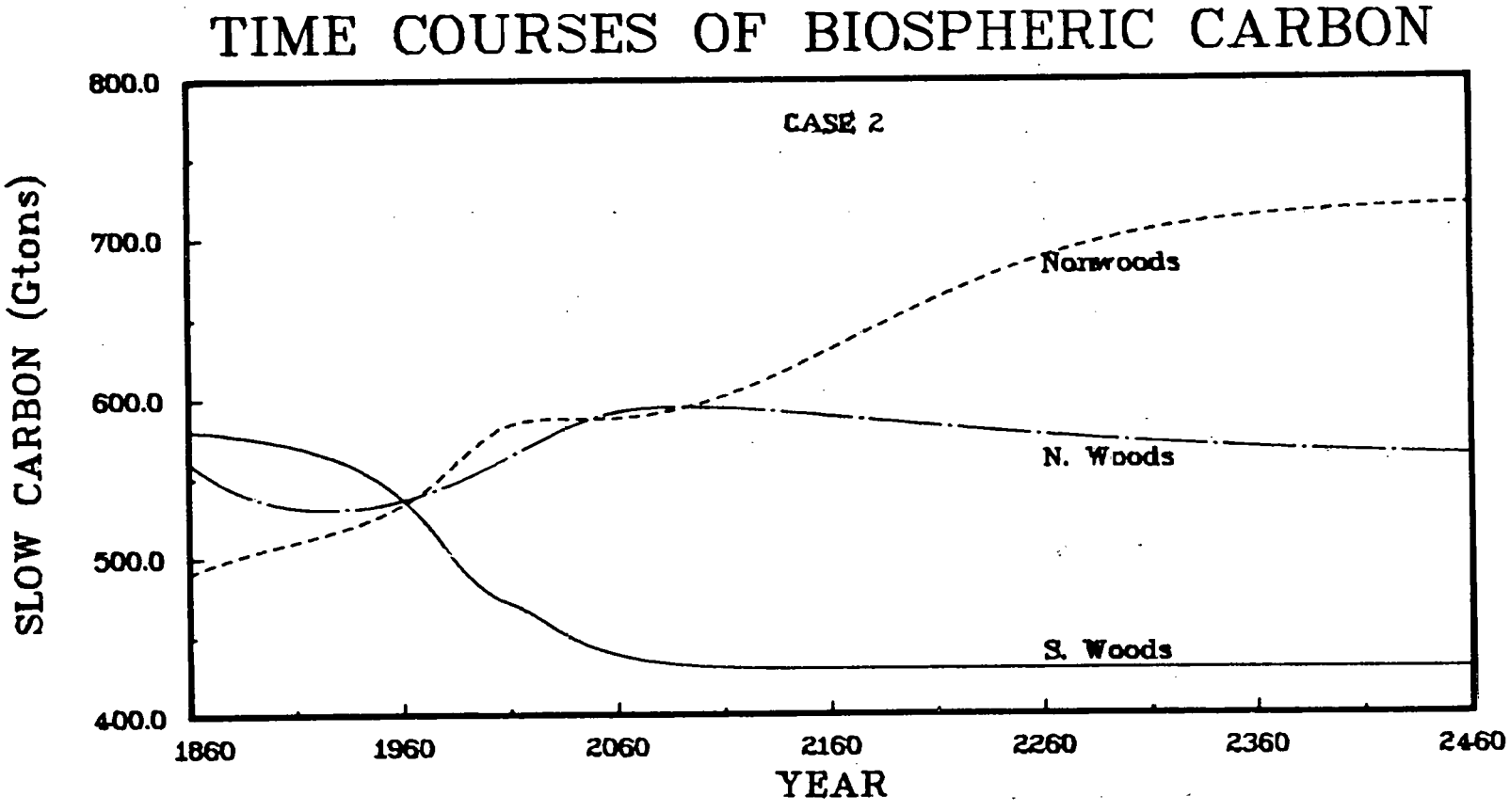

Figure 24. Projections of slowly exchanging carbon in terrestrial ecosystems ("delayed consumption" scenario).

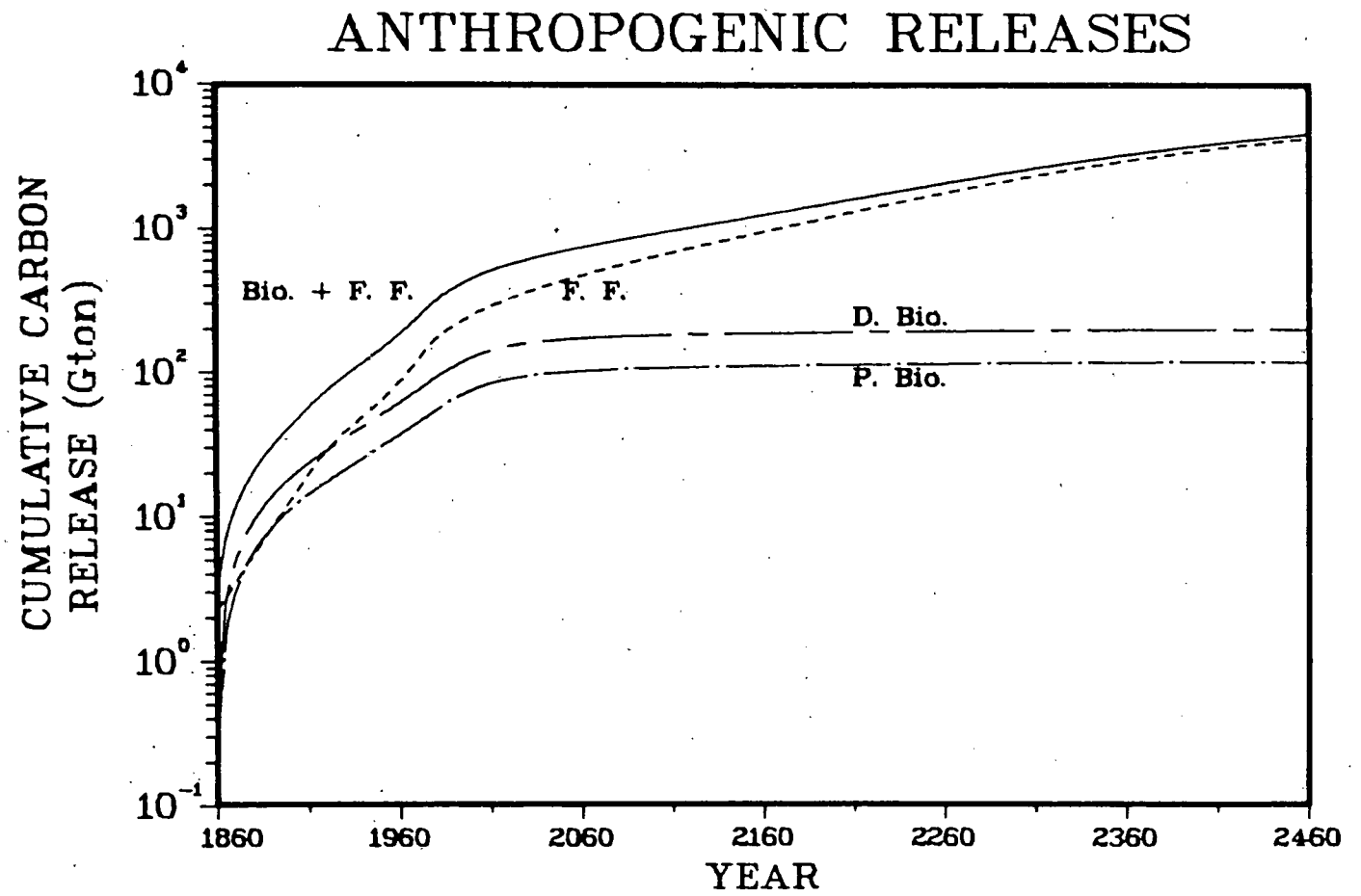

Figure 25. Cumulative release of carbon from deforestation and combustion of fossil fuel ("combination" scenario).

biota shows significant decreases in their mass/area values after 2000 A.D., moreso in the two wood compartments (Figure 26). These observations are also applied to the NPP (Figure 27) except that the slow carbon of the Southern Woods compartment attains a higher NPP in the "combination" case, presumably in response to a subcritical level of atmospheric carbon dioxide and a lower density of carbon in the ecosystems. Most of the excess carbon is now stored in the reactive sediment-deep humus compartment (c_{18}) where the carbon is exchanging very slowly.

Compared with the other scenarios, the amount of rapidly exchanging carbon in the wood compartments in the "combination" scenario (Figure 28) is able to maintain at about the same levels as in the other cases. However, rapid carbon in the Nonwoods compartment decreases to about 60% of that in the "nominal" case in 2100 and only improves to about 70% in 2460. Results obtained from the "combination" scenario (Figure 29) reveal that only the Southern Woods' slow carbon pools are able to store approximately the same amount of carbon as in the "nominal" case, most likely a positive response to the higher NPP. The other two compartments can only store carbon at about 60 to 70% the levels of the "nominal" case. These features essentially suggest implications of using biomass intensively as an alternative energy source to supplement fossil and other energy sources.

If the increased storage of some organic carbon to the deep humus or reactive sediment compartment could be maintained by excellent management on a large scale, the atmospheric carbon concentration might

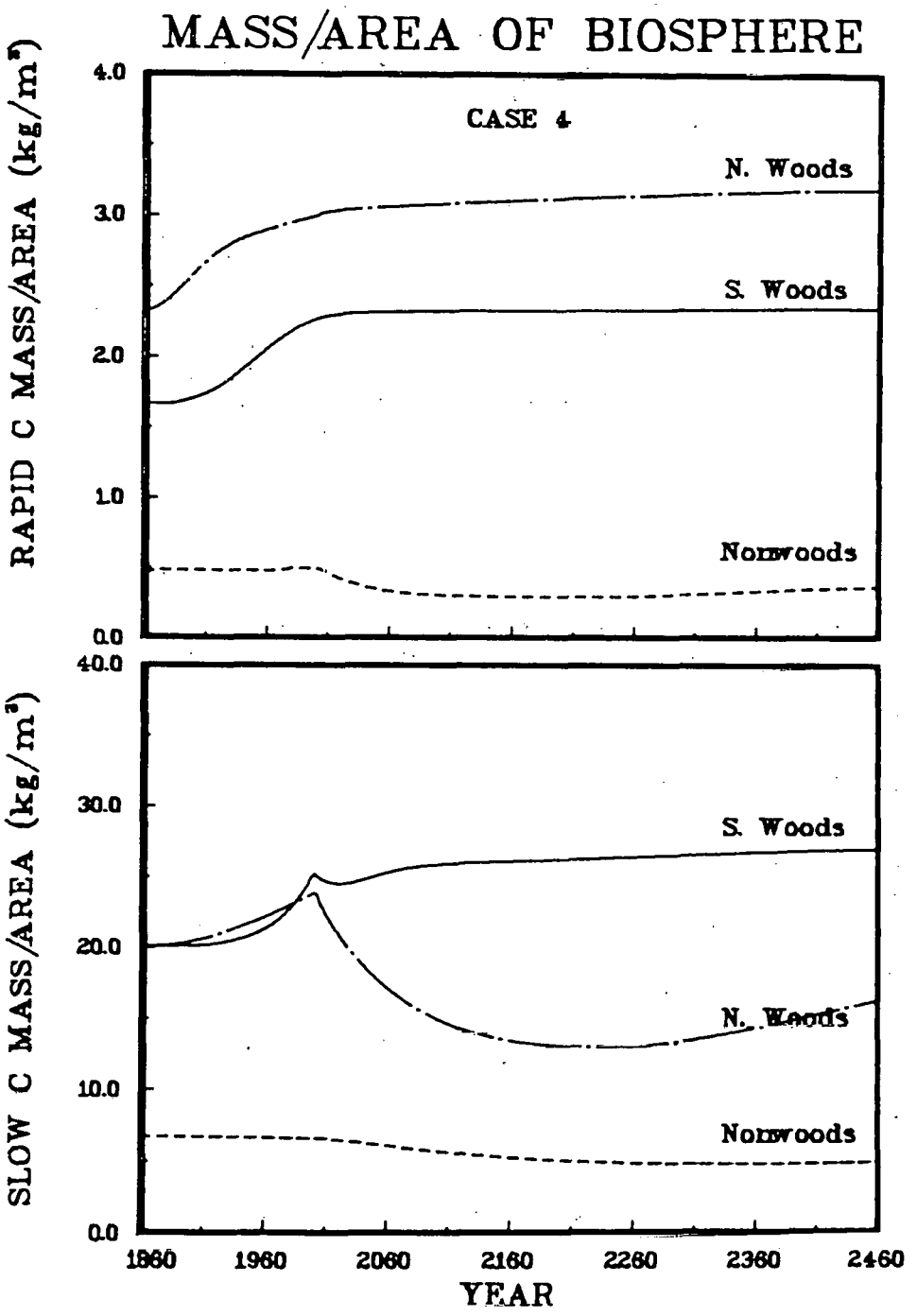

Figure 26. Projected mass density of biospheric carbon ("combination" scenario).

