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Assistant Secretary for Environment
Office of Health and Environmental Research

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Carbon Dioxide Effects Research and Assessment Program

Workshop on the Global Effects of Carbon Dioxide from Fossil Fuels

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United States Department of Energy

Assistant Secretary for Environment
Office of Health and Environmental Research
Washington, D.C. 20545

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Carbon Dioxide Effects Research and Assessment Program

Workshop on the Global Effects of Carbon Dioxide from Fossil Fuels

MIAMI BEACH, FLORIDA
MARCH 7-11, 1977

Edited by
William P. Elliott *and* **Lester Machta**
Air Resources Laboratories
National Oceanic and Atmospheric Administration

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Preface

The growing concern about the long-range consequences of carbon dioxide emissions resulting from ever-increasing fossil fuel combustion prompted the Division of Biomedical and Environmental Research of the Energy Research and Development Administration (ERDA) to undertake a thorough examination of the carbon dioxide problem. As an early step in this assessment, this workshop was convened under the auspices of ERDA's advisory committee, the Study Group on Global Effects of Carbon Dioxide, of which Alvin Weinberg is Chairman.

The workshop was held March 7–11, 1977, at the Konover Hotel, Miami Beach, Florida. Some 75 scientists were assembled to discuss the current knowledge of the CO_2 cycle and the consequences of increases in CO_2 content. They were also asked to identify significant gaps in our understanding and to recommend actions to fill those gaps. The first two days of the workshop were given over to invited (and some impromptu) talks. The speakers have provided written versions of the invited talks and they are reproduced here in Section 2. The third and fourth days were devoted to four separate panel deliberations during which the groups outlined the current state of knowledge and made suggestions for future work. On the fifth day, the workshop reassembled in plenary session to hear preliminary versions of the panel reports and to offer comments. These panel reports were put into final form by the panel chairmen in consultation with their members after the workshop ended. The four panels were:

I. Atmospheric CO_2,
II. Biological Effects,
III. Ocean Geochemistry, and
IV. Climate Effects.

Each panel prepared a report which included discussions of their specific topic. Since no attempt was made to impose a particular format on the panel reports, each is composed in a different style. In order to allow a comparison of the panel reports, the editors have therefore taken the liberty of preparing brief summaries of the major conclusions and recommendations. These summaries precede the actual reports. It should be stressed that our summaries are not to be thought of as substitutes for the reports themselves. The reader should consult the actual products of the panel members for details of their recommendations.

Implicit in all the panel reports is the acceptance of increasing atmospheric CO_2 content, well documented since 1958 and most probably the case since the industrial era began. That this rise has paralleled the increase in fossil fuel usage and is roughly equal to half the CO_2 liberated by industrial activity was also accepted. That fossil fuel usage is the sole cause of the increase, however, is under dispute: some significant fraction *may* be attributable to a decrease in the size of the biosphere. It also seems certain that there is enough fossil fuel still available to raise manyfold the level of atmospheric CO_2, if the current models are anywhere near correct.

It was also accepted that carbon dioxide's radiative properties are well enough known to say that its increase will warm the lower atmosphere. But the interactions and feedback mechanisms within the climate system are so complex that considerable uncertainty exists about the magnitude of the effects.

We believe everyone would agree that there is not yet enough understanding of this very complex issue to state with confidence that increased fossil fuel consumption will bring on catastrophic climate changes. On the other hand, the best current estimates do indicate potential problems; it behooves all to heed the warnings inherent in these calculations and support efforts to reduce the uncertainties of the predictions.

Dr. Göte Ostlund of the Tritium Laboratory of the University of Miami acted as host for the meeting. His efforts and those of his staff were deeply appreciated, especially when a crisis developed over the meeting location. The smooth running of the meeting was made possible by, among others, Barbara Stickney of the Tritium Laboratory and Vivian Normand of the Institute for Energy Analysis, Oak Ridge Associated Universities, Oak Ridge, Tennessee. Betty Wells and Jean Craig, of ARL, NOAA typed and retyped most of the material; their efforts are greatly appreciated.

William P. Elliott
Lester Machta

Air Resources Laboratories
National Oceanic and Atmospheric Administration

1. Panel Reports

Summaries

Panel I: Atmospheric CO_2

INTRODUCTION

This panel reviewed the adequacy of the current instrumentation and monitoring networks which have detailed the rise of atmospheric CO_2 since 1958. Data are needed not simply to record the changes in atmospheric CO_2 but to provide diagnostic and validation data for the modeling efforts which will be used to estimate future levels of CO_2, given the input from industrial and any other sources. In this respect, the reconstruction of past CO_2 concentrations would be of value as well.

MAJOR FINDINGS AND RECOMMENDATIONS

A. Standards and Instrumentation
(Panel I, Section II)

1. New standard reference materials (SRMs) are needed because current SRMs present problems in assuring comparability of measurements made with different instruments.

Recommendations

a. The National Bureau of Standards should develop stable SRMs of CO_2-in-air with relative uncertainty of 0.1% or better.

b. The World Meteorological Organization International Center at Scripps Institution of Oceanography should be continued and supported at least as long as necessary to assure comparability of the Scripps standard and any new SRMs.

2. Independent techniques are needed to verify the precision and accuracy of current instruments. The nondispersive infrared technique is adequate but requires large air samples and frequent comparison with standard gases. There are a number of promising techniques which have not been explored.

Recommendation

Promising techniques, including gas chromatography and laser absorption spectroscopy, should be investigated both for atmospheric concentration and for isotopic composition.

3. There is a need to measure the vertical flux of CO_2 above a number of surface types. Several methods have been used but each needs improvement.

Recommendation

Instrumentation to measure the vertical flux of CO_2 should be developed.

B. Monitoring Network (Panel I, Section III)

1. Measurements of CO_2 at ground level are needed to determine horizontal difference over the globe. The present network needs augmentation at locations isolated from local influences.

Recommendation

A series of 10 to 20 additional ground-level stations should be established for routine monitoring. One or two of these stations should be in the equatorial region.

2. The present network emphasizes remote locations, but additional information on the annual cycle of CO_2 over major biomes would assist the modeling effort. (See also Panel II.)

Recommendation

The benefits and feasibility of establishing monitoring stations with wide capabilities in the major biomes should be investigated.

3. Vertical gradients of CO_2 in the troposphere could be used to estimate broad-scale sources and sinks.

Recommendation

Aircraft soundings of CO_2 concentrations should be attempted over areas such as major forests and oceanic upwelling regions. Improved techniques for measuring vertical fluxes on these scales should also be developed.

C. CO_2 Concentrations Prior to 1958

(Panel I, Section IV)

1. Knowledge of the history of atmospheric CO_2 concentration is needed to support modeling efforts and establish the natural variation of CO_2. Two possible techniques are available: measuring CO_2 trapped in ice of known age and measuring the isotopic composition of carbon from tree rings.

Recommendations

a. Studies on the mechanisms of occlusion and diffusion of CO_2 in ice as well as the isotopic composition of the the ice should be initiated.

b. Isotopic composition of dated tree rings should be determined. Studies are also needed on the temperature dependence of the biological isotopic fractionation to assist in the interpretation of the isotope ratio.

2. It has been found valuable to be able to examine samples taken in earlier periods with instruments available today. Future improvements in instrumentation may permit more accurate measurements on samples taken today.

Recommendation

Consideration should be given to the establishment of a bank for selected samples for possible future reference.

D. Predicting Future Levels of CO_2 Concentration

(Panel I, Section V)

Prediction of future levels of atmospheric CO_2 will depend both on future CO_2 emissions and on models of CO_2 exchange between the major reservoirs. The scenarios of future CO_2 releases, from fossil fuel combustion and other sources, and the sensitivity and confidence in the CO_2 models require continuous assessment.

Recommendation

An active program of model development should be undertaken, including assessment of source terms for atmospheric CO_2 such as fossil fuel combustion, other industrial activities, and deforestation. This effort should include identifying weak points in the prediction scheme to guide research and observational programs.

Panel II: Biological Effects

INTRODUCTION

Predictions of the rise of atmospheric CO_2 depend heavily on knowing how the biosphere will respond to increased CO_2 coupled with climatic changes as yet undetermined. Complicating the picture even more are various land-use practices which some researchers suggested are reducing the size of the biosphere. Others dispute this last point, claiming the biosphere must be growing slightly. Until this point is settled, confidence in the models used to predict future CO_2 levels in the air will be minimal.

Although qualitative estimates of the response of the biota to increased CO_2 and associated climate change can be made, these estimates were felt to offer little guidance for the future. Changes in the biota, particularly agriculture, could be the most profound result of the additional CO_2.

MAJOR FINDINGS AND RECOMMENDATIONS

1. The size and current rate of change of the major biosphere carbon pools is not known. (Panel II, Section III A, C.)

Recommendations

a. The amount of carbon stored in the major biosphere carbon pools and the CO_2 fluxes into and out of them should be determined. Among the possible techniques are:

(1) satellite and high-altitude aircraft coupled with on-the-ground studies;

(2) extensive surveys of world-wide literature, records, and samples;

(3) isotope ratio techniques (see also Panel I, Section II); and

(4) improved instrumentation and monitoring (see also Panel I, Section II).

2. There is evidence that vegetation has the *potential* for increased production if CO_2 is increased, but this response depends on ecosystem type and availability of nutrients and other variables (Panel II, Section III A).

Recommendation

Research on the role of nutrient, water, and sunlight availability in the process by which CO_2 increases affect net carbon fixation should be carried out.

3. There are almost no methods of predicting ecosystem response to increased CO_2 accompanied by climate shifts. (Panel II, Section III B.)

Recommendations

Increased research activity in this area is needed. Factors to be examined include:

a. net productivity changes as both production and respiration change;

b. changes in organic carbon in the soil;

c. changes in effective moisture and any associated hydrologic and shoreline changes; and

d. indirect effects on such organisms as insects.

4. The mid and deep ocean areas appear to be weak sinks for *organic* carbon, but estuaries and near-shore regions may be significant sinks. However, this can be true only if organic material is quickly sequestered rather than recycled. (Panel II, Section III A.)

Recommendations

Research should be carried out to determine better:

a. the rate of oceanic biomass uptake of CO_2;

b. the capacity of present-day oceanic sinks of detritus carbon to absorb future CO_2;

c. the present-day detritus flux of carbon on the continental shelves; and

d. the possible long-term storage of carbon in estuaries and coastal regions.

5. There is the possibility that great increases in atmospheric CO_2 could affect the carbonate chemistry, and hence the biology, of the oceans. (Panel II, Section III A.)

Recommendation

The effect of carbonate chemical conditions, during periods of high atmospheric CO_2, on marine photosynthesis and planktonic species should be investigated.

6. There is a need to express the behavior of the biosphere in quantitative terms. (Panel II, Section III D.)

Recommendation

Greatly improved simulation models on all geographic scales should be developed. Such model development must go hand-in-hand with field research and specific validation tests.

Panel III: Ocean Geochemistry

INTRODUCTION

The oceans will not only be affected by any climate change but, by interacting with the atmosphere, they will help determine what the change will be. The oceans are also likely to be the ultimate repository of much of the excess CO_2. The absorption of this CO_2 is controlled not only by the three-dimensional circulation of the ocean but by the chemical and biological processes occurring within the ocean. These processes in turn are affected by the content and distribution of CO_2 within the oceans.

MAJOR FINDINGS AND RECOMMENDATIONS

A. **Ecological Impacts** [Panel III, Section II; see also Panel II, Section III A (5)]

1. A slight lowering of the pH of the ocean would probably accompany an increase in atmospheric

CO_2. This change is likely to be within the normal tolerance of most marine organisms.

Recommendation

Work should be done to identify those species which might be most sensitive to a lowering of pH and the role of these species in the marine food web.

2. A drop in the carbonate ion concentration in surface waters is to be expected. Such a drop would likely leave the surface waters still highly supersaturated with respect to most carbonate forms.

Recommendations

a. It should be determined whether carbonate precipitation by some organisms depends on a *high* degree of supersaturation.

b. Regions currently exhibiting lower than average carbonate ion content should be identified, since they may be affected early.

3. If significant changes in the global wind patterns occur, equally significant changes in ocean circulation will also occur. These changes might well result in a less vigorous vertical circulation and consequently in reduced nutrient supply to surface waters.

Recommendation

A substantial effort will be necessary to develop accepted ocean-atmosphere circulation models and to determine the implications of these models for biological processes.

B. Ocean Uptake

1. To be able to model the CO_2 cycle, much better knowledge of the horizontal distribution of vertical fluxes of CO_2 across the air-sea interface will be necessary. (Panel III, Section III A.)

Recommendations

a. A strong program to determine the time-space variations of CO_2 flux across the surface should be developed. This will require:
   (1) measurement of p_{CO_2} in the air and water,
   (2) measurements of the gas exchange coefficients,
   (3) measurements of a number of transient tracers in the ocean, and
   (4) measurements of total CO_2 and alkalinity.

b. At least one fully equipped research vessel should be stationed in northern waters for continuous monitoring.

c. An oceanographic expedition should be sent out in about 1985 along tracks of previous expeditions to measure the quantities listed under *a.* above, especially the transient tracers.

d. Remote sensing capabilities should be developed in order to obtain as much coverage as possible.

2. The rate-limiting step for removal of CO_2 from the air is the rate of mixing of surface waters with the deeper waters. (Panel III, Section III B.)

Recommendations

a. Many of the observations recommended in point 1 above will be needed to determine the rate of mixing, especially the study of the transient tracers. In addition, a careful feasibility study of intentional tracer injection should be undertaken.

b. The recommended observations must be accompanied by the development of a hierarchy of simulation models.

3. CO_2 may be entering the ocean more rapidly than current models predict if some of the shallow-water carbonate sediments, particularly high magnesium-calcite cements, are dissolving. (Panel III, Section III C.)

Recommendation

An effort should be undertaken to determine if shallow water sediments are now dissolving and, if so, whether they are dissolving in sufficient quantity to be currently affecting oceanic CO_2 uptake.

4. On the very long time scale it is the reactions with the deep sea sediments which will determine the capacity of the ocean to absorb CO_2. (Panel III, Section III C.)

Recommendations

a. Better thermodynamic and kinetic data for the dissolution of all carbonate in solid phase should be acquired, particularly at the pressures and temperatures of the deep seas.

b. The role of bioturbation, and its extent, in exposing bottom sediments to overlying water should be examined more thoroughly.

c. Greater understanding of the effects of p_{CO_2} on calcium carbonate production and survival of carbonate particles sinking from the surface should be obtained.

Panel IV: Climate Effects

INTRODUCTION

State-of-the-art simulations suggest that doubling the CO_2 content of the atmosphere could cause a 2 to 3°C increase in global mean temperature, with much larger warming in the polar regions. These models also suggest that precipitation would increase somewhat and shifts in the location of rain belts could occur. Reducing the uncertainties in these estimates as well as estimating the environmental (and societal) impacts of any changes is the major goal of the research program recommended.

MAJOR FINDINGS AND RECOMMENDATIONS

A. Assessing the Future Climate (Panel IV, Section III A)

1. Models are the only available tools to estimate the climatic impact of increased CO_2. Further model development is essential.

Recommendation

Climate modeling efforts should be pursued for a hierarchy of modeling approaches with groups devoted to each approach located at several locations.

2. Confidence in any model will be based on validating its sensitivity to CO_2 changes. Since each sensitivity cannot be directly checked against observations, because the CO_2 increase has not yet produced a clean climate signal, other ways must be found to validate the models.

Recommendation

More attention should be paid to validating climate models by comparing their sensitivity to known external factors. These factors could include seasonal forcing, volcanic dust veils, and sea surface temperature anomalies.

3. Cloudiness and precipitation are two of the most important climate variables as well as two of the most difficult to model. Clouds represent a possible negative feedback effect, and precipitation is vital to plant life.

Recommendation

Increased effort to model these variables should be undertaken.

B. Estimating the Consequences of Climate Changes (Panel IV, Section III B, D)

1. Some guidance as to what to expect in regional climate changes if warming does occur could be gleaned from the study of past climate.

Recommendation

Increased effort should be undertaken in paleoclimatology, particularly in past episodes of global warming. This would include studies of episodes in past millenia.

2. Regional climate anomalies do not necessarily follow the global pattern. The effects on local agriculture and water supply cannot be inferred simply from long-term global averages.

Recommendation

Investigation of the relationship of regional climate and climatic anomalies to large-scale patterns could help assess the regional effects of increased CO_2.

C. Monitoring (Panel IV, Section III C)

1. A CO_2 increase is only one of several possible external forcing functions on the climate system.

Recommendation

Studies and measurements of other forcing functions such as volcanic dust veils and variations in solar output should be undertaken to separate these possible causes of climate variability from CO_2.

2. Better knowledge of "normal" climate variability needs to be obtained to determine the noise level of the climate.

Recommendation

Studies of such climate variables as temperature, precipitation, and pressure should be continued, as well as descriptions of the cryosphere and tree ring response.

3. It is desirable to detect the CO_2 "climate signal" as soon as possible to provide timely warnings that CO_2 increases are indeed affecting the climate.

Recommendation

Special studies should be undertaken to discover any precursor signals to CO_2-induced climate changes and to monitor any variables that are so discovered. These precursors are any variables whose statistics are likely to exhibit relatively higher signal-to-noise ratios than other variables normally considered important.

C. Climatic Impacts on Society
(Panel IV, Section III E)

1. We need to know the vulnerability of various sectors of society to climate change.

Recommendation

Interdisciplinary studies of the dependence of our social systems on climate change should be initiated

Panel I: Atmospheric CO_2

Chairman, Lester Machta

Robert Bacastow
David Ballantine
David Gates
Kirby Hanson
Ernest Hughes
Tom Malone
Hans Oeschger
Graeme Pearman
Ralph Rotty
Bert Rust
C. S. Wong

I. INTRODUCTION

The sole evidence for the increase of CO_2 in man's environment derives from the monitoring of atmospheric CO_2 at a limited number of stations. The CO_2 increase is used to test models of future CO_2 concentrations and to uncover discrepancies in the CO_2 budget. The existing monitoring stations should be continued, but deficiencies in the measurement techniques and inadequacies of the network should be remedied. (Canadian ocean weather station "Papa" is threatened. Its loss would be harmful to both atmospheric and oceanic CO_2 monitoring and to studies of air-sea exchange.)

It will be essential to have a sound record of the atmospheric CO_2 concentrations over various parts of the world if it becomes necessary to manage the use of fossil fuels.

II. STANDARDS AND INSTRUMENTS

There are several reasons for recommending new standards and new instrumentation:

- Current standards, which consist of CO_2 in N_2, present problems in assuring accurate comparability of measurements made with different instruments.
- Independent analytical techniques are needed to verify the precision and accuracy of measurements made by the nondispersive infrared technique.
- A critical need exists for new instrumentation to determine the vertical flux of CO_2 in the air.
- Alternate sampling and analysis methods can reduce some operational problems.

A. Standard Reference Materials

The need for continuing long-term measurement of atmospheric carbon dioxide is evident. The availability of standards traceable to a single source throughout the period of measurement is essential to assure analytical control. Standards directly traceable to measurements of mass provide a common link which can be duplicated at any time in the future.

The National Bureau of Standards proposes to issue a set of standard reference materials (SRMs) consisting of carbon dioxide in air at concentrations bracketing the full range of atmospheric concentration. The SRMs will be analyzed by comparison to primary laboratory standards prepared by a gravimetric technique. The total uncertainty in the concentration of carbon dioxide in any sample will not exceed 0.1% relative.

The current measurements are referred to standards prepared and maintained at the Scripps Institution of Oceanography. An orderly integration of any new standards must be accomplished by intensive intercomparison between the new standards and the Scripps standards. A period of several years will elapse during which the primary standards for the

SRMs will be prepared and during which time the stability of the potential SRMs will be established. During this period it is essential that the current program and equipment, especially the manometer, at Scripps Institution of Oceanography be maintained.

Recommendations

The National Bureau of Standards should be urged to develop and make available stable SRMs of carbon dioxide in air with relative uncertainty of 0.1% or better.

The World Meteorological Organization International Calibration Center at Scripps Institution of Oceanography should be continued for at least as long as is necessary to assure comparability of the Scripps and NBS standards.

B. Instrumentation

Almost all background measurements of CO_2 in the gas phase have been made with a nondispersive infrared technique. While this method is adequate, it has several disadvantages which may be avoided by new but untried instruments. Among its shortcomings are the need to dry the relatively large air samples and the need for frequent intercomparison with CO_2 gas standards.

The production of gaseous standard reference materials requires, among other things, a highly precise method for comparing the SRMs to the primary standards. Gas chromatography lends itself quite adequately to the purpose when a detector of sufficient sensitivity is available for the particular species being measured. Until recently, such a detector was not available for carbon dioxide at atmospheric concentrations. However, the newly developed ultrasonic gas detection appears to have sufficient sensitivity to allow measurement of peak areas to better than one part in a thousand. The potential of this detector for the particular application has not been fully evaluated, but such evaluation could be performed in a relatively short period of time.

If gas chromatography proves practical for measuring atmospheric CO_2, several advantages might ensue. First, by proper column selection, the measurement can be independent of the major constituent. Second, the analytical system can be automated for essentially continuous analyses (at ~5-min intervals). Third, the consumption of standard gas is quite small, being about 0.005 liters/min.

A promising technique to measure atmospheric concentrations and the isotopic composition of carbon and oxygen in CO_2 in the same sample is laser absorption spectroscopy. As an example, the small bandwidth of tunable diode lasers allows one to resolve the Doppler-broadened absorption lines of the molecules with different isotopic composition. Another possibility is to use coincidences of emission lines of a CO_2 laser with absorption lines of CO_2. Such methods enable one to avoid interference with other atmospheric constituents; they would also enable one to measure simultaneously the concentration of a variety of minor atmospheric constituents such as CO and CH_4.

Recommendation

Promising techniques for the measurement of carbon dioxide and its isotopic composition should be investigated.

C. Instruments to Measure CO_2 Fluxes

The measurement of CO_2 fluxes in the atmosphere is of paramount importance. There is a great need for better instrumentation for measuring these fluxes.

There are several known methods for obtaining the vertical flux of a gaseous constituent. The aerodynamic method employs a vertical gradient of concentration multiplied by an eddy diffusion coefficient. The autocorrelation technique uses simultaneous measurements of the variable vertical air movements and the gas concentrations, allowing the determination of net vertical transfer of flux of the constituent. Some instruments using a modification of the autocorrelation technique exist in experimental versions for CO_2 flux determination.

Recommendation

Instruments to measure the representative vertical flux of CO_2 should be developed.

III. MONITORING NETWORK

Because of the significance of the buildup of atmospheric CO_2 and its implications concerning climatic change, it is recommended that more stations be established for monitoring atmospheric CO_2 concentrations over long periods of time. Insofar as possible, $^{13}C/^{12}C$ measurements should be made from air samples. Both political and scientific reasons dictate the necessity for the establishment of further stations of three types, as indicated below.

A. Stations Removed from Significant Local Sources

Measurements of CO_2 at ground or sea level are important for determining horizontal difference over broad geographical areas in order to locate the CO_2 sources and sinks. To this end, additional monitoring stations isolated from local influences are needed. In particular, stations located in the equatorial belt would shed light on the variable emission rate of CO_2 connected with equatorial atmospheric and oceanic parameters like sea surface temperature. This need places an extra requirement for interstation comparability.

Recommendation

A series of 10 to 20 additional stations, perhaps using flask sampling procedures, should be established to routinely monitor ground level CO_2 concentrations. One or two of these stations should lie in the equatorial belt, particularly in the Pacific Ocean.

B. Stations Representative of Major Biome Types of Sources and Sinks

The present stations are mostly located at oceanic coasts or islands, except for the South Pole station, and may reflect an annual cycle of CO_2 fluxes to or from the ocean surface as well as the influence of terrestrial vegetation through photosynthesis and respiration. The annual cycles must be better understood in terms of sources and sinks for CO_2.

An understanding of how the differing annual wave shapes contribute to the globally integrated annual waveform will give indications of the relative importance of various mechanisms for removing from or adding to the atmospheric concentrations of CO_2. This may help to resolve the current disagreement between ecologists and oceanographers concerning these matters.

Recommendations

We recommend the establishment of further scientific stations where excellent CO_2 monitoring can be done along with research concerning gradients of CO_2 and the identification of sources and sinks. We urge that at least one station be established within each major terrestrial biome, remote from regions likely to be disturbed by man. These biomes should include a tropical forest, an agricultural region, a boreal forest, a temperate forest, and a desert. We also recommend that additional stations be established within oceanic regions of expected marked CO_2 sources or sinks (e.g., regions of upwelling).

It is important that these stations have considerable scientific competence and that other closely related studies be conducted simultaneously. For example, there is a need to thoroughly investigate the fluxes of a variety of compounds which interact at the organic interface in the biosphere. Such compounds may include hydrocarbons and carbon monoxide, as well as those containing nitrogen and phosphorus.

C. Programs Established Specifically for the Determination of Vertical Gradients

Gradients of CO_2 in the lowest few meters of the atmosphere have been widely used to determine CO_2 exchange rates with vegetation on the microscale. Vertical gradients through a larger part of the troposphere have been used in a limited number of cases to estimate mesoscale surface exchanges. Such techniques could be used to establish the global distribution of CO_2 sources and sinks.

Recommendations

We recommend that attempts be made throughout the world to obtain aircraft soundings of CO_2 concentrations, particularly over the major forest areas and areas of ocean upwelling, and that efforts be made to improve techniques for measuring vertical fluxes on these scales.

IV. HISTORICAL RECORD OF CO_2 CONCENTRATION AND ISOTOPIC RATIOS

The level of CO_2 prior to 1958 and its natural variation is not well established. Efforts should be made to extend our knowledge of atmospheric CO_2 concentrations and isotopic composition back in time prior to 1958. Two possible methods for doing this are available. The first involves the measurement of CO_2 in ice of known age contained in polar ice sheets or glaciers. CO_2 trapped in this way should represent CO_2 conditions during the time of ice formation. The second method uses the isotopic composition of cellulose extracted from the growth rings of trees which are exposed to environments in which air composition is likely to be representative of the atmosphere as a whole. Although difficulties exist in the interpretation of data obtained with both methods, these studies are highly desirable. Collection and archiving of selected air samples for possible future reference may not be of high priority in the context of the broader CO_2 concentration problem but could be invaluable in studies involving the measurement of isotopes of carbon and oxygen and other potentially interesting gases.

Experience has clearly demonstrated the value of being able to examine samples taken in earlier periods.

Recommendations

Studies should proceed on the mechanism of occlusion and diffusion of CO_2 in ice and the concentrations and isotope ratios $^{14}C/^{12}C$, $^{13}C/^{12}C$, and $^{18}O/^{16}O$ of the ice. These data should be used to infer past atmospheric CO_2 concentrations and carbon exchange between major carbon reservoirs.

The isotope ratios $^{14}C/^{12}C$, $^{13}C/^{12}C$, and $^{18}O/^{16}O$ in dendrochronologically dated tree rings should be measured on suitably selected trees. Studies are needed of the temperature dependence of the biological fractionation process, and the data should be interpreted in terms of carbon exchange between major carbon reservoirs.

Consideration should be given to the establishment of a bank of selected atmospheric samples for possible future reference.

V. PREDICTING FUTURE ATMOSPHERIC CO_2 CONCENTRATIONS

A vital link in the CO_2 problem is the incorporation of physics, chemistry, and biology into a predictive model for atmospheric CO_2. Much of the output of the research programs recommended elsewhere in this report provides the basis for such models. The sensitivity and confidence in the predictions of future CO_2 in the formulation of modeling assumptions must be continually assessed.

Recommendation

An active program to develop CO_2 prediction models should be undertaken in order to assess the potential severity of the CO_2 threat and to guide the research program by identifying weak points in the prediction. The program should evaluate and continually update forecasts of future CO_2 releases and predictive models.

Panel II: Biological Effects

Chairman, Edgar Lemon

L. Allen	R. S. Loomis
Dwight Billings	Jerry Olson
Olle Bjorkman	N. J. Rosenberg
T. Bramryd	L. F. Small
C. C. Delwiche	Hans Suess
M. Drosdoff	Robert Watters
C. A. S. Hall	George Woodwell
K. K. Knoerr	P. J. Zinke

I. INTRODUCTION

Those who attended the ERDA Workshop at Miami Beach were acutely aware of the divergent opinions as to whether the biosphere is or is not contributing more CO_2 to the atmosphere than it is absorbing. This, of course, is an important question bearing on future trends and effects of added CO_2 to the atmosphere. Thus, considerable attention was given to this question by the Biological Effects panel at the cost of orderly discussion of all facets of the biosphere carbon budget. The panel focused on those pools and processes most affected by man and/or most significant to the net gain or loss of CO_2 to the atmosphere.

By all agreement the largest biomass pools are the soil organic matter, the tropical forest, and the boreal forests of Canada and Eurasia. Present lively controversy centers on the tropical forests. Is man's cutting of these enormous old pools overwhelming the whole photosynthesis process of the globe? The panel was divided on this issue. Inadequate sampling prevents immediate definitive conclusions. Thus, those who sought clear answers to the orginal basic question of net gain or loss of CO_2 between the biosphere and atmosphere were frustrated. They will be frustrated for some time to come. However, the problem is a clear-cut researchable one amenable to solution with adequate effort and new tools.

II. MAJOR CONCLUSIONS

A. Evidence collected for other purposes (International Biological Programme biome sites) suggests that there have been areas of the biosphere (e.g., temperate forests and prairies, including the soil) which have had a net release of organic carbon to atmospheric CO_2 in the past. Ecosystems such as tundra, taiga, and soils under intensive agriculture are probably accumulating carbon. Although much data exist in the literature, they are so scattered that it is impossible to conclude at present whether the total biosphere is currently accumulating carbon (increasing biomass and soil carbon) or is a source for atmospheric carbon (with a net decrease in biomass and soil). The greatest unknown is the present status of the tropical forests and soils.

B. Physiological evidence strongly indicates that most vegetation has the *potential* for increased production rates in response to increased CO_2 (e.g., 0.5% per 30 μbar CO_2 increase). The extent of such response and the net storage of carbon depend very strongly on ecosystem type and most particularly on available nutrients and other limiting variables.

C. At present we have almost no means of predicting quantitatively how ecosystems will respond or adapt to increasing CO_2 accompanied by temperature changes and/or major shifts in the hydrologic cycle. Given the environmental change, we can make qualitative predictions.

D. Estuaries and near-coastal seas appear to be sinks of unknown, but possibly significant, magnitude at present due to their enrichment by man's activities. However, the biomass produced here may be subject to rapid upwelling and loss from storage unless fecal pellets or sinking phytoplankton are somehow advected quickly out to the depths of the sea or immediately buried in permanent sediments.

E. The mid and deep ocean appear to be very weak sinks for organic carbon. The deep ocean could be an important sink for enhanced atmospheric CO_2, but only if massive new quantities of nutrients are made available.

Recommendations

1. The sizes of the existing biospheric carbon pools and their current rate of change (including sign) must be determined.
    a. Major biospheric carbon pool sizes and CO_2 fluxes need to be studied in detail in different ecological zones, i.e., tropical forests, temperate forests, taiga, tundra, grasslands, deserts, and agricultural systems including irrigated lands and shifting cultivation ("swidden" agriculture).
    b. Monitoring by satellite sensing and high altitude photography should be useful to measure areal extent and changes in vegetation in the different ecological zones. Extensive on-the-ground studies will be necessary to calibrate remote sensing information.
    c. Extensive literature surveys (e.g., of South America and Asia) and examination of existing records and samples (i.e., stored soil samples) would be helpful.
    d. Further development of isotope ratio techniques along with tree ring growth analyses should be encouraged in order to detect and interpret changes in growth or removal rate related to global CO_2 increase.
    e. There is need for better instrumentation and new methods to measure CO_2 flux to and from soils, vegetation, and water surfaces.
    f. There is need to expand the current CO_2 monitoring network.
2. Research will be required on the role of nutrient availability, water availability, and sunlight on the effect of CO_2 increases on plant productivity, particularly on trees.
3. We need to have a better means for assessing how ecosystems will respond and adapt to increased CO_2 accompanied by climate changes. This includes assessing future changes in geographic distributions of areas of major vegetation types including agricultural crops.
4. We need a better assessment of the role of the ocean biomass in CO_2 uptake and in particular the capacity of the oceanic sinks for detritus carbon.
5. Expressing the behavior of the biosphere in quantitative terms calls for much improved simulation models on all geographic scales.
    a. Models must successfully simulate soil-plant-atmosphere and aquatic-atmosphere systems, responses to CO_2 perturbation and climate and land use patterns.
    b. Validation tests must go hand-in-hand with model development.
    c. Modeling teams must be closely coupled with field researchers.
6. To avoid fragmentation and duplication of effort, close collaboration of concerned agencies, such as USDA, NSF, ERDA, and NOAA is imperative. Careful planning and coordinating the research on carbon and climate could well serve these agencies' different objectives simultaneously.

III. DISCUSSION

A. Ecosystems Undergoing Change

It is not yet clear whether or how fast the total world biomass is increasing or decreasing. Many biologists infer that total biomass (live) cannot be increasing, but arguments presented either way require objective evaluation and orderly estimates. CO_2 is probably enhancing photosynthesis and growth in some areas, while other areas are net sources for atmospheric CO_2. It is this mixture of opposing trends, the lack of samples for many areas (e.g., the Amazon basin), and the statistical problems of sampling that leave the balance in doubt.

The greatest uncertainties are the estimates of (1) the biomass of ecosystems, (2) respiration and other turnover in these systems, (3) present rates of photosynthesis, and (4) likely shifts in net balance. These processes need to be related to CO_2 change, to climatic change, and to processes (like succession) which normally shift ecosystem compositions. The rate constants for changes in pool size are particularly critical.

The rates of these processes on an areal basis vary with climatic, soil, relief, and biological conditions. Likewise, the accumulated net production of fixed carbon varies in relation to these factors as well as in relation to human activities. Because of their areal extent, soils and tropical forests are major elements, and they are the least understood in the world carbon budget.

The basic questions are due to both inadequate use of existing data and to deficiencies for present needs.

1. Forests

It has been estimated that 48% to 63% of world biomass is in tropical forests. The major data need with regard to the role of forests follows from high carbon per unit area and the large area covered by forests. Therefore, we need to determine forest types and area (stratified in relation to their efficiency in carbon fixation), actual intensity of process rates, storage quantities, and changes.

Thus, *areal* and *intensity-capacity* data are needed for the forest types in question in order to multiply area by intensity and sum the products for estimated totals.

Existing data for forest land use, forest harvest, stocking, growth rate, and growth potential need to be collected in the context of the carbon cycle. Where gaps exist in these data, efficient means of monitoring forest conditions and areal extent should be suited to the average requirements of the global carbon budget. Further detailed surveys could be extremely expensive; therefore, careful examination of priorities is needed.

Whether forests are sources or sinks of CO_2 depends on the type of forest and age class distribution following previous harvests (if any) or natural disturbances. The existing literature on forest yield and forest succession documents this well in terms of wood products. These data and suitable models could be translated into carbon fixation rates and growth trends expected for various forest conditions; the rates and trends could then be applied to the measured area.

There is a wide divergence of opinion on the possible overall effects of the harvest of tropical forests, ranging from the belief that the process is a source of CO_2 (because of oxidation following harvest) to the belief that it is a sink for CO_2 (and new storage).

The question needs to be resolved by studies carried out through the forest harvest cycle: from mature forest, to recently cut forest, to regrowth, and finally to harvestable old growth. The studies should monitor total carbon storage (including rapidly and slowly exchanging pools) and its changes through the cycle of forest utilization. Where forest is cleared for alternative uses, the corresponding carbon storages in the forest and accompanying soil should be determined to allow predictions to be extended to the areas being converted to pastures, farms, and towns.

2. Soil Organic Carbon and Calcium Carbonate

It has been estimated that there is 3×10^{18} g of organic carbon in soils of the earth's frost-free landmass down to a depth of 1 m. This is several times the amount of organic carbon in the biomass above the soil. The basic question is whether or not this large carbon reservoir is now, and for the foreseeable future, a net source or net sink for atmospheric CO_2 and to what extent. This pool is highly sensitive to changes in moisture and temperature and to variations in the input of organic residues.

If it is assumed that soil organic carbon in most natural ecosystems is in a steady state (additions balanced by losses), then the major loss or gain in soil organic carbon would occur as a result of man's activities, primarily agricultural land use. Whether or not these activities result in a gain or loss of soil organic carbon will depend on a number of factors such as kind of soil, climate, management, etc. With additions of chemical fertilizer, it is likely that the size of this pool is increasing.

When vegetation is cut and burned to prepare the land for agricultural use, the evidence in the literature generally indicates that little or no organic carbon is lost from the soil in the burning process. In fact, there may be a slight gain. Where loss of organic carbon has occurred after burning, it has been due partly to erosion and oxidation following burning of the vegetation.

When soil is planted to crops after the vegetation has been cleared and burned, the loss of organic carbon from the surface soil as a result of cultivation can be substantial, ranging from less than 20% to 50% depending on the kind of soil, climate, and management practices. After a number of years of continuous cultivation, equilibrium is reached and the rate of decomposition of soil organic matter equals the rate of residue addition. The length of time required to reach equilibrium varies primarily with management practices. In many instances, organic carbon content is increased under certain practices: if the soil is cultivated for only a few years and regrowth of vegetation occurs, the organic carbon content generally increases until equilibrium is reached, which is usually at a level below the original level of the soil.

The introduction of irrigation to semiarid regions, especially, and to subhumid regions as well can lead to a very significant increase in soil organic matter (1 to 2 w/v %).

Existing data on soils, litter, and decomposition should be brought together and aggregated in a way to document and extend the soil storage estimates for the various ecosystems.

Soils in arid regions which have precipitated calcium carbonate are potential sinks for CO_2. These constitute a comparatively large portion of the world's soils, but an accurate estimate of their extent and therefore of their potential capacity to absorb CO_2 is difficult. A more detailed mapping of these soils will help in evaluating their significance as a sink for CO_2. Routine determination (in laboratory studies) of the capacity of these soils for CO_2 uptake would be useful. Oxalates in succulent desert plants may be another carbon sink.