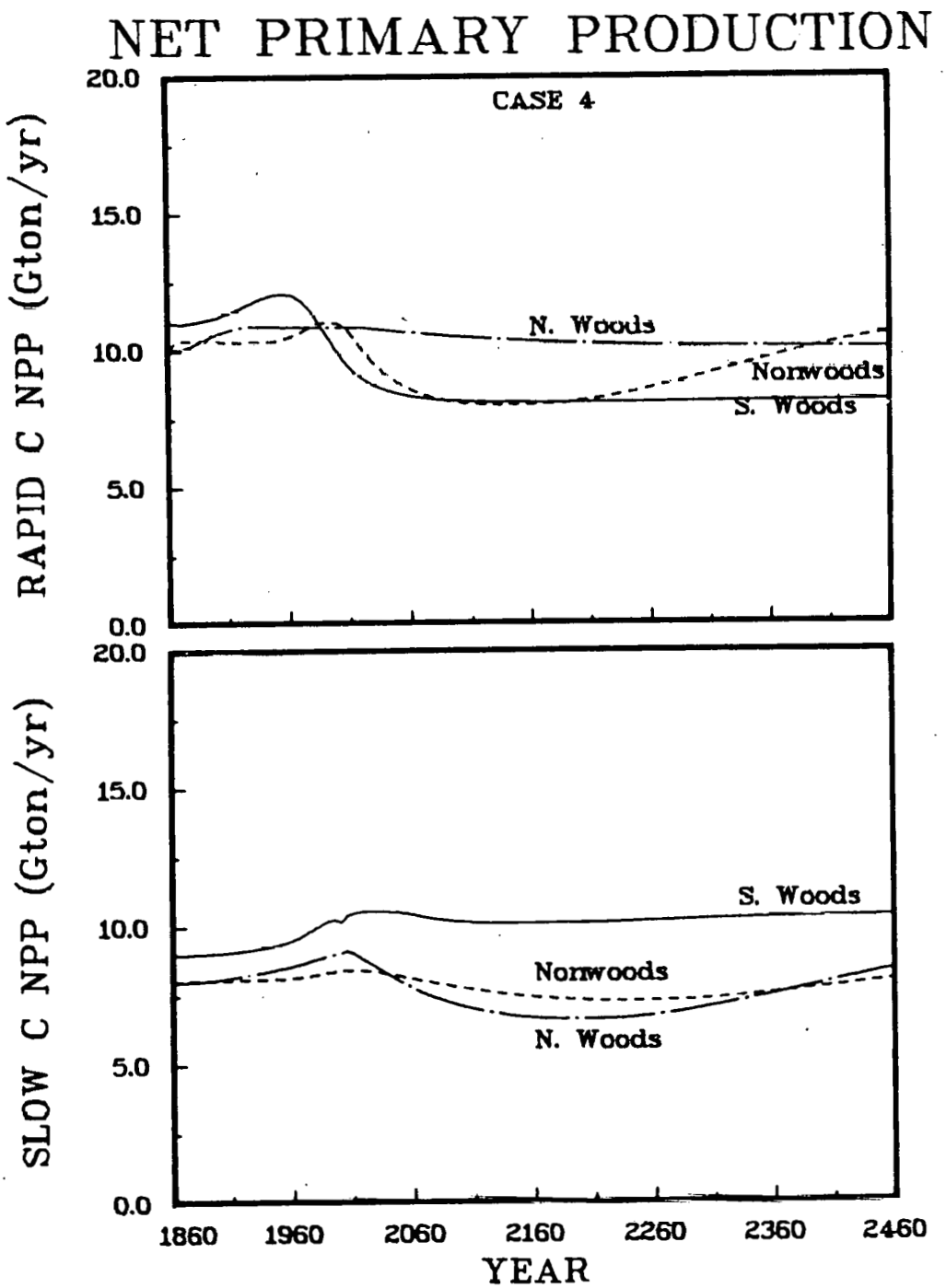

Figure 27. Simulated net primary production of terrestrial ecosystems ("combination" scenario).

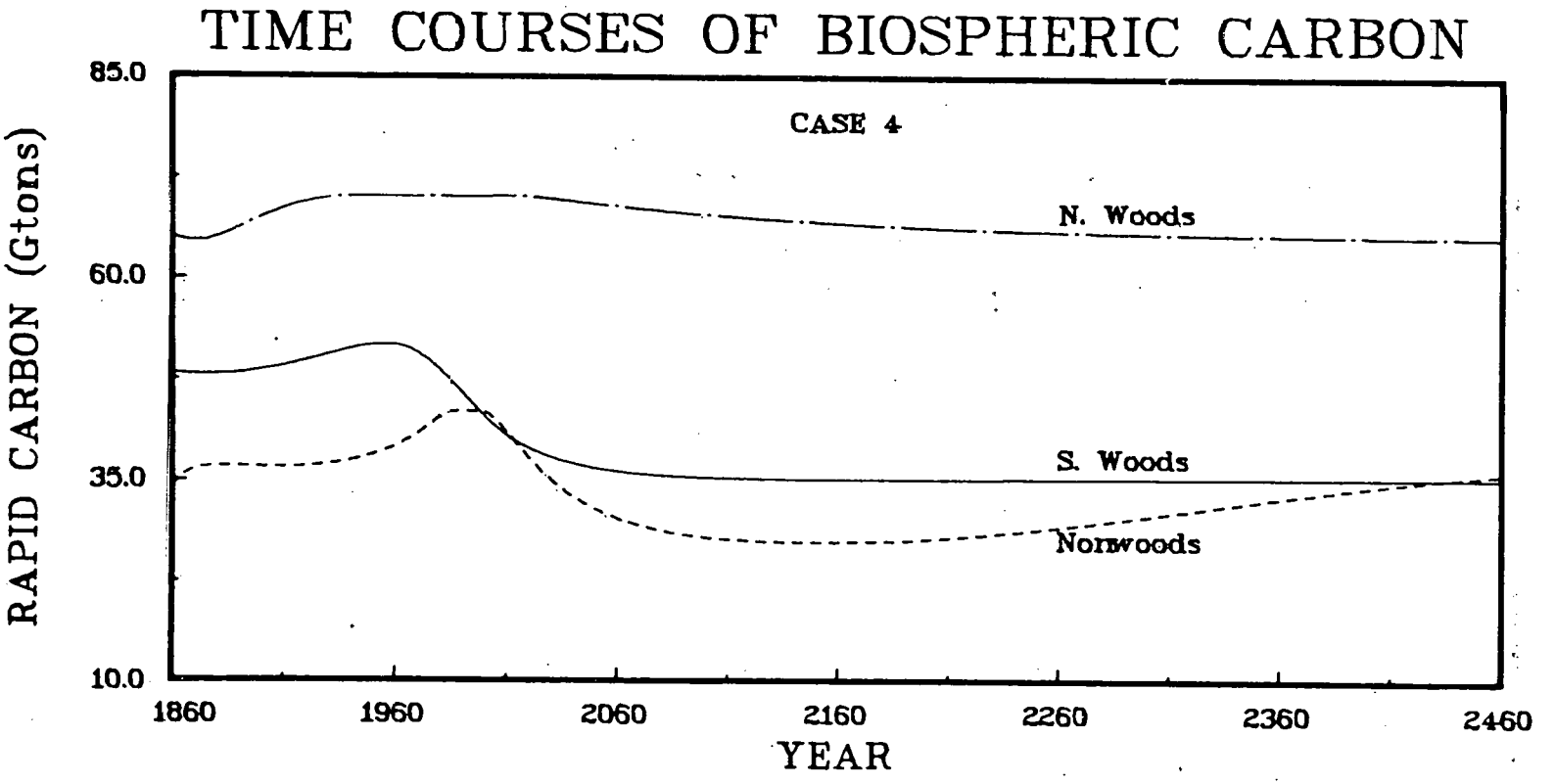

Figure 28. Projections of rapidly exchanging carbon in terrestrial ecosystems ("combination" scenario).

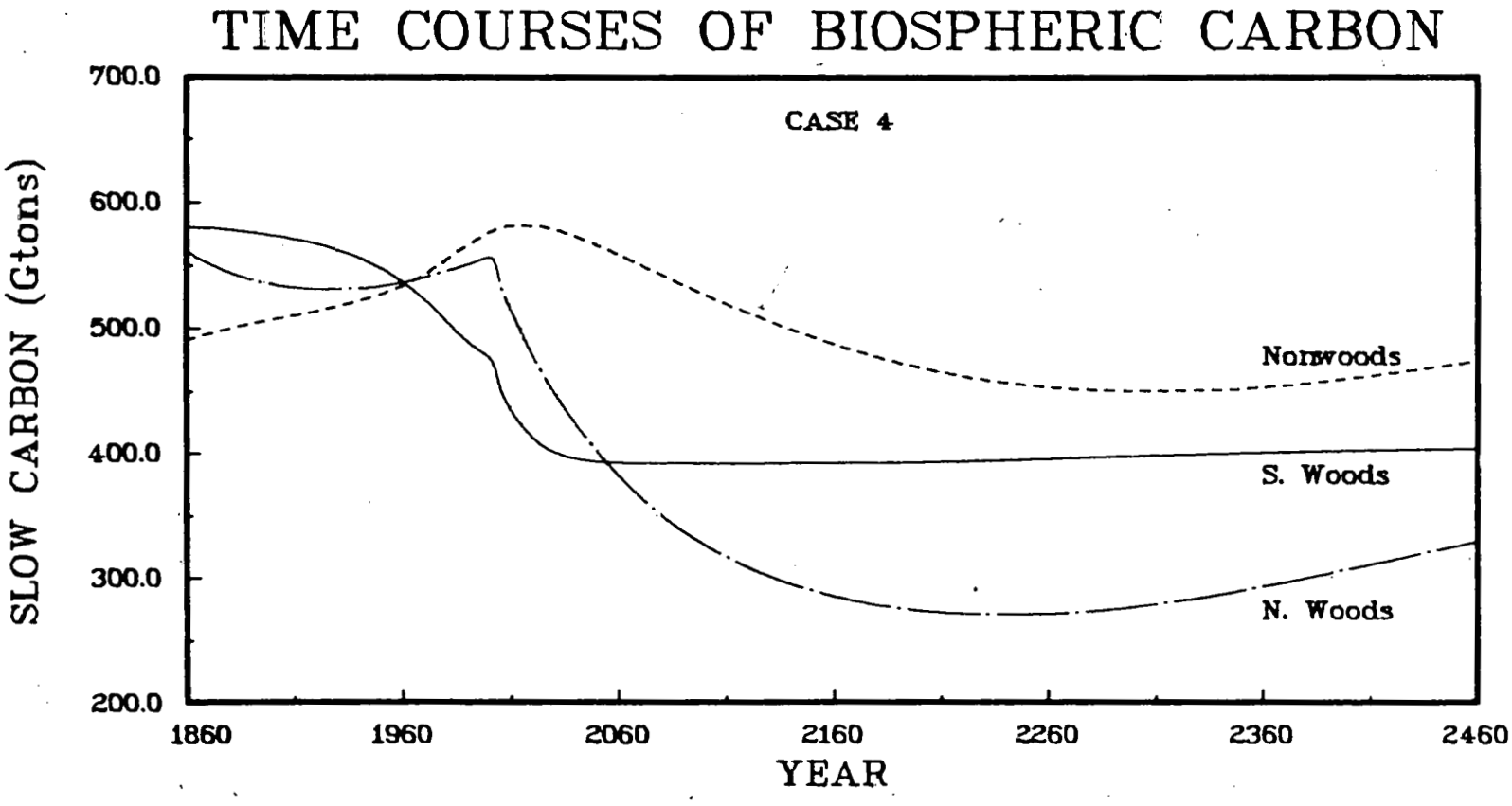

Figure 29. Projections of slowly exchanging carbon in terrestrial ecosystems ("combination" scenario).

be maintained at less than 1.5 times the 1860 value during the next century. The possibilities of technological wood storage to serve the same purpose deserve exploration. The level of CO_2 nevertheless would increase slowly to 1700 Gtons, or slightly below three times the original value (2.5 times the present level), by 2460.

6.3. IMPLICATIONS OF CHANGING FOSSIL AND NONFOSSIL CARBON USE

Several tentative propositions can be observed by comparing the results obtained from different levels of used reserves and from the four scenarios that were used assuming 7500 Gtons reserve. Clearly the former dominate the atmospheric level. There is no substantial reduction in the peak level of atmospheric carbon just by delaying the consumption of fossil fuel for a few more decades. The increased sluggishness of ocean water to absorb carbon dioxide and the probable upper limit of carbon storage reached in the biosphere are the main forces that cause a larger and larger fraction of the excess carbon dioxide to remain in the atmosphere. Lowering of ocean acidity (changing the buffering factor) may make the real situation worse than projected here.

There is a possibility that some of the coal will be kept by necessity as chemical feedstocks or for other uses when oil and gas are exhausted. If mankind elects to limit the ultimate consumption of fossil carbon to about 5000 Gtons releasing at the same initial rate of 4.35%/year, the maximum concentration of atmospheric carbon

would nevertheless increase to five times the 1974 value (3500 Gtons vs. 702 Gtons). This is about the same magnitude obtained in the "slow burner" scenario except that the rise in the next two centuries would be much slower. This scenario reflected the possible ameliorating effects of an alternative policy of utilization of the total fossil-carbon resource over a much longer time span. Clearly, still slower expansion in the next century offers more time for deciding the ultimate level as well as the rate of oxidizing fossil carbon.

An acceptable strategy in the future may have to depend on the proper management of the biosphere. Even the very preliminary current estimates of total NPP suggest this rate is about 10 times faster than the annual release of fossil carbon. In the "nominal" case, these rates are about even in 2100 A.D. when the fossil carbon release was assumed to reach the maximum rate. The NPP modeled in the "slow burner" scenario or the "combination" scenario would still absorb carbon up to four times faster than the emission rate of fossil carbon at the time of peak release. Hence, the biosphere would be able to take up more of the excess carbon in the meantime to compensate for the slow absorption of the ocean under circumstances which remain to be defined. The problem requiring further study is whether more of this carbon flow can be used in lieu of fossil energy, while producing needed food and perhaps increasing carbon storage as well. Some of these objectives may be in conflict (Olson et al., 1978).

Biomass is still the main source of fuel energy in most of the less developed countries (Earl, 1976; Eckholm, 1976). Suggestions have also

been made to generate electrical and fluid energy from biomass in a larger scale in developed nations (e.g., Szego and Kemp, 1973; Rose, 1977; Burgess, 1977; Tillman, 1978). Some advantages of this alternative policy are to replace the organic carbon formed millions of years ago for the recently formed biomass and quickly recycling the carbon in the atmosphere. However, the net effect of this management policy may still produce the same response of a relatively high level of atmospheric carbon as in the case of burning only 5000 Gtons of fossil carbon mentioned earlier. The disrupting impacts from the high carbon level can be reduced provided further consumption of fossil carbon would be drastically reduced or that other nonfossil energy sources could be developed to replace the total dependency on fossil fuel (Niehaus, 1977).

In the final analysis, the basic strategy in the long-term management of the carbon dioxide problem is the reduction of fossil-fuel consumption, keeping part of the fossil carbon as a source of new synthetic materials for the next generations. The management of the biosphere, whether or not it includes the concept of "fuel plantation" or the compaction and storage of organic matter, can be used as the supplementary policy during the transitory period (possibly a century) towards the development of some other sources of energy acceptable to the society.

CHAPTER 7

SUMMARY AND CONCLUSIONS

1. Controversies have arisen recently concerning the role of biospheric organic matter as a net source or sink for the anthropogenic release of atmospheric carbon dioxide. The massive conversion of forests into agricultural lands and possibly accelerated oxidation of humus during the recent past are regarded as the main sources of this excess biospheric carbon dioxide. Whether such nonfossil inputs could conceivably approach or exceed inputs from coal, gas, and oil is a major issue for current research. In the present study, a global carbon model was utilized to simulate the effects of changing ecological carbon balances in addition to changing fossil-fuel consumption on the global carbon cycle between 1860 and 2460.

These uncertainties about biospheric carbon dioxide release should not obscure the dominant future effect of fossil-fuel consumption on atmospheric CO_2 levels—if the use of fuel reserves continues to expand at high rates and then later diminish only slowly. If ultimate usable reserves (near 7500 ± 2500 Gtons of carbon) become oxidized in the next century or two, instead of being conserved partly as feedstocks, then drastic climatic changes are quite credible. Feedback from climatic change to the carbon cycle is beyond the scope of this study.

2. The present compartmental model distinguished rapidly and slowly exchanging carbon pools for both tropical and nontropical forests, and for nonwooded ecosystems. These compartments were coupled with a

two-layer atmosphere, an ocean having three water layers and three organic pools, and five categories of sedimentary and dissolved carbon. Carbon fluxes among these 19 compartments were described as a set of partly nonlinear ordinary differential equations for income minus loss. Quantities of carbon in the terrestrial biospheric compartments were estimated from preliminary projections of biospheric areas and carbon mass per unit area for different ecosystem groups.

Both area and mass/area were treated explicitly as variables in the model. Empirical functions were employed to restrain the net primary production and storage capacity of the biota. Historical data before 1974 of annual release of fossil carbon (and cement) were used as part of the forcing functions for all simulations.

3. Four different carbon utilization scenarios were simulated. The "nominal" and the "delayed consumption" scenarios (cases 1 and 2) assumed that the annual releases of fossil carbon beyond 1974 would follow different forms of the modified logistic equation (Eq. 3.1). These equations assumed that the annual production of fossil carbon would decline slowly as fossil fuel reserves were exhausted. In the "slow burner" and "combination" scenarios (cases 3 and 4), the rate parameter in the logistic equation was considered also to decrease from an initial value of about 0.04/year to about 0.01/year ultimately (Eq. 3.6). In addition, the "combination" scenario also would continuously shift 1% of the slowly exchanging carbon (turnover time > 10 years) in the forest compartments to a very slowly exchanging compartment after 2000 A.D., as if storage could be controlled deliberately.

4. Present climatic models suggest that a doubling of the atmospheric level of CO_2 could cause an increase of the global mean surface temperature by 1.5 to 3 °K. Feedback from climatic change to the biosphere is not yet clear. Assuming the carbon cycle model is valid and that fossil-fuel consumption follows the "nominal" or the "delayed consumption" scenario, a doubling of the preindustrial concentration of atmospheric carbon could occur in 50 to 60 years. By drastic reduction in the consumption of fossil fuel such as in the "slow burner" or the "combination" scenario, the doubling time could be delayed until at least a century later.

5. Results obtained from the simulation runs also reflect the different historic assumptions about the biosphere as a net source of atmospheric carbon over the past century. From 1860 to 1970, the biosphere is here assumed to release about 120 Gtons of carbon to the atmosphere either promptly or with delay of a few years. An equal amount of carbon was added to the atmosphere from the combustion of fossil fuel and cement over the same period, but the estimated prompt contribution of excess CO_2 from the biosphere was larger than the contribution from the fossil carbon from 1860 to 1900. Delayed inputs of nonfossil CO_2 by transfer from slowly to rapidly exchanging pools are still increasing, but are here assumed to be overtaken by fossil-fuel release of carbon around 1930.

6. Projections of biospheric carbon suggested some credible ways by which different parts of the biosphere could serve as source and sink for the excess CO_2 simultaneously. For the next 25 years, the forests

would lose another 100 to 150 Gtons of carbon to the atmosphere due to land clearings alone; at the same time, the nonforest ecosystems, mainly agricultural lands, would increase their total area and mass of carbon and thus, would serve as sinks for a lesser amount of the excess carbon dioxide. Beyond 2050, the terrestrial biosphere could maintain a relatively constant mass of organic carbon, if CO_2 fertilization effect is moderate (the biotic growth factor is about 0.15 to 0.2). The possible decline of photosynthesis at very high level of atmospheric carbon dioxide should be a major factor in studying the long-term response of the global carbon cycle.

7. Recent reports of $^{13}C/^{12}C$ ratios from dated tree rings also suggest that about 120 Gtons of biospheric carbon was released before 1930. They also suggest that the net release of biospheric carbon remained almost stable for the last 30 to 40 years. However, the present model assumes that the biosphere could release a substantial amount of organic carbon to the atmosphere until 2000 A.D. when deforestation rate could decline. In terms of the global carbon cycle, the present model may possibly underestimate the effect of the contribution of biospheric carbon before 1900, while overestimating the same effect for the past 30 to 40 years. The overestimation is probably the cause of the steep rise of atmospheric level of CO_2 in the recent years predicted by the model as compared with the measured increase from Mauna Loa (see Section 5.2). A reappraisal of the contribution of biospheric carbon in each region during the past century is suggested here as one of the topics for continuous research on the global carbon

cycle, so there will be a meaningful basis for anticipating and managing shifts in the net balance of biospheric carbon.

LITERATURE CITED

Adams, J. A. S., M. S. M. Mantovani, and L. L. Lundell. 1977. Wood versus fossil fuel as a source of excess carbon dioxide in the atmosphere: A preliminary report. Science 196:54-56.

Andersen, N. R., and A. Malahoff (eds.). 1977. The Fate of Fossil Fuel CO_2 in the Ocean. Plenum Press, New York.

Aoki, M., and K. Yabuki. 1977. Studies on the carbon dioxide enrichment for plant growth. VII. Changes in dry matter production and photosynthetic rate of cucumber during carbon dioxide enrichment. Agric. Meteorol. 18:475-485.

Arrhenius, S. 1896. On the influence of carbonic acid in the air upon the temperature of the ground. Phil. Mag. 41:237-276.

Attiwill, P. M. 1971. Atmospheric carbon dioxide and the biosphere. Environ. Pollut. 1:249-261.

Bacastow, R. B. 1976. Modulation of atmospheric carbon dioxide by the Southern Oscillation. Nature 261:116-118.

Bacastow, R. B., and C. D. Keeling. 1973. Atmospheric carbon dioxide and radiocarbon in the natural carbon cycle: II. Changes from A.D. 1700 to 2070 as deduced from a geochemical model. pp. 86-135. In Woodwell, G. M., and E. V. Pecan (eds.), Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge, Tennessee.

Baes, C. F., Jr., H. E. Goeller, J. S. Olson, and R. M. Rotty. 1976. The Global Carbon Dioxide Problem. ORNL-5194. Oak Ridge National Laboratory, Oak Ridge, Tennessee.

Baes, C. F., Jr., H. E. Goeller, J. S. Olson, and R. M. Rotty. 1977. Carbon dioxide and climate: The uncontrolled experiment. Am. Sci. 65:310-320.

Bainbridge, A. E. 1971. Atmospheric carbon dioxide variation. Trans. Am. Geophys. Union 52:222.

Baldwin, B., J. B. Pollack, A. Summers, O. B. Toon, C. Sagan, and W. Van Camp. 1976. Stratospheric aerosols and climatic change. Nature 263:551-555.

Baxter, M. S., and A. Walton. 1970. A theoretical approach to the Suess effect. Proc. Roy. Soc. Lond. A. 318:213-230.

Bischof, W. 1970. Carbon dioxide measurement from aircraft. Tellus 22:545-549.

Bohn, H. L. 1976. Estimate of organic carbon in world soils. Soil Sci. Soc. Am. J. 40:468-470.

Bolhár-Nordenkampf, H. R. 1976. Measurements and mathematical models for the CO_2 gas exchange of Phaseolus vulgaris var. nanus L. with special consideration of photorespiration and atrazine effect. Biochem. Physiol. Pflanz. 169:121-161. (In German with English summary.)

Bolin, B. 1970. The carbon cycle. Sci. Am. 223:125-132.

Bolin, B. 1975. A critical appraisal of models for the carbon cycle. pp. 225-235. In Global Atmospheric Research Program, The Physical Basis of Climate and Climate Modeling. Publication 16. World Meteorological Organization - International Congress of Scientific Unions, UNIPUB, New York.

Bolin, B. 1977a. Changes of land biota and their importance for the carbon cycle. Science 196:613-615.

Bolin, B. 1977b. Modeling the oceans and ocean sediments and their response to fossil fuel carbon dioxide emissions. pp. 81-95. In Andersen, N. R., and A. Malahoff (eds.), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York.

Bolin, B., and W. Bischof. 1970. Variations of the carbon dioxide content of the atmosphere in the Northern Hemisphere. Tellus 22: 431-442.

Bolin, B., and E. Eriksson. 1959. Changes in the carbon dioxide content of the atmosphere and sea due to fossil fuel combustion. pp. 130-142. In Bolin, B. (ed.), The Atmosphere and the Sea in Motion. The Rockefeller Institute Press, New York.

Bolin, B., and C. D. Keeling. 1963. Large-scale atmospheric mixing as deduced from the seasonal and meridional variations of carbon dioxide. J. Geophys. Res. 68:3899-3920.

Bolin, B., and H. Rodhe. 1973. A note on the concepts of age distribution and transit time in natural reservoirs. Tellus 25: 58-62.

Botkin, D. B. 1977. Forests, lakes, and the anthropogenic production of carbon dioxide. BioScience 27:325-331.

Borchert, H. 1951. Zür geochemie des Kohlenstuffs. Geochim. Cosmochim. Acta 2:62-75. (In German with English abstract.)

Bray, J. R. 1959. An analysis of the possible recent change in atmospheric carbon dioxide concentration. Tellus 11:220-230.

Brittain, E. G., and R. J. Cameron. 1973. Photosynthesis of leaves of some Eucalyptus species. N. Z. J. of Bot. 1:153-163.

Broecker, W. S. 1963. Radioisotopes and large-scale oceanic mixing. pp. 88-108. In Hill, M. N. (ed.), The Sea, vol. 2. Wiley-Interscience Publishers, London.

Broecker, W. S. 1970. A boundary condition on the evolution of atmospheric oxygen. J. Geophys. Res. 75:3553-3557.

Broecker, W. S. 1971. A kinetic model for the chemical composition of sea water. Quat. Res. 1:188-207.

Broecker, W. S. 1974. Chemical Oceanography. Harcourt Brace Javanovich, Inc., New York.

Broecker, W. S. 1977. The fate of fossil fuel CO_2—a research strategy. pp. 37-41. In Report of the Scientific Workshop on Atmospheric Carbon Dioxide. WMO-474. Secretariat of the World Meteorological Organization, Geneva, Switzerland.

Broecker, W. S., Y.-H. Li, and T.-H. Peng. 1971. Carbon dioxide—Man's unseen artifact. pp. 287-324. In Hood, D. W. (ed.), Impingement of Man on the Oceans. Wiley-Interscience, New York.