3. Wetlands

Wetland areas of the world are variable in area, both as a function of man's activities and as a function of changes in rainfall and other climatic factors.

In general, H_2O-saturated soils are major sinks for CO_2 because they strongly inhibit the decomposition of plant residues. Conversely, when drained they provide potent sources of CO_2 as a result of sudden decomposition of accumulated organic matter. Drained organic soils of this sort can provide an annual output of CO_2 by microbial oxidation to the extent of 2 kg carbon m^{-2} $year^{-1}$ or more depending on their organic content and other factors.

Drainage can be deliberate, such as when wetlands are converted to agricultural or other uses, or it can result from natural processes such as geologic uplifted land or changes in rainfall or other climate factors.

An accurate estimation of the extent of wetlands (by remote sensing or other methods) is a matter of considerable urgency. A continuous monitoring of areas of surface soil saturation would provide information on trends and would therefore make possible an evaluation of these areas as potential sources or sinks for CO_2.

4. River Discharge, Estuaries, Salt Marshes, and Coastal Ecosystems

Potentially, the outward flux of carbon in rivers could be 0.2 to 1.5×10^{15} g carbon/year, although interactions as river water flows through estuaries could be important either way. Estuaries presently appear to be a sink for atmospheric carbon, either directly via deposition in marsh sediments (particularly where the sea is rising against the land), or indirectly via fixation in marsh plants and via export to and deposition in coastal waters. There have been few measurements of the rate of either process.

Should man, in fact, burn all or most of the existing coal resources, or should even a small percentage of the ice caps melt and raise the sea level, our experience indicates that marshes would keep pace and lay down additional peat which would in some small way compensate for the additional CO_2.

Additional nutrients from anthropogenic sources could stimulate coastal productivity, which would have the potential of temporarily taking up considerable CO_2 from the air. However, it is not clear that anthropogenic nutrients are more important than natural cycles.

Analysis of nitrate (almost certainly limiting in coastal environments) from Long Island Sound waters was done in the early 1950s and again in the early 1970s. Comparisons of two comprehensive nitrate data sets show no change in the basic concentration patterns in Long Island Sound in a nearly 20-year interval, although this has been a region severely impacted by man and his activities.

One problem with viewing the coastal region as a sink is that many coastal regions (especially coastal plain regions with moderate to large fresh water advection) have circulation patterns characterized by shoreward-moving deep waters. Thus, any materials "sunk" in coastal regions may be upwelled later near shore unless there is a sedimentation process along the way.

5. The Oceans

The biosphere of the ocean as well as of the land may have to be invoked to account for:

a. seasonal changes of CO_2 observed in the atmosphere, and

b. part of the long-term changes of CO_2 observed in the atmosphere.

Present thinking seems to be that item *a* can be accounted for in terms of the land biosphere. This is a qualitative assessment. One study needed is a quantitative reappraisal of CO_2 injections into the ocean and uptake versus latitude for each month of the year for a "mean" year. It would also be of value to reappraise the influence of changing sea surface temperatures through the solubility effect.

Atmospheric CO_2 increases less rapidly when the tropical Pacific is cold than when it is warm, on a year-to-year basis (i.e., excluding the seasonal cycle from both variables). Some of this year-to-year variation in CO_2 could be interpreted as a response of the marine biosphere to variations in upwelling of nutrient-rich cold water along the equatorial Pacific encouraging photosynthesis at the surface of the ocean. Enhanced photosynthesis would take up more CO_2 from the atmosphere as organisms settle to the deep. The biospheric influence of water temperature change appears more important than direct solubility.

Another aspect of the problem for marine biota is the effect of future CO_2 increase on carbonate chemistry which, in turn, affects biological processes. If the pH of the surface ocean drops to 7.2 and p_{CO_2} increases to 1000 ppm, will the calcareous forms of life be subject to serious distress and might the marine photosynthetic rate be changed?

Future research is needed in the following areas:

a. assessment of the role of the oceanic biomass in the uptake of atmospheric CO_2;

b. the capacity of present-day oceanic sinks of detritus carbon to absorb future CO_2 release;

c. techniques of creating favorable ecological conditions for stimulating biological uptake of CO_2 and its conversion to carbon "storage";

d. effects of carbonate chemical conditions during future periods of maximum fossil fuel burning on marine photosynthesis and planktonic species;

e. effects on the marine biota of a flux of toxic material, e.g., mercury and polyaromatics released during the burning of fossil fuels;

f. an estimation of present-day detritus carbon flux accumulating on the continental shelf using dating techniques, e.g., ^{210}Pb, ^{226}Ra, and organic carbon measurements; and

g. improvement in the modeling of the global marine biota carbon cycle to narrow down the existing lack of understanding of the role of the marine biota and detritus.

Increasing net primary production in the ocean, with the intent of incorporating new fossil CO_2 input, might be accomplished by the introduction of new nutrients (nitrogen, phosphorus, silicon, and possibly some trace metals). However, the new incremental biomass must still sink below surface waters, presumably to the ocean bottom where it can be "stored" or react with $CaCO_3$ in sediments. If the oceanic biomass is to be an important carbon storage compartment, rapid injection into deep waters might be accomplished through zooplankton fecal pellets, since these fecal pellets sink rapidly whereas phytoplankton detritus does not. Furthermore, there is some evidence to suggest that fecal pellets have higher carbon/nitrogen ratios than living phytoplankton (i.e., higher than 7), so that preferential removal of carbon relative to nitrogen is achieved. Finally, as pellets sink, it appears that there is a more rapid remineralization of the nitrogen out of the pellets, relative to the carbon. Thus, there is a conservation of the limiting nutrient (nitrogen) through this process or, relative to the carbon storage problem, a preferential removal of carbon from upper waters.

B. Climatic Impacts, Risks, and Tradeoff

Results from research on carbon cycling and on climatic conditions will raise numerous questions about predicting the impact on the biosphere of changes in CO_2 and climate. Meanwhile, it is only prudent to explore the likely importance of some credible assumed shifts in climate: for example, shifts in climate belts by 1 to 3° C isotherms, with or without changes in moisture. This evaluation must include estimating the geographic extent and nature of changes such as the following:

1. productivity changes (both positive and negative) as production and respiration increase at different rates;
2. soil organic carbon changes;
3. loss of effective moisture related to changes in evaporation and perhaps changed storm tracks;
4. hydrologic and shoreline changes that would follow these changes in water balance;

5. possible direct biological changes among organisms in the biosphere and their interactions (e.g., best problems); and
6. effects on people living in the major climatic zones.

Sociopolitical and moral issues are likely to become extremely controversial, despite present complacency, long before the ultimate balance of benefits and costs or risks is estimated or even committed. The importance of the issue follows from the difference among social sectors, regions, and nations which will incur the risks, or possible benefits.

Analysis of these problems is to some extent clouded by a lack of information regarding the "normal" state of atmospheric CO_2 concentration and other variables relating thereto.

Extensive human (and other) activities such as land clearing for agriculture and deforestation for fuel, coke, and potash production have undoubtedly had their effect on atmospheric CO_2 levels. These questions have not been evaluated and should be.

C. Monitoring

Data on the changing global CO_2 concentration are supplied by observatories isolated from the major terrestrial sources and sinks for CO_2. Little information on the intricacies of biospheric activity is provided by such statistics. It now appears that at least some of the increasing CO_2 input to the atmosphere may be biogenic in origin. It also appears certain that the biosphere (or parts of it) may be a major consumer of the annual CO_2 increments to the atmosphere. Thus, it is essential that the annual patterns of CO_2 fixation (photosynthesis/respiration) be established for the major global ecosystems. Especially desirable for our understanding of the carbon cycle would be a description of the net photosynthesis and respiration terms as a function of water, temperature, and radiation (including above- and below-ground respiratory releases of CO_2). Also, a more accurate estimation of foliage and other important pools and processes is needed for many kinds of vegetation. Particular attention should be given to complete total carbon and nutrient budgets for the following:

1. tropical forests (evergreen and deciduous) and other tropical ecosystems of high productivity and their modifications by swidden management or by more intensive commercial management;
2. the major soil groups that have undergone change due to change in land use;
3. the extensive taiga and tundra of the North, where accumulation has probably prevailed but where net release would become more likely after global climatic warming;
4. agricultural systems, including irrigated agricultural areas which previously were desert or semiarid grasslands; and
5. areas in fairly close proximity to urban and industrial centers (e.g., temperate forests, grasslands, and crop-producing areas) where competing land uses for food and energy sources will be affected by energy technologies such as fossil fuel, biomass use, and other options.

Detailed and continuous analysis of standing biomass and annual cycles of these systems near a few research stations will have value in more accurate modeling. Such analysis is not primarily designed as the main evidence for global budgets, but it will be important in projecting how budgets may shift as CO_2, climate, and land use change. At these observatories regular and continuous measurements should be made of atmospheric CO_2 concentration and the CO_2 exchange between vegetation and soil surfaces and the atmosphere. The measurement of local climatic and biological parameters will be required to adequately interpret the CO_2 measurements in relation to the regional characteristics of the biosphere.

D. Modeling

The issues outlined above indicate the nature of our difficulties in measuring and predicting the response of the biosphere to environmental perturbations. Complexity is inherent in the diversity of ecosystems and in the large number of relevant elements and processes, both biological and environmental, within such systems. In addition, these systems are highly adaptive in structure and in response to external influences. Our need for expressing the behavior of the biosphere in quantitative terms calls for simulation models on local, regional, and global scales.

The broad goals of the modeling effort include:

1. the organization of existing information to identify clearly what we know and what we do not know;

2. the identification of questions requiring additional information and the construction of appropriate models—included here is the identification of specific priorities for research;
3. the study of integrative behavior in soil-vegetation-atmosphere ecosystems and aquatic-atmospheric systems with emphasis on the effects on behavior of variations in carbon dioxide, climate, and management strategies; and
4. full testing of models for prediction and simulation of the influence of CO_2 and climatic change in order to determine the complexity required in the models.

One critical need is for detailed process models of carbon cycling within marine and terrestrial ecosystems. The question of adaptation to changed climate as well as the need to deal with issues beyond our past experience dictate the development of models with a high degree of feedback and physiological explanation. Most "biome models" developed through the International Biological Programme effort are as yet inadequate for this purpose. The ecosystem models should be constructed so as to handle questions about changes in both CO_2 concentration and climate in terms of mechanisms; the models should give emphasis to realism in tracing processes controlling the dynamics of vegetative and soil carbon pools. A second issue relates to modeling of oceanic deposition relative to the deep ocean storage question.

The second identifiable need involves the formulation of global models of two sorts. One of these should deal with the distribution of vegetation types as a function of climate, and the second should deal with global carbon cycling. In these models, accuracy in response to climate, CO_2, and management is critical. This factor, plus the large scale of these models, suggests a structure based on empirical response functions drawn from the more detailed ecosystems model(s).

Past experience with large systems models has shown that modeling teams unfamiliar with the relevant processes can easily overlook critical existing or needed information; therefore, very close coupling of modeling and research is required. These considerations also exist, to a lesser degree, in coupling biosphere and climate modeling efforts.

E. Cooperation

It is necessary to consider seriously the possibility that rising atmospheric CO_2 levels might lead to such severe ecological consequences that we will need to arrest, or even reverse, the growth of atmospheric CO_2 over a short time span. Current proposals to short-circuit the carbon cycle by fertilizing the long-lived terrestrial biota, accelerating the downward flux of moving biological detritus, or physically collecting CO_2 and injecting it into the deep ocean all encounter ~~major~~ major obstacles, but require further evaluation. The massive scale and global cooperation necessary pose difficulties for all such schemes. Creative thinking and better understanding of the carbon cycle might produce more attractive possibilities. Control measures would have to be thought out carefully well in advance to have any hope of success. Clearly any plan for major environmental manipulation would have manifold side effects and should be developed for use only in the face of imminent ecological disaster.

The worst mistake that could be made now is simply to wait and thereby possibly invite disaster before putting the specialized pieces of research together into an integrated assessment. The timetables of decisions of fossil biomass, solar, and nuclear energy research, development, and demonstration cannot await the usual pace and mode of scholarly work. Each of the energy sources requires evaluation related to its implications for CO_2, ^{14}C, and/or climate.

While each specialized area of research moves ahead (carbon cycle, climatic change, climatic impact), it will have to connect to the other areas. In particular, output from improved models of the carbon cycle must serve as input to climate models. Output from climate models must serve as input to models of climatic impact on ecosystems, resources, and society.

The research needed for a realistic assessment of the responses of the biosphere to increases in the atmospheric level of CO_2 and to resulting changes in climate requires a massive effort. The cost of such programs would be high in terms of funding, scientific talent, and manpower. It may seem as if serious competition with other high-priority research areas such as the production of plant materials for food, feed, fiber, and fuel cannot be avoided. However, with careful planning and coordination, the research on carbon and climate could well serve these different objectives simultaneously, since the basic biological and geographic information needed is to a large extent the same. To avoid fragmentation of effort, close collaboration of interested agencies such as the U.S. Department of Agriculture, National Science Foundation, Energy Research and Development Administration, and National Oceanic and Atmospheric Administration is imperative.

Panel III: Ocean Geochemistry

Chairman, Wallace S. Broecker

C. F. Baes	Stephen Kempe
R. Burling	Gregg Marland
W. P. Elliott	Göte Ostlund
Arthur Fairhall	Wolfgang Roether
Robert Garrels	George Saunders
Manuel Fiadeiro	David Schink
Joris Gieskes	Minze Stuiver
Martin Hoffert	Eric Sundquist
William Jenkins	Taro Takahashi

I. INTRODUCTION

Ultimately, virtually all the CO_2 added to the atmosphere by the burning of fossil fuels will take up residence within the sea. We have considered three questions associated with this transfer:

- Will the excess CO_2 taken up by the sea have any adverse effects on marine life?
- For any given fuel use scenario what will be the time history of the distribution of the excess CO_2 between atmosphere and ocean?
- Does any feasible means exist by which the transport of CO_2 from the "smoke stack" to the sea could be accelerated?

The bulk of this report will be devoted to the second question.

II. ECOLOGICAL IMPACTS

In answer to the first question, our panel agreed that the alterations to be experienced by marine life are likely to be smaller than those alterations to be experienced by terrestrial life. Therefore, the land rather than the sea should be the primary focus for ecological studies. The three major influences we foresee on marine systems are as follows:

- A shift in ecology resulting from the global warming.
- A shift in ecology due to the pH change induced by excess CO_2. Two pathways can be envisioned: a direct effect on the metabolism of marine organisms and an indirect effect through a change in the ease with which calcium carbonate–precipitating organisms grow and maintain their skeletons.
- A decrease in total plant productivity caused by a decrease in the rate of vertical mixing within the sea associated with a polar warming. Such a retardation in mixing rate would cause a reduction of the rate of supply of the limiting nutrients to the surface ocean and a corresponding reduction in productivity to all trophic levels.

As marine organisms are short-lived and far more mobile than their terrestrial counterparts, they should adjust more "smoothly" to a change in temperature distribution. Also, the concern with changing water budgets so critical to terrestrial vegetation obviously does not apply to the sea.

The expected pH change (from 8.1 down to 7.8 for a two-fold increase in atmospheric CO_2 and no interaction with solids in sediments) lies well within what is likely the normal tolerance for most aquatic organisms. As exceptions to this generalization may exist, work should be done to identify species which would be adversely affected and then to determine how a decrease in their productivity might alter the makeup of the marine food web.

The carbonate ion change associated with a two-fold increase in atmospheric CO_2 (namely, a drop from 250 μm/kg to about 130 μm/kg) would still leave the surface ocean on the average about three-fold supersaturated with calcite and two-fold supersaturated with aragonite. Only those organisms precipitating high magnesium calcite (i.e., coralline algae, certain benthic foraminifera, etc.) would be affected. A threat to carbonate skeletons of most animals and plants would come only with a four-fold increase in atmospheric CO_2 pressure. However, local areas which now have lower than average CO_3^{2-} ion contents should be identified and considered in this respect. Polar, shelf, and upwelling areas are prime candidates. Also, steps should be taken to determine whether the open ocean carbonate precipitators in any way depend on a *high* degree of supersaturation to grow their shells.

As we understand very few of the details of the mechanisms causing ventilation of the deep sea, and hence the concurrent resupply of nutrients to the sea surface, only guesses can be made with regard to the magnitude of the changes a global warming might induce. A major effort must be made to improve our knowledge in this area.

III. OCEAN UPTAKE

The sea takes up CO_2 from the atmosphere largely because CO_2 reacts with the carbonate ion dissolved in the water, and with the carbonate ions stored in carbonate minerals lining the sea floor, to form highly soluble bicarbonate ion. Three types of kinetic barriers impede this reaction of atmospheric CO_2 with oceanic CO_3^{2-}. These are: (1) transfer of CO_2 across the air-sea interface, (2) transfer of CO_2 from surface waters through the thermocline, and (3) deep sea carbonate dissolution.

A. Gas Exchange Across the Air-Sea Interface

The magnitude of this rate is reasonably well known (±35%) from studies of the distribution of natural radiocarbon between air and sea, from studies of the distribution of bomb-produced ^{14}C between air and sea, and from studies of the deficiency of *in situ* generated radon gas in the oceanic mixed layer. The rate is so high that the sea surface currently maintains on the average about 85% saturation with the excess atmospheric CO_2 content (assuming that the uptake is dominated by reaction with dissolved CO_3^{2-} rather than sedimentary $CaCO_3$). Thus, our ability to do global modeling is currently not inhibited by uncertainties with regard to the gas exchange process. However, as the models evolve beyond their current one-dimensional character, a better knowledge of the *regional* pattern of uptake and release of CO_2 by the sea will become necessary. This will involve studies of both the geographic and annual patterns of surface ocean CO_2 partial pressure and studies of the dependence of gas exchange rates on wind velocity.

One aspect of the exchange problem needing special attention is the situation which exists in the areas where deep waters form (i.e., >45° latitude). The winter water which sinks from these areas may not be at equilibrium with the atmosphere. Since chemical equilibration between the dissolved inorganic carbon in a water column 100 m thick and CO_2 in the overlying air occurs with a time constant of about one year, we must know far more than we do about winter convection processes if we are to state how much excess CO_2 is being carried with this downgoing water.

The situation for isotopes (^{13}C and ^{14}C) is more complex because the isotopic equilibration time is one Revelle factor (i.e., ~10) times greater than that for chemical equilibration. Since the distributions of natural ^{14}C and bomb ^{14}C provide the important calibrations for the mixing models that are used to estimate the transfer of fossil-fuel CO_2 into the deep sea, this difference in response time leads to a possible source of error in the model-based estimates of the extent to which CO_2 has been transferred into the sea.

1. Necessary Measurements

The purpose of the program is to determine the time-space variation of the CO_2 flux into the oceans and to identify major CO_2 sink areas. In order to achieve this goal, the following studies are recommended:

a. measurements of the time and geographic variations of p_{CO_2} in the atmosphere and in the surface ocean,

b. measurement of the dependence of the gas exchange coefficient between sea and air by means of the ^{222}Rn method,

c. measurement of the time, geographic variations, and inventory of the bomb-produced ^{14}C in the atmosphere and ocean water, and

d. measurements of the time and geographic variations of the total dissolved CO_2 and alkalinity in the surface ocean water.

2. Observational Stations

For the selection of the geographic locations for monitoring stations, regions of high influx and efflux of CO_2 are to be considered, as well as the areas where an active reaction between the seawater and magnesium calcite is expected:

a. the source regions of deep water masses,

b. the source region of an intermediate water mass,

c. equatorial regions of high CO_2 efflux, and

d. continental shelf areas rich in magnesium calcite sediments.

Considering those requirements, we recommend the following program:

a. One fully equipped research vessel stationed in the northern North Atlantic Ocean and/or the northern North Pacific for the purpose of continuously monitoring the variables discussed above at the highest precision possible (i.e., p_{CO_2}, ΣCO_2, ALK, radon, primary productivity, etc.).

b. An oceanographic expedition in 1985–86 along the tracks of previous expeditions such as GEOSECS in the Atlantic and Pacific and Luciad Expedition of SIO in the Indian Ocean to observe the global change in the atmospheric and oceanic p_{CO_2}, total CO_2 in seawater, alkalinity, and the bomb ^{14}C in seawater and to further observe global variability in the gas exchange coefficient.

3. Present Status of Experimental Capability

a. Atmospheric p_{CO_2} and ocean surface water p_{CO_2}

The precision of CO_2 measurements achievable is ±0.1% for atmospheric p_{CO_2} and ±1% for surface seawater p_{CO_2}. The magnitude of the difference between the atmospheric and surface ocean water p_{CO_2} is ±20% in the equatorial areas and −20% in the high-latitude areas. The rate for annual increase in atmospheric p_{CO_2} is 0.5%, that for surface water p_{CO_2} should be about the same. Thus, the uncertainty of the surface water p_{CO_2} measurement is about twice the annual increment.

b. Alkalinity and total CO_2

Over the next ten-year period, 0.5 to 0.7% increase in the ΣCO_2 content of surface ocean water is anticipated. The presently achievable precision for ΣCO_2 is ±0.2%, and the accuracy is ±0.5%. Automation is at present difficult. No data for time variation is currently available. Although we do not know what the percentage alkalinity change will be, it cannot exceed twice that for ΣCO_2. The precision and accuracy for alkalinity are about those for total CO_2.

c. ^{222}Rn

^{222}Rn is measured by the alpha-counting method with a precision of about ±3%. A 20-liter water sample is required for a single analysis. However, a continuous ^{222}Rn measurement unit is being developed by W. Roether of the University of Heidelberg, West Germany. This unit will permit a continuous monitoring of ^{222}Rn gas with time.

d. ^{14}C

The method currently used calls for extraction of ^{14}C from a 200-liter water sample on board ship. The ^{14}C sample extracted on board is shipped to a land-based laboratory for analysis. Precision currently achievable for ^{14}C measurements is ±2% and is satisfactory for the purpose of the stated study.

4. Future Technological Development

a. Remote automatic station

Space technology may be applied to this area of study. Although expensive, development of an automated unmanned floating station which transmits data to a laboratory is feasible. Such a station is especially needed in order to obtain complete seasonal coverage in high-latitude regions.

b. Improvement of ion electrodes

The development of selective ion electrodes for p_{CO_2} and CO_3^{2-} would permit the bulky and power-hungry infrared gas analyzer to be replaced by a small, lightweight instrument.

B. Ocean Transport Processes

Currently the rate-limiting step for CO_2 removal from the atmosphere is almost certainly the rate of mixing within the sea. (The only situation in which this would not be true is if shallow water carbonates are dissolving at a very high rate; in this discussion this is assumed not to be the case.) Specifically, it is the rate of ventilation of the main oceanic thermocline (100 to 1000 m) which poses the major barrier. Little fossil-fuel CO_2 has yet reached the deep sea (i.e., >1000 m). The surface layer (i.e., 100 m) is nearly saturated. Our knowledge of this process remains largely descriptive. Only through analogy with the distributions of the bomb tracers, tritium and radiocarbon, have we been able to estimate how much CO_2 has entered this reservoir. Considerable uncertainty exists with regard to the pathways followed.

The only immediate hope for improving our knowledge lies in further measurement of the geographic and temporal distributions of the transient tracers added to the system by man, and perhaps through purposeful tracer injections into the thermocline.

Once the use of fossil fuels peaks and begins to decline, the thermocline will saturate with fossil-fuel CO_2 and the rate-limiting step will become the much slower removal to the deep sea. Saturation of the deep sea will occur on the time scale of hundreds of years rather than decades. Our knowledge of the actual rates comes primarily from the distribution of natural radiocarbon. Fortunately, the GEOSECS program is providing a detailed and highly accurate global map of the distribution of this isotope within the deep sea. However, we still lack adequate models of deep sea ventilation. It is the development of these models that is desperately needed. Until a breakthrough occurs in this area, our predictions of the decline in atmospheric CO_2 during the post-fossil-fuel era will remain semiquantitative.

As mentioned above, at the very time we need the sea's ability to carry away our pollution, we are also threatening to slow its stirring rate by decreasing the contrast between polar and equatorial temperatures. This adds to the difficulties of modeling. The models currently in use are analog models, in the sense that they are designed to reproduce the distribution of natural and man-made radioactivities rather than the physics of circulation. If a climate change slows the rate of ventilation, these models will overestimate the removal rate of CO_2. To avoid this, we will have to develop a new generation of models based on the actual physics of mixing. Coupled ocean-atmosphere models will have to replace the empirically calibrated analog models currently in use.

The resolution of uncertainties in oceanic uptake of anthropogenic CO_2 will require advances in techniques for modeling the transport and dispersal of tracers in various oceanic regimes. The aim of this activity should be to define our quantitative estimates of the anthropogenic CO_2 flux into the deep and intermediate waters to enable us to predict how this flux will evolve in the future.

The ultimate test of these models lies in oceanographic observations. That is, the validity of the models and the constraints under which they are constructed will be tested by the observed distributions of ocean tracers, the most valuable of which are the "transient tracers." Since their spatial distributions are not steady-state, these tracers yield valuable information on the dynamics of ocean transport processes. This also means that the general experimental plan should be concerned with mapping these tracers in time as well as in space. Further, since the tracer transients are currently decaying, it is imperative that tracer mapping should be initiated and continued now, in parallel with model development. Enough is currently known to allow effective experimental design, and feedback between the data, models, and experiments will occur quite naturally.

Specifically, there should be (1) regularly repeated regional surveys (on a five to ten-year repeat basis), (2) smaller more intensive short-term local studies of such episodic processes as water mass renewal, and (3) synoptic, long-term measurements at a few selected ocean sites. Further, less conventional experiments should be considered, such as intentional injections of artificial tracers into the oceanic environment. However, we feel that the political, sociological, and technological ramifications should be thoroughly explored before such a program is undertaken.

Currently, the list of technologically available oceanic tracers is as follows:

1. ^{14}C. Bomb-produced radiocarbon should follow the anthropogenic CO_2 pathways rather closely. It serves to elucidate long-term transport processes (with time scales up to centuries). Sampling and analysis is more difficult than for tritium, but this tracer is very important and should be monitored closely.

2. ^{3}H and ^{3}He. Tritium (^{3}H) is almost wholly bomb-produced. It is applicable to transport processes that occur on a decadal time scale. Its daughter product (^{3}He), when also measured, extends the process time scale down to months and promises to yield most valuable information on transport process physics. It is therefore recommended that both isotopes be measured intensively and in tandem.

3. Freons. These low-molecular-weight halocarbons are anthropogenic, and as far as we know are inert in the oceans. The input function of the freons parallels industrial development and thus will yield information of a different quality than that

from the radionuclides produced by nuclear weapons.

4. ^{85}Kr. This inert gas has an input function similar to that of the freons (being a product of nuclear fuel processing), but it is radioactive, with a half-life of ten years. This tracer has great potential but requires technologic development which we feel should be encouraged.

5. Certain short-lived nontransient tracers hold promise for studying chemical and particulate pathways: ^{228}Ra, ^{228}Th, ^{210}Pb, and ^{210}Po. This list is by no means exhaustive, but these isotopes are the more promising to date.

With regard to modeling, a heirarchy of approaches is possible, each of which aims to describe the transport processes at a different level of detail. These approaches can be separated into box and advection-diffusion models which prescribe ocean transport coefficients (currents, exchange rates, diffusivities, etc.) semiempirically, and dynamically interactive models which can, in principle, yield the circulation and tracer fields simultaneously from the governing equations and boundary conditions.

The box models are differentiated by the numbers of well-mixed reservoirs, their types, and exchange paths; the advection-diffusion models are defined by their dimensionality and their space and time resolution. The level and sophistication of modeling needed for the CO_2 problem is not entirely known at present, so it seems advisable to pursue various approaches simultaneously to determine whether their predictions of CO_2 uptake are substantially different.

C. Sediment Interactions

Roughly half the sediments on the sea floor contain carbonate minerals. When the water in contact with these sediments becomes sufficiently acidified with fossil-fuel CO_2, these carbonate minerals will begin to dissolve at a significant rate. This dissolution will permit an even greater fraction of the fossil-fuel CO_2 to enter the sea than would be the case in the absence of carbonate dissolution. It is likely that this process will take place concurrently with the ventilation of the deep sea. Therefore, we must develop the means to quantify this process and couple it with ocean ventilation models.

Two realms must be separately considered: (1) the surface ocean, which is highly supersaturated with the minerals calcite and aragonite but near saturation with high magnesium calcite cements prevalent in shallow water sediments, and (2) the deep sea, which lies near saturation with both calcite and aragonite. The shallow seas may even now be reacting with some of the carbonate minerals. If so, CO_2 may be entering the ocean somewhat faster than calculated by current models. However, as the area of deposits rich in these sediments constitutes only a few percent of the ocean floor and as the time available for dissolution is as yet small, the likelihood that this has happened is not large. Nevertheless, in view of the unaccounted for residual in fossil CO_2 budgets, this possibility must be given serious study. This can best be done by monitoring the alkalinity of the surface sea. However, because the dissolution of carbonate minerals in the ocean will become apparent only after a decade, laboratory studies of the rates of dissolution of real sediments must be undertaken so that we can soon get at least an order of magnitude estimate as to the possible importance of this process.

On a long time scale it is certainly the carbonates in deep sediments which are important. Since their dissolution is facilitated by the turbation generated by burrowing organisms as well as by molecular diffusion, we must study not only the chemical aspects of the kinetics of dissolution but also the mechanisms and rates of sediment stirring.

Dissolution in the deep ocean proceeds when carbonate phases are in undersaturated waters. As dissolution proceeds, dissolution products will build up in pore waters. Aqueous diffusion counters this buildup, but aqueous diffusion is relatively slow in pore waters and in the overlying boundary film (diffusive sublayer), so the actual zone of dissolution is confined to a thin layer on the top of the sediments. When overlying water is near saturation with the solid carbonate, then only moderate concentration gradients are possible and diffusion is slow.

Increasing the CO_2 content of deep waters reduces the CO_3^{2-} content and makes possible steeper gradients, faster diffusion, and more dissolution. We need quantitative estimates on the speed and extent of this effect. This requires good thermodynamic and kinetic data for the dissolution of all carbonate solid phases. We must have good information on solubility as a function of depth at deep-ocean salinities and temperatures. We must understand how dissolution rates vary with relevant oceanographic conditions, e.g., temperature, pressure, departure from saturation, mineral-grain coatings, surface area of carbonate particles, pore water phosphate concentrations, and other (as yet unknown) factors.

Just above the sea floor interface, a diffusive sublayer impedes the diffusive escape of the dissolution products. In this zone, as in pore waters, molecular diffusion is the dominant transport product. Above this zone, eddy diffusion provides more rapid and more effective dispersal. When dissolution of carbonate is rapid, the diffusive sublayer impedance can determine the extent of dissolution. The thickness of this diffusive sublayer depends on near-bottom physical processes, but the complete set of relationships is poorly known. We need to understand the effect of bottom roughnesses on sublayer thickness; we need to know friction velocities at the sea floor and what determines these conditions.

As carbonate dissolution is accelerated by the excess CO_2, old carbonate sediments will begin to dissolve. To the extent that these sediments contain noncalcareous components, a new "boundary-layer" will be created from the collection of these insoluble materials at the interface. This accumulation would provide a very effective impedance to further dissolution were it not for processes which mix solid phases at the sea floor. Chief among these processes is bioturbation (the disturbance of sediments by organisms). To predict the amount of $CaCO_3$ available to react with excess CO_2, we must know the depth extent and the rate of these sediment mixing processes. These are only poorly known at present.

The flux of $CaCO_3$ to the sea floor plays a critical role in determining the depth distribution and abundance of $CaCO_3$ on the sea floor and thus the availability of solid carbonates for dissolution. Future levels of atmospheric CO_2 may affect the rate at which particulate carbonates form and the fraction which arrives at the ocean floor. We have only a poor understanding of the effects of p_{CO_2} on calcium carbonate production and survival during sinking. We will need to know much more.

Our studies should not be confined to processes as they occur at present. The geological record contains abundant evidence for past variations in important aspects of global carbon geochemistry. Some of these variations may be analogous to changes induced by fossil-fuel CO_2 and may provide important clues concerning global biochemical pathways. By increasing our understanding of past perturbations in the global carbon cycle, we should be able to better anticipate the perturbation caused by fossil-fuel CO_2.

Relevant information required:

1. geographical and depth distribution of carbonates on the sea floor, including the fractions of $CaCO_3$ present and the content of the more soluble magnesium calcites and aragonite (much of these data exist but must be compiled and analyzed);
2. dissolution rate data and solubilities;
3. ocean sediment mixing depths and rates;
4. bottom boundary layer conditions;
5. effect of p_{CO_2} in surface waters (and other changing conditions) on the delivery of $CaCO_3$ to the sea floor; and
6. effect of changes in past times on the dissolution and preservation of $CaCO_3$ in the sedimentary record.

III. ACCELERATION OF THE TRANSFER OF CO_2 TO THE SEA

It is conceivable that the peak in atmospheric CO_2 content produced by consumption of our fossil fuel reserves could be reduced by short-circuiting the transfer of CO_2 from the combustion site to the deep sea. This could be done, as suggested by Marchetti, by piping CO_2 from power plants to the deep sea. It might be done by fertilizing the ocean with the limiting nutrient substances (PO_4^{2-}, NO_3^-, H_4SiO_4, etc.). As roughly 1.4 kg (3 lb) of CO_2 is produced for each half kilogram (pound) of fuel burned, these or any other schemes will be extremely costly and will likely have highly undesirable side effects. However, as we do not now have and will not soon have any firm estimate of the environmental effects of the CO_2-induced climate changes, we must be prepared to cope with a situation in which these effects prove sufficiently harmful that the expense (and adverse side effects) of short-circuiting schemes becomes a secondary consideration. In order to be prepared, we must thoroughly explore all the possible means by which atmospheric CO_2 can be sequestered.

Panel IV: Climate Effects

Chairman, Stephen Schneider

Roger Barry	J. M. Mitchell
A. Bjorkstrom	Reginald Newell
Robert Cess	S. S. Penner
Tzvi Gal-Chen	V. Ramanathan
William Kellogg	Elmar Reiter
John Laurman	Michael Schlesinger
Michael MacCracken	David Slade
Syukuro Manabe	Richard Temkin

I. INTRODUCTION

Carbon dioxide increases observed over the past few decades and projected to continue could cause climatic changes which would be both unprecedented in modern climatic history and long lasting. The consequences of CO_2-induced climate changes on global agriculture, water supply, energy demand, and, not least, global sea level are less well understood than potential climate changes, even though there are reasons to believe that such consequences could be immense. For example, a doubling of atmospheric CO_2 may occur in roughly 50 to 75 years, and the CO_2 concentration in the atmosphere is likely to continue to increase beyond this level if fossil fuel consumption follows current projections.

State-of-the-art estimates based on climate models suggest that, for example, such a doubling of CO_2 can cause an increase in global mean temperature of roughly 2 to 3°C, probably with a larger temperature increase at polar latitudes. Based on these estimates, significant climate changes could occur by the end of the twentieth century. However, uncertainties in both theory and data render this conclusion tentative; it is quite possible that new information could lower or raise the climate model estimates of warming by a substantial factor. These possible climatic changes would occur in conjunction with natural fluctuations of climate of the sort which have been observed in the past.

Although 2 to 3°C changes in global average temperature may seem small, some modeling and observational studies suggest that, as a result of the accompanying changes in atmospheric circulation patterns, some regions would experience changes of considerably greater magnitude. Whereas certain regions on the globe might benefit from the ensuing climatic changes, dislocation of the currently existing climatic and weather patterns to which we have become accustomed could likely cause serious regional economic and agricultural disruptions. Such consequences can be anticipated because significant shifts in precipitation patterns and amounts are also indicated, although with the present state of the art we cannot yet predict what the local climatic changes will be.

Reducing uncertainties in estimates of both CO_2-induced climatic changes and their environmental and societal impacts (which indeed is a major purpose of this document) could ease considerably the decision-making process for energy system planning. Moreover, because of the rapidity of projected CO_2 increases and the complexity of the physical/social environment, it must be recognized that the challenge is to provide more definitive assessments of climatic effects *before* the atmosphere "performs the experiment" itself and to determine the actual timing and impacts of CO_2 increases.

In view of both the potentially large CO_2-induced climatic effects and the large uncertainties in state-of-the-art estimates of these effects, an important conclusion for energy system planners is the need to maintain energy supply and demand options as open as possible so that, as new information becomes available, shifts in energy strategy would cause minimal disruptions.

A brief description of the issues of CO_2-induced climatic changes and some recommendations follow.

II. ASSESSING THE CLIMATIC IMPACTS OF INCREASING CARBON DIOXIDE

A. Some General Climatic Considerations

1. The Climate System

Increases in atmospheric carbon dioxide concentrations have the potential to modify the climate system, which in the general definition includes the atmosphere, land surface, the ice- and snow-covered regions, and the biota, all of which undergo many complex and often simultaneous interactions. The following brief description of the climate system illustrates some of the complex interactions involved.

The system is driven by energy reaching the earth from the sun. Energy absorption occurs both at the surface (about half of the incident solar energy) and in the atmosphere (about 20% of the incident solar energy); the energy not absorbed is reflected to space. Absorption of radiation occurs in a very nonuniform manner dependent upon the types and distribution of various atmospheric constituents. The energy absorbed at the earth's surface is then either reradiated upward, evaporates water, or is directly transported to the atmosphere through heat convection.

Incident solar energy is not itself uniformly distributed over the earth because of the earth's spherical shape, its rotation, and its axial inclination as it travels an elliptical path about the sun. Nonuniform distribution is further influenced throughout the atmosphere and on the surface by the distribution of land and ocean, the height of the land (and ice), and the distribution of cloudiness (affected by the earlier factors as well as by the general circulation and smaller-scale convective processes in the atmosphere). Increases in CO_2 would modify this nonuniform heating pattern.

All of these factors lead to very uneven differential heating in both space and time, which is manifested as temperature gradients. These in turn drive a very complex nonlinear circulation system (nature's attempt to equalize energy distribution). Everyone experiences the results of these meteorological or climatic processes: winds, clouds, rain, snow, etc. These processes are in turn coupled to each other through positive or negative "feedbacks."

The concept of a feedback process is an extremely important one in the study of climatic interactions, since numerous feedback processes exist in the climatic system. Positive feedback processes enhance the effect of a perturbation while negative feedback processes dampen the effect. One example of a positive feedback process is the so-called "snow and ice temperature-albedo" feedback. Here, an increase in temperature, perhaps due to an increase in solar luminosity, would generally decrease the snow and ice cover, thus lowering the regional albedo (fraction of incident light reflected), which in turn allows increased absorption of solar energy, thereby amplifying the original warming. Following the same reasoning, an initial decrease in temperature would be amplified also. Another feedback process in the atmosphere is termed "cloudiness–surface temperature" feedback and is generally thought to be a negative feedback. It still inspires debate as to its arithmetic sign, although recent results suggest that this feedback may be less significant than earlier believed.

The total synergism or antagonism of the many simultaneously acting feedback processes still, unfortunately, remains largely uncertain. Despite our inability to evaluate quantitatively the combined effects of many potential feedback mechanisms, one can nevertheless surmise that a significant disruption of the natural energy flows in the climatic system may well change the climate. Presently, we are able to estimate the potential order of magnitude of the perturbation to the climatic system due to a perturbation in atmospheric CO_2 concentration by comparing the magnitudes of the perturbed and natural energy flows. Often this comparison is facilitated through the use of a mathematical model of the climatic system. These estimates then provide some idea of the size, and perhaps the timing, of a climatic change that might result from a particular perturbation.

Time scales involved in the manifestation of CO_2-induced perturbations to the climate system must be considered when dealing with both the issue of the detectability of these effects (discussed further in Sect. II.C.3) and the time lag of the various climatic perturbations after the actual atmospheric CO_2 perturbation. Examples of two time scales that we might have to consider are (1) changes in sea level perhaps occurring long after the initial perturbation, and (2) changes in the atmospheric (and oceanic) circulation regimes occurring at approximately the same time as the perturbation. Further complicating any estimate of the timing of potential effects, CO_2 production from fossil fuels is essentially a continuing process, but the rate of fossil fuel use can vary widely with changes in social, political, and economic conditions throughout the world.

2. The Climatic Past

Any discussion of climatic effects should place CO_2-induced effects in the context of past climatic fluctuations. A difficult problem in assessing the anthropogenic impact on climate is that the nonanthropogenic climatic changes (and system "noise"), both long-term and short-term, must be identified and separated out from the climatic record.

Over the past several hundred thousand years, which span several major glacial and interglacial stages, indirect evidence for variations in mid-latitude air surface and upper ocean temperatures indicates that there may have been long-term changes on the order of 8 to 10°C. Since the end of the last glacial maximum about 18,000 years ago, data also indicate that there was an increase in mid-latitude surface temperature of several degrees Celsius roughly 4000 to 8000 years ago. The record of just the last 1000 years indicates that mid-latitude temperature fluctuations of some 1 to 2°C have occurred, with a general dip in temperature from about the 16th to the 19th centuries (known as the Little Ice Age) being a prominent feature of that record.

All these temperature records have, of course, been based on indirect evidence. From analyses of instrumental measures of temperature (with coverage largely from the land areas of the northern hemisphere), Mitchell (1972) found about a half degree Celsius warming from 1900 to 1940 followed by a cooling of about half that amount. Although this record is based on instrumental observations taken at many points, there still remain rather large areas of the earth where no, or only a few, measurements exist, and "sampling error bars" on the order of perhaps several tenths of a degree Celsius should probably be imposed (Damon and Kunen, 1976).

Thus, the climatic record shows that many fluctuations in temperature, precipitation, evaporation, etc., either inferred from environmental records or taken from recent observational records, have occurred throughout the past. Since these fluctuations will almost surely continue into the future, there will be a problem in separating and identifying any anthropogenic change (a "signal") from the natural fluctuations ("noise").

3. Overview of Climatic Effects of Increased CO_2

The first-order effects of CO_2 on the atmosphere are associated with the radiative properties of CO_2. Although its concentration in the atmosphere is on the order of 3×10^{-4} parts per volume, its contribution to the radiative energy balance of the earth-atmosphere system is significant.

The radiative properties of carbon dioxide are well established; that is, its absorption bands are reasonably well known. Carbon dioxide absorbs and emits infrared radiation downward towards the earth, where it effectively heats the surface. The CO_2 also reradiates upward, but since atmospheric temperature decreases with height in most of the atmosphere (with radiation proportional to the fourth power of the temperature) effective upward radiation is reduced. Also associated with increased CO_2 are decreased stratospheric temperatures. Removal of all the CO_2 from the atmosphere would correspond to about a 10 to 15°C drop in global surface temperature as indicated from model calculations.

Changes in the atmospheric temperature structure can be expected to induce secondary changes in the atmospheric concentration of various other trace gases. For example, an increase in stratospheric ozone concentrations would probably result from the cooler stratospheric temperatures that would accompany the increase in atmospheric carbon dioxide concentrations. This would follow because the ozone production based on photochemical equilibrium increases with decreasing temperatures. One-dimensional radiative–photochemical equilibrium numerical model calculations by Boughner and Ramanathan (1975) and Luther, Wuebbles, and Chang (1977) indicate that a 2 to 5% increase in total ozone will occur for a doubling of CO_2.

Tropospheric warming may also lead to changes in both tropospheric and stratospheric water vapor. However, there is considerable uncertainty as to whether an increase or decrease will occur in the stratosphere, nor is it known how this might affect stratospheric cloudiness. Changes in stratospheric ozone which would result from changes in water vapor injection into the stratosphere are expected to be relatively small; based on calculations by Luther, Wuebbles, and Chang (1977), a 0.1% decrease in ozone would accompany a 1-ppm increase in water vapor concentration. Such induced changes in atmospheric composition could in turn further affect atmospheric dynamics and other processes. These further implications have received essentially no research attention.

B. Modeling the Climate

1. Purpose of Modeling and Hierarchy of Models

Since historical analogies do not exist for most of the environmental perturbations associated with human energy consumption, many scientists have directed their studies toward mathematical modeling. (Of course, other justifications for climate modeling include: (1) traditional experiments cannot be performed in the climate system; (2) important physical processes can be isolated for further analysis; and (3) models help to interpret observed data.) A number of approaches to climate modeling have been undertaken. A simplified panorama of the hierarchy of climatic modeling approaches is presented here.

At one end of the modeling hierarchy are models which are concerned with the investigation of a single physical mechanism or at most a few coupled mechanisms; these models are usually carried out in one spatial dimension. The purpose of such models is to gain an understanding of how these individual processes might affect climate. At the other end of the hierarchy are the models which incorporate as many of the interacting physical processes as possible. Because such models are generally time-dependent and spatially three-dimensional, they require comparatively large amounts of computer time. The complexity of these models often makes it as difficult to understand their results as it is to understand the real atmosphere. On the other hand, it is often difficult to determine whether the physical mechanisms considered in the simpler models would behave differently if they were considered over other spatial scales or averaging conditions, or in conjunction with other interactive mechanisms.

Dependence on more than one dimension in the simplest models is generally either ignored or "parameterized." That is, those processes in *any* climate model that occur on scales that may not be resolved by the model are either neglected or are treated in the manner of a statistical parametric representation (deemed a "parameterization"). Only a few vertical processes (e.g., radiative transfer) or horizontal processes (e.g., simplified meridional transport) are investigated in the single-spatial-dimension models. Two-dimensional models are usually in the meridional plane (vertical and latitude) but may also be in the horizontal plane (latitude and longitude). When three spatial dimensions and time are considered along with equations of motion, one has a model like the present General Circulation Models (GCMs). In all combinations of spatial dimensions, time may be either neglected for a steady-state approach or modeled as in an actual time-dependent simulation (see Schneider and Dickinson, 1974).

2. Model Verification

The degree to which the output of a model can be verified by being compared to the real system contributes to the confidence one places in a model prediction. One approach to verification involves performing a "simulation" or "variability" experiment. The suggestion here is that a model capable of simulating or reproducing many of the known means and variances in the climatic system is more reliable than a model that reproduces fewer components of the real system.

For instance, one might want a model to reproduce seasonal variations or variations in the strength and location of zonal winds. However, verifying the sensitivity of a model to a given perturbation in a boundary condition (such as an increase in CO_2) may not be possible, since this often requires data for comparison with the model which may not exist before the atmosphere itself "performs" the sensitivity experiment. It is also possible that a model which reproduces poorly some of the means and variances of important climatic variables may still predict the appropriate sensitivity to a particular forcing function. Or the converse may be true. For example, there are some one-dimensional, horizontally-averaged radiative transfer models that estimate a global temperature response to carbon dioxide perturbations which is similar to that of the single GCM experiments, yet the former do not reproduce seasonal cycles. On the other hand, often when a seasonal surface temperature variation occurs in a GCM, it may be the result of artificial forcing such as from prescribed sea surface temperatures and it may not contain the appropriate atmosphere-land-ocean couplings. Thus, confidence in model predictions should be based on a combination of observational evidence, scientific intuition, and the inclusion of sound physical principles verified on the basis of selected (and known) means and variances of the real climatic system.

3. An Attempt at Model Verification

Along the lines of verification, Cess (1976) recently attempted to gauge the basic sensitivity of the climate system based on climatological temperature and cloudiness data together with recent satellite irradiation measurements. Through this approach, he estimated the sensitivity, β, of the global average surface temperature ($\bar{T}_s$) to a solar constant ($\bar{S}$) change for the real climatic system.

$$\beta \equiv \frac{dT_s}{dS} \text{ (}\sim 140^\circ\text{C in the analysis of Cess, 1976).}$$

This simply means that a 1% change in solar constant induces a global average surface temperature change (ΔT_s) of about 1.4°C. This result is twice that which would be anticipated in the absence of climatic feedback mechanisms, suggesting that such feedbacks double the sensitivity of the climatic system to climate forcing mechanisms (e.g., CO_2). Similar results, employing seasonal changes for three latitude zones where the signal-to-noise ratio is high and where no snow and ice albedo feedback exists also corroborate this conclusion. Furthermore, these results provide a global sensitivity value for the climate which agrees well with that of many models, providing a reasonable hope that state-of-the-art models are useful for order-of-magnitude predictions of the global sensitivity of the climate system to change from carbon dioxide increases. Of course, regional changes of the system are not determined from this approach; further modeling and empirical studies are clearly needed.

4. A Radiative Convective Model Approach

Augustsson and Ramanathan (1977) recently reevaluated the increase in global mean surface temperature due to an increase in CO_2, employing a radiative convective model and incorporating several previously neglected absorption bands of carbon dioxide. A study comparing various published results with this analysis indicates that an estimate of the increase in global average surface temperature for a doubling of the CO_2 concentration is $\Delta T_s = 2^\circ C$. This computed value of ΔT_s is, however, extremely sensitive to assumptions within the model concerning climate feedback mechanisms. To cite but one example, two suggested means for the incorporation of clouds within a climate model are (1) that the cloud-top altitude be held fixed and (2) that cloud-top temperature be constant. The model calculations for these two cases are as follows:

fixed cloud-top altitude: $\Delta T_s = 2.0^\circ C$

fixed cloud-top temperature: $\Delta T_s = 3.2^\circ C$.

This clearly indicates a measure of uncertainty within such model endeavors. It must be emphasized, however, that neither of the above cloud prescriptions may be regarded as a limiting or bounding case. They represent merely two possible cloud models. Most prior CO_2 climate studies have employed the fixed cloud-top altitude condition.

5. CO_2 Doubling Results with a GCM

Sensitivity experiments dealing with a CO_2 perturbation have also been performed with a GCM, by Manabe and Wetherald (1975). Their comparison between a standard case and a case where the atmospheric carbon dioxide content was doubled showed a global average surface temperature increase of nearly 3°C. However, polar latitudes showed even larger surface temperature increases. Amplification of temperature changes in polar regions is in qualitative agreement with suggestions from past climatic changes (e.g., Flint, 1971). Polar temperature amplification may have significant implications, since the polar regions contain large continental ice sheets which, if melted, could alter the mean sea level. Snow and ice reflectivity may also provide a strong positive feedback mechanism for temperature fluctuations in the polar regions. In addition, the overall intensity of the hydrological cycle increased in the model, and significant variations in both the amount and patterns of regional precipitation occurred.

Although this model is one of the most detailed in existence to have been applied to the CO_2 problem, several key ingredients were missing: realistic geography, cloud-temperature feedback, aerosol interactions, oceanic heat transport, and seasonal variations. Including these effects could, of course, change the results; but the incorporation of these components into the model might either dampen *or* increase the climatic response obtained from this sensitivity experiment. Thus, considerable uncertainty remains.

C. Empirical Approaches

1. Possible Regional Changes

Regional climatic changes associated with atmospheric CO_2 increase should be investigated using empirical as well as modeling approaches. How global mean warming affects regional atmospheric (and oceanic) circulation and hence temperature, precipitation, and evaporation patterns remains to be answered. Of course, regional changes in the high latitudes could modify the cryosphere (ice and snow). Through the study of both paleoclimatic and more recent records, some indication of the potential changes can be found.

a. Effects in the extra-polar regions

Shifts of the semipermanent standing waves, shifts in the boundaries and characteristics of air masses, and shifts in the location of storm tracks may strongly influence mid-latitude regional climate. Temperature increases in the tropics might increase latent heat release, which in turn might affect the meridional circulation (namely the Hadley Cell), perhaps modifying tropical and subtropical regions. Monsoonal rainfall may well increase with CO_2 warming, as indicated from some paleoclimatic records and general circulation model results. In addition, there is no reason to believe that the variability of the climate (or weather) at a particular place would remain unchanged with increased CO_2. Diurnal, interdiurnal, as well as interannual variability might vary considerably from that of the present day, although considerable additional work is needed to clarify this picture.

b. Effects in the polar regions

Changes in air temperature at high latitudes, as a result of increased CO_2 content, are greatly complicated by the existence in winter of a shallow layer of cold air next to the surface. Surface air temperatures in winter are expected to rise, but the net effect in the arctic basin in summer is less certain, since the pack ice exerts such strong control on temperature.

Warmer winters in high latitudes may be associated with *more* snowfall, as a result of the increase in atmospheric moisture content. However, since the arctic snow cover typically disappears within seven to ten days in early summer, it is unclear whether the increased winter snow depth would be sufficient to shorten significantly the snow-free season. At high elevations on the Greenland and antarctic ice sheets, for example, some increase in net accumulation is possible. In middle latitudes, warming would elevate the snow line, reducing the extent of snow and ice masses.

In the long term (perhaps several centuries), a global warming will tend to thin the arctic pack ice and decrease its extent. If the ice were substantially reduced in extent and thickness, its long-term stability could be upset, leading to a potentially irreversible removal (SMIC, 1971). The climatic consequences of an ice-free arctic could be manifest throughout most of the northern hemisphere.

Geological and botanical evidence suggests that the Thermal Maximum about 5,000 years ago saw a northward advance of the arctic tree line, more arid regions in parts of North America, and more humid conditions in the subtropical desert belt. Global warming would also result in thawing of the permafrost layers underlying most of northern Canada and northern Asia. The effect would be most significant along the southern margins of discontinuous permafrost. Melting of this ground ice would reduce the present tendency to waterlogging of the overlying soil layers, with a potential for changes in the muskeg vegetation.

A serious question is the impact of warming on the major ice sheets of Antarctica and Greenland. Initially, increased winter snowfall could tend to augment their mass balance. However, this trend may subsequently reverse, given a continuing temperature rise, leading to net melting. The last major interglacial interval (about 125,000 years ago) saw the world sea level rise about 5 to 6 m above present; but the time scale involved in this change was of the order of 10,000 years. The 20th century has witnessed sea-level rising at 1 to 2 mm/year; but whether this has been in response to the global warming up to about 1940 is uncertain.

2. "Natural" Variability

The natural interannual variability of the general circulation of the atmosphere under present climatic conditions is quite large. To begin to understand the physical causes of atmospheric variability requires comprehensive refinements of our meteorological data bases, refinements which are not anticipated to come about in the near future. Nevertheless, a number of studies can be initiated to assess the possible shifts of climatic and weather variability from CO_2-induced warming. We can scan historical weather records looking for years with high and low wintertime values of the zonal available potential energy (A_z), a measure related to the pole-to-equator temperature difference. While an assessment of differences in global cloudiness and precipitation between these years may prove to be difficult in view of the poor data base, a study of sea surface temperature anomaly patterns, planetary long-wave behavior, storm tracks, etc., is within the realm of our present meteorological data base.

We may anticipate that global warming due to increased CO_2 will reduce the mean meridional temperature gradients (Manabe and Weatherald, 1975) and hence the mean seasonal level of A_z. Judicious comparison of GCM output and atmospheric flow patterns in years of low wintertime A_z might provide a first crude estimate of local and regional changes in climate to be anticipated as effects of CO_2-induced warming.

3. Detectability of Secular Climatic Change

It is also important to note that because of the problem of identifying climatic signals, certain predictable climatic changes may occur and not be detected for an extremely long time. This problem may arise for two reasons. First, geophysical processes have certain inherent time constants associated with them, which may be as short as a fraction of a second (e.g., molecular relaxation times) or as long as millennia (e.g., glacial size changes). For instance, chlorofluoromethanes released near the earth's surface as aerosol spray propellants may take years to gradually diffuse to the stratosphere where these chlorofluoromethanes will be degraded through the process of photolysis, thereby producing chlorine atoms believed to be involved in ozone destruction. This delay in the actual climatic effect (relative to the time of emission) is governed by the time scales of atmospheric transport and chemistry; thus, the ozone destruction may be delayed for many years after the initial release of the chlorofluoromethanes. Similarly, a glacial advance or retreat might be delayed long after a global cooling or warming event.

A second problem for the scientist wishing to establish the sensitivity of the real climatic system to an external perturbation relates to the time delay involved in establishing the detectability of the perturbation. Because of the inherent variability, an important signal of climatic change may not be evident for a long time because of the noise in the climatic system, again implying a latent period before the signal strength relative to the noise is sufficiently large as to be judged "detectable." For a signal of constant strength, statistical techniques (including ordinary time-averaging) do exist for increasing the signal-to-noise ratio, but sufficient data from observations or model experiments are still required. A case in which the signal-to-noise ratio would be expected to increase might be the gradual amplification of a global temperature increase from carbon dioxide which is monotonically accumulating in the atmosphere from fossil fuel combustion. (if, in actuality this dominates over other potential anthropogenic or nonanthropogenic changes). Even if this is the case, the ratio may not be judged statistically significant until the effect has reached what may be called a "threshold of detectability."(Perhaps "threshold" is too vague a term here and the appropriate concept should be expressed as a probabilistic confidence limit.) It is this "threshold of detectibility," which occurs at some chosen value of signal-to-noise ratio, above which the reality of climatic change is highly probable. The magnitude of the threshold would certainly depend on such factors as the climatic time under study, the inherent variability of the climatic system, the judgement of threshold by the observer, and the accuracy and spatial coverage of the instrumental observations of climatic variables.

One may compare this type of threshold to an "effects threshold" (a concept often used in toxicology), below which no effect is present as a result of a particular agent or insult. An effects threshold may exist for some anthropogenic insults or changed boundary conditions that perturb the climatic system sufficiently to excite some atmospheric or climatic instability (such as ice melting or cloud formation). However, it is more likely that any *large-scale* climatic changes caused by a continuously varying boundary condition or a continuously increasing insult (e.g., an anthropogenic emission of CO_2) will be characterized by a continuously increasing climatic change (with perhaps some latency delay). Thus, the concept of a threshold of detectability becomes more useful for the purpose of determining a climatic change than the more classical concept of an effects threshold.

Other time delays further complicate the recognition of changing climatic signals. For instance, the post–World War II cooling trend in the northern hemisphere is now believed to have bottomed out around 1965, but it was necessary for climatologists to observe the temperature trend for several years beyond 1965 to minimize the possibility that spurious annual variations were not being observed rather than a consistent trend spanning at least half a decade.

An additional delay in detecting temperature trends is created by the sheer volume of data which presently must be analyzed by a small number of persons and with limited facilities in order to establish hemispheric or global averages. This lag in analysis adds an additional delay in trend recognition.

III. RECOMMENDATIONS FOR RESEARCH STRATEGY

A. Modeling

Rationale. It is clear that the ability to predict the climate consequences of increased CO_2 will depend on the development of climate theory in general and climatic models in particular. Models are the *only* tools available to estimate the potential climatic consequences of increased CO_2, since such increases are unprecedented in modern history of climate and their effects cannot be predicted by empirical means. Since the present generation of climatic models is not able to provide *definitive* estimates of CO_2 effects, further model development is essential. Several comprehensive analyses of the problems and prospects for climatic modeling, including extensive lists of recommended research topics, have been completed recently: Report of the Stockholm GARP Study Conference, 1975; U.S. National Academy of Sciences, 1975; and World Meteorological Organization, 1977. These reports contain comprehensive reviews of the large number of specialized disciplinary studies that are required in the development of improved climatic models; they need to be recognized in the formulation of a CO_2/climate-oriented research program. In addition, however, recommendations specifically addressed to such a program are required. The latter are developed below and follow upon the general modeling research recommendations in GARP (1975) and NAS (1975). Such model development will contribute directly to improve estimates of CO_2 effects.

1. Climate models are used to estimate the effects of CO_2 increases on climate by a "sensitivity experiment" which compares the climatic statistics of a model that includes increased CO_2 with the statistics of an unperturbed control experiment. Since the climatic effects of the CO_2 increase to date have yet to be unequivocally identified from observations (which is also consistent with present model estimates), there cannot yet be direct validation of a model's sensitivity to future CO_2 increase. Confidence in most models has therefore been based on simulation experiments which test the model's ability to reproduce known features of the climate. Such "simulation experiments," however, do not ensure the applicability of the model for sensitivity experiments. The physical complexity of the model is not in itself sufficient to justify confidence in model predictions, since it is the sensitivity to CO_2 perturbations over a spectrum of time and space scales that should be validated. Therefore, more attention should be paid to validating climate models by comparing their sensitivity to external factors (other than CO_2) that have already occurred in nature and for

which some observational data on the real climatic responses are also available. Responses to a seasonal forcing, volcanic dust veils, sea surface temperature anomalies, or insolation regime changes (and the glacial-interglacial "response") are examples of known or suspected forcings and responses. Such sensitivity validation procedures need to be applied across the hierarchy of climate models, where the validation data are compared to model data on a scale greater than, or equal to, the smallest scales resolvable in the model under validation.

2. Cloudiness and precipitation are two of the most difficult variables to model; some models do not even include a hydrological cycle calculation. Vast gaps exist in the observations needed for validation. It is recommended that further testing of the precipitation performance of models be conducted, both through simulation and sensitivity experiments and validations. This is of particular importance to the CO_2 question since changes in precipitation, in addition to their obvious importance for the earth's energy budget, would have a major impact on society through effects on agriculture and water supply.

3. In view of the large uncertainties in climate model estimates resulting from gaps in knowledge of both data and theory, it is important to maintain sufficient intellectual diversity in the climate modeling community to ensure that the most promising avenues of approach to model development will be explored. Additionally, intercomparisons of results from different modeling groups will help to identify points of consensus as well as to point out areas for priority research. We recommend that climatic modeling efforts be pursued for a hierarchy of modeling approaches, with groups devoted to each approach being present at least at several separate locations. Only by such diverse and independent efforts can the uncertainties in the estimates of today's models be reduced. In this connection the desirability of centralized direction and coordination at both national and international levels of the diverse research groups working on the CO_2 question needs immediate study.

B. Paleoclimatic Studies

The very important question of the geographical patterns of climatic change that may accompany a general planetary warming attributable to CO_2 can be investigated through paleoclimatic studies of climatic changes in past centuries and millennia.

The period of climatic warming between 8,000 and 4,000 years ago (the so-called Thermal Maximum that occurred after the end of the last glacial period) can be studied by means of paleoecological and snow/ice stratigraphic analyses to reveal many details of the pattern of temperature and precipitation changes that took place over North America, Europe, the western USSR, and other continental areas.

A number of shorter warming episodes within the past millennium can be studied in the same way, taking advantage of many kinds of human recordings and proxy climatic indicators, to learn if the geographical patterns of climatic change during general warmings are similar from one warming episode to another. To whatever extent similarity is seen in geographic patterns of change during different episodes of warming, our confidence can be strengthened that a future general warming (such as one induced by CO_2 increases) might be associated with similar patterns of change.

C. Monitoring

Rationale. Analogously to the recommendations on climate modeling, we view monitoring as part of the overall research strategy for a climate program (e.g., as outlined in GARP, 1975 or NAS, 1975). Our recommendations here are thus intended only to supplement those more general plans with particular emphasis on assessing CO_2 implications on climate.

1. CO_2 increase is an external forcing on the climatic system; other external forcings such as volcanic dust veils or variability in solar energy output need to be monitored if a climatic signal from CO_2 is to be unambiguously recovered from climatic data.

2. Studies of climate data (e.g., temperature, rainfall, and pressure) from as many stations as possible over the last 50 to 75 years, as well as recent satellite data, and related environmental parameters (e.g., snow cover, sea ice extent, ground ice, mountain glaciers, and tree rings) should be continued, so that the natural variability of climate can be determined. Until this "noise level" is known, it will not be possible to ascribe any significance to the "signal" (or climate change) due to increasing atmospheric CO_2

content. In addition, clues for the reasons and mechanisms of natural climatic variations can be found by examining historical data. Such information will also benefit work on model formulation.

Although the recommendations embodied here could also be classified under the following category of "diagnostic studies," they are listed here because their chief intent is to recommend that we look for the possibility of discovering "precursor indicators" of CO_2-induced climatic change, that is, variables whose statistics are likely to exhibit relatively higher signal-to-noise ratios than most other variables normally considered to be important climatic variables.

D. Diagnostic Studies

The recommendations listed above in Sects. A.1, A.2, B, and C.2 are already, to a considerable extent, suggestions for special diagnostic studies. For example, investigation of the relationship of regional climates and climatic anomalies to the large-scale patterns that are calculated by most models may lead to important additional information on potential regional climatic changes due to increased CO_2 concentrations. The regional consequences of such changes are of obvious importance to local agriculture and water supply, as the severe climatic anomalies such as those of 1972, 1974, and 1977 have already demonstrated.

E. Climatic Impacts on Society

Regardless of the causes of climatic change, an essential aspect of assessing CO_2 impacts on society is the determination of the vulnerability of various sectors of society (e.g., agriculture, water supply, energy demand, or coastal demography) to any kind of climatic change. We recommend that interdisciplinary studies be undertaken on the dependence of our physical/social systems on climatic change, with emphasis on possible CO_2-related changes.

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2. Invited Papers

Introduction

This section contains the texts of the invited papers given at the workshop. In some cases the material is slightly different from the actual presentation, since the authors revised their material after the meeting.

In addition, some participants presented some of their own work either to the whole group or to one of the panels. Much of this work has been published elsewhere or is being prepared for submission; space limitations prevent the presentation of the material here. The titles and authors are given below, and further information on this work can be obtained from the authors directly.

T. Bramryd, "Drainage and Utilization of Peatland—Possible Effects on the Carbon Cycle"
"The Effects of Garbage Destruction on the Carbon Cycle"

Y. H. Chan and J. S. Olson, "Limits on the Storage of Carbon from Burning Fossil Fuels"

J. S. Olson, H. A. Pfuderer, and Y. H. Chan, "Changes in the Global Carbon Cycle and the Biosphere"

S. S. Penner, "On the Influence of Major OTEC Development on Atmospheric Cooling and Atmospheric CO_2 Concentration"
"A Monitoring Program to Measure Changes in the Mean Atmospheric Temperature of the Earth"

C. S. Wong, "Atmospheric CO_2 Input from Non-Fossil Wood Burning"

Present and Future Production of CO_2 from Fossil Fuels—A Global Appraisal

Ralph M. Rotty, Institute for Energy Analysis,
Oak Ridge Associated Universities, Oak Ridge, Tennessee

ABSTRACT

The level of carbon dioxide in the atmosphere is a problem of worldwide proportions. Unilateral action by any one nation in planning alternatives to fossil fuel use will most likely be ineffective in controlling carbon dioxide. Energy growth in the past has been largely based on fossil fuels, and consequently annual carbon dioxide production has increased steadily at a rate of 4.3%. In 1976, global carbon dioxide production contained more than 5 billion metric tons of carbon. Of this, 27% was a result of activity in the United States. By 2025 the total will have grown more than fivefold, and the developing countries and communist Asia will be producing over half the global total. The challenge to the United States is to develop energy supply systems not based on fossil fuels which can and will be used by developing nations.

The concentration of carbon dioxide in the atmosphere is a global problem which depends on emissions from all parts of the world. Unilateral actions taken by the United States are of limited value largely because of the changing fuel requirements of the developing areas of the globe. It is important in planning multinational policies to recognize where the fossil fuel burning is now occurring and is likely to occur in the next several decades.

As in the case of earlier work (Keeling 1973; Rotty 1973, 1976), the carbon dioxide emissions from fossil fuels have been calculated for fuel production data given in the United Nations (UN) *World Energy Supplies* series (United Nations, 1976). Although this 1976 publication gives preliminary fuel production values for 1975, these have been slightly revised here in accordance with later information from the UN Statistical Office (Ramsdell, 1977).

In 1973, Keeling estimated the amount of carbon released by the burning of fossil fuel, basing his work on the fuel production data reported by the UN at that time. The emission of CO_2 from fossil fuel sources based on more recent data is shown graphically in Fig. 1. The calculations here follow the procedures of Keeling and also add a small contribution from the manufacture of cement and use estimates of flared natural gas (Rotty 1973, 1974). With the publication of *World Energy Supplies, 1950-1974* (United Nations, 1976), fuel data were revised, updated, and made consistent within the series; these data were used in CO_2 production estimates presented at the Office of Naval Research Conference on Fate of the Fossil Fuel Carbonates in January 1976 (Rotty, 1976). Data for the last few years have been revised slightly and the record extended through 1976 in Fig. 1.

Production for 1976 has been estimated on the basis of data published by the UN (1977) for the first three quarters of 1976. Since there is a seasonal variation in fuel production as well as fuel use, estimates for the fourth quarter of 1976 were based on a ratio between third quarter and fourth quarter production developed for each fuel (coal, lignite, crude petroleum, and natural gas) for the three-year period 1973 through 1975.

For example, coal production for the third quarters of each year—1973, 1974, and 1975—totaled 1325 million metric tons, and during the fourth quarters it totaled 1372 million metric tons. Production for the fourth quarter of 1976 was then

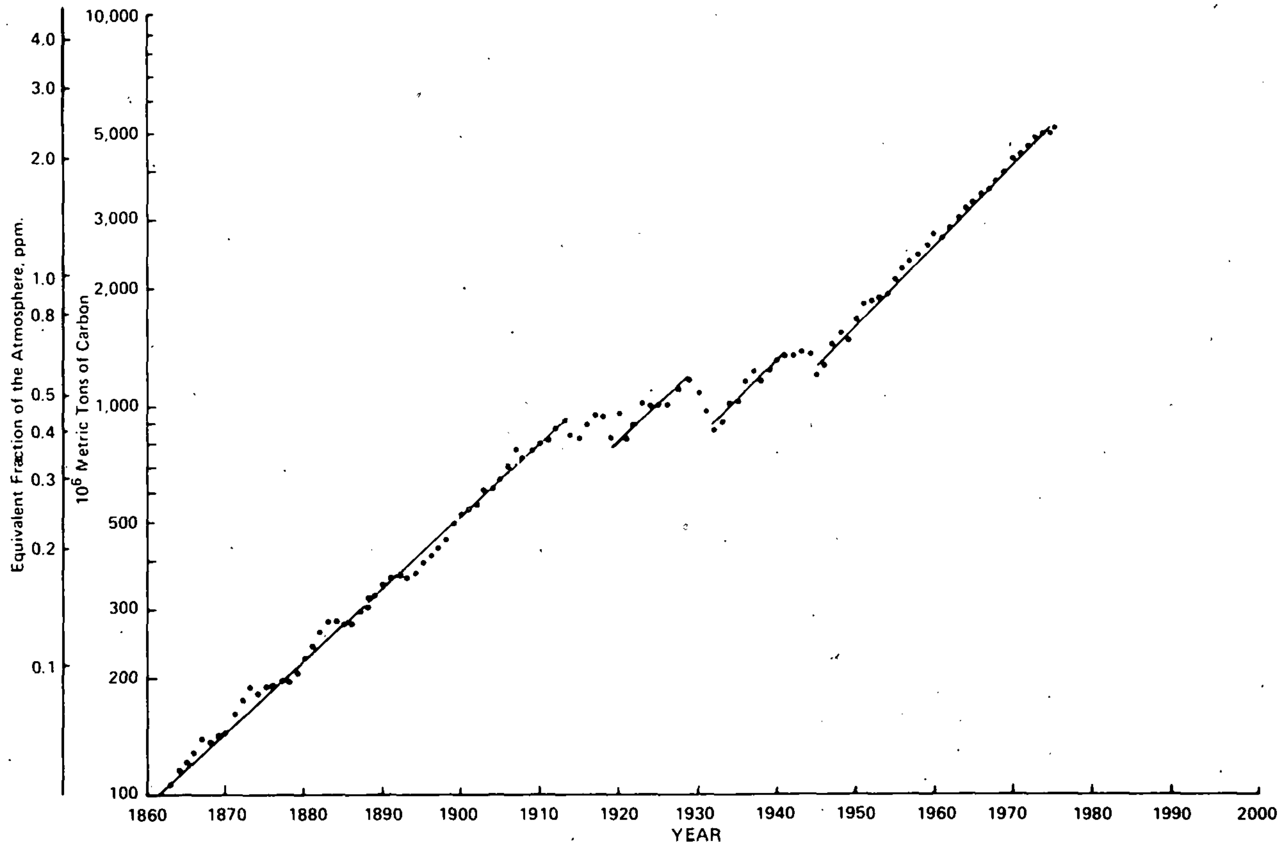

Fig. 1. CO_2 production from fossil fuels and cement.

estimated from third quarter 1976 production, which was 467 million tons, by simple arithmetic:

$$\frac{1372}{1325} \times 467 = 484 \text{ million tons} .$$

Values obtained in this way (based on statistics in the UN *Monthly Bulletin of Statistics*) do not include fuels produced in China, but do include all other countries of the world. A global total was then obtained by adding an estimated value for Chinese fuel production. The UN Statistical Office (Ramsdell, 1977) provided such values for each fuel. The global 1976 coal production (in millions of metric tons) was then estimated by adding:

First Quarter	+	Second Quarter	+	Third Quarter	+	Fourth Quarter	+	China	=	Total
489	+	488	+	467	+	484	+	490	=	2418

Comparative values for 1975 are as follows:

474	+	473	+	447	+	483	+	470	=	2347

Similar procedures were followed in estimating crude petroleum, natural gas, and lignite values for 1976. In the case of lignite, UN data in the *Energy Supplies* series include peat used as a fuel in Ireland and the USSR, but the data in the *Monthly Bulletin* do not include peat. Since this is a small quantity, extrapolating the 1975 peat production value to 1976 introduces an extremely small error possibility. China has never reported mining any lignite.

The exponential trend in CO_2 production is quite evident, and except for the world wars and the great economic depression of the 1930s, a growth rate of 4.3% per year provides an excellent fit to the data. A magnification of the current decade offers some additional insight. Table 1 shows a breakdown on the calculations for 1970 through 1976, and the slowed economic growth of the period 1973 to 1975 is evident in the results.

The upper part of Fig. 2 shows these production amounts. The lower part of the figure shows the annual rate of energy use in the United States. The change in global CO_2 production is proportionally

Table 1. Calculation of CO_2 emissions. (All entries in 10^6 tons except natural gas and natural gas flared, which are in 10^9 m^3.)

	1970	1971	1972	1973	1974	1975	1976
Coal produced	2160	2134	2163	2203	2237	2347	2418
CO_2 produced	1497	1479	1499	1527	1550	1626	1676
Lignite produced	855	864	876	886	887	898	940
CO_2 produced	237	239	243	246	246	249	260
Crude petroleum produced	2270	2402	2532	2777	2791	2650	2801
Natural gas liquid produced	74	77	83	88	87	90	91
CO_2 produced	1803	1907	2011	2203	2213	2107	2224
Natural gas produced	1040	1113	1172	1233	1257	1269	1288
CO_2 produced	545	583	614	646	659	665	675
Natural gas flared	167	173	186	204	209	198	210
CO_2 produced	90	94	100	110	113	105	113
Cement produced	598	639	678	690	692	728	723
CO_2 produced	82	88	93	95	95	100	99
Total CO_2 produced (carbon)	4253	4389	4559	4826	4876	4852	5047
(ppm)	2.00	2.07	2.15	2.27	2.29	2.28	2.38

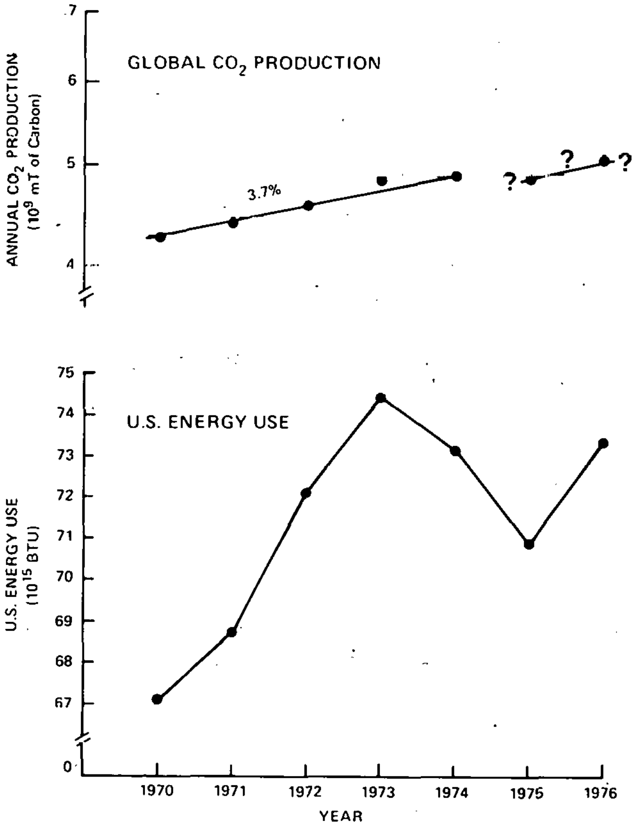

Fig. 2. Global CO_2 production and U.S. energy use.

much less than the change in U.S. energy use. Perhaps of greater importance is the change in the growth rate. The indicated rate of 3.7% for the best fit for 1970 to 1974 does not represent a statistically significant change from the 4.3% for the long-term growth. For the change during the single year 1975-1976, the growth was 3.9%. The period 1970 to 1975, considered as a whole, gives a growth rate of 2.9%, which shows a statistically significant difference from 4.3%. Whether one draws the line with the significant difference in slope, or whether one indicates two segments with essentially the same slope with a break as indicated in the upper half of Fig. 2, is an open question. In either view, a significant change is suggested. I prefer the representation shown in Fig. 2, because there is a historical precedent for resumption of growth after a recession.