Brooks, H. 1977. Potentials and limitations of societal response to long-term environmental threats. pp. 241-252. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Brown, C. W., and C. D. Keeling. 1965. The concentration of atmospheric carbon dioxide in Antarctica. J. Geophys. Res. 70:6077-6085.

Brünig, E. F. 1977. The tropical rain forest—A wasted asset or an essential biospheric resource? Ambio 6:187-191.

Bryan, W., and R. D. Wright. 1976. The effect of enhanced CO_2 levels and variable light intensities on net photosynthesis in competing mountain trees. Am. Midl. Nat. 95:446-450.

Bunt, J. S. 1975. Primary productivity of marine ecosystems. pp. 169-183. In Lieth, H., and R. H. Whittaker (eds.), Primary Productivity of the Biosphere. Springer-Verlag, Inc., New York.

Burgess, R. L. 1978. Potential of forest fuels for producing electrical energy. J. For. 76:154-157.

Callendar, G. S. 1938. The artificial production of carbon dioxide and its influence on temperature. Q. J. Roy. Meteorol. Soc. 64:223-237.

Callendar, G. S. 1940. Variations of the amount of carbon dioxide in different air currents. Q. J. Roy. Meteorol. Soc. 66:395-400.

Callendar, G. S. 1958. On the amount of carbon dioxide in the atmosphere. Tellus 10:243-248.

Craig, H. 1957. The natural distribution of radiocarbon and the exchange time of carbon dioxide between atmosphere and sea. Tellus 9:1-17.

Craig, H. 1958. A critical evaluation of radiocarbon techniques for determining mixing rates in the oceans and the atmosphere. pp. 358-363. In Proceedings of the Second United Nations International Conference on Peaceful Uses of Atomic Energy, vol. 18. United Nations, Geneva.

Craig, H. 1963. The natural distribution of radiocarbon: Mixing rates in the sea and residence times of carbon and water. pp. 103-114. In Geiss, J., and E. D. Goldberg (eds.), Earth Science and Meteoritics. North Holland Publishing Co., Amsterdam.

Cramer, J., and A. L. Myers. 1972. Rate of increase of atmospheric carbon dioxide. Atmos. Environ. 6:563-573.

Darby, H. C. 1956. The clearing of the woodland in Europe. pp. 183-216. In Thomas, W. L., Jr., C. O. Sauer, M. Bates, and L. Mumford (eds.), Man's Role in Changing the Face of the Earth. The University of Chicago Press, Chicago.

Delwiche, C. C., and G. E. Likens. 1977. Biological response to fossil fuel combustion products. pp. 73-88. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Dugas, D. J. 1968. Increase of Exchangeable Carbon in the Earth's Reservoirs from Combustion of Fossil Fuels. P-3990. The RAND Corporation, Santa Monica, California.

Dumas, M. 1841. On the chemical statics of organized beings. Philos. Mag. (S. 3) 19:337-347, 456-469.

Earl, D. E. 1975. Forest Energy and Economic Development. Clarendon Press, Oxford.

Echols, J. R. 1976. Population vs. the environment: A crisis of too many people. Am. Sci. 64:165-173.

Eckholm, E. P. 1976. Losing Ground. W. W. Norton & Co., Inc., New York.

Ekdahl, C. S., and C. D. Keeling. 1973. Atmospheric carbon dioxide and radiocarbon in the natural carbon cycle: I. Quantitative deductions from records at Mauna Loa Observatory and at the South Pole. pp. 51-85. In Woodwell, G. M., and E. V. Pecan (eds.), Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge, Tennessee.

Elliot, W. P., and L. Machta (eds.). In press. Proceedings of the ERDA Conference on Global Effects of Increased Carbon Dioxide, Miami Beach, Florida, March 7-11, 1977.

Eriksson, E. 1971. Compartment models and reservoir theory. Ann. Rev. Ecol. Syst. 2:67-84.

Eriksson, E., and T. Rosswall. 1976. Man and biogeochemical cycles: Impacts, problems and research needs. pp. 11-16. In Svensson, B. H., and R. Söderlund (eds.), Nitrogen, Phosphorus and Sulphur—Global Cycles. SCOPE Report 7. Ecological Bulletin 22. Swedish Natural Science Research Council, Sweden.

Eriksson, E., and P. Welander. 1956. On a mathematical model of the carbon cycle in nature. Tellus 8:155-175.

Fairhall, A. W. 1973. Accumulation of fossil carbon dioxide in the atmosphere and the sea. Nature 245:20-23.

Fairhall, A. W., A. W. Young, and P. A. Bradford. 1973. Radiocarbon in the sea. pp. 226-240. In Rafter, T. A., and T. Grant-Taylor (eds.), Proceedings of the Eighth International Radiocarbon Dating Conference, vol. 1. Royal Society of New Zealand, New Zealand.

Farmer, J. G., and M. S. Baxter. 1974. Atmospheric carbon dioxide levels as indicated by the stable isotope record in wood. Nature 247:273-275.

Fogg, G. E. 1975. Primary productivity. pp. 385-453. In Riley, J. P., and G. Skirrow (eds.), Chemical Oceanography, vol. 2. Second edition. Academic Press, London.

Frejka, T. 1973. The prospects for a stationary world population. Sci. Am. 228:15-23.

Gaastra, P. 1959. Photosynthesis of crop plants as influenced by light, carbon dioxide, temperature, and stomatal diffusion resistance. Meded. Landbouwhogesch. Wageningen 59:1-68.

Garratt, J. R., and G. I. Pearman. 1973a. Large-scale CO_2 fluxes in the Southern Hemisphere troposphere. Nature (London) Phys. Sci. 242:54-56.

Garratt, J. R., and G. I. Pearman. 1973b. CO_2 concentration in the atmospheric boundary-layer over south-east Australia. Atmos. Environ. 7:1257-1266.

Garrels, R. M., and A. Lerman. 1977. The exogenic cycle: Reservoirs, fluxes and problems. pp. 23-32. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Garrels, R. M., A. Lerman, and F. T. Mackenzie. 1976. Controls of atmospheric O_2 and CO_2: Past, present, and future. Am. Sci. 64: 306-315.

Garrels, R. M., and F. T. Mackenzie. 1972. A quantitative model for the sedimentary rock cycle. Mar. Chem. 1:27-41.

Garrels, R. M., F. T. Mackenzie, and C. Hunt. 1975. Chemical Cycles and the Global Environment: Assessing Human Influences. William Kaufmann, Inc., Los Altos, California.

Gear, C. W. 1971. Numerical Initial Value Problems in Ordinary Differential Equations. Prentice-Hall, Inc., Englewood Cliffs, New Jersey.

Goel, N. S., S. C. Maitra, and E. W. Montroll. 1971. On the Volterra and other nonlinear models of interacting populations. Rev. Mod. Phys. 43:231-276.

Gómez-Pompa, A., C. Vásquez-Yanes, and S. Guevara. 1972. The tropical rain forest: A nonrenewable resource. Science 177:762-765.

Goodland, R. J. A., and H. S. Irwin. 1975. Amazon Jungle: Green Hell to Red Desert? Elsevier Scientific Publishing Company, Amsterdam.

Gowdy, C. M., R. J. Mulholland, and W. R. Emanuel. 1975. Modelling the global carbon cycle. Int. J. Syst. Sci. 6:965-976.

Green, K., and R. Wright. 1977. Field response of photosynthesis to CO_2 enhancement in Ponderosa pine. Ecology 58:687-692.

Gullikesen, S., R. Nydal, and K. Lövseth. 1973. Further calculations on the C-14 exchange between the ocean and the atmosphere. pp. 287-296. In Rafter, T. A., and T. Grant-Taylor (eds.), Proceedings of the Eighth International Radiocarbon Dating Conference, vol. 1. The Royal Society of New Zealand, New Zealand.

Hamilton, L. S. 1976. Tropical Rainforest Use and Preservation: A Study of Problems and Practices in Venezuela. Sierra Club, Washington, D.C.

Hamilton, R. E. 1977. Present and predicted level of consumption of fossil fuels. pp. 155-164. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Hellmers, H. 1964. An evaluation of the photosynthetic efficiency of forests. Q. Rev. Biol. 39:249-257.

Hobbs, P. V., H. Harrison, and E. Robinson. 1974. Atmospheric effects of pollutants. Science 183:909-915.

Hoffert, M. I. 1974. Global distributions of atmospheric carbon dioxide in the fossil-fuel era: A projection. Atmos. Environ. 8:1225-1249.

Hofstra, C., and J. D. Hesketh. 1975. The effects of temperature and CO_2 enrichment on photosynthesis in soybean. pp. 71-80. In Marcelle, R. (ed.), Environmental and Biological Control of Photosynthesis. Dr. W. Junk b.v., Publishers, The Hague.

Hubbert, M. K. 1971. The energy resources of the earth. Sci. Am. 224: 60-70.

Hubbert, M. K. 1973. Survey of world energy resources. Can. Min. Metall. Bull. 66:37-54.

Hurt, P., and R. Wright. 1976. CO_2 compensation point for photosynthesis: Effect of variable CO_2 and soil moisture levels. Am. Midl. Nat. 95:450-455.

Hutchinson, G. E. 1954. The biochemistry of the terrestrial atmosphere. pp. 371-433. In Kuiper, G. P. (ed.), The Earth as a Planet. University of Chicago Press, Chicago.

International Business Machines (IBM). 1975. Continuous Systems Modeling Program III (CSMP III). Program Reference Manual, third edition. SH19-7001-3. International Business Machines Corporation, New York.

Integrated Software Systems Corporation. 1970. DISSPLA (Display Integrated Software System and Plotting Language). Beginner/Intermediate Manual. Integrated Software Systems Corporation, San Diego.

Jørgensen, S. E., and H. Mejer. Modelling the global cycle of carbon, nitrogen and phosphorus and their influence on the global heat balance. Ecol. Model. 2:19-31.

Junge, C. E. 1963. Air Chemistry and Radioactivity. Academic Press, New York.

Junge, C. E. 1977. Stable isotope fractionation in geochemical and environmental cycles. pp. 33-44. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Junge, C. E., and G. Czeplak. 1968. Some aspects of the seasonal variation of carbon dioxide and ozone. Tellus 20:422-434.

Junge, C. E., M. Schidlowski, R. Eichmann, and H. Pietrek. 1975. Model calculations for the terrestrial carbon cycle: Carbon isotope geochemistry and evolution of photosynthetic oxygen. J. Geophys. Res. 80:4542-4552.

Keeling, C. D. 1973a. The carbon dioxide cycle: Reservoir models to depict the exchange of atmospheric carbon dioxide with the oceans and land plants. pp. 251-329. In Rasool, S. I. (ed.), Chemistry of the Lower Atmosphere. Plenum Press, New York.

Keeling, C. D. 1973b. Industrial production of carbon dioxide from fossil fuels and limestone. Tellus 25:174-198.

Keeling, C. D., J. A. Adams, Jr., C. A. Ekdahl, Jr., and P. R. Guenther. 1976a. Atmospheric carbon dioxide variations at the South Pole. Tellus 28:552-564.

Keeling, C. D., and R. B. Bacastow. 1977. Impact of industrial gases on climate. pp. 72-95. In Energy and Climate. National Academy of Sciences, Geophysics Research Board, Geophysics Study Committee, Washington, D.C.

Keeling, C. D., R. B. Bacastow, A. E. Bainbridge, C. A. Ekdahl, Jr., P. R. Guenther, L. S. Waterman, and J. F. S. Chin. 1976b. Atmospheric carbon dioxide variations at Mauna Loa Observatory, Hawaii. Tellus 28:538-551.

Kellogg, W. W. 1978. Is mankind warming the earth? Bull. At. Sci. 34: 10-19.

Kerr, R. A. 1977. Carbon dioxide and climate: Carbon budget still unbalanced. Science 197:1352-1353.

Kester, D. R., and R. M. Pytkowicz. 1977. Natural and anthropogenic changes in the global carbon dioxide system. pp. 99-120. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Killough, G. G. 1977. A Diffusion-Type Model of the Global Carbon Cycle for the Estimation of Dose to the World Population from Release of Carbon-14 to the Atmosphere. ORNL-5269. Oak Ridge National Laboratory, Oak Ridge, Tennessee.

Kira, T. 1975. Primary production of forests. pp. 5-40. In Cooper, J. D. (ed.), Photosynthesis and Productivity in Different Environments. IBP 3. Cambridge University Press, London.

Kira, T., M. Ando, and H. Sato. 1973. Environmental disruption and terrestrial plants. Jpn. J. Ecol. (Nippon Seitai Gakkaishi) 23: 65-73. (In Japanese.)

Lemon, E. 1977. The land's response to more carbon dioxide. pp. 97-130. In Andersen, N. R., and A. Malahaff (eds.), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York.

Leopold, A. C., and P. E. Kriedemann. 1975. Plant Growth and Development. Second edition. McGraw-Hill Co., New York.

Li, Y.-H. 1972. Geochemical mass balance among lithosphere, hydrosphere, and atmosphere. Am. J. Sci. 272:119-137.

Lieth, H. 1963. The role of vegetation in the carbon dioxide content of the atmosphere. J. Geophys. Res. 68:3887-3898.