For us to address the questions, *Why should we be concerned? Won't any shift in the mix of energy supply systems in the United States only cause minor changes in the global total of CO_2?* it is informative to determine the global distribution of fossil fuel CO_2 emissions. As long as the global total is the end result, fuel *production* data are easiest and probably the most accurate base from which to proceed. However, if it is the purpose to "assign the blame" to particular countries, it is more logical to use fuel *consumption* data. Such a procedure requires an analysis of imports and exports and, in the case of petroleum, a distinction between imports of crude petroleum and imports of finished petroleum products.

I have followed these procedures:

1. The world was divided into 12 political-economic sectors consisting of: United States; Canada; Western Europe (by definition this means the European Economic Community and the European Free Trade Association); USSR; Centrally Planned Economy (CTP) Europe (minus USSR, but including some other parts of Europe: Gibraltar, Greece, Malta, Spain, and Yugoslavia); Japan; Australia and New Zealand; developing America; developing Africa; developing Asia; developing Middle East; and CTP Asia.
2. In the consumption statistics, all solid fuel data are lumped together, based on 7000 cal/g. (This does not account for some coal with heating values higher than 7000 cal/g). Since the heating value of solid fuels is closely related to the amount of unoxidized carbon in the fuel, this provides an adequate base for estimating the amount of carbon present. This gives a global total approximately 1% lower than the way I computed CO_2 from production data.
3. In assigning liquid fuel consumption, I used the equation,

   $$EPC = AS + NGL + 1.0905\,(PROD_{IMP-EXP}) ,$$

   where:

   - EPC is the equivalent crude petroleum consumed by a user country;
   - AS is the apparent refinery supply which is the domestic production of crude, plus the excess of imported crude over exported crude;
   - NGL is the natural gas liquids which are normally added to the petroleum at some stage in the refining process. There is very little international trade in natural gas liquids, so I used its domestic production for each country (with the exception of the United States and Canada, this is small);
   - 1.0905 is the scaling factor to get mass of products back to mass of crude. It is the ratio of crude supply to refineries over total liquid energy products. I have used the values for 1974: 2743.9/2516.3 = 1.090;
   - $PROD_{IMP-EXP}$ is the excess of imports over exports of finished petroleum products.
4. "Bunkers" (fuel consumed in international transportation—ships and aircraft) were charged to the country which supplied the fuel.

Using these procedures, I computed the CO_2 produced from fossil fuel combustion (not including cement) for each of the 12 geographical sectors for 1974. The sum of the results gave a figure 4% lower than the one tabulated in Table 1. Two percent (half the difference) is accounted for by the omission of cement manufacture, 1% is accounted for in the lumping together of solid fuel data (essentially a discounting of some of the USSR coal), and the other 1% is unaccounted for, perhaps lost in the bookkeeping of international trade, in oil spills, and so on.

The results of all these calculations are shown in Fig. 3. Clearly North America, Western Europe, and CTP Europe including USSR are responsible for nearly three-fourths of the global total fossil fuel CO_2. The United States alone is responsible for over one-quarter.

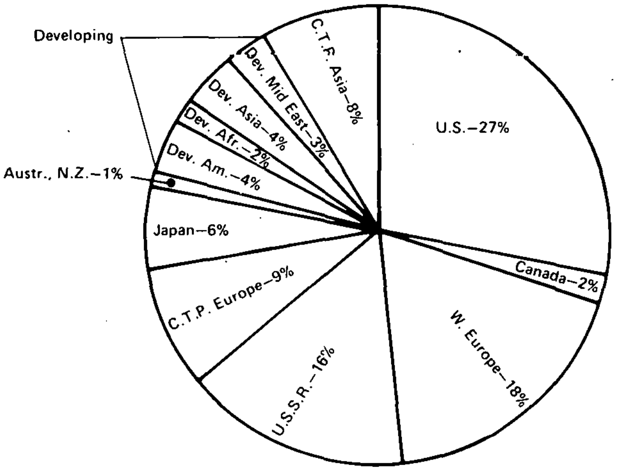

Fig. 3. Global CO_2 production by world segments, 1974.

What might this look like in the future? I can suggest a scenario for 50 years from now. For this purpose I have divided the world into six sectors, corresponding to combinations of the 12 sectors used in the 1974 analysis. I have used the following assumptions:

1. For the United States I have assumed an energy growth to 125 quads (1 quad = 10^{15} Btu), from 73 quads in 1974. Further, 15% of the 125 quads will be nonfossil.

2. Western Europe will experience a 2% per year growth in energy use. This is sufficient to bring Western Europe to nearly the same total as the United States. (Western Europe now uses about 60% as much energy as the United States.) This represents "low-growth," less than half the rate of the recent past. Further, 15% will be nonfossil.
3. For CTP Europe (including USSR) the forecast is based on 4% growth versus 5.1% for the past 15 years. Again, I assumed 15% will be nonfossil.
4. Japan and Australia will experience the same 2% per year growth as Western Europe, with 15% of the total being nonfossil.
5. CTP Asia (largely China) will continue to expand energy growth at its historical 5.1% rate; this will be mostly fossil energy. The low per capita energy use now makes something close to this growth rate necessary for world stability.
6. The developing world has the most serious problem of continued rapid population growth. I have assumed an average population growth of 2% and a 50-year increase in the annual per capita energy use to 53×10^6 Btu/year—the 1970 global average, an amount which all of us would find unacceptably low even today. (The U.S. average is now over six times this 2025 goal!) While this requires a 5.5% average growth rate for 50 years, these countries have achieved an 8% annual growth during the past decade or so.

Based on these assumptions, my calculations result in an annual fossil fuel carbon dioxide release containing 26×10^9 tons of carbon. This is a fivefold increase over the 1974 amount. Figure 4 shows the global distribution of fossil fuel CO_2 emissions as projected for 2025. The circles (1974 and 2025) are proportioned so that the areas on the figure represent the total quantity of carbon released from fossil fuel.

Obviously, the global problem becomes that of providing fuel for the developing countries to assure their progress without such heavy dependence on fossil fuels. Perhaps this is an area in which the United States can make a contribution through research and development on new energy supply systems and on small (decentralized) nonfossil systems.

One might suspect that growth to the magnitude indicated in Fig. 4 for the year 2025 will heavily tax the fossil fuel reserves of the world. This is simply not true; *recoverable* fossil fuels (and shale oil) contain 7.3×10^{12} tons of carbon.

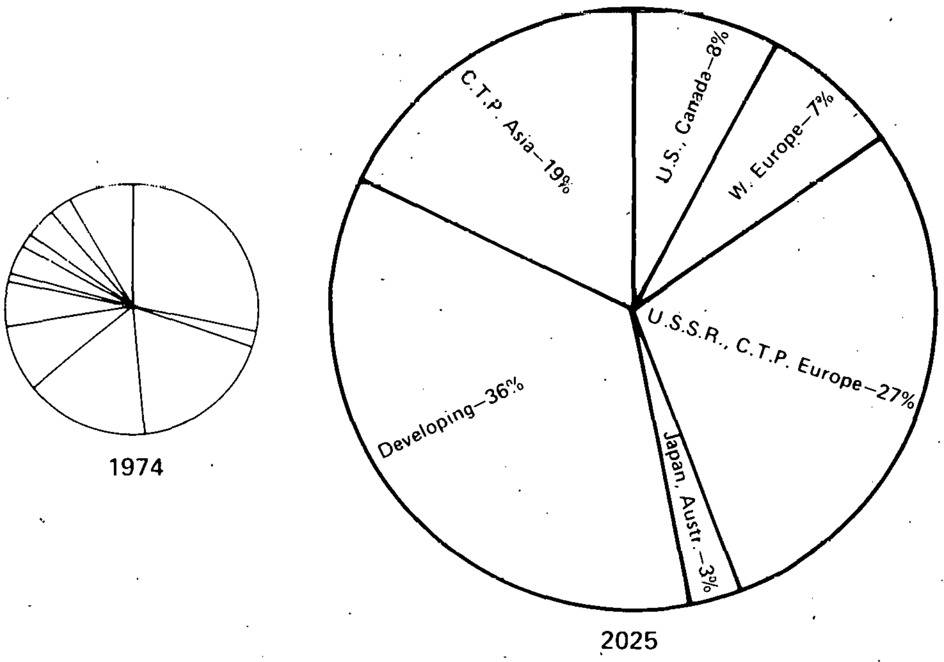

Fig. 4. Global CO_2 production by world segments, 2025. 1974 production is shown to the same scale for comparison.

Several possible patterns for future CO_2 production from fossil fuel may be determined from an assumed mathematical expression:

$$\frac{1}{N}\frac{dN}{dt} = 0.043\left(1 - \frac{N}{N_\infty}\right)^A ,$$

where:

N is a function of time t and represents the total cumulative amount of CO_2 produced from fossil fuel use up to that time; N_∞ is the quantity of CO_2 which would be produced from all the fossil fuel ultimately recoverable (~7.3×10^{12} tons carbon); and A is a parameter to place more or less emphasis on the fraction of recoverable fossil fuels remaining (to be used). The quantity $(1 - N/N_\infty)^A$, which is always between 0 and 1, enables the cumulative use function to reflect a reduced rate of use as the resource approaches depletion and the costs rise.

In Fig. 5, the historical fossil fuel use (indicated by the open circles) falls on the curve; since $(1 - N/N_\infty)$ is so nearly equal to 1.0 up to 1975, the exponential fit is excellent. The curve with A = 1 might represent "free and easy" access to remaining reserves along with early exhaustion of those readily available. The curve with A = 4 might represent lowered demand scenarios for fossil fuels resulting from early extensive reliance on solar, nuclear, and other sources, or from high prices resulting from governmental actions (such as cartels). The vertical barred line in Fig. 5 gives the range of CO_2 production in 2025 as predicted by scenarios of future energy demand previously developed at the Institute for Energy Analysis; the heavy black circle corresponds to the quantity included in Fig. 5.

In Fig. 6, cumulative production of CO_2 since 1958, as calculated above, is shown by the series of solid circles, and the projections along the lines A = 1, A = 2, and A = 4 correspond to the fuel use scenarios of Fig. 5. The projected future atmospheric concentration of CO_2, starting with observations represented by the open circles, is based on 50% of the cumulative production remaining in the atmosphere. The size of this fraction and how the fraction may change with time are, of course, among the major points now being addressed, but for illustrative purposes the assumed 50% should suffice. The atmospheric concentration increases rapidly after the year 2000; for the scenario developed for Fig. 4, it approaches a 200 ppm increase (over 1958) in CO_2 concentration

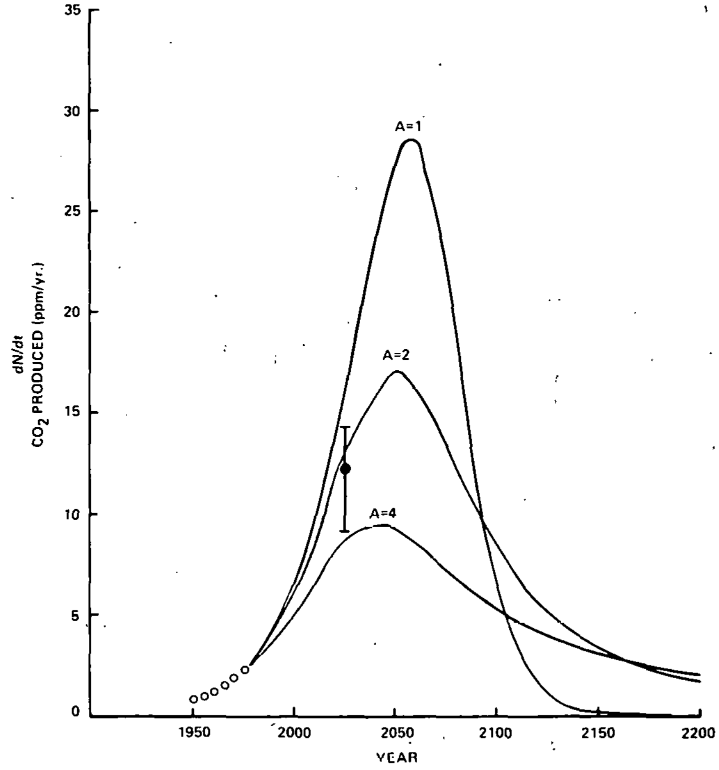

Fig. 5. Fossil fuel use scenarios.

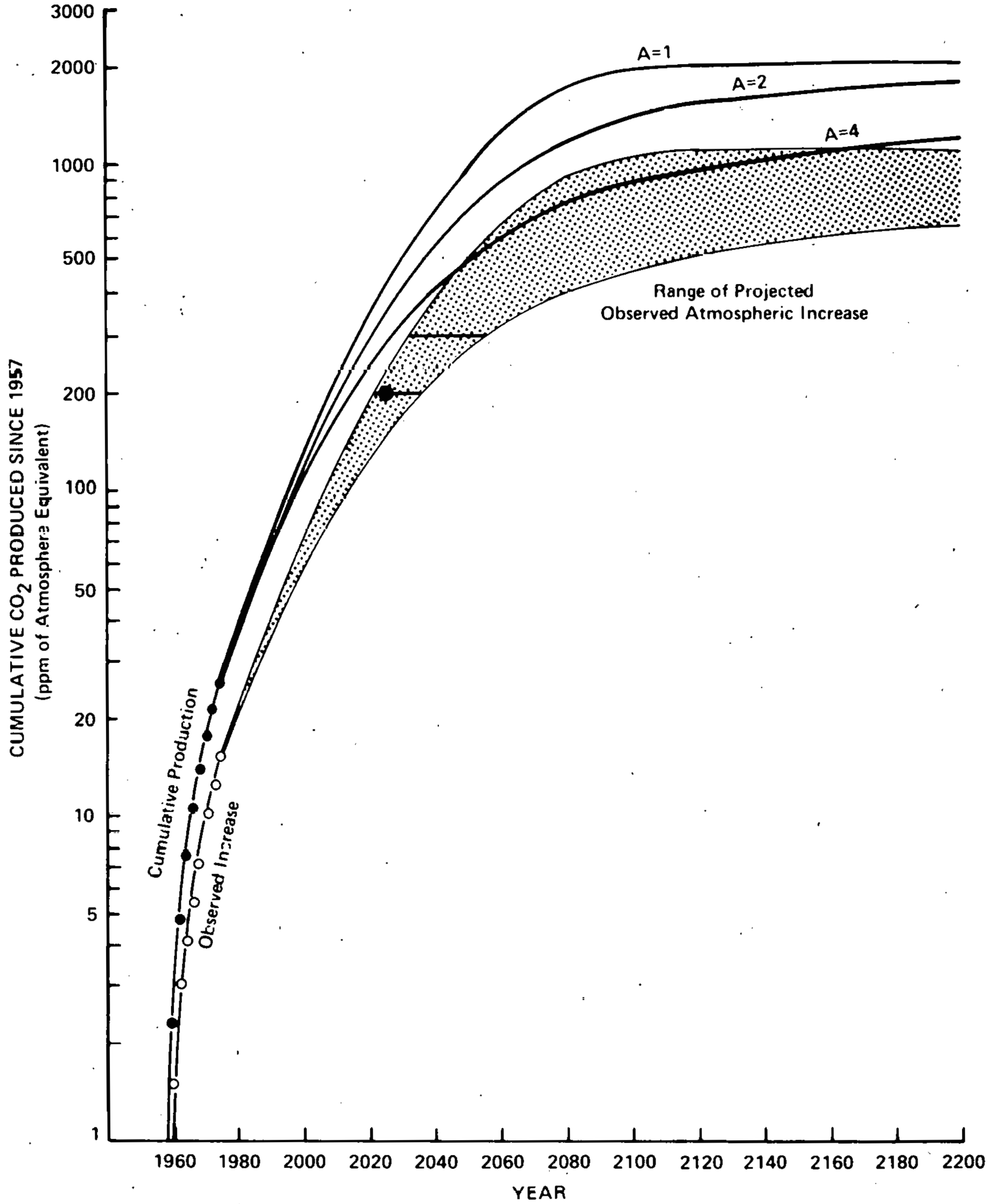

Fig. 6. Projected cumulative CO_2 production and projected observed atmospheric CO_2 increase since 1957.

in 2025. The relatively short horizontal distances in the shaded area of Fig. 6 at additional concentrations of 200 and 300 ppm suggest that almost any low-growth scenario (such as A = 4) will result in observable climate change during the first half of the next century.

The problem is clearly serious, global, and most directly affected by what happens in the developing countries in the next 50 years. It also provides the United States an opportunity to supply guidance and leadership. The need for development of nonfossil, benign energy supplies is rapidly becoming critical.

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Atmospheric Measurements of Carbon Dioxide

Lester Machta, Air Resources Laboratories, National Oceanic
and Atmospheric Administration

INTRODUCTION

Perhaps the most relevant piece of information pertaining to the CO_2 problem is the long record of atmospheric measurements obtained at Mauna Loa, Hawaii. This observatory, started during the International Geophysical Year, 1958, by Dr. C. D. Keeling of Scripps Institution of Oceanography and the late Dr. Harry Wexler of the U.S. Weather Bureau, is located some 3400 m above the sea within the trade wind zone. The disturbance to the CO_2 record by volcanic emissions has not proven to present a problem.

Figure 1 presents the mean monthly concentrations of CO_2 from the beginning of the station's operation through 1975 (Keeling and Bacastow, 1977). Concentrations are shown as open circles connected by a solid line; a dashed line between data points indicates missing data. The horizontal dashed lines locate mean annual concentrations. The numbers at the bottom of the figure show the annual change of concentration from one year to the next. The parentheses denote the absence of data for one or more months during the year for which the average was derived; an adjustment has been made for missing months. The concentrations appearing on this figure may be about 4 ppm higher than on earlier versions of this graph. The change results from a new standard established by Dr. Keeling (noted on the figure) and is due to low readings caused by CO_2-in-N_2 standards in lieu of CO_2-in-air standards.

Two types of variations are immediately obvious: first, the upward growth of CO_2 concentration over the 18 years on record; second, the annual cycle displaying a drop during the summer growing season.

THE ANNUAL CYCLE

The annual cycle is characteristic of almost all CO_2 concentration records. Figure 2 shows the amplitude of variation of the several monitoring stations plotted at the stations' latitudes (Machta, Hanson, and Keeling, 1977; Machta, 1972). Each station is identified, with an indication of the number of years of available records. It is clear that the number of stations with more than half a dozen years is very limited.

The vertical lines are calculations of the expected amplitude using a meteorological model of horizontal and vertical air transport of CO_2 (Machta, 1974). The underlying assumption of the calculation argues that the seasonal variation of atmospheric CO_2 is controlled solely by seasonal variation in the uptake and release of CO_2 by the land biosphere. The observed points (the dots) coincide reasonably well with the theory in all parts of the globe except near the North Pole area and on Long Island. The latter does not qualify as a background monitoring station (and was not so intended); a strong wintertime maximum results from human activities releasing fossil fuel CO_2, making the amplitude larger than would be expected from a solely biospheric source. The higher-than-expected amplitudes in high northern latitudes, it is speculated, stem from the seasonal variation in ice cover (Sanderson, 1975). The late summer minimum CO_2 concentration which enhances the amplitude may be due in part to the minimum ice cover; the exposed water, which is both cold and biologically active, can take up, locally, large amounts of atmospheric CO_2. It is argued from Fig. 2 that the major features of the seasonal variation in atmospheric CO_2 can be

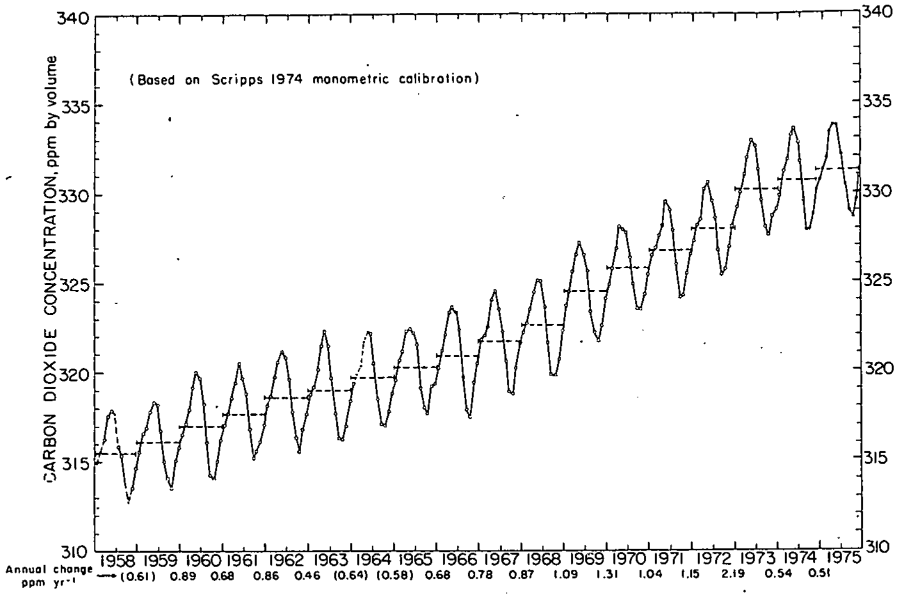

Fig. 1. **Mean monthly carbon dioxide concentrations at Mauna Loa.** Annual changes in parentheses are based on incomplete records.

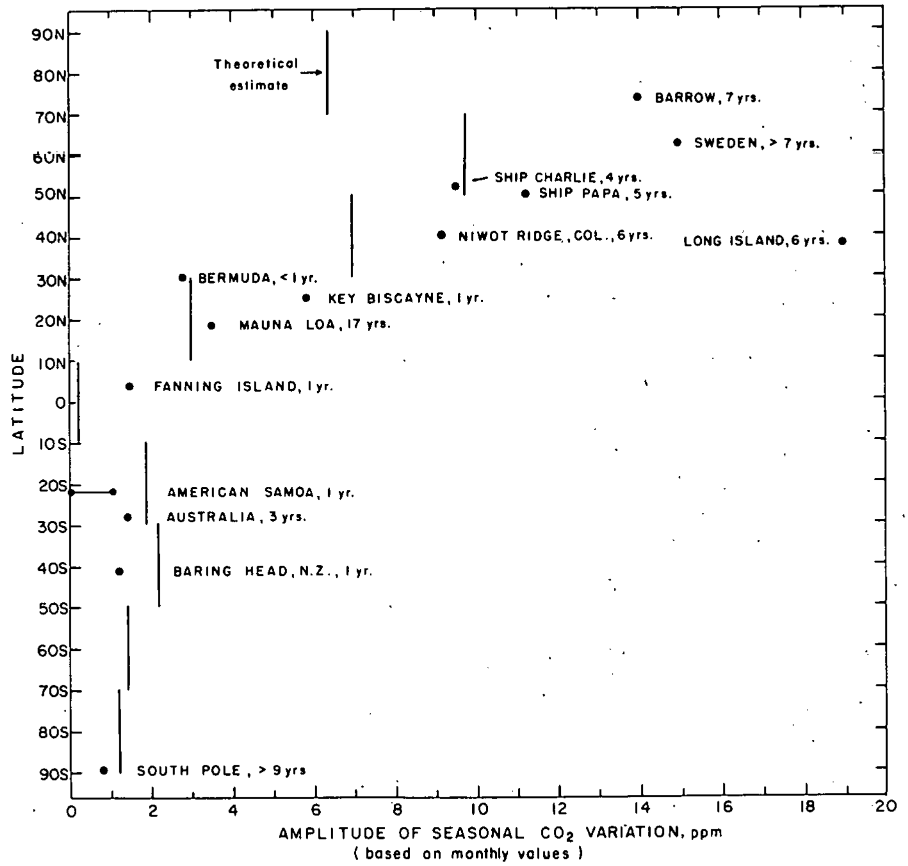

Fig. 2. **Variation of amplitude of monthly carbon dioxide concentration in the course of a year, averaged over all available years, with latitude (dots).** The number of years of records follows the identification of the sampling point. The vertical lines provide a theoretical estimate of the same amplitude using uptake and release of carbon dioxide by the land vegetation as the sole seasonal forcing function.

accounted for by the land biosphere, except at locations which have local influences of fossil fuel CO_2 or lie near a variable oceanic ice pack.

The second point of interest is the possibility of a long-term change in the seasonal amplitude which might reflect a change in the biosphere. Figure 3 shows the time history of the amplitude at Mauna Loa. No systematic long-term increase or decrease in the amplitude is evident. A more sophisticated analysis by Hall, Ekdahl, and Wartenberg (1975) comes to the same conclusion: there is no evidence from the Mauna Loa record for a change in the biosphere, assuming that the Mauna Loa seasonal variation mainly reflects the biospheric uptake and release of CO_2. However, it should be noted that this technique for monitoring the biosphere is not sensitive to changes in the size of the biosphere.

LONG-TERM TRENDS

Evidence for a long-term upward trend of about the same magnitude as seen in Fig. 1 for Mauna Loa appears in all other records at background stations. Figure 4 presents data from all of the stations with relatively long histories: Mauna Loa and three others. The evidence is unequivocal, despite small year-to-year differences among the stations. Indeed it is this high-quality data in contrast to highly uncertain measurements taken prior to 1957 which has highlighted the existence of a potential CO_2 problem.

Figure 5 relates the Mauna Loa record to fossil fuel emissions. In the figure, it is assumed that the fossil fuel CO_2 each year immediately mixes throughout the atmosphere but does not leave it. The dots joined by dashed lines (measured on the left-hand ordinate) indicate the CO_2 increase that would be observed at the end of each year given the above assumption. Secondly, if it is assumed that the CO_2 concentration at Mauna Loa is representative of the entire atmosphere, its annual increase can measure the percentage of fossil fuel CO_2 which remains airborne. The airborne percentage may be read on the right-hand scale of the figure. It is seen to be highly variable, ranging from 25% to over 100%. Were the global average increase derived from more stations than just Mauna Loa, the airborne fraction would be more uniform. As will be seen below, the Mauna Loa record has just recently been brought up to date through 1976; the airborne fraction derived from this information suggests a 40% value from 1975 to 1976.

The airborne fraction possesses special significance for this workshop. That fraction of the fossil fuel CO_2 which does not remain airborne must obviously transfer to another CO_2 reservoir; it is accepted that this transfer is to the oceans. Many

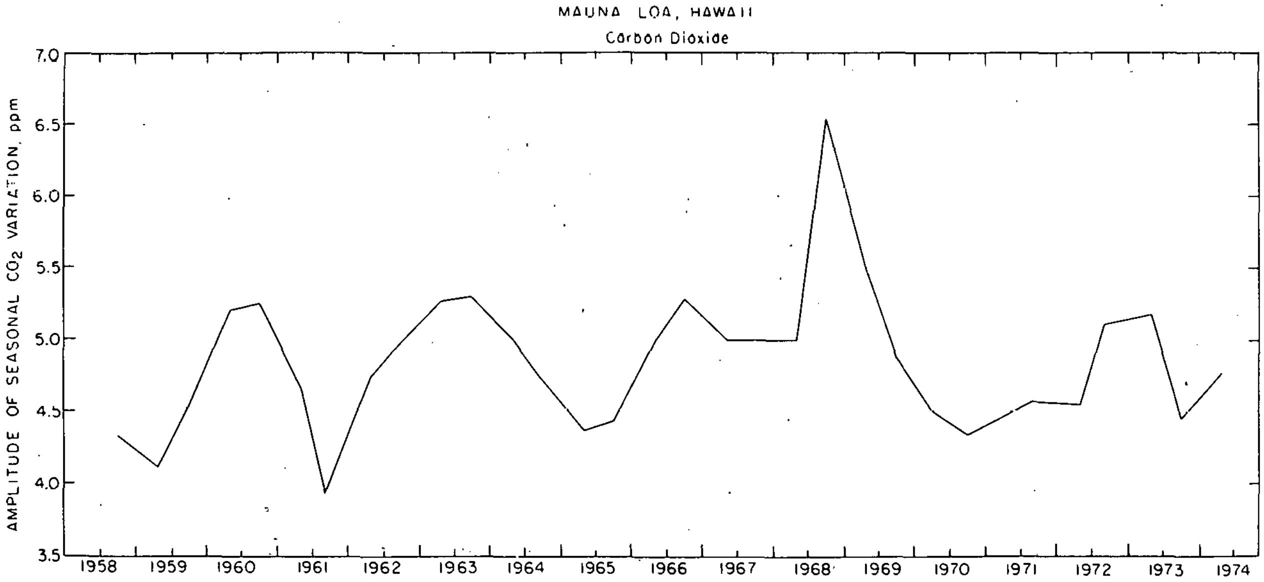

Fig. 3. The amplitude of carbon dioxide concentration within each year at Mauna Loa. The amplitude is obtained by subtracting the minimum three-month (usually September, October, and November) average from the mean of the maximum concentrations (usually March, April, and May) before and after the maximum. This amplitude is plotted in October. A similar calculation is performed by subtracting the mean of the two adjacent minima from the maximum concentrations and plotting at April of each year.

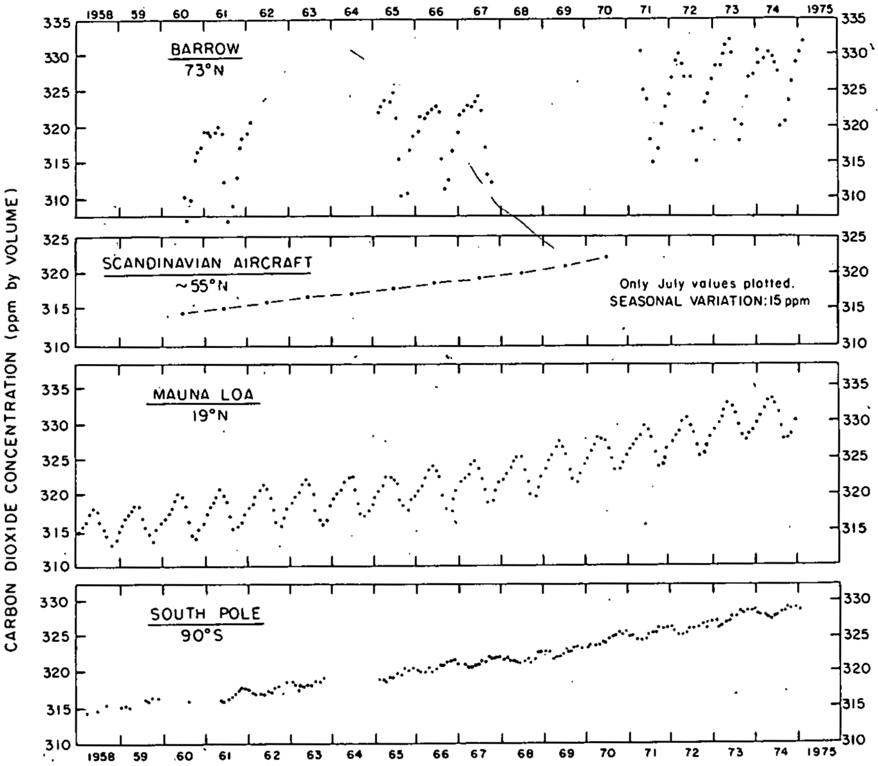

Fig. 4. **The growth of atmospheric carbon dioxide concentration at four locations.** Except for the Scandinavian aircraft data, the points denote monthly average concentrations. In Scandinavia, the values have been referred to July, using an average pattern of seasonal variability. Some data are based on intermittent flask sampling; others, including the entire record at Mauna Loa, are based on frequent in situ analyses.

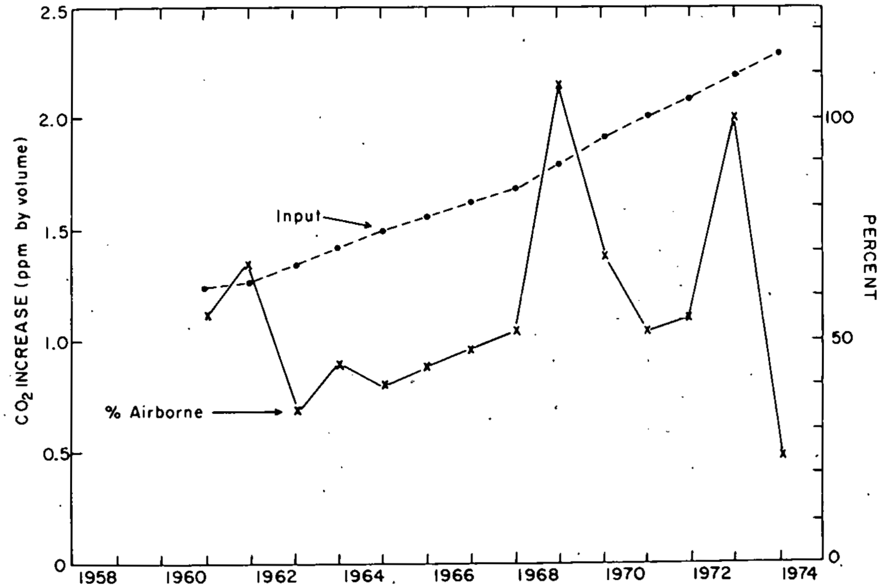

Fig. 5. **The percent of man-made fossil fuel carbon dioxide remaining airborne each year (lower curve), assuming that the observed Mauna Loa carbon dioxide growth represents world growth.** The right-hand ordinate is for the percent airborne. The upper curve shows the amounts of increase expected from fossil fuels (and other man-made sources of carbon dioxide), assuming that this carbon dioxide mixes rapidly with the entire atmosphere.

sets of calculations have been made which either directly or implicitly use a 50 to 60% airborne fraction as the basis for prediction of future CO_2 concentrations. This may be challenged on a number of grounds. First, is there any reason to suppose that the fossil fuel CO_2 transfer to the oceans or the biosphere can be represented in the future by the average of conditions during the 17 years displayed in Fig. 5? Second, the figure and analysis also assume that the only new airborne CO_2 derives from the combustion of fossil fuels (and cement manufacture). This assumption, as will be noted below, cannot now be confirmed.

Figure 6 updates some of the monitoring station records. The squares joined by straight lines, read on the right-hand scale, repeat information presented earlier by Dr. R. M. Rotty in this workshop. A leveling off of the CO_2 emissions in 1974 and 1975 is evident, with a recovery back to the 4 to 5% per year growth rate in 1976. The remaining curves display mean annual concentrations of CO_2 at the indicated stations (the points are to be read from the scale on the left). The annual changes in concentration at the four stations appear on the upper part of the figure. These are changes per year even when some intermediate years are missing. The reduced growth rate at Mauna Loa of about 0.5 ppm per year from 1973 through 1975 has increased to a value somewhat over 1 ppm per year from 1975 to 1976. The growth rates at other stations are as variable as those at Mauna Loa.

Finally, there is evidence of a growth of CO_2 in the atmosphere before 1958, but the numerical values of the long-term change are on very shaky grounds. Figure 7 presents many of the earlier estimates of CO_2 concentration (Machta and Telegadas, 1974) and one guess as to the preindustrial concentration, 290 ppm. The very wide scatter of the data more likely reflects the

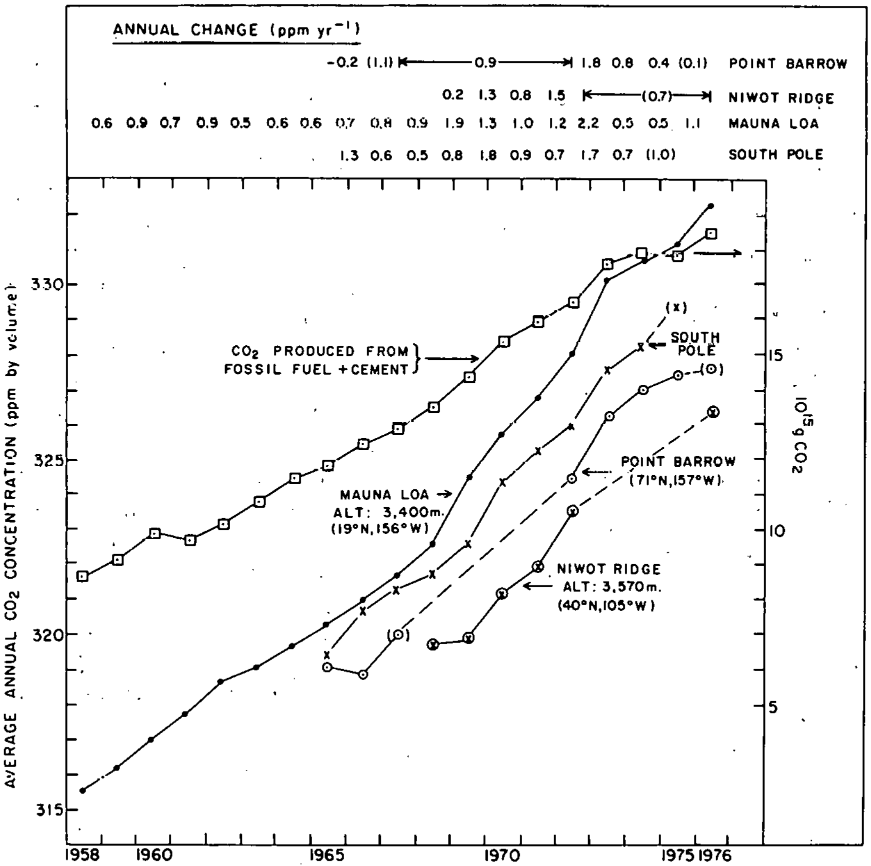

Fig. 6. **A summary of some recent trends in annually averaged atmospheric CO_2 at Mauna Loa, the South Pole, Point Barrow, and Niwot Ridge.** The upper part of the figure lists the year-to-year change in ppm; numbers in parentheses are based on incomplete information. The straight line joining the squares indicates the expected yearly changes in atmospheric CO_2 expressed in 10^{15} g CO_2/year, if all of the CO_2 remained airborne. Note the slowing down in the growth rate of this latter curve between 1973 and 1975.

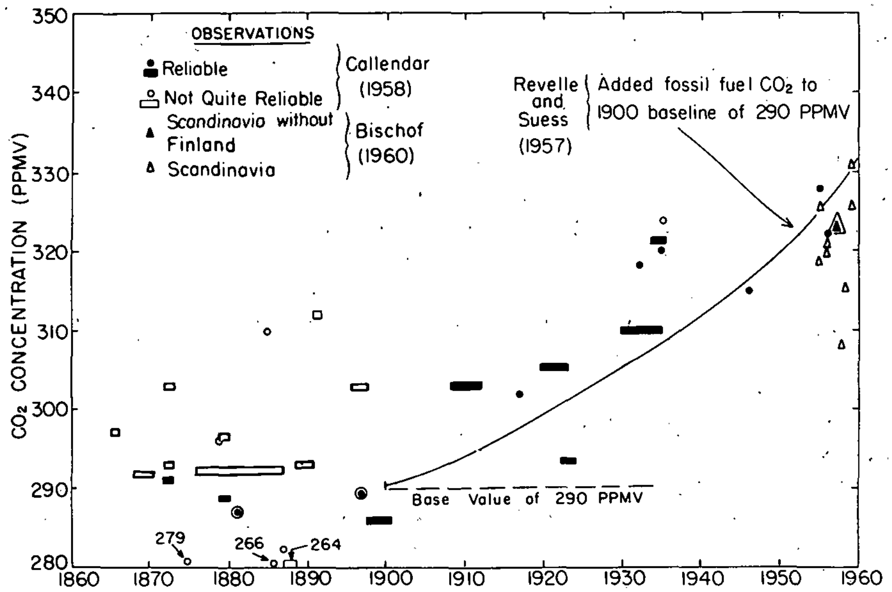

Fig. 7. **The secular increase of carbon dioxide.** The solid line represents the expected increase due to fossil fuel consumption since 1900 if all carbon dioxide remained airborne.

inaccuracy of the measurements than the contribution of natural variation. Based on the data, it seems that a concentration in the range of 285 to 305 ppm was likely before man added significant amounts of fossil fuel CO_2 in the nineteenth century.

THE GLOBAL PICTURE

The relative annual changes between reservoirs contrasted with the annual addition of fossil fuel CO_2 to the atmosphere is summarized in Table 1.

Table 1. Global sources and sinks for atmospheric CO_2 on a time scale of years or less

	Source (ppm/year)	Sink (ppm/year)
Ocean	40–45	40–45
Steady-state biosphere		
Photosynthesis		25
Net primary production		
Organic decay, fires, etc.	25	
Disturbed biosphere	0–2(?)	
Fossil fuel	2	0
Others:		
Variable ocean ice cover	(variable sink?)	
Volcanic eruptions	(source)	
Rain, snow	(sink)	

Thus, in 1975 the annual increase expected from the introduction of fossil fuel CO_2, assuming that this CO_2 remains in the atmosphere, is about 2 ppm. The oceans, it is believed, take up and release about 20 times this amount of CO_2 (mainly natural) on a global scale. In an equilibrium state about the same amount of CO_2 leaves the oceans as returns to it. The biospheric photosynthesis takes up about 25 ppm/year, or 10 times more than the 2 ppm/year from the air; but through decay, fires, and other processes, the same amount is returned back to the atmosphere. However, it is possible that man is disturbing the biosphere; some scientists estimate that as much as 2 ppm/year will be added to the air by this disturbance in the mid-1970s. Although not shown in the table, other scientists feel that the disturbed biosphere may, on the average, actually take up more CO_2 per year than it releases, and therefore that it acts as a sink. Part of the rationale for this view is the known increase in the rate of photosynthesis with greater availability of atmospheric CO_2, provided there is no limitation of sunlight, soil moisture, and nutrients. Another factor is the possibility that, in some regions of the world, there has been a return from marginal agriculture back to woodlands.

The lower section of Table 1 lists three other possible sources and sinks for atmospheric CO_2. The annual advance and recession of ice appears to emphasize the annual cycle of atmospheric CO_2.

But of potentially greater significance over a long-term period might be the greater ice-free uptake of CO_2 should the arctic ice pack disappear. This would act as a negative feedback mechanism but its magnitude is very uncertain.

Over geological time atmospheric CO_2 has presumably derived from volcanic emissions. Should there be an epoch of intense volcanism, this source could add to atmospheric CO_2. However, more than CO_2 spews out of volcanoes; dust, for example, might play a cooling role in the climate.

Precipitation will deposit carbonates on the soil and in the oceans. Estimates of rainwater carbonate content suggest that this sink for the atmosphere is negligibly small compared to the direct uptake by the biosphere and oceans.

RECENT STUDIES IN THE TROPICS

In the past several months, three papers have statistically related changes in atmospheric CO_2 to conditions in the equatorial Pacific Ocean. In the first, Bacastow (1976) showed that a departure from the long-term increase appeared to fluctuate with the Southern Oscillation Index, a measure of the trade wind speeds in the Pacific Ocean. Machta, Hanson, and Keeling (1977) presented a similar statistical correlation with the appearance of the El Nino, a phenomenon during which the equatorial Pacific Ocean near South America is warmer than usual (and the rainfall is anomalous in some of the islands of the tropical Pacific Ocean). They also presented arguments derived from transport models showing that a CO_2 anomaly in the equatorial belt will produce anomalies at Mauna Loa, Australia, and the South Pole which agree well with those observed. Finally, Newell and Weare (1977) used departures from a steady growth rate at the South Pole, which had been observed by Bacastow, to demonstrate a statistical relationship between changes in atmospheric CO_2 and equatorial Pacific sea surface temperatures.

Although the evidence is statistical in nature, it appears likely that some property of the equatorial Pacific Ocean, most likely the sea surface temperature (directly or through its influence on biological activity in the surface waters) modulates Mauna Loa, Australian, and South Pole atmospheric CO_2. The amplitude of the modulation is about 1.5 ppm, but presumably the amplitude at a location in the equatorial Pacific Ocean would be larger.

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CO_2 and the Biosphere

R. S. Loomis, Agronomy and Range Science,
University of California, Davis

Two major areas of concern are evident in assessments of the interactions between atmospheric CO_2 and the biosphere. The first of these, the biosphere as a source and sink for gaseous CO_2, involves biosphere pool sizes and transfer processes such as photosynthesis and respiration. The second concerns changes in the character and activity of the biosphere that may result from changes in CO_2 concentration and climate. Time limits the issues that can be explored here; my approach is that of a critic of the principal lines of evidence and argument as presented in the literature.

THE ROLE OF THE BIOSPHERE IN CARBON CYCLES

The Biomass Pool

Living organisms accumulate significant quantities of carbon. Current estimates place more than 99% of this in terrestrial ecosystems, principally in forests, where large amounts of carbon are stored in wood. The turnover rate of the oceanic biomass is much more rapid, and the ratio of organic production on terrestrial sites to that in the oceans is more nearly 2:1 than 99:1. Detailed estimates of standing biomass have been attempted only recently (Rodin and Bazilevich, 1967, 1968; Whittaker and Woodwell, 1971; Whittaker and Likens, 1973, 1975). The main features of Whittaker and Likens' (1975) summary are presented in Table 1. These estimates are the best current values. Forests, particularly tropical forests, are shown to be important reservoirs of carbon.

It is important that we understand how these numbers were obtained. Estimates for the temperate forests apparently involve the most detail. The number is drawn as a mean of intensively studied sites in the Great Smoky Mountains, for example, and at Brookhaven, New York and Oak Ridge, Tennessee (Whittaker, 1966; Whittaker and Woodwell, 1967), where dimensional analysis procedures were employed. In sample quadrats, trunk length and diameter, the number, length, and diameter of sample branches of trees and shrubs, and sample weights were used to formulate allometric ("double logarithmic") regressions between basal measurements and aboveground tree and shrub volume, mass, foliage density, and estimates of net annual production. Root estimates, a difficult task, were made in some cases. Biomass values for sampled trees were considered to fall within 10 to 20% of the regression estimates.

Biomass estimates from these North American sites were arrayed with estimates from other sites here and in Europe and Asia. A normal range and a subjective mean were adjudged, and the mean was multiplied by the global estimate of land area in temperate forests to obtain a total for the world.

The approach is reasonable. The individual sites were measured carefully, but the number of sites was small and mapping uncertain, so that the final number for carbon in the temperate forest biomass can be viewed as only tentative. Documentation for the table has not been published and the possible extent of the error cannot be determined; ± 30% may be a reasonable guess. Temperate forests are highly variable, not only in age and structure but in site factors such as slope, moisture supply, and length of season. Greater accuracy will require a wider sampling of forests, leading to an area-weighted total. Critical issues will be the degree of detail in the mapping and the representativeness of the sampled quadrats, particularly the extent of grassy forest openings.

The other major ecosystems presented in Table 1 have been mapped and sampled with much less detail. There is considerable uncertainty in these other

Table 1. Estimates of standing biomass and related characteristics of the biosphere

Ecosystem type	Area (10^6 km^2)	Biomass: Range (metric tons/ha)	Biomass: Mean (metric tons/ha)	Biomass: Total (10^9 metric tons)	Biomass: As carbon[a] (10^9 metric tons)	Net primary production: Range (metric tons ha^{-1} year^{-1})	Net primary production: Mean (metric tons ha^{-1} year^{-1})	Net primary production: Total (10^9 metric tons year^{-1})	Net primary production: As carbon[a] (10^9 metric tons year^{-1})
Tropical rainforest	17.0	60–800	450	765	344	10–35	22	37.4	16.8
Tropical seasonal forest	7.5	60–600	350	262	118	10–25	16	12.0	5.4
Temperate forest	12.0	60–2000	320	384	173	6–25	13.5	15.0	6.8
Boreal forest	12.0	60–400	200	240	108	4–20	8	9.6	4.3
Woodland and shrub	8.5	20–200	60	51	23	2.5–12	7	6.0	2.7
Savannah	15.0	2–150	40	60	27	2–20	9	13.5	6.1
Temperate grassland	9.0	2–50	16	14	6.3	2–15	6	5.4	2.4
Tundra and alpine	8.0	1–30	6	5	2.3	0.1–4	1.4	1.1	0.5
Desert scrub	42.0	1–40	3	13	5.8	0–2.5	0.4	1.7	0.8
Cultivated land	14.0	4–120	10	14	6.3	1–40	6.5	9.1	4.1
Wetlands	2.0	30–500	150	30	13.5	8–60	30	6.0	2.7
Lakes and streams	2.0	0–1	0.2	0.04	0.02	1–15	4	0.8	0.4
Total terrestrial	149	1–2000	122	1338	827		7.8	118	53
Total marine	361	0–40	0.1	3.9	1.8		1.6	56	25.2
Total biosphere	510		36	1842	829			174	78

[a] Whittaker and Likens assumed that 45% of the biomass is carbon in their paper Carbon in the biota. Pages 281–302 *in* G. M. Woodwell and E. V. Pecan (Eds.). *Carbon and the biosphere.* U.S. AEC, Springfield, Va.

Source: Adapted from Whittaker, R. H. and G. E. Likens. 1975. The biosphere and man. Pages 305–328 *in* H. Lieth and R. H. Whittaker (Eds.). *Primary productivity of the biosphere.* Springer-Verlag, New York.

values, particularly for tropical forest systems. The estimates in Table 1 are apparently based on a small number of Asian and African sites, ignoring values of 530 and 860 metric tons/ha given by Rodin and Bazilevich for Amazon sites (1967). (In 1968 Rodin and Bazilevich reported these forests as 1060 and 1720 metric tons. The difference seems to be in their definition of the centner as a unit of weight, and I have used the 50-kg centner as defined in their 1967 book.) Forests of this mass may exist in the western Amazon, but much lighter vegetation than the means in Table 1 is indicated by my own experiences (Williams, Loomis, and Alvim, 1972; personal communication, V.H.F. Moraes and M. Pires, Instituto de Pesquisas e Esperimentação Agropecuárias do Norte (IPEAN), Belem, Brazil) and reports of Instituto Nacional de Pesquisas de Amazónia (INPA), Manaus and by IPEAN and Museu Goeldi, Belem.

For example, Rodriques (1963) at two good sites in the eastern Amazon found a mean of 350 m^3 of commercial wood in trees greater than 15 cm diameter at breast height (dbh). This translates to approximately 350 metric tons/ha of total biomass for the forest (cubic meters × 0.6 for density × 1.7 for total biomass). The current Food and Agriculture Organization (FAO) survey is yielding much smaller values: the mean of over 1000 hectare-size plots sampled in a 1500-km transect parallel to the Amazon river is only 180 m^3 of timber in trees greater than 25 cm dbh (or about 200 metric tons/ha in biomass). This is a mean, unweighted for area, of 19 regional means which ranged from 104 to 271 m^3 of timber (F. Wadsworth, personal communication). Thus, the forest biomass estimates shown in Table 1 may be too large by a factor of 2.

In addition, the area given to heavy forests appears inflated. Large parts of the Amazon basin are given to lower-biomass, poverty "catanga," "igapo," "campino", and "cerrado" savannah vegetations (see reports and Boletim of INPA, IPEAN, and Museu Goeldi). Large areas of savannah on the Rio Branco, for example, have the aspect of short-grass prairie, yet all of the basin is mapped by Lieth (1975) as very high in net productivity (>15 metric tons ha^{-1} yr^{-1}) and is given a mean biomass of 350 to 450 metric tons/ha by Whittaker and Likens (1975). The tropical forest areas (7.5×10^6 km^2 for seasonal forest and 17×10^6 km^2 for rain forest), their mean biomass, and total biomass (262 and 765 × 10^9 metric tons, respectively) must be viewed as extremely tenuous figures.

An additional difficulty appears in translating biomass to carbon equivalents. Whittaker and Likens (1975) use 45% as the carbon fraction of biomass. That may be a reasonable average for herbaceous material; but wood, with its high lignin content, is closer to 50% carbon (USDA, 1952). There are large variations in the content of ash (0.5 to 10% of biomass), organic acids (low percent carbon), highly reduced lipids and terpenes (high percent carbon), and other constituents. Few elemental analyses have been made of carbon in biomass. Again, weighted means would seem necessary.

Soil Carbon

The most common global estimate of reduced carbon in soils (710×10^9 metric tons) was traced by Bohn (1976) to a 1915 textbook (by Lyon, Fippin, and Buckman) which reported carbon analyses for nine U.S. soils. (More commonly, soil carbon is estimated indirectly from nitrogen content with assumptions about the carbon to nitrogen ratio.) Bohn's new estimate is a global extrapolation of the FAO soil map of South America with certain assumptions about bulk density, carbon to nitrogen ratio, and depth. A depth of 1 m, considered to include all of the "active" soil, yielded an estimate of $3000 \pm 500 \times 10^9$ metric tons of active soil carbon. The actual total pool size may be much greater, since most soils carry organic matter below 1 m and most peat deposits are much deeper than 1 m. For example, the Sacramento–San Joaquin Delta of California extends over 1200 km^2 with peat deposits up to 20 m in depth. At the deepest point, using Bohn's factor of 200 kg carbon/m^3, we could anticipate 4000 kg carbon/m^2. This and similar areas are underestimated in Bohn's approach. Schlesinger (1977) has also estimated soil carbon but from a narrower data base than Bohn, limiting the analysis to the better quality observations. His total for the biosphere, 1459×10^9 metric tons, is half of Bohn's. Much of the difference occurs in the area and intensity factors for wetlands.

The Biosphere as Source or Sink

Woodwell and Houghton (1977) presented an argument that the biosphere is currently a net source of as much as 5 to 10×10^9 metric tons of carbon as CO_2. Their case rests on man's disturbance of the tropical forests and wet soils. Soil disturbance is taken as a corollary of forest-cutting.

Much of the evidence of cutting is drawn from Hamilton's (1976) study of the Western Llanos region of Venezuela, where cutting was estimated at about 1.3% per year. That rate is derived from Viellon's appendix to Hamilton's report, which gives 21% as the forested area of that region in 1825, 45% in 1950, and 30% in 1975. This is a populated region, and the mixed vegetation (tall grass–galleria forest) apparently has long been subject to cutting and regrowth. The report notes that if proposed cutting is carried out, a similar rate might apply in the future to much of South America. Woodwell and Houghton apparently use that statement as a basis to suggest that the worldwide cutting rate *is* 1 to 2%. That such has not yet happened in Venezuela is revealed by examining Hamilton's report in more detail: 85% of Venezuela's forests lie in other provinces south of the Orinoco River, and Hamilton observes that no one knows how fast those forests are being altered because of lack of commercial and scientific contacts. Those provinces are essentially unpopulated, and logging and shifting agriculture are clearly far less intense.

Woodwell and Houghton's 1 to 2% rate also does not apply to the Amazon basin. Zinke (personal communication) finds from radar maps that less than 0.5% of that basin has been disturbed during the past ten years. Brazilian agricultural development is centered mostly on the cerrado vegetation south of the basin. Muthoo (1977) provides data for all Brazilian forest lands, 1958 and 1973, from which Wadsworth (personal communication) estimates that 39 × 10^6 ha (6.1% of forest) containing 2.1 × 10^9 metric tons carbon (5.0% of total biomass carbon as estimated from quadrat samples) has been cut in that 15-year period. That corresponds to 0.3% of biomass per year. Correction for regrowth of forest or pasture and for remaining stumps (a large fraction) would reduce that rate.

The rate at which the global forest biomass is being cut appears to be significantly less than 1 to 2% per year, and the rate of conversion of such lands to other uses is considerably less than the cutting rate. Caution is also required in developing estimates of the secondary effects of forest cutting: associated losses of soil carbon generally appear to be overestimated, while forest regrowth and soil carbon storage are frequently underestimated. Nye and Greenland (1960) offer some of the best data sets available on such matters for tropical forests.

Bolin (1977) has presented an argument similar to that of Woodwell and Houghton but with more detail on temperate and boreal forests. He concludes that no definite conclusions can be drawn, but that it is most likely that man's intervention has increased the biosphere-to-atmosphere flux, principally through tropical forest disturbance, cultivation of organic soils, and an accelerated use of fuel wood. Despite the absence of tropical data, he feels that the net release may amount to 0.5 to 2.0 × 10^9 metric tons carbon/year.

Such analyses are quite dependent upon the time period explored: 1840 to 1970 saw the opening of the American prairies to agriculture, with the release (from 1840 to 1950) of perhaps 1 to 3 × 10^9 metric tons carbon from soils. That was perhaps the most concentrated disturbance of high-carbon soils in the history of man. Over the same period, forest biomass in that region may have increased slightly through fire control. Since 1950, it seems likely that the long-term trend in soil carbon has been reversed through irrigation, legume rotations, fertilizers, and soil conservation—all practices that enhance inputs of organic matter to soil. For example, if one makes the reasonable assumption that 25% of the 40 × 10^6 metric tons of fertilizer nitrogen used annually throughout the world is retained in the soil at a carbon to nitrogen ratio of 10, the flux could be 0.1 × 10^9 metric tons carbon/year worldwide from this source until new steady-state levels are attained. A 1% change in global soil carbon represents ±30 × 10^9 metric tons; such changes might occur easily with climate variations. I will explore this point later. Individually, many of these changes are small. Collectively, they may be significant, particularly if forest numbers are actually found to be smaller than those shown in Table 1.

Biosphere Role in Isotope Exchange

The Suess effect (the recent decline in the specific activity of ^{14}C in the atmosphere) and Stuiver's (1977) observation of a decline of ^{14}C in tree wood between 1850 and 1950 can both be ascribed in part to dilution of the atmosphere by fossil carbon. Fossil carbon is fully depleted of the radioisotope ^{14}C. Organic carbon, both fossil and recent, has a lower content of the stable ^{13}C isotope than does the atmosphere, as a result of discrimination in metabolism. A net decline in global organic carbon would dilute the ^{13}C content of the atmosphere, enhancing the ^{13}C depletion of newly formed wood.

The total release of fossil carbon (with $\delta\,^{13}C = -25$ per mill) is known, and the isotopic fraction (25%) retained in the atmosphere can be estimated from the Suess effect, assuming that limestone quarries and fossil fuel represent the anthropogenic sources of ^{14}C-

free carbon. Stuiver followed this calculation for the change in $\delta^{13}C$ of wood formed between 1850 and 1950 and concluded that the biosphere was a net source of about 1.2×10^9 metric tons carbon/year during that period. His approach is perhaps the only one that has the potential of dealing with the globe as a whole. But the calculation is uncertain beyond 1950 because of isotopic disturbance from nuclear testing.

In sharpening that approach, it may be appropriate to consider other sources of isotopic exchange. Deep-well waters in the American West contain 5 to 7 meq/liter of HCO_3^-. A conservative estimate of 0.75 m/year of irrigation from wells on 10^5 km^2 gives 4×10^6 metric tons of released fossil carbon per year. To the extent that groundwater carbon was derived by soil respiration it would also be depleted in ^{13}C. Stuiver further assumes that all biospheric carbon is relatively young (i.e., 200 years or less). That may not be the case with organic matter from virgin soils and peat, sources that then could be a significant contributor to the Suess effect.

Soil carbonates might serve also as a very significant source/sink for carbon and isotopic exchange. Taking 0.4% as a representative calcium content for humid soils and 1.1% for arid-zone soils, 2000 to 7000 metric tons carbon could theoretically be neutralized as HCO_3^- per square kilometer in the active 1-m depth; an exchange of 50 metric tons/km^2 on 14×10^6 km^2 of cultivated land equals $\pm 0.7 \times 10^9$ metric tons carbon.

The enormous size of the soil organic carbon pool means that small changes in the rates of organic inputs or humus oxidation can lead to very large carbon exchanges with the atmosphere. Jenny's (1930) classic study of soil nitrogen led to the generalization:

$$N = 0.55e^{-0.08T}(1 - e^{-0.005H}),$$

where N is in percent of soil weight, T is mean annual temperature in °C, and H is a constant for a given moisture regime. Using Jenny's factors with $H = 400$, N is 0.121% at 17°C and 0.116% at 17.5°C; and the carbon to nitrogen ratio is 10.2 at 17°C and 10.0 at 17.5°C. Assuming a global mean soil nitrogen content of 0.116%, the active soil mass at 17.5°C is 2.586×10^{14} metric tons $[(3000 \times 10^9$ metric tons carbon$)/(10 \times 0.00116)]$. A 0.5°C decrease in temperature would increase the steady-state soil carbon content of this mass by 155×10^9 metric tons.

This crude calculation is made only to show that the sensitivity of soil carbon to the climate change of the last 40 years might cause soils to serve as a net carbon source or sink on the order of 300 to 400×10^9 metric tons/degree. Associated humidity changes could negate or enhance this effect. These calculations are based rather dangerously on a presumed normality in the carbon to nitrogen composition of lignoprotein complexes in humus. The apparent rate constants for gain or loss of humus in the surface 30 cm of tropic soils lie in the range of 2 to 5% of the standing amount per year (Nye and Greenland, 1960). The constants are variable with climate, disturbance, and past history, but they are large enough that global changes approaching 1% per year might occur with climate changes.

Carbon Transfers in the Biosphere

Estimates are even more uncertain for global net primary production than for biomass. The production, assimilation, and accumulation of reduced carbon in an ecosystem is an enormously complex business (Batzli, 1974). Green plants, through their photosynthetic activities, hold the key position.

At present, we have no satisfactory means for estimating true gross production by these organisms. Gaseous carbon exchange in photosynthesis cannot be distinguished from dark respiration in the photosynthetic organ or from the flow of reductant to other acceptors such as NO_3^-, SO_4^{2-}, and incompletely reduced carbon acceptors. We must usually be content with measurements of the net CO_2 exchange of leaves, net energy gain, or net biomass gain. Since organic substances are transported rapidly from leaves, energy and biomass measurements must involve the whole plant, and thus also the respiratory activities associated with biosynthesis, transport, and maintenance (Penning de Vries, Brunsting, and van Laar, 1974). For the whole plant, then, we can speak only in terms of "net production." (As photosynthates are used in the growth of higher plants, the reduction level of the carbon increases and the associated respiratory quotients (CO_2 released/O_2 consumed) are greater than 1.0. The overall respiratory quotient for the complete oxidation of biomass through grazer and decomposer activity must then be less than 1.0.)

Despite these difficulties, we have considerable understanding of production processes in plant communities (e.g., Loomis and Williams, 1969; Loomis, Williams, and Hall, 1971; Loomis and

Gerakis, 1975). Irradiance level and CO_2 level are critical to the efficiency of photosynthesis. Their influence is dependent upon environment (e.g., radiation geometry, atmospheric turbulence, temperature), canopy factors (e.g., aerodynamic porosity, angles and density of leaf display, optical properties of leaves, and the distribution of nonphotosynthetic structures), physiological and morphological capability (e.g., maximum quantum efficiency and stomatal behavior), and the physiological state of the system (nutrient and water status and the degree of physiological and morphological adjustment to the current environment).

Important factors in translating this complexity into annual net primary production are the extent and duration of foliage cover and of a favorable environment, and the amount of living biomass. As living biomass increases, the maintenance respiration load also increases. Two full-cover communities that have similar photosynthetic properties but are widely different in biomass, for example, a pasture and a forest, may have similar rates of photosynthesis, though pasture may greatly exceed forest in net primary production because of lower maintenance respiration (McCree, 1970; Kira, 1975; Loomis and Gerakis, 1975). In addition, growth respiration per unit biomass would be greater for trees because of the more reduced state of their final products.

The foliage cover can be limiting to production. Light interception and net photosynthesis of the community increase in a diminishing-returns relation to increasing leaf-area index (m^2 of leaf/m^2 of ground) because the degree of mutual shading also increases. Thus, "net assimilation rates" (or "mean productive efficiencies," net production expressed per unit leaf area or unit chlorophyll) compared for different communities give rather meaningless results since the net production rate increases as leaf area increases whereas net assimilation rate declines. A grass community with a leaf-area index of 4 to 6 can provide complete interception and a maximum production rate. In forests, wide vertical separation of the leaf elements relative to their size allows greater penumbral scattering, so more leaves may be necessary for net photosynthesis to be maximum.

Some plant ecologists accept the erroneous conclusion that a massive and complex community is necessary for high net production. In fact, large biomass is more a *result* of vegetation type and/or high production than the cause. The linear correlation between net annual production and leaf-area duration (time-integrated leaf-area index) (Kira, 1975) is much more meaningful than the association between production and biomass (Whittaker and Woodwell, 1971). If moisture is favorable for most of the growing season, trees will survive on most sites. Their greater stature favors success in competition with grass, and a forest of large biomass results. That is the situation in the moist tropics, where long growing seasons and favorable moisture are only partly offset by short days and low radiation (humid atmospheres). Net annual production is also estimated to be large. If one reinterprets biomass as a measure of "site index," the correlation simply shows that a heavy biomass accumulates in long-season environments that have favorable temperature and moisture.

There are a number of reasons for being very cautious about the production estimates for tropical forests in Table 1. The best data came from destructive sampling of regrowth forests (allowing standing biomass to be averaged over the regrowth period; Nye and Greenland, 1960). Such values are undoubtedly larger than the rates for steady-state mature forests burdened with maintenance respiration. Because of the lack of seasons in the tropics, determination of net annual increment in stem diameter can be done by growth-ring analysis for only a few species, and repeated measurements of stem diameter are required for application of the dimensional analysis technique. That has been done at only a few sites (e.g., IPEAN, Belem, Brazil).

It seems to have been more common to measure leaf fall and then multiply by 2.5 to 3 for an estimate of total production (Murphy, 1975). That assumption is drawn from Rodin and Bazilevich (1967), Bray and Gorham (1964), and other sources who give stem production as about 60% of net primary production and 75% of biomass.

Consider the implications of that with the mean values given for tropical rainforests in Table 1. An annual net production of 20 metric tons/ha might allow 12 metric tons ha^{-1} $year^{-1}$ to accrue aboveground as stem tissue. Taking 340 metric tons of the 450 metric tons/ha standing crop as stem tissue and dividing by 12, gives 28 years as the average longevity of biomass. With 300 trees per hectare, 11 must fall each year. More realistically, with 100 dominant trees containing 80% of the biomass and 200 understory trees accounting for 20%, 3 to 4 large trees fall each year. Other numbers are possible, and stem growth may be as low as 6 metric tons ha^{-1} $year^{-1}$. But, either way, such forests would be for hard hats only. In Amazon forests at least, the fall of a dominant tree is very rare. The event is probably more important for the release of suppressed seedlings than for turnover. Again, we need better numbers and more of them before we can be very explicit about tropical forests.

Nutrient Cycling

The supply of available nutrients is a critical factor for maintaining high rates of net production when water and temperature are not limiting. For example, herbaceous plant material commonly contains 2% or more of nitrogen on a dry basis, and net primary productivity as high as 20 metric tons ha^{-1} $year^{-1}$ would require 400 kg nitrogen as new or recycled supply. Intensive agriculture has achieved four- to fivefold greater net primary productivity, although only with massive nitrogen fertilization (up to 2000 kg nitrogen/ha) to saturate the growth rate. In natural systems, the soil nitrogen pool may be much larger than 2000 kg but annual cycling rates are much smaller.

Ulrich and Gersper's (1977) view is that the tundra is severely limited for nitrogen and that net primary productivity is held well below potential. Trees, on the other hand, are less limited by nutrient cycles. As they increase in girth, through cambial activity, the protoplasm (high in nitrogen) of the xylem elements in new wood is largely reassimilated by living tissues. The dead heartwood has a nitrogen content of only 0.1 to 0.5%. Thus, a relatively small pool of active nitrogen in the cambium can accumulate large amounts of wood biomass with only small inputs.

Returning to a tropical forest with 7 metric tons ha^{-1} $year^{-1}$ of leaf litter (1% nitrogen) and 12 metric tons ha^{-1} $year^{-1}$ of new stem (0.2% nitrogen), at least 100 kg nitrogen/ha must cycle or be added each year, whereas 40 metric tons ha^{-1} $year^{-1}$ net primary productivity would require 200 kg nitrogen; 100 to 200 kg nitrogen is a reasonable possibility. The marked difference in nitrogen requirement between herbs and trees may account in part for the forest's advantage over grass. Productivity may also be limited by other nutrients, and the highly weathered oxisols of wet tropics are commonly low in phosphorus and potassium. Nutrient availability is particularly limiting to oceanic productivity.

BIOSPHERE RESPONSE IN INCREASED ATMOSPHERIC CO_2

Direct Effects of CO_2

The diminishing-returns relationship between CO_2 uptake by a leaf in normal air and increasing leaf irradiance is depicted in Fig. 1. In vegetation, the irradiance to each leaf depends upon radiation geometry and exposure. The slope of this curve (moles CO_2 absorbed/moles quanta, 400 to 700 nm) provides us with the quantum efficiency, ϕ. The variable ϕ is maximum with light flux limiting and declines in high irradiance as the flux of CO_2 to fixation sites becomes limiting.

With the most common, "C_3," photosynthetic pathway (Calvin-Benson carbon pathway; 99% of vegetation), both regions of the curve are enhanced by increasing CO_2 concentrations outside the leaf. In C_3 plants, the carboxylating enzyme can also act as an oxygenase and release CO_2 in the process known as photorespiration; the kinetics are analogous to competitive inhibition of CO_2 fixation by O_2. Increasing CO_2 enhances ϕ at all light levels; changes in ϕ_{max} are illustrated in Fig. 2 (Ehlringer and

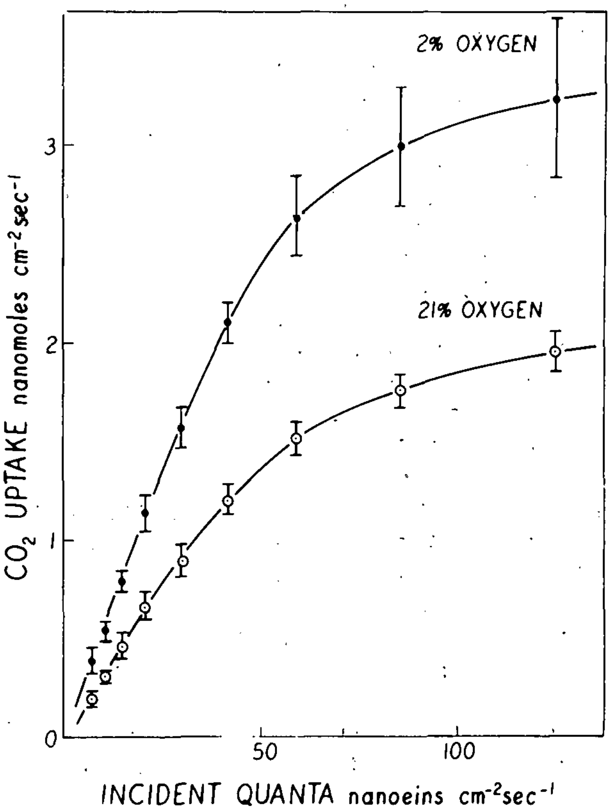

Fig. 1. Net photosynthesis of sugar beet leaves as a function of irradiance, typifying the behavior of sun-adapted C_3 leaves. The measurements were made in low O_2 (photorespiration repressed) and normal 21% O_2. A low external CO_2 concentration of 225 μbar accentuated the inhibition by O_2. At 124 nano einsteins $cm^{-2}sec^{-1}$ of photosynthetically active radiation (400 to 700 nm; light saturation). the response to increasing CO_2 was approximately linear to over 400μbar: CO_2 uptake $= -0.52 + 0.11\ (CO_2)_{ext}$. (Hall and Loomis, 1972.)

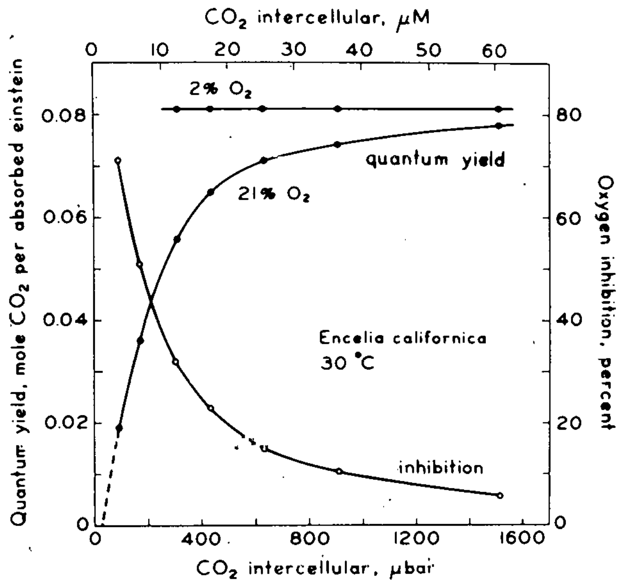

Fig. 2. Quantum yield for CO_2 uptake in *Encelia californica* (a coastal zone shrub of Southern California; C_3 species) as a function of intercellular CO_2 concentration (i.e., corrected for leaf diffusion resistance) at 21% and 2% O_2. O_2 inhibition is calculated from the difference in quantum yields at the two O_2 tensions. (From Ehlringer and Björkman, 1977.)

Björkman, 1977). With external CO_2 at 325 μbar, ϕ_{max} for this species is 0.052 moles CO_2 per mole quanta, or nearly the same as for C_4 species. C_3 plants become increasingly more efficient as CO_2 increases ($\phi_{21\%} = \phi_{2\%} - 10\phi_{2\%}/C_i$; Fig. 2), whereas C_4 plants do not. The situation with high irradiance is more complex. Increases in CO_2 enhance the diffusion gradient for CO_2 flux through stomates and the leaf protoplasm complex, but that process tends to be offset by a decrease in stomatal conductance (Lemon, 1976).

The effect of these changes is to increase photosynthesis by individual leaves and foliage arrays. Water use efficiency (net primary productivity/transpiration rate) is also increased, a significant point considering the extent to which annual net primary production is limited by moisture supply.

Much less is known about the influence of CO_2 concentration as a regulator of growth and development. CO_2 concentration can markedly alter root growth (Radin and Loomis, 1969), and Purohit and Tregunna (1974) report altered sensitivity of photoperiodic processes. At the same time, plants are highly adaptable physiologically and, over time, genetically. That is true for photosynthesis as well as for other traits. With significant CO_2 increases, one might expect small changes in the relative ecological performance of particular species, and thus some modification in the composition of vegetation. For example, genotypes in which increased CO_2 accelerated development so that flowering and maturity occurred earlier might have an increased advantage where season length was limited by temperature or moisture stress. In localities with long growing seasons, however, the same genotype might be at a greater disadvantage than before. My feeling is that genetic shifts would generally cover such situations and that the main response of vegetation would be increased productivity.

Increased production, clearly, should yield a greater standing biomass and a higher steady-state level of soil carbon. It is not clear where and to what extent we might observe such increases; water-limited regions appear as prime candidates. Where the vegetation is limited by nutrient supply (e.g., tundra) little response could be expected unless some of the production could be channeled to greater N_2 fixation (symbionts or free-living) or to nutrient cycling.