Lieth, H. 1975. Modeling the primary productivity of the world. pp. 237-263. In Lieth, H., and R. H. Whittaker (eds.), Primary Productivity of the Biosphere. Ecological Studies 14. Springer-Verlag, New York.

Lister, R. A., and E. R. Lemon. 1976. Interactions of atmospheric carbon dioxide, diffuse light, plant productivity and climate processes—Model predictions. pp. 112-135. In Atmosphere-Surface Exchange of Particulate and Gaseous Pollutants. Proceedings of a Symposium, Richland, Washington, September 4-6, 1974. U.S. Energy Research and Development Administration, Washington, D.C.

Loomis, R. S. In press. CO_2 and the biosphere. In Elliot, W. P., and L. Machta (eds.), Proceedings of the ERDA Conference on Global Effects of Increased Carbon Dioxide, Miami Beach, Florida, March 7-11, 1977.

Lotka, A. J. 1924. Elements of Physical Biology. The Williams and Wilkins Co., Inc., New York.

Machta, L. 1972a. The role of the ocean and biosphere in the carbon dioxide cycle. pp. 121-145. In Dyrssen, D., and D. Jagner (eds.), The Changing Chemistry of the Oceans. Proceedings of the 20th Nobel Symposium. Almqvist and Wiksell, Stockholm.

Machta, L. 1972b. Mauna Loa and global trends in air quality. Bull. Am. Meteorol. Soc. 53:402-420.

Machta, L. 1973. Prediction of CO_2 in the atmosphere. pp. 21-31. In Woodwell, G. M., and E. V. Pecan (eds.), Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge, Tennessee.

Machta, L., K. Hanson, and C. D. Keeling. 1977. Atmospheric carbon dioxide and some interpretations. pp. 131-144. In Andersen, N. R., and A. Malahoff (eds.), The Fate of Fossil Fuel CO_2 in the Ocean. Plenum Press, New York.

Mackenzie, F. T. 1975. Sedimentary cycling and the evolution of sea water. pp. 309-364. In Riley, J. P., and G. Skirrow (eds.), Chemical Oceanography, vol. 1. Second edition. Academic Press, London.

Mackenzie, F. T., and R. M. Garrels. 1966. Chemical mass balance between rivers and oceans. Am. J. Sci. 264:507-525.

Mackenzie, F. T., and R. Wollast. 1977a. Sedimentary cycling models of global processes. pp. 739-785. In Goldberg, E. D., I. N. McCave, J. J. O'Brien, and J. H. Steele (eds.), The Sea, vol. 6. Marine Modeling. John Wiley & Sons, Inc., New York.

Mackenzie, F. T., and R. Wollast. 1977b. Thermodynamic and kinetic controls of global chemical cycles of the elements. pp. 45-59. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Manabe, S., and R. T. Wetherald. 1975. The effects of doubling the carbon dioxide concentration on the climate of a general circulation model. J. Atmos. Sci. 32:3-15.

Martin, D. F. 1970. Marine Chemistry, vol. 2. Theory and Application. Marcel Dekker Inc., New York.

Meijer, W. 1973. Devastation and regeneration of lowland Dipterocarp forests in Southeast Asia. BioScience 23:528-533.

Minderman, G. 1968. Addition, decomposition and accumulation of organic matter in forests. J. Ecol. 56:355-362.

Miyake, Y., and K. Saruhasi. 1973. The radio-carbon model of the ocean. Pap. Meteorol. Geophys. 24:263-271.

Möller, F. 1963. On the influence of changes in the CO_2 concentration in air on the radiation balance of the earth's surface and on the climate. J. Geophys. Res. 68:3877-3886.

Moore, A. L. 1977. Photosynthesis and photorespiration. Nature 267: 307-308.

Mulholland, R. J., C. L. Cole, and C. M. Gowdy. 1977. Modeling the effect of atmospheric carbon dioxide on the global radiative heat balance. IEEE Trans. Syst. Man, and Cybern. SMC-7:803-805.

National Academy of Sciences (NAS). 1977. Energy and Climate. Geophysics Research Board, Geophysics Study Committee, National Academy of Sciences, Washington, D.C.

Nef, J. U. 1977. An early energy crisis and its consequences. Sci. Am. 237:140-151.

Niehaus, F. 1975. Long-Term Aspects of the Environmental Burden from Energy Production: Carbon Dioxide and 3-Hydrogen. BKEJAS (1165). Kernforschungsanlage, Jülich. (In German with English Abstract.)

Niehaus, F. 1976. A Nonlinear Eight Level Tandem Model to Calculate the Future CO_2 and C-14 Burden to the Atmosphere. RM-76-35. Research Memorandum of the International Institute for Applied Systems Analysis, Laxenburg, Austria.

Noggle, G. R., and G. J. Fritz. 1976. Introductory Plant Physiology. Prentice-Hall Inc., New York.

Nydal, R. 1967. On the transfer of radiocarbon in nature. pp. 119-128. In IAEA Radioactive Dating and Methods of Low-level Counting. IAEA, Vienna.

Nydal, R. 1968. Further investigation on the transfer of radiocarbon in nature. J. Geophys. Res. 73:3617-3635.

Nydal, R., and K. Lövseth. 1970. Prospective decrease in atmospheric radiocarbon. J. Geophys. Res. 75:2271-2278.

Oeschger, H., U. Siegenthaler, U. Schotterer, and A. Gugelmann. 1975. A box diffusion model to study the carbon dioxide exchange in nature. Tellus 27:168-192.

Olson, J. S. 1970a. Carbon cycles and temperate woodlands. pp. 226-241. In Reichle, D. E. (ed.), Analysis of Temperate Forest Ecosystems. Springer-Verlag, New York.

Olson, J. S. 1970b. Continental ecosystem patterns and reconstructed living carbon prior to Iron Age. Back flyleaf. In Reichle, D. E. (ed.), Analysis of Temperate Forest Ecosystems. Springer-Verlag, New York.

Olson, J. S., and G. G. Killough. 1977. Atmosphere carbon exchanges modeled with subdivided biospheric pools and a diffusive ocean. Trans. Am. Geophys. Union 58:808.

Olson, J. S., M. F. Olson, and W. T. Swank. 1974. Man and stability of some ecosystems. p. 226. In Proceedings of the First International Congress of Ecology: Structure, Functioning and Management of Ecosystems. Center for Agricultural Publishing and Documentation, Wageningen, The Netherlands.

Olson, J. S., H. A. Pfuderer, and Y.-H. Chan. 1978. Changes in the Global Carbon Cycle and the Biosphere. ORNL/EIS-109. Oak Ridge National Laboratory, Oak Ridge, Tennessee.

Pales, J. C., and C. D. Keeling. 1965. The concentration of atmospheric carbon dioxide in Hawaii. J. Geophys. Res. 70:6053-6076.

Palmer, B. J. 1973. A review paper on the effect of carbon dioxide and aerosols on climate modification. Environ. Lett. 5:249-265.

Pearman, G. I., and J. R. Garratt. 1972. Global aspects of carbon dioxide. Search 3:67-73.

Pearman, G. I., and J. R. Garratt. 1973. Space and time variations of tropospheric carbon dioxide in the Southern Hemisphere. Tellus 25: 309-311.

Perry, H., and H. H. Landsberg. 1977. Projected world energy consumption. pp. 35-50. In Energy and Climate. National Academy of Sciences, Geophysics Research Board. Geophysics Study Committee, Washington, D.C.

Persson, R. 1974. World Forest Resources. Review of the World's Forest Resources in the Early 1970's. Research Note No. 17. Department of Forest Survey, Royal College of Forestry, Stockholm.

Plass, G. N. 1956. The carbon dioxide theory of climatic change. Tellus 8:140-154.

Plass, G. N. 1972a. Relationship between atmospheric carbon dioxide amount and properties of the sea. Environ. Sci. Technol. 6:736-740.

Plass, G. N. 1972b. The influence of the combustion of fossil fuels on the climate. pp. 24-46. In Bockris, J. O'M. (ed.), Electrochemistry for Cleaner Environments. Plenum Press, New York.

Platt, T., and D. V. Subba Rao. 1975. Primary production of freshwater microphytes. pp. 249-280. In Cooper, J. P. (ed.), Photosynthesis and Productivity in Different Environments. IBP 3. Cambridge University Press, London.

Plesset, M. S., and D. J. Dugas. 1967. Decay of a disturbance in the natural distribution of carbon-14. J. Geophys. Res. 72:4131-4137.

Plesset, M. S., and A. L. Latter. 1960. Transient effects in the distribution of carbon-14 in nature. Proc. N. A. S. 46:232-241.

Putnam, P. C. 1953. Energy in the Future. Van Nostrand Co., Inc., New York.

Pytkowicz, R. M. 1967. Carbonate cycle and the buffer mechanism of recent oceans. Geochim. Cosmochim. Acta 31:63-75.

Pytkowicz, R. M. 1972. The chemical stability of the oceans and the CO_2 system. pp. 147-152. In Dyrssen, D., and D. Jagner (eds.), The Changing Chemistry of the Ocean. Almquist and Wiksell, Stockholm.

Pytkowicz, R. M. 1973. The carbon dioxide system in the oceans. Schweiz. Z. Hydrol. 35:8-28.

Pytkowicz, R. M., and L. F. Small. 1977. Fossil fuel problem and carbon dioxide: An overview. pp. 7-31. In Andersen, N. R., and A. Malahoff (eds.), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York.

Reichle, D. E. (ed.). In press. Dynamic properties of forest ecosystems. IBP Synthesis Series. Cambridge University Press, London.

Reiners, W. A., L. H. Allen, Jr., R. Bacastow, D. H. Ehalt, C. S. Ekdahl, Jr., G. Likens, D. H. Livingston, J. S. Olson, and G. M. Woodwell. 1973. A summary of the world carbon cycle and recommendations for critical research. pp. 368-382. In Woodwell, G. M., and E. V. Pecan (eds.), Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge, Tennessee.

Revelle, R. 1974. Food and population. Sci. Am. 231:161-182.

Revelle, R. 1976. The resources available for agriculture. Sci. Am. 235:165-178.

Revelle, R., and W. Munk. 1977. The carbon dioxide cycle and the biosphere. pp. 140-158. In Energy and Climate. National Academy of Sciences, Geophysics Research Board, Geophysics Study Committee, Washington, D.C.

Revelle, R., and H. S. Suess. 1957. Carbon dioxide exchange between the atmosphere and the ocean, and the question of an increase in atmospheric CO_2 during the past decades. Tellus 9:18-27.

Robinson, E., and R. C. Robbins. 1972. Emissions, concentrations, and fate of gaseous atmospheric pollutants. pp. 1-93. In Strauss, W. (ed.), Air Pollution Control. Wiley-Interscience, New York.

Rodin, L. E., N. I. Bazilevich, and N. N. Rozov. 1975. Productivity of the world's main ecosystems. pp. 13-26. In Reichle, D. E., J. F. Franklin, and D. W. Goodall (eds.), Productivity of World Ecosystems. National Academy of Sciences, Washington, D.C.

Rose, D. W. 1977. Cost of producing energy from wood in intensive cultures. J. Environ. Manage. 5:23-35.

Rostlund, E. 1956. The outlook for the world's forests and their chief products. pp. 633-671. In Haden-Guest, S., J. K. Wright, and E. M. Teclaff (eds.), A World Geography of Forest Resources. The Ronald Press Co., New York.

Rotty, R. H. 1973. Commentary on and extension of calculative procedure for CO_2 production. Tellus 25:508-517.

Rotty, R. H. 1974. First estimates of global flaring of natural gas. Atmos. Environ. 8:681-686.

Rotty, R. H. 1975. A note updating carbon dioxide production from fossil fuels and cement. IEA(M)-75-4. IEA Research Memorandum. Institute for Energy Analysis, Oak Ridge Associated Universities, Oak Ridge, Tennessee.

Rotty, R. M. 1976. Global Energy Demand and Related Climate Change. IEA(M)-76-5. IEA Research Memorandum. Institute for Energy Analysis, Oak Ridge Associated Universities, Oak Ridge, Tennessee.

Rotty, R. M. 1977. Global carbon dioxide production from fossil fuels and cement, A.D. 1950-A.D. 2000. pp. 167-181. In Andersen, N. R., and A. Malahoff (eds.), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York.

Rotty, R. M., and A. M. Weinberg. 1977. How long is coal's future? Clim. Change 1:45-57.

Russo, J. M., and W. M. Knapp. 1976. A numerical simulation of plant growth. Int. J. Biometeorol. 20:276-285.

Ryabchikov, A. 1975. The Changing Face of the Earth. The Structure and Dynamics of the Geosphere, Its Natural Development and the Changes Caused by Man. Progress Publishers, Moscow.

Schidlowski, M., R. Eichmann, and C. E. Junge. 1975. Precambrian sedimentary carbonates: Carbon and oxygen isotope geochemistry and implications for the terrestrial oxygen budget. Precamb. Res. 2:1-69.

Schindler, D. W., and E. J. Fee. 1975. The roles of nutrient cycling and radiant energy in aquatic communities. pp. 323-343. In Cooper, J. P. (ed.), Photosynthesis and Productivity in Different Environments. IBP 3. Cambridge University Press, London.

Schlesinger, W. H. 1977. Carbon balance in terrestrial detritus. Ann. Rev. Ecol. Syst. 8:51-81.