Climate Shift

Climatic changes associated with CO_2 increases appear likely to have a much greater impact than any direct effect of CO_2 in photosynthesis, although interactions will also occur. Returning to Fig. 2, Ehlringer and Björkman (1977) found in C_3 plants that ϕ_{max} is temperature-dependent, following the relation at 325 μbar CO_2:

$$\phi_{21\%,t} = \phi_{1\%} - \exp[(E_a/RT) + 9.79] ,$$

where E_a, the apparent energy of activation, equaled −8.1 kcal/mole. A rise in mean temperature from 20° to 25°C would decrease ϕ_{max} by 7%, from 0.0623 to 0.0579, about equal to the gain achieved by raising CO_2 by 10% while increasing the rate of photosynthesis in saturating light. Extremely complex interactions also will occur through temperature effects on growth rate, thus influencing the expansion and longevity of leaf area and the activity of photosynthetic sinks.

A glance at the landscape reveals that subtle changes in climate can cause dramatic shifts in natural vegetation, agriculture, and forestry. Here we are at the mercy of the climatologists: detailed predictions are needed of seasonal shifts in temperature, moisture, and cloudiness. The rapidity of change and the variability of the new climate are particularly important, since vegetation composition and geographic extent are strongly determined by the occasional severe event.

Given a climate prediction, a number of methods are available for predictions of change. The simplest is to compare ecosystems along existing environmental gradients—for example, with similar temperatures but differing moisture supplies. While it is not always easy to quantify such gradients and to recognize controlling rare events, qualitative predictions about the direction in vegetation shifts are then possible. Similar predictions about global biomass and its exchange of CO_2 are beyond our present capability.

Here again, we can expect genetic shifts within existing populations as well as expansion and contraction of territories. Agriculture is actually rather well insulated from climate shifts if you are willing to accept geographical dislocations. A very wide range of germ plasm exists in most cultivated species. In addition, production of particular crops would simply move to a new location that better suited their culture. If the old location became warmer and drier, we might find sorghum in place of corn; wheat would serve if it became cooler and drier, and potatoes if it was cooler and wetter. The sufficiency of the new patterns would depend upon land areas, yield levels, and demand.

Quantitative Predictions

The possibility of changes in atmospheric CO_2 poses a series of difficult questions in mensuration and quantitative prediction of biosphere response. The required degree of resolution on net changes and response cannot be obtained from inventories such as that given in Table 1, even if they are highly detailed. Such inventories are nevertheless extremely important, and every effort should be made to sharpen their focus on both pool sizes and transfers. More promise for monitoring is found in Stuiver's (1977) approach and in kinetic analysis of the fluctuations in atmospheric CO_2 as attempted by Hall, Ekdahl, and Wortenberg (1975) and Verma and Rosenberg (1976).

The prediction questions are multivariate problems of great complexity. Experiments are prohibitive in cost and scope, or impossible, and the only reasonable alternative is through predictive simulation models. Such models are still inadequate for the tasks. Two basic types of approaches have been used: multivariate models based on system-level associations, and hierarchic, dynamic simulators based on physiological processes.

Associative models have been employed extensively for climate analysis in agriculture (e.g., Robertson, 1973; Thompson, 1969) and to a lesser extent in forestry and natural systems. Lieth's (1975) effort at a global productivity model is particularly relevant. My conclusion is that such models lack sufficient structure to deal with the issues at hand. The historical data base may not include the experiences from which to predict future interactions, and the relationships are easily confounded and too readily forced into dependence on the chosen independent variables. In short, obvious answers and wrong answers are readily achieved.

The "biome models" constructed by large systems teams for representative ecosystems are generally intermediate in nature. Some, like the grassland model, sacrifice detail and realism in productivity estimates to include more information on lower trophic levels, thus achieving a total-system model. The simpler tundra system allowed the modelers to give greater emphasis to the physiology of production and growth.

The intermediate forest simulator developed by Botkin, Janak, and Wallis (1972, 1973) predicts that a 100% increase in tree growth (unknown CO_2 increase) will lead eventually to a 50% greater basal diameter, and thus a considerable increase in carbon storage, when the forest is at steady state. The model structure, however, is highly empirical in its treatment of photosynthesis (annual insolation × leaf density factor × growing-season degree-days × moisture factor × photosynthesis constant), and specific CO_2 levels cannot be simulated. The model has not been validated, and its treatment of competition, microclimate, moisture stress, tree growth, and other issues is based on very general associative functions with limit constraints. Lemon (1976) and Woodwell and Houghton (1977) place more value on its predictions than seems merited.

SPAM-type vegetation models (Shawcroft, Lemon, and Stewart, 1973) can deal much more realistically with the influence of CO_2 on photosynthetic production, since they consider solar geometry, canopy architecture, and physiological responses including transpiration (hence leaf temperature), stomatal resistance, and photosynthesis. SPAM predicts a 22% increase in net photosynthesis of C_3 foliage as CO_2 is increased from 315 to 400 ppm in a temperate environment (Lemon, 1976). A key point is that the model is structured to handle additional detail, such as Ehlringer and Björkman's (1977) findings, and nonrandom leaf distributions (Fukai and Loomis, 1977).

However, SPAM does not deal with growth and development and competition, which the Botkin model attempts to approach. Growth simulators of an advanced nature have been developed (ELCROS: deWit, Brouwer, and Penning de Vries, 1970; SUBGOL: Loomis, Ng, and Hunt, 1976) but so far they deal only with monocultures of uniform age. Considerable further effort, and very large computing capabilities, will be required for a sufficient vegetation model that couples microclimate simulations with SPAM-ELCROS models in a form that handles mixed species competition at uneven ages. Such models, used with representative vegetations, will be of enormous power in monitoring and prediction.

CONCLUSIONS

It is important to recognize the great diversity and adaptive nature of vegetation. While broad generalizations can be drawn about the behavior of terrestrial and aquatic ecosystems, we presently lack the information base and tools necessary for a careful assessment of the biosphere's response to changes in atmospheric CO_2. Particular difficulties are posed by unmapped and unsampled patchiness; our ignorance of soils, of tropical forest ecology, and of physiology in general; and the rapid changes in vegetation under natural and human influences. Soils are shown here as potentially the most significant source/sink in the biosphere.

The problems are largely human. I have confidence in the ability of the biosphere, or should we say *a* biosphere, to adapt to and survive climate and CO_2 changes. Whether those changes will be tolerable to man and whether their effects can be ameliorated are the issues. My chief concern is that we lack much of the necessary basic biological and environmental information as well as the means with which to address the problem.

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Carbon Dioxide Chemistry in Ocean Water

Taro Takahashi, Lamont-Doherty Geological Observatory
of Columbia University, Palisades, New York

1. THE REVELLE FACTOR

The partial pressure of CO_2, p_{CO_2}, exerted by seawater is a function of temperature, total dissolved CO_2 species, and alkalinity. The total CO_2, T_{CO_2}, and the alkalinity, ALK, are defined as:

$$T_{CO_2} = (CO_2)_{aq} + (HCO_3^-) + (CO_3^{2-})$$

$$\begin{aligned} ALK = {} & (Na^+) + (K^+) + 2(Mg^{2+}) + 2(Ca^{2+}) \\ & - [(Cl^-) + (Br^-) + 2(SO_4^{2-})] \\ & - (HCO_3^-) + 2(CO_3^{2-}) + (H_2BO_3^-) \\ & + (OH^-) - (H^+)\,, \end{aligned}$$

where the parentheses indicate the concentration of respective ionic species. The alkalinity is the sum of excess anionic charges carried by weak acids.

The Revelle factor, R, is a factor which relates the CO_2 partial pressure to the total CO_2 concentration at constant alkalinity, temperature, and salinity. It is defined as:

$$\begin{aligned} R &= (\Delta p_{CO_2}/p_{CO_2})/(\Delta T_{CO_2}/T_{CO_2}) \\ &= [\partial(\ln p_{CO_2})/\partial(\ln T_{CO_2})]_{ALK,T,S}\,, \end{aligned}$$

where Δ indicates a small change in p_{CO_2} or total CO_2. Thus, the Revelle factor is a useful parameter for a study of CO_2 distribution between the atmosphere and ocean water. For instance, normal surface ocean water has a Revelle factor of about 10. This means that if the atmospheric p_{CO_2} was increased by 10%, the total CO_2 in seawater would be increased by 1% at equilibrium.

The values of the Revelle factor which have been computed using the CO_2 solubility of Weiss (1973), the apparent dissociation constants of carbonic acid in seawater of Mehrbach et al. (1975), and the apparent dissociation constant of boric acid in seawater of Lyman (1953) are shown in Fig. 1 for 35 parts per thousand salinity and a total borate concentration of 0.41 millimoles/kg. Curves for two

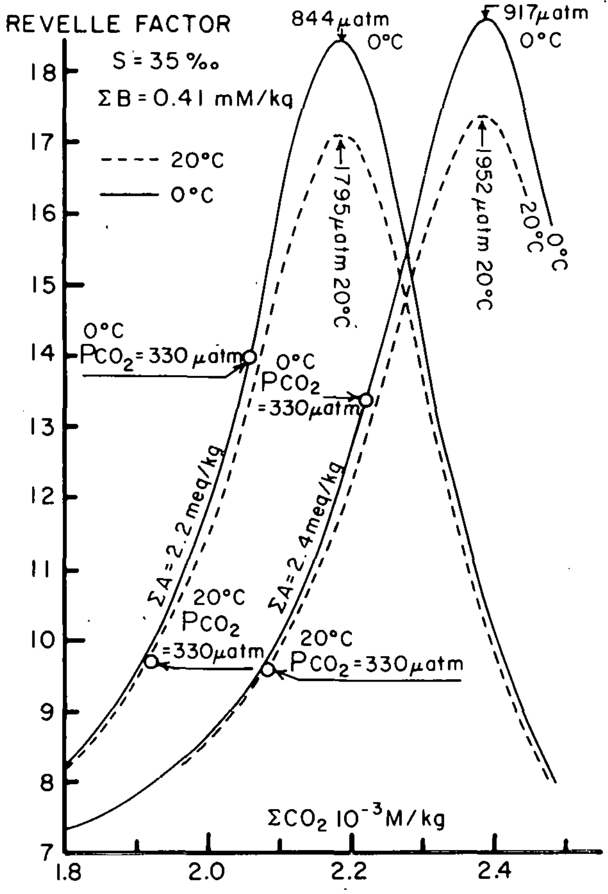

Fig. 1. The Revelle factor as a function of total CO_2 in seawater.

alkalinity values of 2.2 and 2.4 meq/kg, and for two temperature values of 0° and 20° C, which cover most of the oceanic range, are shown. The most interesting feature in these curves is that first the Revelle factor increases rapidly with increasing total CO_2 concentration, then it goes through a maximum, and finally it decreases to unity at large total CO_2 concentrations. For 0°C and an alkalinity value of 2.2 meq/kg, the maximum Revelle factor of 18.5 is located at a p_{CO_2} of 844×10^{-6} atm and a total CO_2 concentration of 2.19 millimoles/kg.

The conditions at which p_{CO_2} is 330×10^{-6} atm and hence equal to the present atmospheric p_{CO_2} are marked with circles in Fig. 1. The area defined by these four circles indicates the present-day oceanic range of the Revelle factor, total CO_2, and alkalinity.

As total CO_2 increases to an extremely large value, seawater becomes acidic (due to carbonic acid formed by reaction with water), and hence the predominant species in the dissolved CO_2 is CO_2(aq). In such a condition, total CO_2 concentration becomes nearly proportional to p_{CO_2} (or obeys Henry's Law). Therefore, the Revelle factor approaches unity as total CO_2 is increased.

During the Atlantic and Pacific phases of Geochemical Ocean Sections Study (GEOSECS) expeditions, alkalinity and total CO_2 values for a large number of samples have been determined. The Revelle factors for the Atlantic and Pacific surface ocean water have been computed using these data and are plotted as a function of temperature in Fig. 2. Although the Revelle factor has been shown to be a weak function of temperature as demonstrated in Fig. 1, it appears to be well correlated with the temperature of water in the Pacific and Atlantic oceans, and can be expressed by:

$$R = 13.6 - 0.18t\ (°C)$$

with an error root mean square of 0.5 in the Revelle factor. For an average of the world oceans a Revelle factor value of 10 is appropriate.

2. GEOGRAPHICAL DISTRIBUTION OF THE PARTIAL PRESSURE OF CO_2 EXERTED BY SURFACE OCEAN WATER

The distribution of surface water p_{CO_2} determined during the GEOSECS expeditions (1972 to 1974) in the Atlantic and Pacific Oceans is shown in Figs. 3 and 4. The average p_{CO_2} in the atmosphere during the same period was about 320×10^{-6} atm. Thus, the oceanic areas which have surface water p_{CO_2} values smaller than 320×10^{-6} atm are acting as a CO_2 sink, whereas those areas where the surface water p_{CO_2} exceeds the atmospheric value are exhaling CO_2 to the atmosphere. In the Atlantic (Fig. 3), the area north of 55° N (including the Labrador, Greenland, and Norwegian Seas) is listed as a major sink. A less intense and localized sink is observable east of the Falkland Islands in the southwestern Atlantic. In the Pacific, no major sink area has been found. Two weak sink areas are observed in a narrow strip corresponding to the Kuroshio extension in the western North Pacific, and east of New Zealand in the South Pacific. In both oceans, the equatorial areas are exhaling CO_2. The general pattern of GEOSECS surface water p_{CO_2} in the Pacific is entirely consistent with that presented by Keeling (1968).

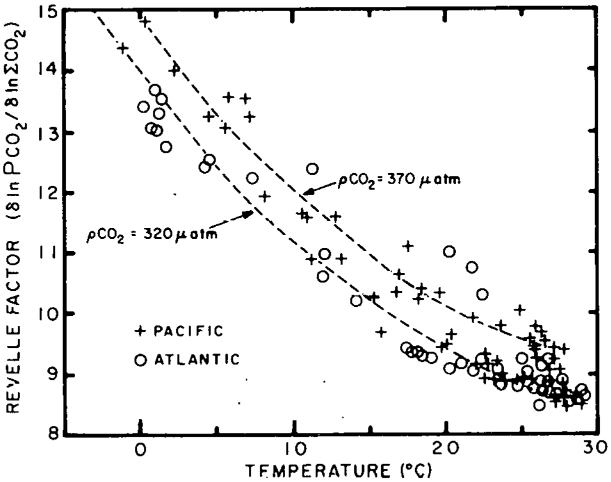

Fig. 2. The Revelle factor as a function of the temperature of the Atlantic and Pacific surface water. The Revelle factor was computed from the GEOSECS alkalinity and titration total CO_2 data using the dissociation constants of Mehrbach et al. (1973).

3. TIME VARIATION OF SURFACE WATER CO_2 PARTIAL PRESSURES IN THE ATLANTIC

In the western Atlantic, the surface water p_{CO_2} was determined by Takahashi (1961) during the 1957 to 1958 International Geophysical Year (IGY) expedition, and the results are compared with the GEOSECS data in Fig. 5. Note that the 1972 GEOSECS curve is displaced upward and northward from the 1957 to 1958 IGY curve. The northward

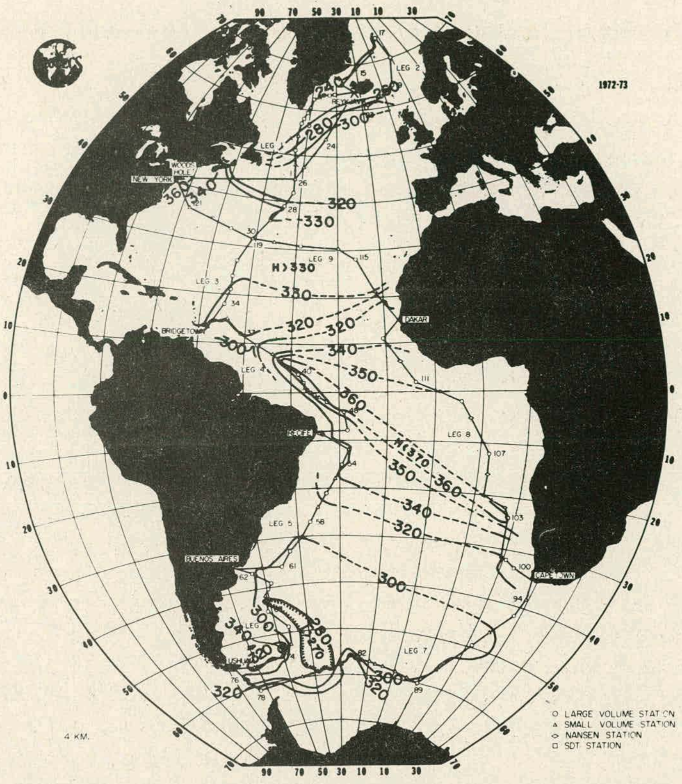

Fig. 3. Distribution of the surface water p_{CO_2} in the Atlantic Ocean, July 1972 to March 1973.

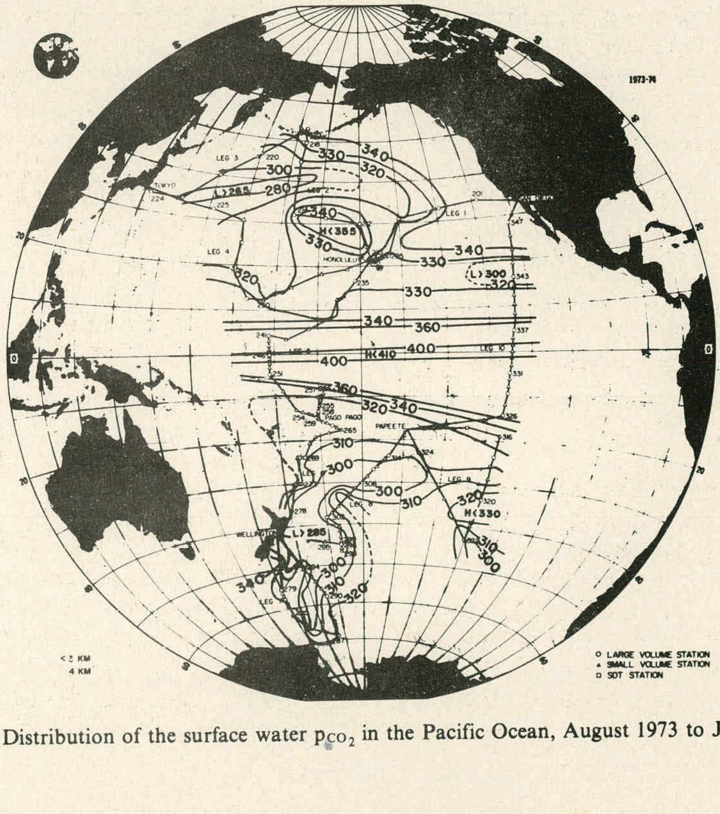

Fig. 4. Distribution of the surface water p_{CO_2} in the Pacific Ocean, August 1973 to June 1974.

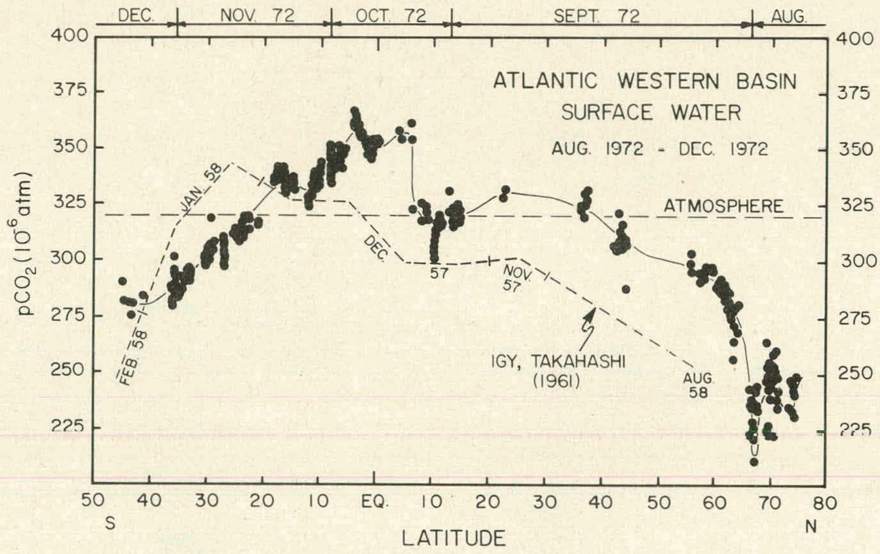

Fig. 5. A north-south profile of p_{CO_2} in the surface seawater and in the atmosphere over the western Atlantic basin.

displacement is likely due to the seasonal effect: the IGY expedition (November 1957 to March 1958) took place about three months later in the season than the GEOSECS expedition (August 1972 to December 1972). The upward displacement appears to be due to the increase in p_{CO_2} taking place during the 15-year period. Since the surface water p_{CO_2} values are sensitively affected by water temperature, the IGY p_{CO_2} values have been corrected to the temperatures of the GEOSECS samples at the same latitude using a temperature coefficient of 4.5% per degree Celsius. The pertinent data are summarized in Table 1. Although the temperature correction applied does not necessarily correct for geographic differences in the samples, no other correction has been made because of the lack of a reliable means for correcting the difference. However, correction for geographic variation appears to be small in the areas within a gyre system and away from boundary currents.

The mean annual rate of surface water p_{CO_2} increase observed in the western and southern Atlantic Ocean thus obtained is $1.8(\pm 0.4) \times 10^{-6}$ atm/year. This compares with the rate of atmospheric CO_2 concentration increase of 1.2 to 1.4 ppm (by volume) per year observed by Keeling and his associates at the Mauna Loa Observatory, Hawaii over the same 15-year period. The data indicate that the rate of p_{CO_2} increase in the surface ocean water is nearly equal to that in the atmosphere.

4. TIME SCALE FOR AIR-SEA CO_2 EXCHANGE

The time constant for surface ocean water to absorb CO_2 from the atmosphere is given by:

$$\tau_{CO_2} = \frac{(\text{total } CO_2 \text{ in mixed layer})}{(\text{Revelle factor})(CO_2 \text{ exchange flux})}$$

(Broecker, personal communication). Assuming that the mean CO_2 concentration in the mixed layer of the ocean is 2 moles m^3, the mean thickness of the mixed layer 150 m, the mean Revelle factor 10, and the CO_2 exchange flux 17 ± 5 moles CO_2 m^{-2} $year^{-1}$, we obtain

$$\tau_{CO_2} = 1.80 \pm 0.7 \text{ years}.$$

This means that mean surface water p_{CO_2} should lag behind atmospheric p_{CO_2} by one to two years. Since the mean rate of atmospheric CO_2 increase observed by Keeling at the Mauna Loa Observatory is about 1.3 ppm (by volume) per year, mean oceanic p_{CO_2} should be 1 to 3×10^{-6} atm smaller than the mean atmospheric p_{CO_2}. However, as shown in Figs. 3, 4, and 5, the geographic and seasonal variation of the oceanic p_{CO_2} is on the order of 200×10^{-6} atm and is 100 times greater than the expected difference between atmospheric and oceanic p_{CO_2} values at a given time. Therefore, it is

Table 1. Summary of surface water p_{CO_2} values obtained during the 1957 to 1958 IGY and 1972 GEOSECS expeditions in the western and southern Atlantic Ocean

Latitude range	1957–1958		1972		Temperature change, ΔT (°C)	Δp_{CO_2} (10^{-6} atm)
	Temperature (°C)	p_{CO_2} (10^{-6} atm)	Temperature (°C)	p_{CO_2} (10^{-6} atm)		
60–50°N	8.3	254 ± 10[a]	9.2	296 ± 13[a]	0.9	32
30–20°N	25.5	302 ± 2	27.4	331 ± 3	1.0	3
20–10°N	26.6	299 ± 4	27.8	318 ± 6	1.2	3
10–0°N	27.3	300 ± 6	27.5	339 ± 17	0.2	36
0–10°S	27.2	326 ± 8	26.6	350 ± 8	–0.6	33
10–20°S	27.5	327 ± 3	26.0	335 ± 4	–1.5	30
20–30°S	25.3	345 ± 6	23.4	312 ± 6	–1.9	–3
30–40°S	23.2	315 ±7	18.0	291 ± 5	–4.8	44
40–50°S	16.6	249 ± 10	13.0	284 ± 5	–2.4	62
50–60°S	4.1	294 ± 5	2.3	317 ± 18	–1.8	47
30–40°S(E)	17.2	256 ± 5	22.3	328 ± 9	5.1	13
40–50°S(E)	8.3	283 ± 8	8.5	294 ± 2	0.2	8
50–60°S(E)	3.5	289 ± 21	1.1	300 ± 5	–2.4	42

Mean increase in surface water $p_{CO_2} = 27(\pm 6) \times 10^{-6}$ atm, where (±6) is the standard deviation of the mean.

Mean annual rate of increase in the 15-year period = $1.8(\pm 0.4) \times 10^{-6}$ atm/year, where (±0.4) is the standard deviation of the mean.

[a] One standard deviation.

not possible to determine the amount of p_{CO_2} lag on the basis of direct observations. Although the oceanic lag in p_{CO_2} cannot be directly determined, observational data obtained in the Atlantic over the 15-year period indicate that the mean oceanic p_{CO_2} increases at a similar rate as atmospheric p_{CO_2} increases. Thus, no unexpected kinetic barrier appears to exist in the CO_2 exchange processes between the oceanic and atmospheric CO_2, and the CO_2 exchange flux estimated from the natural as well as the bomb-produced C-14 distributions appears to be consistent with the p_{CO_2} data.

5. SIGNIFICANCE OF THE NORTH ATLANTIC OCEAN

In Sect. 2, it was pointed out that surface water p_{CO_2} values observed north of 55°N in the North Atlantic are as much as 30% lower than the atmospheric values. Since the observation was made in July–August when the water temperature is highest, the p_{CO_2} in this region of the ocean is expected to be considerably lower during the winter months. Thus, atmospheric CO_2 is being actively absorbed into the surface water.

The tritium data recently obtained for the western Atlantic by Ostlund and his associates at the University of Miami are reproduced in Fig. 6. Their tritium (half-life of 12 years) data show that a downward penetration of tritium to 4000 m deep in the northern end has taken place since the thermonuclear bomb tests of 12 years ago. The data also indicate active sinking of the cold and dense surface water outflow over the Iceland-Scotland sill located at about 65°N. The data for bomb-produced C-14 obtained by Ostlund and his associates in this region also show an active transport of the surface water to the deep water regime. Since the dense and low-p_{CO_2} surface water from the Greenland and Norwegian Seas absorbs atmospheric CO_2 by the time it reaches the source area for deep water, the sinking water transports atmospheric CO_2 including fossil fuel CO_2 into the deep water regime. The volume of the Iceland-Scotland overflow has been estimated to be about 5×10^6 m^3/sec. If this water was fully equilibrated with atmospheric CO_2 before sinking, it could transport about 8×10^{12} moles/year CO_2. This rate corresponds to 2% of the 1975 fossil fuel CO_2 production rate. Since about 10% of present-day atmospheric CO_2 is of fossil fuel

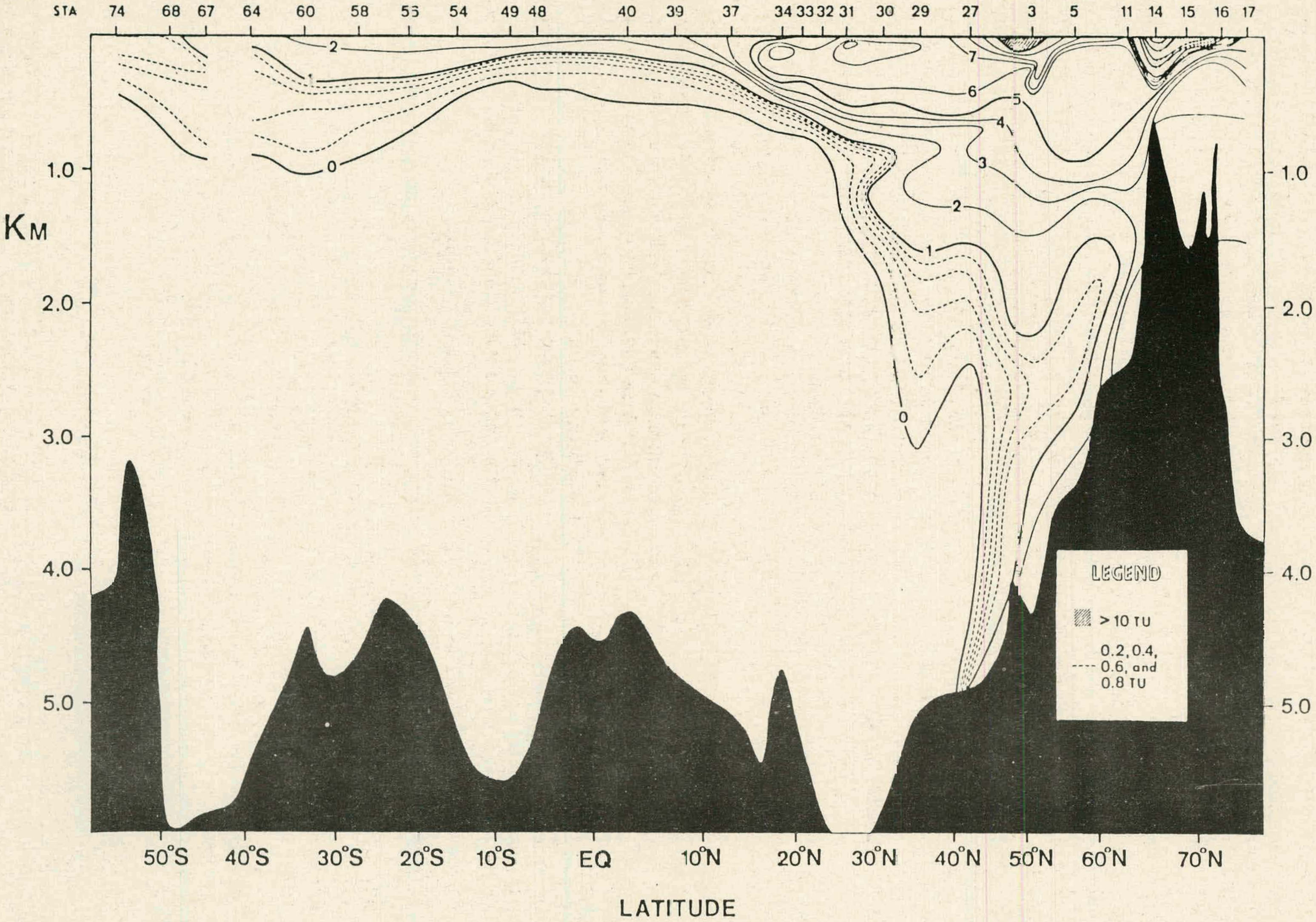

Fig. 6. The distribution of tritium in the Western Atlantic Ocean, 1972 to 1973. (From: Ostlund, H. G., H. G. Dorsey, and R. Brescher. December 1976. *GEOSECS Atlantic radiocarbon and tritium results (Miami)*. Tritium Laboratory Data Report No. 5, Rosenstiel School of Marine and Atmospheric Sciences, Univ. of Miami, Fla.)

origin, 0.2% of fossil fuel CO_2 is being lost to the deep sea regime in the North Atlantic.

The newly formed North Atlantic deep water contains a greater CO_2 concentration since it was exposed to atmosphere containing a greater CO_2, and thus it should be more acidic than older water. This acidic deep water will eventually sink deep enough to react with the calcareous deep sediments, and it will be neutralized by $CaCO_3$. A preliminary model for the neutralization process has been discussed by Broecker and Takahashi (in press). They estimate that the time scale for the deep sea neutralization of fossil fuel CO_2 is about 1500 years or shorter. Although this process is not important for short-range forecasting of the fossil fuel CO_2 level in the atmosphere, it must be taken into account for a long-range prediction.

6. COMMENT ON CO_2 MODELS

As an increasing amount of observational data is being accumulated in recent years, a number of models which deal with the distribution of CO_2 in the atmosphere/ocean/biosphere have been developed by Craig (1957), Revelle and Suess (1957), Bolin (1960), Broecker, Li, and Peng (1971), Fairhall (1973), Bacastow and Keeling (1973), Keeling (1973), Ekdahl and Keeling (1973), and Oeschger et al. (1975). With the exception of Oeschger et al. (1975), Broecker, Li, and Peng (1971), and Fairhall (1973), these workers have employed various box models. On the other hand, Oeschger et al. (1975) considered that a well-mixed surface ocean layer is coupled to a diffusive deep ocean. Although Broecker, Li, and Peng (1973) and Fairhall (1973) treated the transfer of CO_2 between the surface mixed layer and the deep ocean in terms of eddy diffusion, their treatment ended up basically the same as the box models due to the simplifying assumptions used. Thus, it is worthwhile to briefly discuss the significance of the model developed by Oeschger et al. (1975).

Oeschger and his colleagues considered four reservoirs: the atmosphere, the mixed layer of the ocean, the deep ocean, and the biosphere. The atmosphere with a homogeneous CO_2 distribution is coupled to the homogeneously mixed layer by means of CO_2 evasion and invasion rate constants. The CO_2 in the mixed layer is transferred to the deep ocean via eddy diffusion. They estimated the mean eddy diffusivity in the deep ocean using the steady-state natural C-14 distribution observed in the Pacific and Atlantic Oceans. The vertical eddy diffusivity thus obtained is 1.27 cm^2/sec (3987 $m^2/year$). It is noteworthy to point out that this value is consistent with the 1.3 cm^2/sec estimated by Broecker and his associates using data on tritium (in thermocline depths), radon-222, and radium-228. The CO_2 transfer coefficient from the atmosphere to the mixed layer and that from the mixed layer to the atmosphere have also been determined to be 1/7.7 and 1/10.0 year, respectively.

Using the transfer coefficient, the vertical diffusivity, and the total fossil fuel CO_2 input to the atmosphere, Oeschger et al. computed the atmospheric CO_2 concentration and the Suess effect starting in 1860. The calculated values are in good agreement with Keeling's Mauna Loa atmospheric CO_2 data and the observed Suess effect (C-14 in the atmosphere) (Fig. 7). They also noted that the observed atmospheric CO_2 increase and the Suess effect can be accounted for without assuming growth of the biosphere, although a biosphere growth rate up to 40% of the atmospheric CO_2 increase rate is permitted.

The model of Oeschger et al. thus appears to explain the major features of atmospheric CO_2 and C-14, and it is considered to be the most reliable model which can be used for short-term prediction. However, by its own nature, this model is a perturbation model and is lacking in a formulation of various oceanographic processes. Therefore, caution must be exercised against its use as a long-term predictive model.

One of the most critical parameters in this model is the vertical eddy diffusivity in the deep ocean, which has been assumed to be constant. Using the

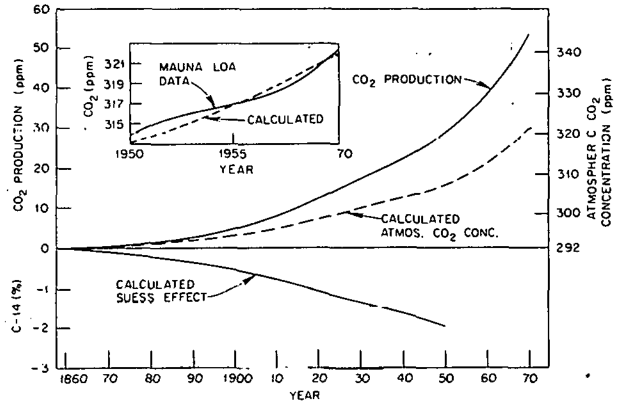

Fig. 7. The atmospheric CO_2 and C-14 concentrations calculated by Oeschger et al. (1975) and their comparison with the observed data. The total CO_2 production estimated by Keeling (1973) is used as input.

tritium, radon-222, and radium-228 data, Broecker and his associates (e.g., Sarmiento et al., 1976; Quay, 1977) have shown that the vertical eddy diffusivity in oceans and lakes is inversely proportional to the square of the Brunt-Väisälä buoyancy frequency (or proportional to the vertical density gradient) as shown in Fig. 8. Such a relationship is valuable for further improvement of CO_2 exchange models.

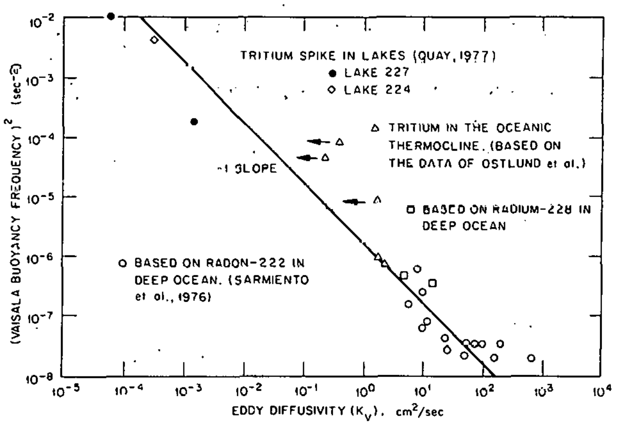

Fig. 8. The relationship between the vertical eddy diffusivity and the square of the Väisälä buoyancy frequency (or the density gradient) in the lakes and oceans. (After W. S. Broecker, Lamont-Doherty Geological Observatory, Columbia Univ., Palisades, New York.)

7. SUMMARY AND CONCLUSIONS

1. The Revelle factor, the ratio of the fractional change in p_{CO_2} exerted by seawater to that in total CO_2 dissolved in seawater, is an important parameter in the air/sea CO_2 exchange study. It has been computed over a wide range of total CO_2 values and has been shown to increase rapidly from 8 to 18 with increasing total CO_2. In an oceanic alkalinity range of 2.2 to 2.4 meq/kg, it passes a maximum at a total CO_2 value 10 to 20% greater than the mean total CO_2 value of 2.00 millimoles/kg in the surface ocean water. For oceanic alkalinity and total CO_2 range, the Revelle factor increases from 10 to 18 as the p_{CO_2} increases from 330 to 900 × 10^{-6} atm. This indicates that the surface ocean water tends to absorb proportionally less CO_2 from the atmosphere until the atmospheric CO_2 concentration is raised to three times the present value.

2. The p_{CO_2} data obtained in the Atlantic and Pacific Oceans during the GEOSECS expeditions (1972 to 1974) identify the northern North Atlantic as the active sink area for atmospheric CO_2.

3. A comparison of the Atlantic surface water p_{CO_2} data obtained during the IGY (1957 to 1958) and the GEOSECS expeditions (1972 to 1973) shows that, during the 15-year period, the surface water p_{CO_2} in the Atlantic increased at an average rate of $1.8(\pm 0.4) \times 10^{-6}$ atm/year. This rate is comparable to the atmospheric CO_2 increase rate of 1.2 to 1.4 ppm (by volume) per year observed by Keeling at the Mauna Loa Observatory, Hawaii. It appears the oceanic p_{CO_2} increases at a similar rate as the atmospheric CO_2.

4. The box diffusion model of CO_2 distribution in the atmosphere/ocean system formulated by Oeschger et al. (1975) yields the observed atmospheric CO_2 increase rate and the Suess effect. Thus, among various models which have been proposed, this model appears to be the most realistic. It may be used reliably for short-term forecasting of atmospheric CO_2 concentration. However, for a reliable longer-term forecasting method, more oceanographic features, such as regional differences in the vertical eddy diffusivity, the effect of fossil fuel CO_2 neutralization by marine carbonates, particulate carbon flux, and vertical flow velocity should be incorporated into the model.

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Models to Predict Future Atmospheric CO_2 Concentrations

Robert B. Bacastow and Charles D. Keeling, Scripps Institution of Oceanography,
University of California at San Diego, La Jolla, California

INTRODUCTION

From current knowledge of the geochemical reservoirs which store and transfer carbon on a global basis, we have attempted to predict future concentrations of atmospheric CO_2 for various assumed patterns of fossil fuel consumption. To carry out these predictions, we have depended principally on simple nondimensional chemical tracer reservoir or "box" models, but we have compared these models with a model in which the ocean is assumed to mix tracers vertically by diffusion. Calculations of atmospheric CO_2 increase using reservoir models have been made by Revelle and Suess (1957), Bolin and Eriksson (1959), Machta (1973a and b), Cramer and Myers (1972), Keeling (1973a), Bacastow and Keeling (1973), Zimen and Altenhein (1973), Hoffert (1974), Gowdy, Mulholland, and Emanuel (1975), Niehaus (1976), Keeling and Bacastow (1977), and Revelle and Munk (1977). The diffusive model was developed by Broecker, Li, and Peng (1971), and Oeschger et al. (1975).

LINEAR MODELS

A basic assumption in these tracer models is that there formerly existed a preindustrial steady state which would exist today had not fossil-fuel-derived CO_2 been added to the atmosphere since the advent of the industrial revolution. The inferred preindustrial carbon abundances of the various reservoirs are disregarded by subtracting them from the time-dependent carbon abundances. Thus only perturbations in abundance are considered.

Because these perturbations, at least until now, are relatively small, they can be reasonably well described by linear equations. Also, as Rotty has indicated in this workshop, the past input of industrial CO_2 derived from fossil fuel, at least since 1945, is satisfactorily represented by an exponential function with a characteristic, or *e*-fold, time of approximately 22 years. A linear reservoir model with an exponential tracer input predicts that a constant fraction, of the input will remain in each reservoir after times long compared to the exponential time constant of the input (see Appendix B). Thus, as long as the carbon system behaves in a nearly linear manner, a continued exponential growth in the use of fossil fuel will result in a nearly exponential rise in atmospheric CO_2.

AIRBORNE FRACTION

We shall define the airborne fraction of industrial CO_2 at a fixed location as the ratio of the observed increase in atmospheric CO_2 concentration at that location to the increase which would have occurred there if the industrial CO_2 produced over the same period had all remained airborne and been evenly distributed throughout the atmosphere. The airborne fraction since 1958 has been estimated from observations of atmospheric CO_2 at the Mauna Loa Observatory, Hawaii, and at the South Pole (Keeling et al., 1976a and b; Keeling, 1977), combined with data on industrial CO_2 production from fossil fuel and cement manufacture (Keeling, 1973b; Rotty, 1973).

The airborne fraction has been found to vary both spatially and temporally. Monthly averages of CO_2 at Mauna Loa and the South Pole indicate a seasonal variation and a long-term, or "secular," trend. Before the airborne fraction at a station can be calculated, the secular trend must be established. For Mauna Loa, the curve in Fig. 1 illustrates our analysis of the original monthly observations (shown by dots). This curve is derived by a least squares fit to these data of a function which expresses the sum of a seasonal effect, assumed to be the same each year,

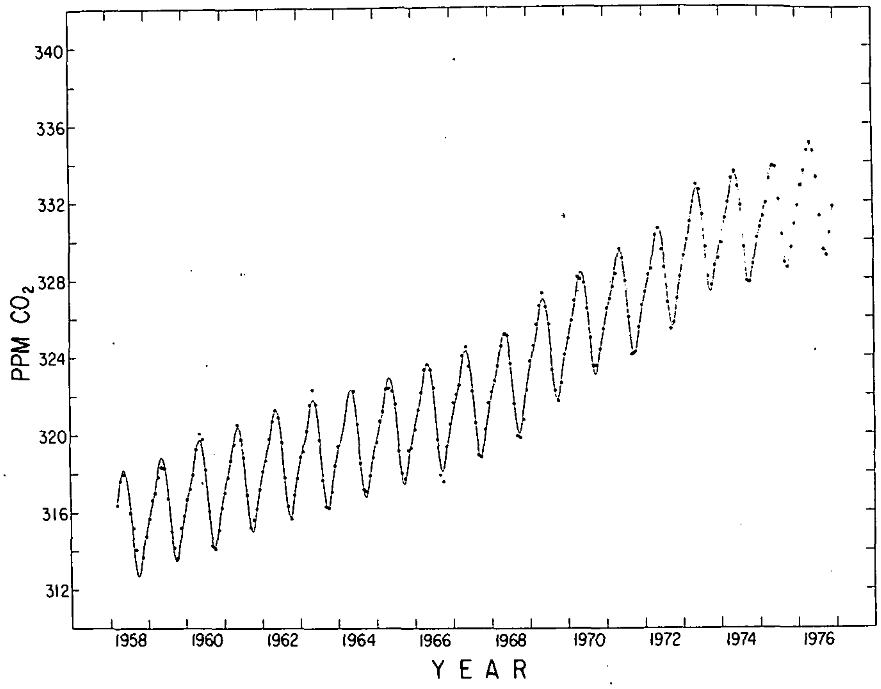

Fig. 1. Concentration of atmospheric CO_2 at Mauna Loa Observatory, Hawaii, at 19.5°N, 155.6°W. The dots indicate monthly averages based on continuous measurements. The curve is a fit to the monthly averages of the sum of the average annual variation and a spline function representation of the seasonally adjusted secular trend. The stiffness of the spline function (Reinsch, 1967) was set so that the curve fits the seasonally adjusted data within the precision of the monthly means.

and a secular trend, represented by a spline function. The secular trend is shown in Fig. 2. Monthly averages for the South Pole (Fig. 3) reveal a similar secular trend (Fig. 4). The Mauna Loa record was obtained from a continuous infrared analyzer operated by National Oceanic and Atmospheric Administration (NOAA) personnel in collaboration with the Scripps Institution of Oceanography. The South Pole record was obtained principally from flasks collected twice monthly by NOAA personnel and sent back to Scripps for analysis.

The secular CO_2 increase observed at Mauna Loa and the South Pole is not the same from year to year, but exhibits small irregular oscillations with periods of approximately four years. We have found that these oscillations, expressed as departures from the trend which would produce a constant airborne fraction, correlate with a large-scale meteorological and hydrological fluctuation called the Southern Oscillation (Bacastow, 1976). Their time derivatives correlate even more strongly. The Southern Oscillation involves winds, ocean currents, sea surface temperatures, and barometric pressures. The mechanism by which it influences atmospheric CO_2 might be variation in wind intensity or variation in tropical sea surface temperature (Bacastow, 1976; Machta, Hansen, and Keeling, 1977; Newell and Weare, 1977).

The Mauna Loa and South Pole records, in addition to showing a four-year periodicity, exhibit a longer-term slowdown in secular increase from 1963 to 1969 (Fig. 5). The slowdown is more obvious for Mauna Loa but not altogether lacking for the South Pole. The six-year interval corresponds to a period of reduced atmospheric transmittance of solar radiation attributed to dust thrown into the stratosphere by the eruption of Mt. Agung in March 1963 (Ellis and Pueschell, 1971). The dust cloud spread all over the world, but the turbidity effect was greater over the southern hemisphere than over the northern hemisphere (Volz, 1970a and b). Widespread reduced transmittance might be expected to cause reduced sea surface temperature. Data analyzed by Namias (1973) for the Pacific Ocean north of 20°N indeed indicates an average sea surface temperature reduction of about 0.5°C from 1964 to 1967. Newell and Weare (1976) find evidence for a reduction in tropical tropospheric temperature of about 0.5°C between 1964 and 1967, after the variance correlating with the Southern Oscillation is removed from the temperature data. Using the six-reservoir model of the CO_2

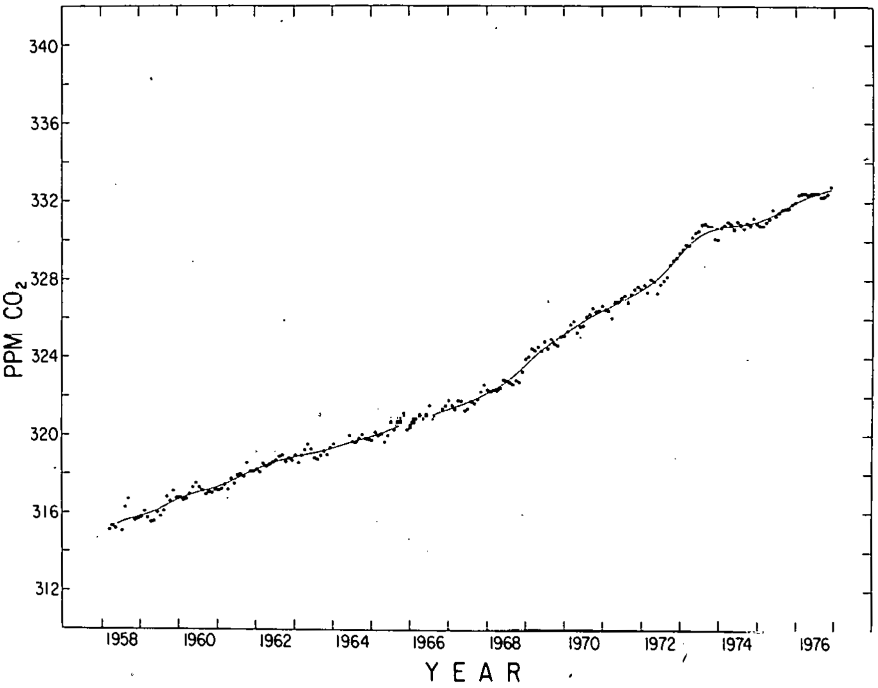

Fig. 2. Seasonally adjusted concentration of atmospheric CO_2 at Mauna Loa Observatory, Hawaii. Dots indicate seasonally adjusted monthly averages. The smooth curve is the spline function used to derive the curve in Fig. 1.

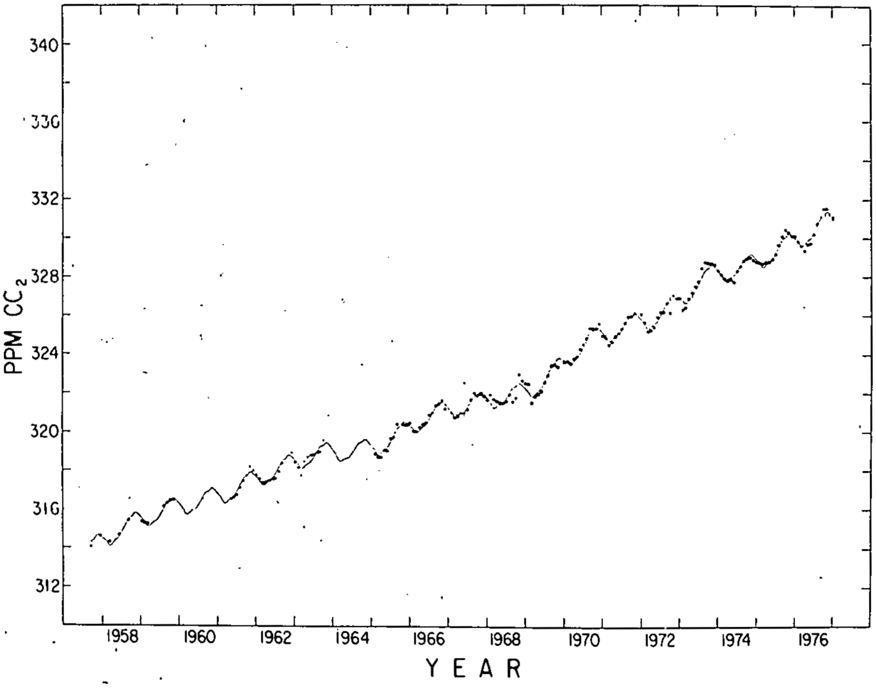

Fig. 3. Concentration of atmospheric CO_2 at the South Pole. The dots indicate monthly averages based on flask analyses, and the crosses (between 1961 through 1963) indicate monthly averages based on continuous measurement. The curve is a fit to the monthly averages of the sum of the average annual variation and a spline function representation of the seasonally adjusted trend as in Fig. 1.

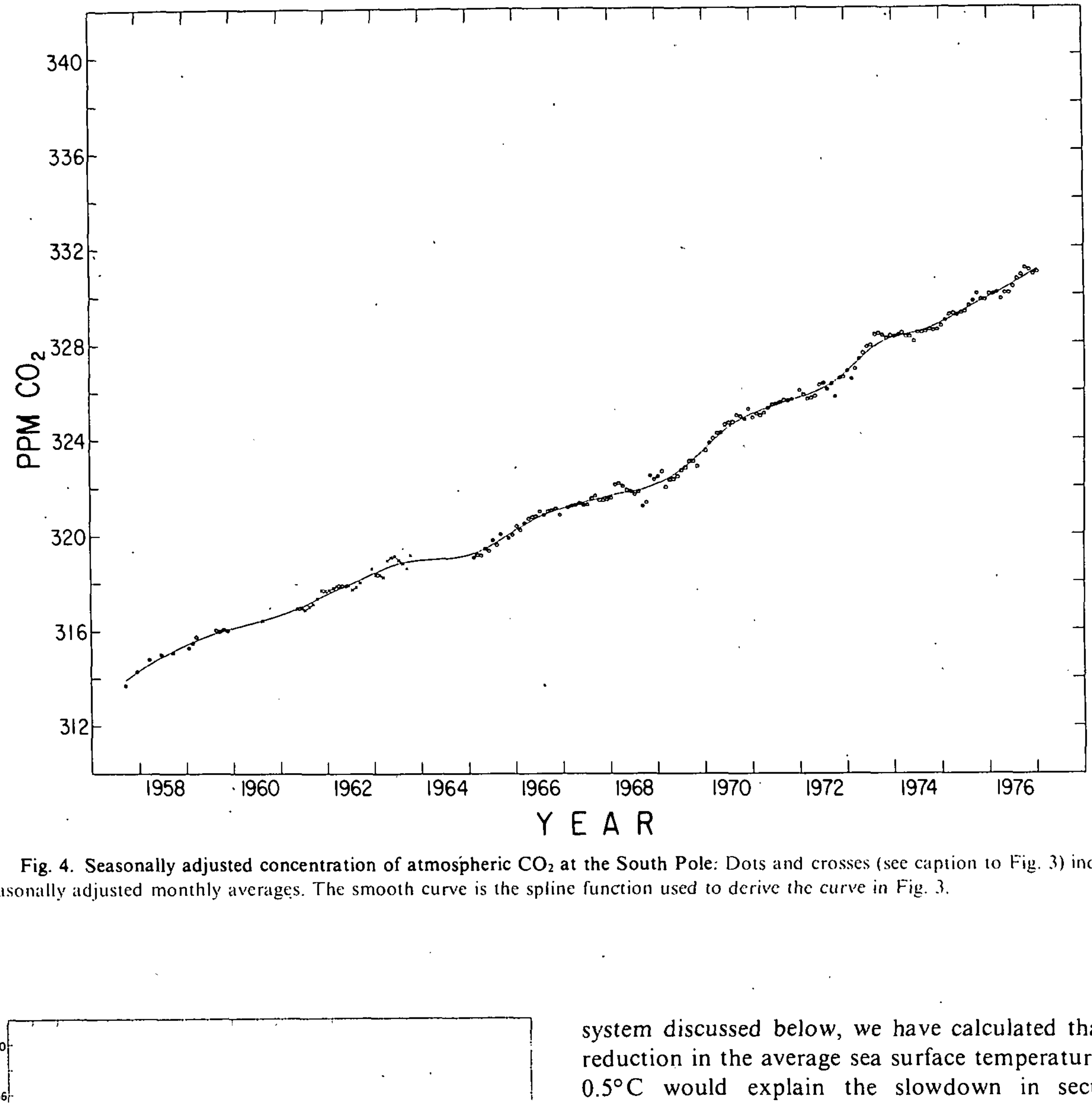

Fig. 4. Seasonally adjusted concentration of atmospheric CO_2 at the South Pole: Dots and crosses (see caption to Fig. 3) indicate seasonally adjusted monthly averages. The smooth curve is the spline function used to derive the curve in Fig. 3.

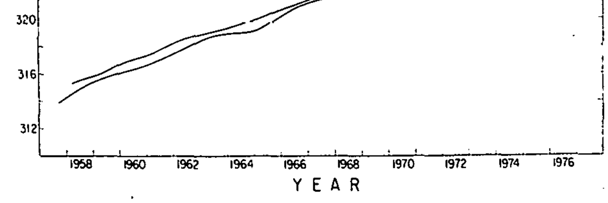

Fig. 5. Comparison of seasonally adjusted trends in atmospheric CO_2 for Mauna Loa Observatory, Hawaii, and the South Pole.

system discussed below, we have calculated that a reduction in the average sea surface temperature of 0.5°C would explain the slowdown in secular increase at Mauna Loa (Fig. 6). It is not obvious, however, why the reduction in secular increase should be less evident in the South Pole data than in the Mauna Loa data, when the Mt. Agung effect was presumably larger in the southern hemisphere than in the northern hemisphere.

Charles A. S. Hall (personal communication) has suggested that the reduced secular increase might be caused by enhanced uptake of CO_2 by the land biota. The land biota carbon pool in the northern hemisphere is far larger than in the southern hemisphere. Diffuse solar radiation, which may have increased after the eruption, is more effective for photosynthesis than direct solar radiation even though the diffuse radiation is reduced in quantity

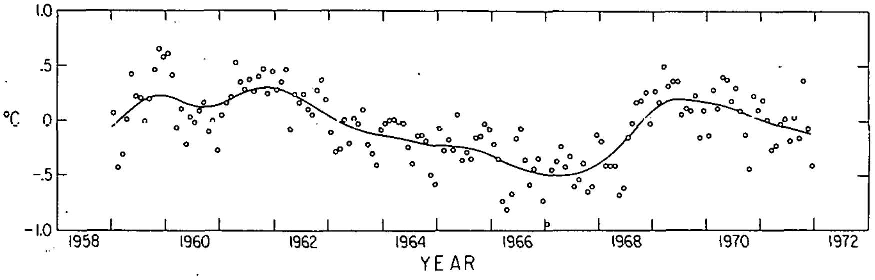

Fig. 6. Sea surface temperature anomaly required to reproduce the observed trend at Mauna Loa, assuming that no net temperature change occurred between 1959 and 1969. The trend was obtained from a 12-month moving average of the atmospheric CO_2 data. Scatter of the points occurs because the moving average was not smoothed. The curve is a spline fit to the points.

relative to the direct radiation (Leon H. Allen, Jr., in comments to Ekdahl and Keeling, 1973.)

Since the yearly secular increase in atmospheric CO_2 has varied considerably, so has the yearly airborne fraction. The models of the CO_2 cycle which we have used for long-range predictions do not attempt, however, to explain short-term phenomena such as the Southern Oscillation. Therefore, to adjust these models to observed data, we have used an average airborne fraction. The effects of the Southern Oscillation and other short-term phenomena have been eliminated, in so far as possible, by averaging between corresponding phases of the observed quasi-regular oscillation over as long a section of record as possible. Recent changes in the yearly increase in fossil fuel production from the mean value of 4.5% have been too small to have any noticeable effect on the airborne fraction.

The long-term airborne fraction from 1959 to 1974 at Mauna Loa Observatory is 61%. The corresponding long-term airborne fraction at the South Pole is 52%. The average of the two stations is 56.5%. These fractions are uncertain, owing to systematic errors in the fossil fuel consumption data of the order of 15%. Uncertainty in the CO_2 measurements contributes a smaller additional error.

The differences between the long-term airborne fractions calculated at Mauna Loa and the South Pole are probably related to the exchange of air between the northern and southern hemispheres. If we assume a reservoir model in which the atmospheric reservoir is divided at the equator, and we assume that the sinks of industrial CO_2 are uniformly distributed between the hemispheres, the resulting atmospheric exchange time between the hemispheres is computed to be 2.1 years. This calculation takes into account that the industrial CO_2 source is 96% in the northern hemisphere, as indicated by data on fuel consumption (United Nations, 1975), and employs the recent time constant for exponential growth in fossil fuel production of 22 years. A time for transport of two years between the hemispheres is rather long in comparison to 320 days estimated by Newell, Vincent, and Kidson (1969) from meteorological observations. This disparity suggests that the sink in the southern hemisphere is larger than that in the northern hemisphere, since such a difference in sinks would be consistent with a longer computed exchange time.

Even if the hemispheric exchange time is as long as two years, it is short compared to the response times of interest in predicting global atmospheric CO_2 variations. Thus, we introduce little error in modeling global variations if we neglect this feature and assume that the atmosphere is homogeneous with respect to atmospheric CO_2.

Continued exponential growth in fossil fuel input at the 1945 to 1975 average rate of 4.53% per year and maintenance of an airborne fraction of 56.5% (based on the average of Mauna Loa and South Pole data) would lead to a doubling of the preindustrial CO_2 level (assumed to be 290 ppm) in the year 2027, possibly with an attendant atmospheric temperature increase of as much as 2.5°C, as predicted by the three-dimensional model of Manabe and Wetherald (1975 and this workshop). Observable climate effects owing to CO_2 buildup might occur sooner. Indeed, a world temperature increase of 1°C is substantial and might be expected to occur as early as the year 2000, when the CO_2 increase above the preindustrial value is predicted to be about 33%.

FACTORS THAT WILL CAUSE THE AIRBORNE FRACTION TO CHANGE

The global average airborne fraction will probably not remain near 56% in the future for several reasons. First, because fossil fuel resources are finite, exponential growth in fossil fuel consumption cannot long continue. Well before fossil fuel resources approach exhaustion, fuel production will reach a maximum, after which increasing difficulties in recovering these diminishing resources will force a decline. To simulate future industrial CO_2 production over the entire remainder of the fossil fuel era, we have used a modified logistics function:

$$\frac{dN}{dt} = r\,N\left[1 - \left\{\frac{N}{N_\infty}\right\}^n\right] , \qquad (1)$$

where N denotes the total CO_2 input (in grams of carbon) up to time t years, N_∞ denotes the ultimate total CO_2 input, r is a constant growth factor, and n is an arbitrary parameter that permits variability in the combustion pattern. The initial value of N and the value of r are determined from recent data. If $n = 1$, Eq. 1 is the logistics function; n less than 1 results in a slower rate of production than the logistics function; and n greater than 1 results in a faster rate. With $n = 0.5$ and $N_\infty = 8.2$ times the preindustrial amount of CO_2 in the atmosphere, an input function is obtained that closely approximates a recent "best" estimate of Perry and Landsberg (1977). For the period before 1976, however, we use actual historic data from a data table (Keeling, 1973b; Rotty, 1973). The method by which we effected a smooth transition from this data table to Eq. 1 is indicated by Keeling and Bacastow (1977).

Second, if continued CO_2 production occurs, the perturbations in the global carbon system will become large, and interactions will probably become more and more nonlinear. This is especially likely in the surface layer of the ocean. Carbon dioxide reacts there with the carbonate ion in sea water. Only about 10% of the ocean's dissolved carbon is in the form of ionic carbonate, and as this species becomes depleted by reaction with industrial CO_2, the CO_2 partial pressure exerted by the water will increase so as to oppose further uptake, that is, the so-called buffer factor will increase (Bacastow and Keeling, 1973). The buffer factor will not increase as rapidly if dissolution of carbonate sediments occurs in deep water and adds to the supply of carbonate ion, and it will remain almost constant if surface sediment dissolution occurs so rapidly as to produce chemical equilibrium between surface sea water and a carbonate phase. The latter case would result in an almost linear response, but such rapid dissolution is unlikely, as discussed below.

The response of the land biota to CO_2 is also likely to become increasingly nonlinear. The principal storage of carbon in the biota is in forests, particularly in the tropics and northern hemisphere temperate and boreal regions. Carbon storage may now be increasing, as a result of CO_2 fertilization (brought about by the increase in CO_2 concentration in the atmosphere), but it is unlikely to increase without limit. Demands of an increasing human population for food and shelter are likely to force continued harvesting of the forests and maximum conversion of lands to agricultural use, causing less storage of carbon in the affected soil and vegetative cover.

We thus believe it unlikely that undisturbed lands or lands specifically managed to remove atmospheric CO_2 will be able to produce a large increase in the total land biota carbon pool. Consequently, for most of our calculations of future atmospheric CO_2 levels we prescribe in our models that the biota carbon reservoir ceases to expand after the year 2025. To avoid an abrupt change, we linearly extinguish a biota growth factor (for definition see Eq. A.17, Appendix A below) in our model beginning in the year 2000. The alternative of a large biota carbon increase is considered in some calculations, however.

PREDICTIONS OF THE SIX-RESERVOIR MODEL

We have investigated the possible future increase in atmospheric CO_2 using a six-reservoir model which subdivides the atmosphere, oceans, and land biota each into two reservoirs of the box model type (Bacastow and Keeling, 1973; Keeling and Bacastow, 1977; and Fig. 7). Subdivision in all three cases was made to satisfy salient aspects of the observational data. Further subdivision did not seem warranted on the basis of available data. The six reservoirs are: the surface ocean layer (or "mixed layer"), deep ocean (or "deep sea"), troposphere, stratosphere, short-cycled biota, and long-cycled biota. The stratosphere, with an assumed exchange time of two years, and the short-cycled biota, with an exchange time of 2.5 years, were found to have little effect on the model predictions of industrial CO_2 dispersal. Thus, except when the time-dependent radiocarbon cycle is considered, the model may be reduced to four reservoirs with only the oceans subdivided.

Since a reservoir model avoids explicit expression of dynamic transport mechanisms, the assigning of

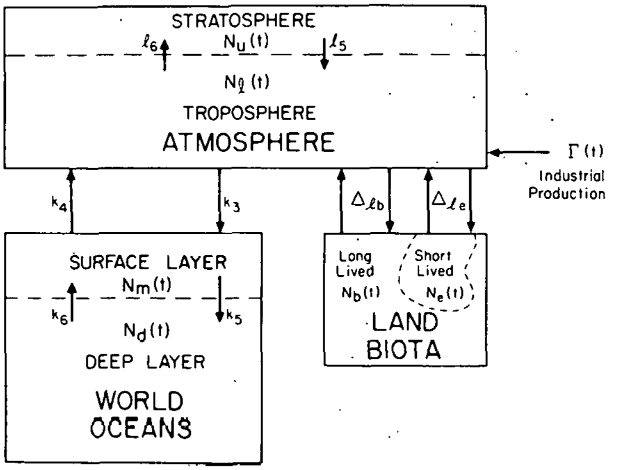

Fig. 7. Six-reservoir model of the carbon cycle (Bacastow and Keeling, 1973). The mass of carbon in each reservoir is represented by $N_j(t)$, and the transfer coefficients between reservoirs are given by ℓ_j and k_j. The magnitudes of the initial values of $N_j(t)$ and of the values of k_j and ℓ_j are listed in Appendix A in Table A.1, and $\Delta\ell b$ and $\Delta\ell e$ are defined in Appendix A, Eqs. A.45 and A.46.

boundaries such as the water depth separating surface and subsurface ocean water is left up to the modelist. As a basis for choosing this depth, we have considered the distributions of a number of tracers, but especially temperature and radiocarbon. Over much of the ocean a thin, wind-mixed layer of rapidly circulating relatively warm water is underlain by a transition layer, beneath which lies a deep layer of slowly circulating colder water. The depth of the wind-mixed layer varies considerably (mainly as a function of latitude), with an average of about 75 m. The transition zone or "main thermocline," below this layer, extends to 800 or 1000 m, or almost one-fourth of the distance to the average oceanic depth of about 3800 m.

If the oceans are to be divided into two layers for purposes of modeling, it is more reasonable to set the boundary within the transition zone than at its top as done in previous models. We have chosen the boundary by requiring that the model predictions for radiocarbon agree with the Suess effect (see Keeling and Bacastow, 1977). The Suess effect is defined as the reduction in the radiocarbon-to-total-carbon ratio in the atmosphere resulting from addition of radiocarbon-free fossil fuel CO_2 to the atmosphere. The effect has been measured in tree rings that grew before about 1954. After that year radiocarbon from atomic bomb tests complicates the interpretation. The Suess effect is uncertain, to the extent that the model boundary calculated from it may lie anywhere from about 120 to 700 m. The boundary clearly is found, however, to lie in the transition zone.

The biota growth factor has been adjusted so that the predicted atmospheric CO_2 increase agrees with the observed increase from 1959 to 1974. In this way, the observed Suess effect and the observed atmospheric increase, even though for different time periods, are used to apportion fossil fuel uptake between the oceans and the biota.

The atmospheric CO_2 concentration resulting from any of the four input curves given in Fig. 8 rises to a peak value between the years 2100 and 2300 and then falls slowly (Fig. 9). As might be expected, the

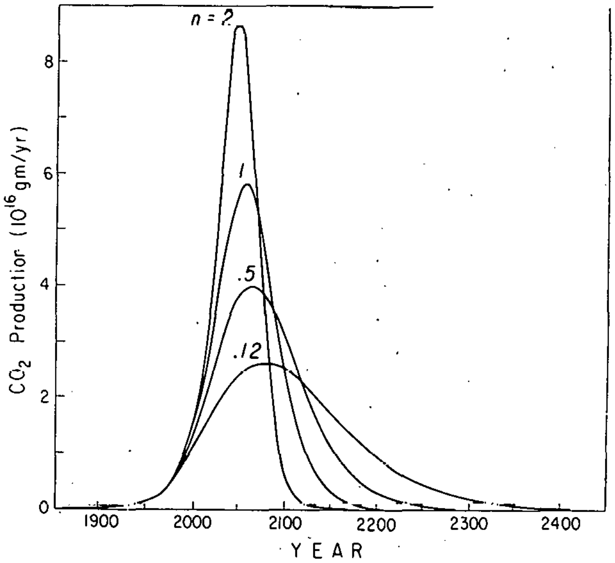

Fig. 8. Industrial CO_2 production for various assumed patterns of fossil fuel consumption. The ultimate production of industrial CO_2 is fixed at 8.2 times the amount of CO_2 in the preindustrial atmosphere. The parameter n is defined by Eq. 1 of the text.

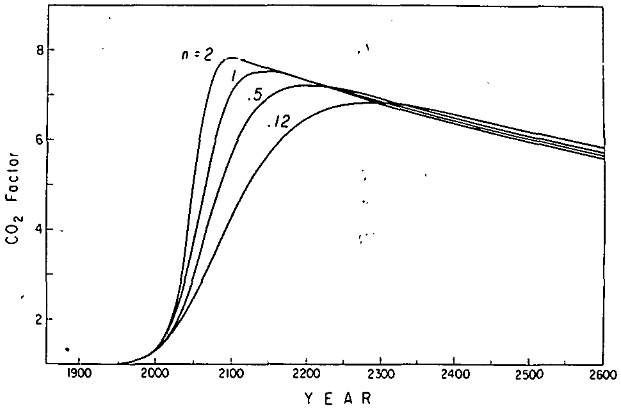

Fig. 9. Predicted increase in atmospheric CO_2. Curves are shown for the four simulated fossil fuel consumption patterns shown in Fig. 8. The CO_2 factor is the ratio of the amount of CO_2 in the atmosphere to the amount in the preindustrial atmosphere.

peak level is delayed if the period of combustion is stretched out. The peak value, however, is only slightly reduced by slower production of industrial CO_2. For example, the two extreme input curves ($n = 0.12$ and 2.0) differ in maximum production rate by 327%, but the peaks of the atmospheric response differ by only .15%. Both of these inputs affect the model like comparatively rapid impulses because of the long tracer exchange time associated with the subsurface ocean. The slow drop in atmospheric CO_2 level after peaking is also due to this long oceanic exchange time.

Basic Model with Varied Biota Uptake

If the biota in our six-reservoir model is allowed to grow without limit under the assumed influence of CO_2 fertilization, the biota carbon pool is predicted almost to triple in size and to reduce atmospheric CO_2 to approximately the preindustrial level within 300 years (lower curve in Fig. 10). As stated earlier, we do not expect this to occur, and we have provided a cutoff of biota growth in the model. Our chosen cutoff is arbitrary, however, so we have considered a range in linear cutoff intervals all beginnning in the year 2000 (Fig. 10). In the standard case, in which the biota growth factor β is reduced to zero in 2025, the biota carbon pool increases relative to its preindustrial size by 6%. In contrast, a final cutoff to zero in the year 2150 leads to a biota increase of 42%, in the year 2250 to 98%, and in the year 2350 to 165%. Provided that the biota carbon pool no more than doubles, the atmospheric curves are similar to the standard case, except lowered by the amount the biota takes up.

Basic Model plus Warming Surface Water

Warming of the atmosphere caused by increased CO_2 would be expected to cause a warming of the surface layer of the oceans. This warming in turn would cause the CO_2 partial pressure in the water to rise and release more CO_2 to the atmosphere. We have calculated the magnitude of this feedback process using our six-reservoir model. We assume that the average temperature increase of the ocean surface layer at time t is:

$$\Delta T = \frac{2.5}{\ln 2} \ln \left(\frac{P}{P_0}\right) , \qquad (2)$$

where P is the atmospheric CO_2 partial pressure (or concentration) at time t, P_0 is the preindustrial CO_2 partial pressure, and the constant 2.5 is the Manabe-Wetherald (1975) estimate of the increase in average temperature for a doubling of the CO_2 concentration. The temperature feedback effect is found to be insignificant compared to uncertainties in the prediction (Fig. 11).

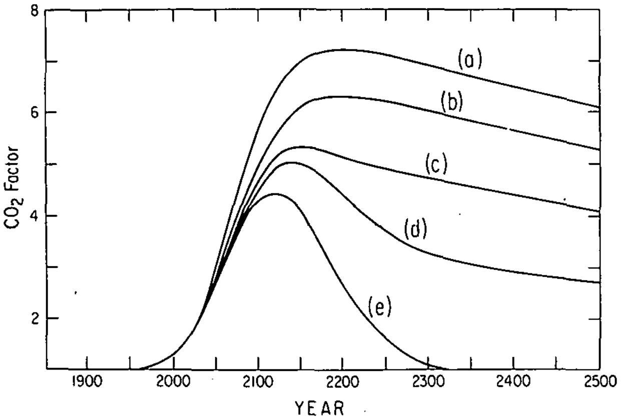

Fig. 10. Effect of biota growth on the predicted increase in atmospheric CO_2: (a) Standard case, same as the $n = 0.5$ curve in Fig. 9. The biota growth factor was set linearly to zero between the years 2000 and 2025. The ultimate biota increase is 6% relative to its preindustrial value. (b) Biota growth factor set linearly to zero between 2000 and 2150. Ultimate biota increase is 42%. (c) Biota growth factor set linearly to zero between 2000 and 2250. Ultimate biota increse is 98%. (d) Biota growth factor set linearly to zero between 2000 and 2350. Ultimate biota increase is 165%. (e) Biota grows without a cutoff. The biota carbon pool continues to increase until all of the fossil fuel input is in the biota. Ultimate biota increase is 308%.

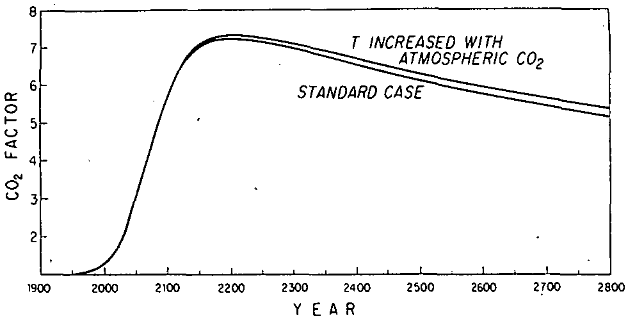

Fig. 11. Predicted increase in atmospheric CO_2 if the temperature T of the surface ocean is assumed to increase with increasing atmospheric CO_2 concentration according to Eq. 2 of the text. This case is compared to the standard case, shown as the curve with $n = 0.5$ in Fig. 9. The CO_2 factor has the same meaning as for Fig. 9.

Implicit in this calculation is the assumption that the circulation times of the natural carbon dioxide cycle, such as the emission of CO_2 in equatorial waters and absorption at higher latitudes, would be unchanged by an atmospheric temperature increase. How the natural cycle would change if the temperature of the earth increased is not well enough known to be reliably modeled at this time.

MODEL WITH A DIFFUSIVE DEEP SEA

Recently Oeschger et al. (1975) have predicted industrial CO_2 uptake with a "box-diffusion" model similar to our six-reservoir model except that they represent the ocean by a wind-mixed layer 75 m thick, coupled with a deep ocean in which transport is by vertical eddy diffusion. The vertical diffusivity coefficient, K of their model, is set to be consistent with radiocarbon data believed to represent steady-state conditions, as is the exchange time for deep ocean in our model. An advantage of the diffusive model is that the predicted CO_2 uptake is rather insensitive to the assumed depth of the surface layer. Also, oceanic tracer concentrations, such as that of radiocarbon, vary more or less smoothly in the vertical and, consequently, the box-diffusion model seems physically more plausible than a box model. The model, however, is still a parameterization because, in reality, horizontal motions and vertical advection taken together are probably as important to CO_2 transport as vertical diffusion.