Schneider, S. H. 1975. On the carbon dioxide-climate confusion. J. Atmos. Sci. 32:2060-2066.

Sears, P. B. 1956. The importance of forests to man. pp. 3-12. In Haden-Guest, S., J. K. Wright, and E. M. Teclaff (eds.), A World Geography of Forest Resources. The Ronald Press Co., New York.

Siegenthaler, U., and H. Oeschger. 1978. Predicting future atmospheric carbon dioxide levels. Science 199:388-395.

Sioli, H. 1973. Recent human activities in the Brazilian Amazon region and their ecological effects. pp. 321-334. In Meggers, B. J., E. S. Ayensu, and W. D. Duckworth (eds.), Tropical Forest Ecosystems in Africa and South America: A Comparative Review. Smithsonian Institution Press, Washington, D.C.

Skirrow, G. 1975. The dissolved gases-carbon dioxide. pp. 1-192. In Riley, J. P., and G. Skirrow (eds.), Chemical Oceanography, vol. 2. Second edition. Academy Press, London.

Small, L. F., and P. L. Donaghay, and R. M. Pytkowicz. 1977. Effects of enhanced CO_2 levels on growth characteristics on two marine phytoplankton species. pp. 183-204. In Andersen, N. R., and A. Malahoff (eds.), The Fate of Fossil Fuel CO_2 in the Oceans. Plenum Press, New York.

Smil, V., and D. Milton. 1974. Carbon dioxide: Alternative futures. Atmos. Environ. 8:1213-1223.

Sommer, A. 1976. Attempt at an assessment of the world's tropical moist forests. Unasylva 28:5-24.

Spofford, W. O., Jr. 1971. Decision making under uncertainty: The case of carbon dioxide buildup in the atmosphere. pp. 537-556. In Mathews, W. H., W. W. Kellogg, and G. D. Robinson (eds.), Man's Impact on the Climate. MIT Press, Cambridge, Massachusetts.

Steemann Nielsen, E. 1975. Marine Photosynthesis. Elsevier Oceanography Series Volume 13. Elsevier Scientific Publishing Co., Amsterdam.

Stengel, E., and C. J. Soeder. 1975. Control of photosynthetic production in aquatic ecosystems. pp. 645-660. In Cooper, J. P. (ed.), Photosynthesis and Productivity in Different Environments. IBP 3. Cambridge University Press, London.

Strain, B. R. 1977. Effects of Global Carbon Dioxide Enrichment on Plants: A Partially Annotated Bibliography. Department of Botany, Duke University, Durham, North Carolina.

Study of Critical Environmental Problems (SCEP). 1970. Man's Impact on the Global Environment: Assessment and Recommendations for Action. MIT Press, Cambridge, Massachusetts.

Study of Man's Impact on Climate (SMIC). 1971. Inadvertent Climate Modification. Report of 1971 Study of Man's Impact on Climate, hosted by Sweden's Royal Academies of Sciences and Engineering. MIT Press, Cambridge, Massachusetts.

Stuiver, M., 1973. The ^{14}C cycle and its implications for mixing rates in the ocean-atmosphere system. pp. 6-20. In Woodwell, G. M., and E. V. Pecan (eds.), Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge Tennessee.

Stuiver, M. 1978. Atmospheric carbon dioxide and carbon reservoir changes. Science 199:253-258.

Stumm, W. (ed.). 1977. Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Svensson, B. H., and R. Söderlund (eds.). 1976. Nitrogen, Phosphorus and Sulphur—Global Cycles. SCOPE Report 7. Ecol. Bull. 22. Swedish National Science Research Council, Sweden.

Swift, M. J. 1977. The ecology of wood decomposition. Sci. Prog., Oxf. 64:175-199.

Szego, G. C., and C. C. Kemp. 1973. Energy forests and fuel plantations. Chem. Technol. 3:275-284.

Tait, R. V. 1972. Elements of Marine Ecology. An Introductory Course. Second edition. Butterworth & Co., London.

Tillman, D. A. 1978. Wood as an Energy Resource. Academic Press, New York.

Tomović, R. 1964. Sensitivity Analysis of Dynamic Systems. McGraw-Hill, New York.

Totsuka, T. 1966. Relationships between CO_2 concentration of air and dry matter production of higher plants. Bot. Mag. (Tokyo) 79: 51-60. (In Japanese.)

Tuan, Y. F. 1968. Discrepancies between environmental attitude and behavior. Can. Geogr. 12:176-191.

Tuan, Y. F. 1970. Our treatment of the environment in ideal and actuality. Am. Sci. 58:244-249.

Turner, M. E., Jr., E. L. Bradley, Jr., K. A. Kirk, and K. M. Pruitt. 1976. A theory of growth. Mathematical Biosciences 29:367-373.

Wagener, K., and H. Förstell. 1972. Relaxation phenomena in the biological carbon cycle under conditions of variable atmospheric CO_2 content. Z. Naturforsch. Teil A 27:812-815.

Walton, A., M. Ergin, and D. D. Harkness. 1970. Carbon-14 concentrations in the atmosphere and carbon dioxide exchange rates. J. Geophys. Res. 75:3089-3098.

Weinberg, A. M., and R. M. Rotty. 1977. Alternative long-range energy strategies. pp. 225-239. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Whitfield, M. 1974. Accumulation of fossil CO_2 in the atmosphere and in the sea. Nature 247:523-525.

Whittaker, R. H. 1975. Communities and Ecosystems. Second edition. Macmillan Publishing Co., New York.

Whittaker, R. H., and G. E. Likens. 1973. Carbon in the biota. pp. 281-302. In Woodwell, G. M., and E. V. Pecan (eds.), Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge, Tennessee.

Whittaker, R. H., and G. E. Likens. 1975. The biosphere and man. pp. 305-328. In Lieth, H., and R. H. Whittaker (eds.), Primary Productivity of the Biosphere. Ecological Studies 14. Springer-Verlag, New York.

Wilson, A. T. 1978. Pioneer agriculture explosion and CO_2 levels in the atmosphere. Nature 273:40-41.

Wong, C. S. 1978. Atmospheric input of carbon dioxide from burning wood. Science 200:197-199.

Woodwell, G. M. 1978. The carbon dioxide question. Sci. Am. 238:34-43.

Woodwell, G. M., and R. A. Houghton. 1977. Biotic influences on the world carbon budget. pp. 61-72. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

Woodwell, G. M., and E. V. Pecan (eds.). 1973. Carbon and the Biosphere. CONF-720510. AEC Symposium Series 30. Technical Information Center, Oak Ridge, Tennessee.

Woodwell, G. M., R. H. Whittaker, W. A. Reiners, G. E. Likens, C. C. Delwiche, and D. B. Botkin. 1978. The biota and the world carbon budget. Science 199:141-146.

World Meteorological Organization (WMO). 1977. Report of the Scientific Workshop on Atmospheric Carbon Dioxide. WMO-474. Secretariat of the World Meteorological Organization, Geneva, Switzerland.

Wright, R. D. 1974. Rising atmospheric CO_2 and the photosynthesis of San Bernardino Mountain plants. Am. Midl. Nat. 91:360-370.

Young, D. B. 1976. A wood famine? The question of deforestation in old regime France. Forestry 49:45-56.

Young, J. W., J. W. Brewer, and H. M. Jameel. 1972. An engineering systems analysis of man's impact on the global carbon cycle. pp. 31-39. In Sworder, D. D. (ed.), Systems and Simulation in the Service of Society. Simulation Council, Inc., La Jolla, California.

Zelitch, I. 1971. Photosynthesis, Photorespiration, and Plant Productivity. Academic Press, New York.

Zimen, K. E., and F. K. Altenhein. 1973a. The future burden of industrial CO_2 on the atmosphere and the oceans. Naturwissenschaften 60:198-199.

Zimen, K. E., and F. K. Altenhein. 1973b. The future burden of industrial CO_2 on the atmosphere and the oceans. Z. Naturforsch. Teil A 28:1747-1752.

Zimen, K. E., P. Offermann, and G. Hartmann. 1977. Source functions of CO_2 and future CO_2 burden in the atmosphere. Z. Naturforsch. Teil A 32:1544-1554.

Zinke, P. J. 1977. Man's activities and their effect upon the limiting nutrients for primary productivity in marine and terrestrial ecosystems. pp. 89-98. In Stumm, W. (ed.), Global Chemical Cycles and Their Alterations by Man. Dahlem Konferenzen, Berlin.

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APPENDIX

CSMP PROGRAM

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*
** SIMULATION OF LAND-USE PATTERNS AFFECTING THE GLOBAL CARBON CYCLE **
*
*
RENAME    TIME=YEAR
FIXED   KASE
*
*****THE FOLLOWING MACRO COMPUTES THE ENVIRONMENTAL EFFECT (EF) OF NPP.
*
MACRO     EF = RESPON (P,GM,OM,VA)
  PROCEDURAL
      IF (VA.LE.GM)GO TO 90
      EF = 0.0
      GO TO 95
  90  IF (VA.GT.OM) GO TO 91
      EF = 1.0
      GO TO 95
   91 EF = TANH (P*(GM-VA)/(GM-OM))
  95  CONTINUE
ENDMACRO
INITIAL
*
*****ANNUAL PRODUCTION OF FOSSIL CARBON (GT/YR) FROM 1860 TO 1974.

*****FROM KEELING (1973) FOR PERIOD 1860 TO 1949.
*****FROM ROTTY (1977) FOR PERIOD 1950 TO 1974.
*
FUNCTION  CO2= (1860.,0.0933),(1861.,0.0987),(1862.,0.0984),...
               (1863.,0.1060),(1864.,0.1151),(1865.,0.1219),...
               (1866.,0.1287),(1867.,0.1379),(1868.,0.1367),...
               (1869.,0.1418),(1870.,0.1450),(1871.,0.1619),...
               (1872.,0.1759),(1873.,0.1884),(1874.,0.1838),...
               (1875.,0.1892),(1876.,0.1916),(1877.,0.1960),...
               (1878.,0.1965),(1879.,0.2076),(1880.,0.2271),...
               (1881.,0.2444),(1882.,0.2625),(1883.,0.2800),...
               (1884.,0.2821),(1885.,0.2764),(1886.,0.2787),...
               (1887.,0.2977),(1888.,0.3219),(1889.,0.3285),...
               (1890.,0.3497),(1891.,0.3654),(1892.,0.3687),...
               (1893.,0.3616),(1894.,0.3771),(1895.,0.3986),...
               (1896.,0.4116),(1897.,0.4315),(1898.,0.4546),...
               (1899.,0.4973),(1900.,0.5249),(1901.,0.5403),...
               (1902.,0.5529),(1903.,0.6064),(1904.,0.6134),...
               (1905.,0.6466),(1906.,0.6961),(1907.,0.7712),...
               (1908.,0.7366),(1909.,0.7690),(1910.,0.8048),...
               (1911.,0.8218),(1912.,0.8662),(1913.,0.9290),...
               (1914.,0.8384),(1915.,0.8308),(1916.,0.8948),...
               (1917.,0.9453),(1918.,0.9320),(1919.,0.8289),...
               (1920.,0.9589),(1921.,0.8280),(1922.,0.8906),...
               (1923.,1.0053),(1924.,0.9985),(1925.,1.0064),...
               (1926.,1.0060),(1927.,1.0975),(1928.,1.0909),...
               (1929.,1.1719),(1930.,1.0775),(1931.,0.9682),...
               (1932.,0.8738),(1933.,0.9188),(1934.,0.9965),...
               (1935.,1.0317),(1936.,1.1464),(1937.,1.2262),...
               (1938.,1.1614),(1939.,1.2329),(1940.,1.3004),...
               (1941.,1.3371),(1942.,1.3344),(1943.,1.3640),...
               (1944.,1.3522),(1945.,1.2036),(1946.,1.2705),...
               (1947.,1.4215),(1948.,1.5175),(1949.,1.4696),...
               (1950.,1.6652),(1951.,1.8060),(1952.,1.8387),...
               (1953.,1.9062),(1954.,1.9151),(1955.,2.0984),...
               (1956.,2.2372),(1957.,2.3353),(1958.,2.4215),...
               (1959.,2.5605),(1960.,2.7130),(1961.,2.6732),...
               (1962.,2.8103),(1963.,2.9738),(1964.,3.1492),...
               (1965.,3.2865),(1966.,3.4564),(1967.,3.5178),...
               (1968.,3.7420),(1969.,3.9485),(1970.,4.2337),...
               (1971.,4.3780),(1972.,4.5466),(1973.,4.7928),...
               (1974.,4.8907)
*
CONSTANT  KASE=1,     HRV=0.0
*
*****BIOLOGICAL GROWTH FACTORS FOR NONWOODS (O), NORTHERN WOODS (N) AND
***** SOUTHERN WOODS (S).
```

```
*
CONSTANT  BETAO=0.20, BETAN=0.15, BETAS=0.20
*
*****MULTIPLES OF INITIAL MASS/AREA.
*
CONSTANT THETA =1.5
*
******POOL SIZES IN UNITS OF GIGATON (=10E15 G) AT STEADY STATE.
*
INCON XIC1 = 508.0,        XIC2 = 90.0,        XIC3 = 1.5
INCON XIC4 = 29.0,         XIC5 = 1620.0,      XIC6 = 592.0
INCON XIC7 = 650.0,        XIC8 = 6600.0,      XIC9 = 31820.0
INCON XIC10 = 34.0,        XIC11 = 490.0,      XIC12 = 65.0
INCON XIC13 = 560.0,       XIC14 = 48.0,       XIC15 = 580.0
INCON XIC16 = 1290.0,      XIC17 = 30.0E6,     XIC18 = 1000.0
INCON XIC19 = 6.6E6
*
*****FLUXES IN UNITS OF GIGATON PER YEAR AT STEADY STATE.
*
PARAMETER F0102=22.5,     F0107=90.0,     F0110=10.0,     F0111=8.0
PARAMETER F0112=10.0,     F0113=8.0,      F0114=11.0,     F0115=9.0
PARAMETER F0201=22.5,     F0304=23.0,     F0307=8.0,      F0405=1.65
PARAMETER F0407=21.5,     F0508=1.0,      F0509=0.62,     F0519=0.03
*
PARAMETER F0607=0.32,     F0701=90.4,     F0703=31.0,     F0708=19.0
PARAMETER F0709=17.0,     F0716=0.65,     F0807=38.0,     F0809=40.0
PARAMETER F0908=58.0,     F1001=10.0,     F1101=7.85,     F1118=0.15
PARAMETER F1201=10.0,     F1301=7.85,     F1318=0.15,     F1401=11.0
PARAMETER F1501=8.86,     F1518=0.14,     F1609=0.38,     F1617=0.29
PARAMETER F1701=0.04,     F1707=0.23,     F1716=0.02,     F1804=0.08
PARAMETER F1806=0.29,     F1819=0.07,     F1904=0.07,     F1906=0.03
*
*****PARAMETERS FOR MODIFIED LOGISTIC EQUATION TO PROJECT FUTURE
***** RELEASE OF FOSSIL CARBON.
*
CONSTANT  PINF=7.5E3,   P0=137.259,    U1974=4.8907
CONSTANT   NLT = 137.259
*
NCSCRT
*
*****CALCULATE THE EXPONENT OR THE RATE IN THE MODIFIED LOGISTIC CURVE.
***** CASE 1: NOMINAL CASE (HIGH RATE OF FOSSIL FUEL CONSUMPTION)
*
      IF (KASE.GT.1) GO TO 1
      RATE = 0.0435
      EXPN = ALOG(1.0-U1974/(P0*RATE))/ALOG(P0/PINF)
       WRITE (25,901)
 901  FORMAT ('NAMEFOR19.DAT')
      GO TO 3
*
***** CASE 2: DELAYED FOSSIL FUEL CONSUMPTION
*
  1   IF (KASE.GT.2) GO TO 2
      EXPN = 0.1
      RATE = U1974/(P0*(1.0-(P0/PINF)**EXPN))
       WRITE (26,902)
 902  FORMAT ('NAMEFOR26.DAT')
  3   CONTINUE
      TEMP1=PINF**EXPN
      TEMP2=P0**EXPN
      TEMP3=(TEMP1-TEMP2)/TEMP2
      GO TO 4
  2   CONTINUE
*
***** CASE 3: SLOW BURNER--DECREASING RATE OF CONSUMPTION
***** CASE 4: COMBINATION SCENARIO--CASE 3 + BIOMASS MANAGEMENT
*
       IF (KASE.EQ.3) WRITE(27,906)
       IF (KASE.EQ.4) WRITE(28,907)
 906  FORMAT('NAMEFOR27.DAT')
 907  FORMAT('NAMEFOR28.DAT')
      RATE = 0.0435
```

```
      EXPN = ALOG(1.0-U1974/(P0*RATE))/ALOG(P0/PINF)
  4   CONTINUE
*
*****THE VOLUME TO SURFACE RATIO.
*
      B = 2.0/3.0
*
*****TRANSFER RATES AT STEADY STATE.
*
      A0102 = F0102/XIC1
      A0107 = F0107/XIC1
      A0201 = F0201/XIC2
      A0304 = F0304/XIC3
      A0307 = F0307/XIC3
      A0405 = F0405/XIC4
      A0407 = F0407/XIC4
      A0508 = F0508/XIC5
      A0509 = F0509/XIC5
      A0519 = F0519/XIC5
      A0607 = F0607/XIC6
      A0701 = F0701/XIC7
      A0708 = F0708/XIC7
      A0709 = F0709/XIC7
      A0716 = F0716/XIC7
      A0807 = F0807/XIC8
      A0809 = F0809/XIC8
      A0908 = F0908/XIC9
      A1001 = F1001/XIC10
      A1101 = F1101/XIC11
      A1118 = F1118/XIC11
      A1201 = F1201/XIC12
      A1301 = F1301/XIC13
      A1318 = F1318/XIC13
      A1401 = F1401/XIC14
      A1501 = F1501/XIC15
      A1518 = F1518/XIC15
      A1609 = F1609/XIC16
      A1617 = F1617/XIC16
      A1701 = F1701/XIC17
      A1707 = F1707/XIC17
      A1716 = F1716/XIC17
      A1804 = F1804/XIC18
      A1806 = F1806/XIC18
      A1819 = F1819/XIC18
      A1904 = F1904/XIC19
      A1906 = F1906/XIC19
*
      A0101 = A0102+A0107
      A0202 = A0201
      A0303 = A0304+A0307
      A0404 = A0405+A0407
      A0505 = A0508+A0509+A0519
      A0606 = A0607
      A0707 = A0708+A0709+A0716
      A0808 = A0807+A0809
      A0909 = A0908
      A1010 = A1001
      A1111 = A1101+A1118
      A1212 = A1201
      A1313 = A1301+A1318
      A1414 = A1401
      A1515 = A1501+A1518
      A1616 = A1609+A1617
      A1717 = A1707+A1716
      A1818 = A1804+A1806+A1819
      A1919 = A1904+A1906
*
*****TOTAL CARBON IN THE SYSTEM.
*
      TOTOC=XIC1+XIC2+XIC3+XIC4+XIC5+XIC6+XIC7+XIC8+XIC9+XIC10+XIC11...
          +XIC12+XIC13+XIC14+XIC15+XIC16+XIC17+XIC18+XIC19
*
```

```
      TRUN=0.0
      COUNT=0.0
      CFFRL = 2.145
      CFFRLT= 2.0537
      CCRL  = 0.0
      CDRL  = 0.0
      BICR12 = 0.0
      BIOR13 = 0.0
      BIOR14 = 0.0
      BICF15 = 0.0

***LAND AREA (IN MILLION SQUARED KM) AT STEADY STATE.
***TOTAL LAND AREA EQUALS 130.9 MILLION SQUARED KM.

      AC0 = 74.0
      AN0 = 27.9
      AS0 = 29.0

***CARBON MASS PER UNIT AREA OF TERRESTRIAL ECOSYSTEMS.

      Q100 = XIC10/AC0
      Q110 = XIC11/AC0
      Q120 = XIC12/AN0
      Q130 = XIC13/AN0
      Q140 = XIC14/AS0
      Q150 = XIC15/AS0

      ATM0 = XIC1+XIC2
      ATMLT = ATM0
      C3LT=XIC3
      C4LT=XIC4
      C5LT=XIC5
      C6LT=XIC6
      C7LT=XIC7
      C8LT=XIC8
      C9LT=XIC9
      C10LT=XIC10
      C11LT=XIC11
      C12LT=XIC12
      C13LT=XIC13
      C14LT=XIC14
      C15LT=XIC15
      C16LT=XIC16
      C17LT=XIC17
      C18LT=XIC18
      C19LT=XIC19

DYNAMIC
 NOSORT

'***COMPUTE CHANGE IN LAND AREA OF NORTHERN WOODS ACCORDING TO
'***  A SIMPLE LOGISTIC EQUATION.

      RAN = 0.00481814
      ANT = 20.0/(1.0+(-0.28315412*EXP(-RAN*(YEAR-1860.0))))
      DAN = RAN*ANT*(1.0-ANT/20.0)

'***COMPUTE CHANGE IN LAND AREA OF SOUTHERN WOODS
'***  1860 TO 1980 -- SECOND ORDER EXPONENTIAL EQUATION,
'***  1980 TO 2000 -- EXPONENTIAL EQUATION,
'***  2000 TO 2460 -- SIMPLE LOGISTIC EQUATION.

      IF (YEAR.GT.1980.0) GO TO 5
      RAS = -2.222112E-4*EXP(0.03*(YEAR-1860.0) )
      AST = AS0*EXP(-2.222112E-4*(EXP(0.03*(YEAR-1860.0))-1.0)/0.03)
      DAS =  RAS*AST
      GO TO 7
5     IF(YEAR.GT.2000.0) GO TO 6
      RAS = -0.008132538
      AST = 22.27841494*EXP( RAS*(YEAR-1980.0))
      DAS = RAS*AST
      GO TO 7
```

```
  6   FAS =  0.031007427
      AST = 15.0/(1.0+(-0.20778092*EXP(-FAS*(YEAR-2000.0))))
      DAS = FAS*AST*(1.0-AST/15.0)
   7  CONTINUE
      DAO = ABS(DAN)+ABS(DAS)
*
*****COMPUTE ANNUAL CO2 PRODUCTION FROM FOSSIL FUELS.
*****LINEAR INTERPOLATION OF VALUES BEFORE 1974.
*****AFTER 1974 AD, PROJECTION FROM SOURCE FUNCTION.
*
      IF(YEAR.GT.1974.0) GO TO 10
      U   =AMGEN(CO2,YEAR)
      GO TO 20
  10 IF (KASE.LT.3) GO TO 15
*
*****COMPUTE CARBON RELEASED FROM FOSSIL FUELS FOR CASES 3 AND 4.
*
      R=1.0/(74.0*(1.0-EXP(-0.00271*NLT)))
      CFFRL=FINF/(1.0+((FINF/F0 )**EXPN-1.0)*EXP(-R*EXPN*...
         (YEAR-1974.0)))**(1.0/EXPN)
      U=CFFRL*R*(1.0-(CFFRL/FINF)**EXPN)
      NLT = CFFRL
      GO TO 20
*
*****COMPUTE CARBON RELEASED FROM FOSSIL FUELS FOR CASES 1 AND 2.
*
  15  CFFRL=FINF/(1.0+TEMP3*EXP(-EXPN*RATE*(YEAR-1974.0)))**(1.0/EXPN)
      REFF = RATE*((TEMP1-CFFRL**EXPN)/TEMP1)
      U    = CFFRL*REFF
  20  CONTINUE
*
  SCRT
      ATM = C1+C2
*
*****COMPUTE CARBON DIOXIDE ENRICHMENT EFFECT ON NPP.
*
  PROCEDURE FEO,FEN,FES = BOUND1(ATM,ATM0,BETAO,BETAN,BETAS)
      EF1 = RESPON (3.1416,10679.25,2135.85,ATM)
      FEO = (1.0+BETAO*ALOG(ATM/ATM0))*EF1
      FEN = (1.0+BETAN*ALOG(ATM/ATM0))*EF1
      FES = (1.0+BETAS*ALOG(ATM/ATM0))*EF1
  ENDPROCEDURE
*
*****COMPUTE MASS DENSITY EFFECT ON NPP.
*
  PROCEDURE G10,G11,G12,G13,G14,G15 = BOUND2(Q10,Q11,Q12,Q13,Q14,Q15,...
           Q100,Q110,Q120,Q130,Q140,Q150)
      Q10GM = THETA*Q100
      Q11GM = THETA*Q110
      Q12GM = THETA*Q120
      Q13GM = THETA*Q130
      Q14GM = THETA*Q140
      Q15GM = THETA*Q150
      Q10OM =      Q100
      Q11OM =      Q110
      Q12OM =      Q120
      Q13OM =      Q130
      Q14OM =      Q140
      Q15OM =      Q150

      EF10 = RESPON (3.1416,Q10GM,Q10OM,Q10)
      EF11 = RESPON (3.1416,Q11GM,Q11OM,Q11)
      EF12 = RESPON (3.1416,Q12GM,Q12OM,Q12)
      EF13 = RESPON (3.1416,Q13GM,Q13OM,Q13)
      EF14 = RESPON (3.1416,Q14GM,Q14OM,Q14)
      EF15 = RESPON (3.1416,Q15GM,Q15OM,Q15)
      G10 = ((Q10/Q100 )**B)*EF10
      G11 = ((Q11/Q110 )**B)*EF11
      G12 = ((Q12/Q120 )**B)*EF12
      G13 = ((Q13/Q130 )**B)*EF13
      G14 = ((Q14/Q140 )**B)*EF14
      G15 = ((Q15/Q150 )**B)*EF15
```

```
  ENDPROCEDURE
*
*****RELEASE OF ORGANIC CARBON FROM DEFORESTATION.
*
      DM12 = ABS(DAN)*Q12
      DM13 = ABS(DAN)*Q13
      DM14 = ABS(DAS)*Q14
      DM15 = ABS(DAS)*Q15
*
*****LAND AREA OF NONWOODS.
*
      ACT = 130.9-(ANT+AST)
*
*****COMPUTE OCEAN BUFFER FACTOR ACCORDING TO REVELLE & MUNK (1977).
*
      ZETA = 9.0+4.0*((ATM -ATM0)/ATM0)
      FL0107 = A0107*C1
      FL0701 = A0701*(ZETA*C7-(ZETA-1.0)*XIC7)
*
*****NET PRIMARY PRODUCTION OF TERRESTRIAL BIOTA.
*
      FL0110 = F0110*FEO*G10
      FL0111 = F0111*FEO*G11
      FL0112 = F0112*FEN*G12
      FL0113 = F0113*FEN*G13
      FL0114 = F0114*FES*G14
      FL0115 = F0115*FES*G15
      FLAB   = FL0110+FL0111+FL0112+FL0113+FL0114+FL0115
*
*****SHIFT OF ORGANIC CARBON AFTER DEFORESTATION.
*
      FL1201 = DM12*0.3
      FL1210 = DM12*0.7
      FL1301 = DM13*0.2
      FL1310 = DM13*0.3
      FL1311 = DM13*0.5
      FL1401 = DM14*0.4
      FL1410 = DM14*0.6
      FL1501 = DM15*0.2
      FL1510 = DM15*0.35
      FL1511 = DM15*0.45
*
*****MODEL EQUATIONS.
*
      C1  = INTGRL(XIC1 ,DC1 )
      C2  = INTGRL(XIC2 ,DC2 )
      C3  = INTGRL(XIC3 ,DC3 )
      C4  = INTGRL(XIC4 ,DC4 )
      C5  = INTGRL(XIC5 ,DC5 )
      C6  = INTGRL(XIC6 ,DC6 )
      C7  = INTGRL(XIC7 ,DC7 )
      C8  = INTGRL(XIC8 ,DC8 )
      C9  = INTGRL(XIC9 ,DC9 )
      C10 = INTGRL(XIC10,DC10)
      C11 = INTGRL(XIC11,DC11)
      C12 = INTGRL(XIC12,DC12)
      C13 = INTGRL(XIC13,DC13)
      C14 = INTGRL(XIC14,DC14)
      C15 = INTGRL(XIC15,DC15)
      C16 = INTGRL(XIC16,DC16)
      C17 = INTGRL(XIC17,DC17)
      C18 = INTGRL(XIC18,DC18)
      C19 = INTGRL(XIC19,DC19)
      Q10 = INTGRL(Q100,DQ10)
      Q11 = INTGRL(Q110,DQ11)
      Q12 = INTGRL(Q120,DQ12)
      Q13 = INTGRL(Q130,DQ13)
      Q14 = INTGRL(Q140,DQ14)
      Q15 = INTGRL(Q150,DQ15)
*
      DCHT10=FL0110+FL1210+FL1310+FL1410+FL1510 -A1010*C10
      DCHT11=FL0111+FL1311+FL1511 -A1111*C11
```

```
      DCHT12=FL0112 -FL1201-FL1210-A1212*C12
      DCHT13=FL0113 -FL1301-FL1310-FL1311-A1313*C13 -HRV*C13
      DCHT14=FL0114 -FL1401-FL1410-A1414*C14
      DCHT15=FL0115-FL1501-FL1510-FL1511-A1515*C15 -HRV*C15
*
      DQ10 =  (AOT*DCHT10-C10*DAO)/(AOT*AOT)
      DQ11 =  (AOT*DCHT11-C11*DAO)/(AOT*AOT)
      DQ12 =  (ANT*DCHT12-C12*DAN)/(ANT*ANT)
      DQ13 =  (ANT*DCHT13-C13*DAN)/(ANT*ANT)
      DQ14 =  (AST*DCHT14-C14*DAS)/(AST*AST)
      DQ15 =  (AST*DCHT15-C15*DAS)/(AST*AST)
*
      DC1 = 0+FL0701+A0201*C2+A1001*C10+A1101*C11+A1201*C12+A1301*C13...
            +FL1201+FL1301+FL1401+FL1501                                ...
            +A1401*C14+A1501*C15+F1701 -A0101*C1-FLAB
      DC2 = A0102*C1 -A0201*C2
      DC3 = F0703 -A0303*C3
      DC4 = A0304*C3+A1804*C18+A1904*C19 -A0404*C4
      DC5 = A0405*C4 -A0505*C5
      DC6 = A1806*C18+A1906*C19 -A0606*C6
      DC7 = A0107*C1+A0307*C3+A0407*C4+A0607*C6+A0807*C8+A1707*C17   ...
            - FL0701-F0703-A0707*C7
      DC8 = A0508*C5+A0708*C7+A0908*C9  -A0808*C8
      DC9 = A0509*C5+A0709*C7+A0809*C8+A1609*C16 -A0909*C9
      DC10 = AOT*DQ10+Q10 *DAO
      DC11 = AOT*DQ11+Q11 *DAO
      DC12 = ANT*DQ12+Q12 *DAN
      DC13 = ANT*DQ13+Q13 *DAN
      DC14 = AST*DQ14+Q14 *DAS
      DC15 = AST*DQ15+Q15 *DAS
      DC16= A0716*C7+A1716*C17 -A1616*C16
      DC17= A1617*C16 -A1717*C17-F1701
      DC18= A1118*C11+A1318*C13+A1518*C15 -A1818*C18 +HRV*(C13+C15)
      DC19= A0519*C5+A1819*C18 -A1919*C19 +0
*
  NOSORT
*
*****CHECK RCONDING ERROR.
*
      TRUN=TRUN+1.0
      IF (KEEP.NE.1) GO TO 50
      COUNT = COUNT+1.0
      CPTC=C1+C2+C3+C4+C5+C6+C7+C8+C9+C10+C11+C12+C13+C14+C15 ...
           +C16+C17+C18+C19
      RESIDU=TOTOC-CPTC
      PERF = RESIDU/COUNT
      IF(SAMPLE(1860.0,2460.0,1.0).NE.1.0) GO TO 55
*
*****COMPUTE AIRBORN FRACTIONS OF EXCESS CARBON FROM ANNUAL RELEASE.
*
      IF (YEAR.LE.1974) CFFRL = CFFRL+0
      BINC=C10+C11+C12+C13+C14+C15-XIC10-XIC11-XIC12-XIC13-XIC14-XIC15
      ACRL  = FL1201+FL1301+FL1401+FL1501
      ADRL  = FL1210+FL1310+FL1410+FL1510
      APRL  = 0+ACRL
      AARL  = APRL+ADRL
      ABPA  = (ATM-ATMLT)/0
      APPA  = (ATM-ATMLT)/APRL
      AFDA  = (ATM-ATMLT)/AARL
*
*****PARTITION OF EXCESS CARBON FROM ANNUAL PRODUCTION.
*
      P3= (C3-C3LT)/AAPL
      P4= (C4-C4LT)/AAPL
      P5= (C5-C5LT)/AAPL
      P6= (C6-C6LT)/AAPL
      P7= (C7-C7LT)/AAPL
      P8= (C8-C8LT)/AAPL
      P9= (C9-C9LT)/AAPL
      P10= (C10-C10LT)/AAPL
      P11= (C11-C11LT)/AAPL
      P12= (C12-C12LT)/AAPL
```

```
      P13=(C13-C13LT)/AARL
      P14=(C14-C14LT)/AARL
      P15=(C15-C15LT)/AARL
      P16=(C16-C16LT)/AARL
      P17=(C17-C17LT)/AARL
      P18=(C18-C18LT)/AARL
      P19=(C19-C19LT)/AARL
   61 ACUT  = DM12+DM13+DM14+DM15
      ATMLT = ATM
      C3LT=C3
      C4LT=C4
      C5LT=C5
      C6LT=C6
      C7LT=C7
      C8LT=C8
      C9LT=C9
      C10LT=C10
      C11LT=C11
      C12LT=C12
      C13LT=C13
      C14LT=C14
      C15LT=C15
      C16LT=C16
      C17LT=C17
      C18LT=C18
      C19LT=C19
      CFFRLT = CFFRL
*
*****CUMULATIVE RELEASES OF BIOSPHERIC CARBON FROM DEFORESTATION.
*
      BIOR12 = BICR12 + DM12
      BIOR13 = BICR13 + DM13
      BIOR14 = BIOR14 + DM14
      BIOR15 = BIOR15 + DM15
      CCUT = BICR12+BICR13+BIOR14+BIOR15
*
*****COMPUTE AIRBORN FRACTIONS OF CUMULATIVE EXCESS CARBON.
*
      CCRL  = CCFL+ACRL
      CDRL  = CDRL+ADRL
      CPRL  = CFFFL+CCRL
      CARL  = CPRL+CDRL
   65 AFFC  = (ATM-ATM0)/CFFRL
      AFDC  = (ATM-ATM0)/CARL
      AFBC  = (BINC+CCRL+CDRL)/CARL
      AFPC  = (ATM-ATM0)/CPRL
*
*****PARTITION OF THE CUMULATIVE EXCESS CARBON.
*
      P3C  = (C3-XIC3)/CARL
      P4C  = (C4-XIC4)/CARL
      P5C  = (C5-XIC5)/CARL
      P6C  = (C6-XIC6)/CARL
      P7C  = (C7-XIC7)/CARL
      P8C  = (C8-XIC8)/CARL
      P9C  = (C9-XIC9)/CARL
      P10C = (C10-XIC10)/CARL
      P11C = (C11-XIC11)/CARL
      P12C = (C12-XIC12)/CARL
      P13C = (C13-XIC13)/CARL
      P14C = (C14-XIC14)/CARL
      P15C = (C15-XIC15)/CARL
      P16C = (C16-XIC16)/CARL
      P17C = (C17-XIC17)/CARL
      P18C = (C18-XIC18)/CARL
      P19C = (C19-XIC19)/CARL
   55 CONTINUE
      PARMOD = AMOD(YEAR,OUTDEL)
      IF (PARMOD.NE.0.0) GO TO 50
      IF (KASE.EQ.1)WRITE(25,903)YEAR,ATM,C1,C2,C7,C8,C9,C10,C11,...
       C12,C13,C14,C15,Q10,Q11,Q12,Q13,Q14,Q15,U,CFFRL,ACRL,ADPL,...
       CCRL,CDRL,CARL,FLO110,FLO111,FLO112,FLO113,FLO114,FLO115
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      IF (KASE.EQ.2) WRITE(26,903) YEAR,ATM,C1,C2,C7,C8,C9,C10,C11,...
       C12,C13,C14,C15,Q10,Q11,Q12,Q13,Q14,Q15,U,CFFRL,ACRL,ADRL,...
       CCRL,CDRI,CARI,FL0110,FL0111,FL0112,FL0113,FL0114,FL0115
      IF (KASE.EQ.3) WRITE(27,903) YEAR,ATM,C1,C2,C7,C8,C9,C10,C11,...
       C12,C13,C14,C15,Q10,Q11,Q12,Q13,Q14,Q15,U,CFFRL,ACRL,ADRL,...
       CCRL,CDRI,CARI,FL0110,FL0111,FL0112,FL0113,FL0114,FL0115
      IF (KASE.EQ.4) WRITE(28,903) YEAR,ATM,C1,C2,C7,C8,C9,C10,C11,...
       C12,C13,C14,C15,Q10,Q11,Q12,Q13,Q14,Q15,U,CFFRL,ACRL,ADRL,...
       CCRL,CDRI,CARI,FL0110,FL0111,FL0112,FL0113,FL0114,FL0115
  903 FORMAT (F5.0,5F15.5/5X,5F15.5/2F15.5,5F10.4/8F10.4/6F10.4)
   50 CONTINUE
*
TERMINAL
  METHOD STIFF
  TIMER YEAR=1860.0,FINTIM=2460.0,OUTDEL=5.0,DELMIN=1.E-12,DELMAX=1.0
  OUTPUT  ATM, U, CFFRL
  OUTPUT  C1, C2, C3, C4, C5, C6, C7, C8, C9
  OUTPUT  C10, C11, C12, C13, C14, C15, C16, C17, C18, C19
  OUTPUT  AFFA,AFFA,AFDA,AFFC,AFPC,AFDC,AFBC
  OUTPUT  P3,P4,P5,P6,P7,P8,P9,P10,P11,P12,P13,P14,P15,P16,P17,P18,P19
  OUTPUT  P3C,P4C,P5C,P6C,P7C,P8C,P9C,P16C,P17C,P18C,P19C
  OUTPUT  P10C,P11C,P12C,P13C,P14C,P15C
  OUTPUT  ABFA,AFPA,AFDA,AFPC,AFPC,AFDC,AFBC
  OUTPUT  ZETA,FL0107,FL0701
  OUTPUT  ACUT, CCUT,CCRL,CDFL, DM12, DM13, DM14, DM15
  OUTPUT  FL0110,FL0112,FL0114,FL0111,FL0113,FL0115
  OUTPUT  Q10,Q11,Q12,Q13,Q14,Q15
  OUTPUT  TRUN,COUNT,CPUC,RESIDU,RERR
  RERUN KASE
  END RERUN
*
  CONSTANT KASE = 2
  RERUN KASE
  END RERUN
*
  CONSTANT KASE = 3
  RERUN KASE
  END RERUN
*
  CONSTANT KASE = 4
  TIMER FINTIM=2000.05
  END CONTINUE
  TIMER FINTIM=2460.0
  CONSTANT HRV= 0.01
  CONSTANT BETAO=0.2 , BETAN=0.2, BETAS=0.25
  END
STOP
ENDJOB
```

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