Oeschger et al. (1975) claim that CO_2 uptake by their model agrees with observed data, while uptake by an ocean model consisting of two reservoirs does not. However, they have compared the two models with equal 75-m mixed layers. This comparison is unfair because, if the ocean is to be modeled with two reservoirs, it is physically more plausible, as noted earlier, to set the level separating the reservoirs somewhere near the center rather than at the top of the main thermocline.

The box-diffusion model and our ocean model, when adjusted to fit the same steady-state radiocarbon data and the atmospheric CO_2 increase from 1959 to 1974, give similar predictions over the next 1500 years (Fig. 12). The peak atmospheric CO_2 level is about 7% lower for the box-diffusion model, which is hardly significant in view of other uncertainties in prediction.

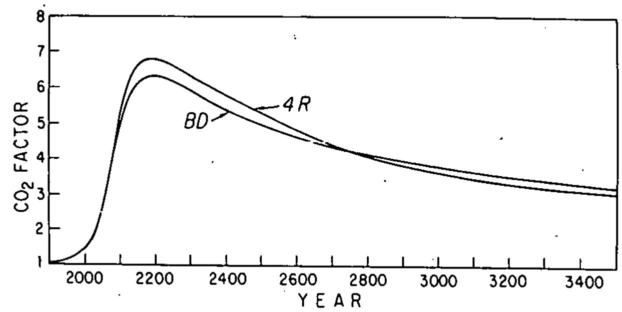

Fig. 12. Comparison of predictions of the box-diffusion model (BD) and the four-box (4R) model as described in the text. The CO_2 factor has the same meaning as for Fig. 9.

In this comparison the box-diffusion model is as described by Oeschger et al. (1975), except that their representation of the long-cycled biota parameterization was modified slightly to agree with that used by Bacastow and Keeling (1973). Also, the stratosphere and the short-cycled biota reservoir were omitted from both models. (Our model was thus reduced to a four-reservoir model.) The long-cycled biota carbon pool was assumed not to increase in size after 2025 so that the comparison between the models would emphasize differences in oceanic parameterization. The deep ocean exchange time of the four-reservoir model was set at 1208 years to agree with the vertical average of the radiocarbon profile predicted by the box-diffusion model with $K = 3987\ m^2/year$, whereas we used 1500 years in our other calculations. In the box-diffusion model the deep ocean was divided into 42 boxes following the approximation to the diffusion equation used by Oeschger et al. (1975). We found the resulting model equations somewhat "stiff" (Gear, 1971), and consequently many very small steps were required for integration by stepwise numerical methods.* The ocean surface layer depth of the four-reservoir model was adjusted to a value such that the atmospheric CO_2 increase between 1959 and 1974 would agree with the observed increase. This value, 284 m, is only slightly larger than the approximately 264 m used as a preferred depth in our six-reservoir model. Keeling and Bacastow (1977) further discuss the comparison of the models.

SIX-RESERVOIR MODEL WITH CARBONATE DISSOLUTION

In all of the cases considered above, an absence of carbon exchange between sea water and solid carbonates has been assumed. If dissolution of

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*A set of initial value differential equations is said to be stiff if their solution is approximately a sum of exponentials with time constants which differ greatly in magnitude, and the interesting behavior is dominated by the longer time constants. Unless the step size in time in the numerical solution is made small compared to the smallest time constant, large errors or instability are to be expected. Division into more boxes would result in even stiffer equations.

marine $CaCO_3$ deposits occurs, one would expect an increase in CO_2 uptake by the oceans. Although the surface ocean is believed to be several-fold supersaturated with respect to calcite and aragonite, more soluble high-magnesium carbonates may exist, as noted by R. M. Garrels at this workshop. In the deep ocean, dissolution of carbonates will almost certainly occur below the carbonate saturation depth, where sea water is undersaturated with respect to calcium carbonate. Deep ocean carbonate deposits lying below the saturation depth, however, are isolated from the atmosphere by the slow exchange time of the deep ocean with the surface layer. Dissolution will be further slowed by diffusive processes associated with transport in the sediment.

As a first approximation to calculating carbonate dissolution, we have modified our six-reservoir model to include chemical equilibrium between sea water and calcium carbonate in the deep ocean and, in some calculations, in the surface layer as well. Thus we assume the limit of very fast dissolution and may exaggerate the expected effect.

If only deep water dissolution occurs, the atmospheric CO_2 level for several centuries is only slightly lower than if no dissolution occurs (Fig. 13). After 1500 years, however, the CO_2 level will have fallen to 2.6 times the preindustrial level instead of 3.9 times as otherwise predicted. Calcium carbonate will have been removed from sediments on the ocean floor to an average depth corresponding to 3 cm of pure calcium carbonate.

If equilibrium between sea water and calcium carbonate, brought about by dissolution, is assumed to have occurred in the shallow water of the ocean surface layer (as well as in deep water) since the beginning of the fossil fuel era, the atmospheric CO_2 increase is noticeably lower and its rate of fall is faster. After 1500 years, the CO_2 level will be only 1.7 times the preindustrial level. Calcium carbonate will have been dissolved from shallow water sediments to an average depth corresponding to 40 cm of pure calcium carbonate. Since far less than half of the area of shallow seas contain carbonate-rich sediment, several meters of such sediment on average will have been stripped of carbonate. In this case, the model almost surely overemphasizes dissolution, since if shallow water dissolution occurs, the penetration of industrial CO_2 into the sediment layer itself will be the rate-determining process. Thus the prediction shown as the lower curve in Fig. 13 seems likely to exaggerate dissolution even if undersaturation actually exists in the surface water.

The calculations of calcium carbonate equilibrium dissolution were made by assuming invariance of a special alkalinity, A′, defined so as to be uninfluenced by dissolution of $CaCO_3$ as well as addition of CO_2 (compare Bolin and Eriksson, 1959; Plass, 1972). That is, we let

$$A' = A - 2([Ca^{2+}] - [Ca^{2+}]_{m0}) \quad , \tag{3}$$

where A denotes the combined carbonate and borate alkalinity and $[Ca^{2+}]$ denotes the calcium concentration for a given reservoir. The preindustrial calcium concentration in the ocean surface layer, $[Ca^{2+}]_{m0}$, has been included, as a third term, for convenience. The

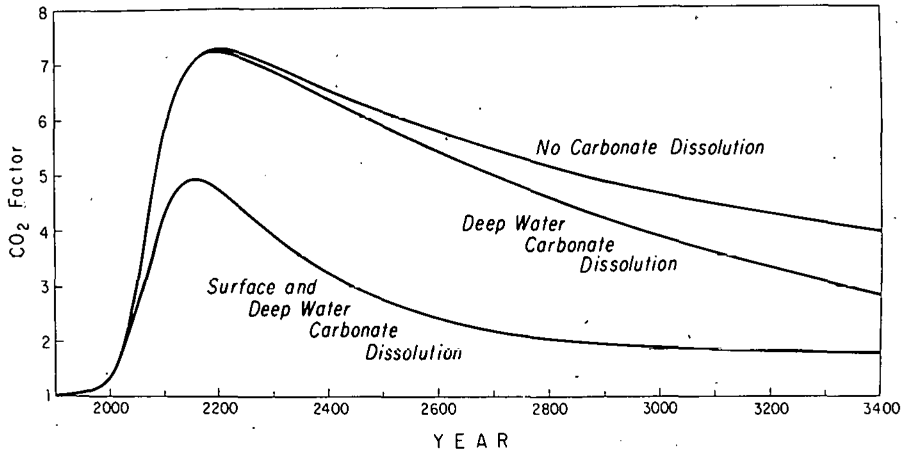

Fig. 13. Predicted increase of atmospheric CO_2 with and without dissolution of carbonate sediments. The pattern of fossil fuel consumption is as shown by the case n = 0.5 of Fig. 8. The CO_2 factor has the same meaning as for Fig. 9.

chemical parameters in the model were evaluated at each numerical step of the differential equations. The perturbation in transport of carbonate alkalinity between reservoirs is prescribed by exchange equations similar to those for total dissolved CO_2 (see Eqs. A.30 and A.31, Appendix A). The model assumes that initially the special alkalinity is the same in both the ocean surface layer and the deep ocean. Since this quantity remains constant, the model assumes that no transport of special alkalinity occurs between the oceanic reservoirs. This is equivalent to assuming that the flux of carbonate by gravitational setting of particulate matter from the ocean surface layer is unaffected by industrial CO_2.

CRITICAL ASSUMPTIONS

Since the climatic effects resulting from high atmospheric CO_2 levels may impact human institutions for several thousand years, we should ask how firm are the observations and assumptions critical to the predictions of the model, especially the sharp rise and slow fall in atmospheric CO_2 concentration in response to the various CO_2 input patterns. The most important oceanic feature on which this prediction depends is not a matter of dispute, however: much of the oceans' water is known from radiocarbon dating to be old and consequently to exchange CO_2 with the atmosphere only slowly.

A more controversial assumption is that the biota carbon pool will not increase in response to CO_2 fertilization by so much that the predicted atmospheric CO_2 increase falls relatively rapidly after a peak value is reached. If we leave out of consideration very long cyclic soil humus and detrital matter with steady-state cycling times predominately greater than 1000 years, we estimate that the biota carbon reservoir contains today no more than about $2^1/_2$ times the amount of carbon stored in the atmosphere. Even if the long-cycled and short-cycled biota were both to double in mass, high atmospheric CO_2 levels for thousands of years would still be predicted by all the models we have considered.

Conceivably, a considerable amount of carbon could be taken up by the very long-cycled carbon pool during the few centuries which we ascribe to the fossil fuel era, in spite of this pool having a steady-state turnover time of the order of thousands of years. But until more is known about the response of soils and deep humus to high inputs of detrital carbon, this possibility seems to us too speculative to incorporate into a global model of industrial CO_2 uptake.

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APPENDIX A

Here we list explicitly the equations and parameters of the six-reservoir model as an aid to the discussion in the main text. Except for changes related to calcium carbonate dissolution, the six-reservoir model is as described in *Carbon and the biosphere* (Bacastow and Keeling, 1973).

Carbon System – Six-Reservoir Model

Material balance equations. Equations A.1 through A.6 express material balances for each of the reservoirs (Fig. 7).

Upper atmosphere:

$$dN_u/dt = k_{\ell u}N_\ell - k_{u\ell}N_u \, . \tag{A.1}$$

Long-cycled land biota:

$$dN_b/dt = F_{\ell b} - F_{b\ell} \, . \tag{A.2}$$

Short-cycled land biota:

$$dN_e/dt = F_{\ell e} - F_{e\ell} \, . \tag{A.3}$$

Lower atmosphere:

$$\begin{aligned} dN_\ell/dt = {} & k_{u\ell}N_u - k_{\ell u}N_\ell + F_{e\ell} - F_{\ell e} + F_{b\ell} - F_{\ell b} \\ & + k'_{ma}P_m - k_{am}\,(N_{a0}/N_{\ell 0})\,N_\ell + \gamma_\ell \, . \end{aligned} \tag{A.4}$$

Ocean surface layer:

$$\begin{aligned} dN_m/dt = {} & k_{am}\,(N_{a0}/N_{\ell 0})\,N_\ell - k'_{ma}P_m + k_{dm}N_d \\ & - k_{md}N_m - F_g + \gamma_m \, . \end{aligned} \tag{A.5}$$

Deep ocean:

$$dN_d/dt = k_{md}N_m + F_g - k_{dm}N_d + \gamma_d \, . \tag{A.6}$$

Water conservation leads to a relationship between the rate constants k_{md} and k_{dm}:

$$0 = k_{md}W_m - k_{dm}W_d \, , \tag{A.7}$$

where

W_m, W_d = volumes (assumed to be time invariant) of the ocean surface layer and deep ocean. (A.8)

In these equations we use the following symbols:

Reservoir exchange coefficients

$k_{u\ell}$ = upper-atmosphere to lower-atmosphere transfer coefficient, (A.9)

$k_{\ell u}$ = lower-atmosphere to upper-atmosphere transfer coefficient [the preindustrial equation leads to $k_{\ell u} = k_{u\ell}\,(N_{u0}/N_{\ell 0})$], (A.10)

k_{am} = air to sea transfer coefficient [defined in terms of the carbon mass of the total atmosphere; e.g., the preindustrial (steady-state) flux is $k_{am}\,N_{a0}$], (A.11)

k'_{ma} = sea to air transfer coefficient (the preindustrial flux $k'_{ma}P_{m0}$ is numerically equal to $k_{am}N_{a0}$), (A.12)

k_{md}, k_{dm} = surface layer to deep ocean and deep ocean to surface layer transfer coefficients, respectively. (A.13)

(All transfer coefficients, k_{ij}, are assumed to be time invariant, except k'_{ma}.)

Reservoir masses

$N_u, N_b, N_e,$
N_ℓ, N_m, N_d = mass of carbon in each designated reservoir, (A.14)

$N_a = N_u + N_\ell$ = mass of carbon in the atmosphere as a whole, (A.15)

$N_{u0}, N_{b0}, N_{e0},$
$N_{\ell 0}, N_{m0}, N_{d0},$
N_{a0} = masses of carbon in the designated reservoir before the industrial era. (A.16)

Reservoir fluxes

$F_{\ell b} = F_{b0}[1 + \beta \ln(N_\ell/N_{\ell 0})]\,(N_b/N_{b0})$ = flux from lower atmosphere to long-cycled land biota, (A.17)

F_{b0} = flux between lower atmosphere and long-cycled land biota before industrial era in either direction, (A.18)

β = land-biota growth factor, (A.19)

$F_{b\ell} = F_{b0}(N_b/N_{b0})$ = flux from long-cycled land biota to lower atmosphere, (A.20)

$F_{\ell e} = F_{e0}[1 + \beta \ln(N_\ell/N_{\ell 0})]\,(N_b/N_{b0})$ = flux from lower atmosphere to short-cycled land biota (taken as proportional to N_b/N_{b0} rather than N_e/N_{e0}), (A.21)

F_{e0} = flux between lower atmosphere and short-cycled land biota before the industrial era in either direction, (A.22)

$F_{e\ell} = F_{e0}\,(N_e/N_{e0})$ = flux from short-cycled land biota to lower atmosphere, (A.23)

γ_ℓ = production rate of carbon in the atmosphere from industrial sources, (A.24)

γ_m = production rate of carbon in mixed ocean surface layer from dissolution of $CaCO_3$, (A.25)

γ_d = production rate of carbon in deep ocean from dissolution of $CaCO_3$, (A.26)

$F_g = F_{g0}$ = gravitational flux of inactive carbon from ocean surface layer to deep ocean (assumed to be time invariant). (A.27)

Reservoir pressures

P_m = CO_2 partial pressure in ocean surface layer, (A.28)

P_{m0} = value of P_m before the industrial era. (A.29)

Calcium System

Changes in the calcium content of the ocean surface layer and the deep ocean in this model arise from dissolution of calcium carbonate and the subsequent exchange of seawater between the ocean reservoirs. The transport of calcium corresponds to the transport of alkalinity, and the calcium balance equations could be replaced by similar alkalinity balance equations. The calcium equations are similar to those for the inactive carbon system.

Ocean surface layer:

$$dC_m/dt = -k_{md}C_m + k_{dm}C_d - J_g + \gamma_m/R\,. \tag{A.30}$$

Deep ocean:

$$dC_d/dt = -k_{dm}C_d + k_{md}C_m + J_g + \gamma_d/R\,, \tag{A.31}$$

where:

C_m, C_d = masses of calcium in ocean surface layer and deep ocean, respectively, (A.32)

$J_g = J_{g0}$ = gravitational flux of calcium from ocean surface layer to deep sea (assumed to be time invariant), (A.33)

R = molecular weight of carbon/molecular weight of calcium. (A.34)

Perturbation Equations

For numerical calculations, the material balance equations A.1 through A.6, A.30, and A.31 are reformulated in terms of changes from the assumed preindustrial state. The new variables are:

$$n_i \equiv N_i - N_{i0}\,, \tag{A.35}$$

$$c_i \equiv (C_i - C_{i0})\,R\,. \tag{A.36}$$

The variables n_i and c_i refer to changes in the mass of carbon and calcium, respectively, in reservoir i, where i is u, b, e, ℓ, m, or d. The inclusion of R in Eq. A.36 is equivalent to expressing the mass of calcium in each oceanic reservoir in units of the corresponding amount of carbon, i.e., each mole of calcium is assigned a mass of 12.011 g.

Upper atmosphere:

$$dn_u/dt = -\ell_5 n_u + \ell_6 n_\ell\,. \tag{A.37}$$

Long-cycled biota:

$$dn_b/dt = \Delta_{\ell b}\,. \tag{A.38}$$

Short-cycled biota:

$$dn_e/dt = \Delta_{\ell e}\,. \tag{A.39}$$

Lower atmosphere:

$$dn_\ell/dt = \ell_5 n_u - \ell_6 n_\ell - \Delta_{\ell b} - \Delta_{\ell e} - k_3 n_\ell + k_4 n_m + \gamma_\ell\,. \tag{A.40}$$

Ocean surface layer:

$$dn_m/dt = k_3 n_\ell - k_4 n_m + k_6 n_d - k_5 n_m + \gamma_m\,, \tag{A.41}$$

$$dc_m/dt = k_6 c_d - k_5 c_m + \gamma_m\,. \tag{A.42}$$

Deep ocean:

$$dn_d/dt = -k_6 n_d + k_5 n_m + \gamma_d\,, \tag{A.43}$$

$$dc_d/dt = -k_6 c_d + k_5 c_m + \gamma_d\,. \tag{A.44}$$

New symbols introduced above are:

$$\Delta_{\ell b} = \beta F_{b0}\ [1 + (n_b/N_{b0})]\ \ln[1 + (n_\ell/N_{\ell 0})]\ , \qquad (A.45)$$

$$\Delta_{\ell e} = \beta F_{e0}\ [1 + (n_b/N_{b0})]\ \ln[1 + (n_\ell/N_{\ell 0})] + F_{e0}\,[(n_b/N_{b0}) - (n_e/N_{e0})]\ , \qquad (A.46)$$

$$\ell_5 = k_{u\ell}\ , \qquad (A.47)$$

$$\ell_6 = k_{u\ell} N_{u0}/N_{\ell 0}\ , \qquad (A.48)$$

$$k_3 = k_{am} N_{a0}/N_{\ell 0}\ , \qquad (A.49)$$

$$k_4 = k_{am}(N_{a0}/N_{\tilde{m}0})\xi\ , \qquad (A.50)$$

$$k_5 = (W_d/W_m)k_{dm}\ , \qquad (A.51)$$

$$k_6 = k_{dm}\ , \qquad (A.52)$$

where ξ is the CO_2 evasion factor defined by Eq. A.83 and $N_{\tilde{m}0}$ is the inorganic carbon in the ocean surface layer (see Table A.1).

CaCO₃ Saturation

If the ocean surface layer or deep ocean is exactly saturated with respect to $CaCO_3$, then the calcium concentration in that reservoir is determined by the carbon concentration. For this case we introduce a new parameter, θ, which is evaluated from seawater parameters and equilibrium constants:

$$\theta_i \equiv dc_i/dn_i = d[Ca^{2+}]_i/d[\Sigma C]_i\ , \qquad (A.53)$$

where i is m or d, and

$[Ca^{2+}]_i$ = calcium concentration in reservoir i in gram atoms per liter, (A.54)

$[\Sigma C]_i$ = total inorganic carbon concentration in reservoir i in gram atoms of carbon per liter. (A.55)

Consequently,

$$dc_i/dt = \theta_i(dn_i/dt)\ . \qquad (A.56)$$

Because, to a good approximation at saturation,

$$CaCO_3 + CO_2 + H_2O \rightarrow 2HCO_3^- + Ca^{2+}\ , \qquad (A.57)$$

the value of θ is close to (but slightly smaller than) ½. Precise values of θ_i were evaluated numerically as the calculation proceeded.

Calcium carbonate is assumed to dissolve in the ocean surface layer if undersaturation or saturation prevails, but not to precipitate if supersaturation occurs. Consequently, the source term γ_m is prescribed to be zero if the surface layer is supersaturated. The deep ocean is assumed to be in equilibrium with solid calcium carbonate, except in those models that do not include calcium carbonate at all. If an oceanic reservoir is saturated with calcium carbonate, the calcium and carbon balance equations can be combined to eliminate the source term. Calcium and carbon balance equations

Table A.1. Six-reservoir model parameters

Initial values of carbon (10^{18} g)		
N_{u0}	CO_2 in stratosphere	0.092
$N_{\ell 0}$	CO_2 in troposphere	0.523
N_{b0}	carbon in perennial biota	1.56
N_{e0}	carbon in annual biota	0.075
N_{m0}	carbon in ocean surface layer	2.46
$N_{\tilde{m}0}$	inorganic carbon in ocean surface layer	2.37
N_{d0}	inorganic carbon in deep ocean water	36.2
Initial fluxes (10^{18} g carbon/year)		
F_{b0}	preindustrial flux to perennial biota	0.026
F_{e0}	preindustrial flux to annual biota	0.030
Steady-state transfer coefficients (per year)		
ℓ_5	stratosphere to troposphere	1/2.00
ℓ_6	troposphere to stratosphere	1/11.3
k_3	troposphere to surface ocean layer	1/5.92
k_4	surface ocean layer to troposphere	(1/26.86)ξ
k_5	surface ocean to deep ocean	1/113
k_6	deep ocean to surface ocean	1/1500
Additional model quantities		
$I(t)$	industrial CO_2 production *1700 to 1860:* assumed exponential increase at 4.35% per year with 1860 production of 9.53×10^{13} g of carbon (Keeling, 1973b) *1860 to 1969:* based on annual records as cited by Keeling, 1973b *1970 to 1975:* based on annual records as cited by Rotty, 1977 *after 1975:* as predicted by Eq. 1 of text.	
N_∞	ultimate production of fossil fuel = $8.2\,(N_{\ell 0} + N_{u0})$	
β	biota growth factor = 0.266 (if no calcium carbonate dissolution) = 0.2666 (if calcium carbonate dissolution occurs in deep ocean) = −0.0052 (if calcium carbonate dissolution occurs in both the deep ocean and surface ocean layer)	
	Unless otherwise stated, the growth factor, β, was set to zero linearly between the years 2000 and 2025.	

in the surface layer and deep sea for the three possibilities considered are:

(1) Surface layer and deep ocean supersaturated (or no solid carbonate phase present):

$$dn_m/dt = k_3 n_\ell - k_4 n_m + k_6 n_d - k_5 n_m \,, \qquad (A.58)$$

$$dc_m/dt = k_6 c_d - k_5 c_m \,, \qquad (A.59)$$

$$dn_d/dt = -k_6 n_d + k_5 n_m \,, \qquad (A.60)$$

$$dc_d/dt = -k_6 c_d + k_5 c_m \,. \qquad (A.61)$$

(2) Ocean surface layer supersaturated, deep ocean saturated:

$$dn_m/dt = k_3 n_\ell - k_4 n_m + k_6 n_d - k_5 n_m \,, \qquad (A.62)$$

$$dc_m/dt = k_6 c_d - k_5 c_m \,, \qquad (A.63)$$

$$(1 - \theta_d)\, dn_d/dt = -k_6(n_d - c_d) + k_5(n_m - c_m) \,. \qquad (A.64)$$

(3) Ocean surface layer and deep ocean both saturated:

$$(1 - \theta_m)\, dn_m/dt - k_3 n_\ell \quad k_4 n_m + k_6(n_d - c_d) - k_5(n_m - c_m) \,, \qquad (A.65)$$

$$(1 - \theta_d)\, dn_d/dt = -k_6(n_d - c_d) + k_5(n_m - c_m) \,. \qquad (A.66)$$

Calculation Procedure with $CaCO_3$ Dissolution

In the computations in which calcium carbonate was assumed to be present, the solubility constant in the deep ocean was adjusted to a value such that saturation was prescribed to exist at preindustrial conditions. The deep ocean was then assumed to remain saturated during the fossil fuel era, and the ensuing calcium concentration therein was determined from the adjusted solubility constant.

The solubility constant in the ocean surface layer was adjusted as for the deep ocean, so that saturation prevailed at preindustrial conditions, or else it was set according to data in the literature (Table A.2) for calcite or aragonite. If data for calcite were used, the preindustrial product of calcium concentration and carbonate concentration was larger than K_{sp} by a factor of 4.8, and saturation was never reached with the N_∞ = 8.2 and n = 0.5 input described in the main text. However, if data for aragonite were used, saturation occurred near the year of peak fossil fuel input.

Table A.2. Average chemical parameters assumed for the oceans. [The equilibrium constants K_0, K_1, K_2, K_B and K_W were calculated from equations summarized by Bacastow and Keeling (1973). K_{sp} is from Edmund and Gieskes (1970).]

Surface ocean layer:	
Temperature, °C	19.59
Chlorinity, parts per thousand	19.24
Borate, millimoles/liter	0.41
Calcium, millimoles/liter (initial value)	10.20
Total CO_2, millimoles/liter (initial value)	2.057
Partial pressure CO_2, ppm (initial value)	290.
Alkalinity, millimoles/liter (initial value)	2.435
$K_0 = [CO_2]/P_{CO_2}$, moles/liter-atm	0.03347
$K_1 = [H^+]\,[HCO_3^-]/[CO_2]$, moles/liter	9.747×10^{-7}
$K_2 = [H^+]\,[CO_3^{2-}]/[HCO_3^-]$, moles/liter	8.501×10^{-10}
$K_B = [H^+]\,[B(OH)_4^-]/[B(OH)_3]$, moles/liter	1.881×10^{-9}
$K_W = [H^+]\,[OH^-]$, (moles/liter)^2	6.463×10^{-15}
K_{sp} (aragonite) = $[Ca^{2+}]\,[CO_3^{2-}]$ (moles/liter)^2	9.445×10^{-7}
K_{sp} (calcite) = $[Ca^{2+}]\,[CO_3^{2-}]$ (moles/liter)^2	5.944×10^{-7}
Deep ocean:	
Temperature, °C	1.5
Chlorinity, parts per thousand	19.21
Borate, millimoles/liter	0.41
Calcium, millimoles/liter (initial value)	10.32
Total CO_2, millimoles/liter (initial value)	2.365
Alkalinity, millimoles/liter (initial value)	2.214
K_1	6.511×10^{-7}
K_2	5.086×10^{-10}
K_B	1.191×10^{-9}
K_W	1.304×10^{-15}

The ocean surface layer was allowed to change from supersaturated to saturated, or vice versa, only at the end of a time integration step. Since two steps per year were employed for calculations up to 1975, and eight steps per year thereafter, this procedure resulted in very small errors. To detect a change from saturated to supersaturated conditions, the program checked the source term to find if it had become negative.

Numerical integration was by a fourth-order Runge-Kutta procedure (Carnahan et al., 1969, p. 363).

Chemical Variables with $CaCO_3$ Dissolution

Charge conservation in the system containing sea-water, calcium carbonate, and CO_2 is equivalent to conservation of

$$A - 2[Ca^{2+}] \,, \qquad (A.67)$$

where

$$A \equiv [HCO_3^-] + 2[CO_3^{2-}] + B(OH^-)_4 + [OH^-] - [H^+] \tag{A.68}$$

is the combined carbonate and borate (titration) alkalinity and $[Ca^{2+}]$ is the calcium concentration. We define a special alkalinity

$$A' \equiv A - 2([Ca^{2+}] - [Ca^{2+}]_{m0}) , \tag{A.69}$$

where $[Ca^{2+}]_{m0}$, the initial (preindustrial) calcium concentration in the ocean surface layer, is included in the definition for convenience. Consequently, in the ocean surface layer, the special alkalinity is equal to the initial titration alkalinity:

$$A'_m = A_{m0} . \tag{A.70}$$

We also assume that the special alkalinity in the deep sea equals the special alkalinity in the mixed layer:

$$A'_d = A'_m . \tag{A.71}$$

Since both A'_d and A'_m remain constant, the model prescribes zero transport of special alkalinity between the ocean surface layer and the deep ocean, although there is transport of ordinary alkalinity. The initial titration alkalinity in the deep ocean is set to be 10% higher than the titration alkalinity in the ocean-surface layer (compare Broecker, 1974, p. 44). As a result of these assumptions, the initial calcium concentration in the deep ocean is given by:

$$[Ca^{2+}]_{d0} = [Ca^{2+}]_{m0} + (0.05)A_{m0} . \tag{A.72}$$

For a particular oceanic reservoir, the total dissolved inorganic carbon, ΣC, is the sum of the concentration of the dissolved (and hydrated) carbon dioxide, $[CO_2]$, and the concentrations of the bicarbonate and carbonate ions:

$$\Sigma C = [CO_2] + [HCO_3^-] + [CO_3^{2-}] . \tag{A.73}$$

In terms of equilibrium constants, listed in Table A.2,

$$\Sigma C = [CO_2] \; \{1 + K_1/[H^+] + K_1K_2/[H^+]^2\} \tag{A.74}$$

$$= [CO_2] \; \{1 + \sqrt{K_1/K_2} \; X + X^2\} , \tag{A.75}$$

where

$$X \equiv \sqrt{K_1K_2}/[H^+] \tag{A.76}$$

is a convenient (and dimensionless) parameter. Similarily, the total borate, ΣB, is

$$\Sigma B = [B(OH)_3] + [B(OH)_4^-] \tag{A.77}$$

$$= [B(OH)_4^-] \; \{1 + [H^+]/K_B\} \tag{A.78}$$

$$= [B(OH)_4^-] \; \{1 + \sqrt{K_1K_2}/(K_BX)\}. \tag{A.79}$$

The alkalinity then becomes

$$A = [CO_2] \; \{K_1/[H^+] + 2K_1K_2/[H^+]^2\} + [B(OH^-)_4] + K_W/[H^+] - [H^+] \tag{A.80}$$

$$= [CO_2] \; \{\sqrt{K_1/K_2} \; X + 2X^2\} + (\Sigma B)X/(X + \sqrt{K_1K_2}/K_B) + (K_W/\sqrt{K_1K_2}) \, X - \sqrt{K_1K_2}/X . \tag{A.81}$$

If the seawater of the ocean reservoir is assumed to be saturated with calcium carbonate, then

$$[Ca^{2+}] = K'_{sp}/[CO_3^{2-}] = K'_{sp}/([CO_2] \; X^2) , \tag{A.82}$$

where K'_{sp} is the calcium carbonate solubility constant, perhaps adjusted so that saturation holds for preindustrial conditions, as discussed above.

The stepwise integration of the model equations requires the repeated evaluation of the evasion factor

$$\xi = \left[\frac{(P_m - P_{m0})/P_{m0}}{(\Sigma C - \Sigma C_0)/\Sigma C_0}\right]_{A'}$$

$$= \left[\frac{([CO_2]_m - [CO_2]_{m0})/[CO_2]_{m0}}{(\Sigma C - \Sigma C_0)/\Sigma C_0}\right]_{A'} \tag{A.83}$$

for the ocean surface layer, and of the derivative

$$\theta = [d[Ca^{2+}]/d[\Sigma C]]_{A'} \tag{A.84}$$

for the deep ocean. If the surface ocean is saturated with calcium carbonate, another value of θ is determined for the surface ocean layer. These parameters were evaluated numerically by solving Eqs. A.69, A.81, and, if the seawater was saturated, A.82, for X (and $[CO_2]$) for given values of ΣC. Newton's Method (Carnahan et al., 1969, p. 171) was employed, with the derivative dA'/dx, which is required by this method, evaluated numerically.

With no $CaCO_3$ dissolution in the ocean surface layer, ξ was initially 8.87; with $CaCO_3$ dissolution, it was

2.43. As CO_2 was added to the model, however, ξ increased. The derivative θ, however, was always near 0.47 for both ocean reservoirs.

The values of equilibrium constants and ocean parameters used in the calculations are summarized in Table A.2.

APPENDIX B

We show below that a linear model of the carbon cycle with constant coefficients and an exponentially rising input of CO_2 to the air will result in a constant airborne fraction after times long compared to the time constant of the exponential input. Ekdahl and Keeling (1973) have shown that such a model will result in constant airborne fraction after times long compared to both the exponential input and the fundamental time constants of the model. This difference is not trivial: the industrial revolution is perhaps 200 years old, long compared to the recent exponential time constant of approximately 22 years, but less than the longest fundamental time constant of the model (typically about 300 years).

The equations of a linear m-reservoir model with constant coefficients can be written:

$$dn/dt = -Kn + \Gamma \,, \qquad \text{(B.1)}$$

where n(t) is an m-dimensional vector of perturbations in each reservoir, K is an m by m dimensional matrix with constant elements, and $\Gamma(t)$ is the source term, an m-dimensional vector with all elements zero except that element corresponding to the atmosphere. For the atmospheric source element we postulate:

$$\Gamma_a = \gamma_0 e^{rt} \,, \qquad \text{(B.2)}$$

where γ_0 is a constant.

Equation B.1 can be put in the form:

$$(1/p)\,(d/dt)\,(pn) = \Gamma \,, \qquad \text{(B.3)}$$

where p is an integrating factor. Expansion of Eq. B.3 and comparison with B.1 allows p to be calculated:

$$dn/dt + [(1/p)\,(dp/dt)]\ n = \Gamma \,, \qquad \text{(B.4)}$$

$$(1/p)\,(dp/dt) = K \,, \qquad \text{(B.5)}$$

$$p = e^{Kt} \,. \qquad \text{(B.6)}$$

Equation B.3 can now be integrated. We assume the initial condition

$$n(0) = 0 \,, \qquad \text{(B.7)}$$

and find:

$$n = e^{-Kt} \int_0^t e^{K\tau}\, \Gamma(\tau)\, d\tau \,. \qquad \text{(B.8)}$$

The interpretation of e^{Kt}, where K is a matrix, requires a representation in which K is diagonal.

We assume the existence of a complete set of eigenvectors of K and a set of vectors reciprocal to the eigenvectors. Let the eigenvectors be v_α and the reciprocal vectors be u_α, where the index α runs from 1 to m. Then

$$Kv_\alpha = \lambda_\alpha v_\alpha \,, \qquad \text{(B.9)}$$

where λ_α is the eigenvalue of K corresponding to the eigenvector v_α, and

$$\langle u_\alpha, v_\beta \rangle = \delta_{\alpha\beta} \,, \qquad \text{(B.10)}$$

where $\delta_{\alpha\beta}$ is the Kroniker delta. The left-hand side of Eq. B.10 is a scaler product. As a consequence of Eq. B.9,

$$e^{Kt} v_\alpha = \exp(\lambda_\alpha t)\, v_\alpha \,. \qquad \text{(B.11)}$$

Because the set of v_α is complete,

$$\Gamma = \sum_{\alpha=1}^{m} v_\alpha \langle u_\alpha, \Gamma \rangle \,. \qquad \text{(B.12)}$$

After substitution of Eqs. B.11 and B.12 into B.8,

$$n = \int_0^t \sum_{\alpha=1}^{m} \exp[-\lambda_\alpha (t-\tau)]\, v_\alpha \langle u_\alpha, \Gamma \rangle\, d\tau \,. \qquad \text{(B.13)}$$

Since Γ is zero except for the atmospheric component, the CO_2 increase in the atmosphere is:

$$n_a = \sum_{\alpha=1}^{m} \Big\{ (v_\alpha)_a (u_\alpha)_a \times \int_0^t \exp[-\lambda_\alpha (t-\tau)]\ \gamma_0 e^{r\tau}\, d\tau \Big\} \,, \qquad \text{(B.14)}$$

$$n_a = \sum_{\alpha=1}^{m} \{ (v_\alpha)_a (u_\alpha)_a \gamma_0 e^{rt} / (\lambda_\alpha + r) \times [1 - \exp(-(\lambda_\alpha + r)t)] \} \,. \qquad \text{(B.15)}$$

The total CO_2 input is:

$$T = \int_0^t \gamma_0 e^{rt} = (\gamma_0/r)\,(e^{rt} - 1) \,, \qquad \text{(B.16)}$$

and after a time long compared to 1/r,

$$T \approx (\gamma_0/r)\, e^{rt} \,. \tag{B.17}$$

The airborne fraction is then:

$$F = n_a/T = \sum_{\alpha=1}^{m} \{(v_\alpha)_a (u_\alpha)_a\, r/(\lambda_\alpha + r) \times [1 - \exp(-(\lambda_\alpha + r)t)]\} \,. \tag{B.18}$$

This fraction will be approximately constant after times long compared to 1/r because all the λ_α must be positive (or zero) or have a positive (or zero) real part.

The real part of each λ_α must be positive or zero because CO_2 is conserved. To demonstrate this, consider the response of the system to a unit impulse of CO_2 at time zero. The response can be determined from Eq. B.13, and is a sum of exponentials with time constants $1/\lambda_\alpha$:

$$n = \sum_{\alpha=1}^{m} C_\alpha \exp(-\lambda_\alpha t) \,, \tag{B.19}$$

where the C_α are constants, independent of time. If any λ_α had a negative real part, CO_2 in the system would grow without limit, according to Eq. B.19, in contradiction to CO_2 being conserved. This can be formally (and relatively easily) proven for systems of equations of the type that arise in box models (Siegert, 1949).

Since the preindustrial CO_2 level is only poorly known, the airborne fraction is usually calculated during the period of the Mauna Loa and South Pole records:

$$F = [n_a(t_2) - n_a(t_1)]/[T(t_2) - T(t_1)] \,. \tag{B.20}$$

Substitution from Eqs. B.15 and B.16 then results in

$$F = \sum_{\alpha=1}^{m} \frac{(v_\alpha)_a (u_\alpha)_a\, r}{(\lambda_\alpha + r)} \times \left\{ 1 - \exp(-(\lambda_\alpha + r)t_1) \left[\frac{\exp(-\lambda_\alpha(t_2 - t_1)) - 1}{\exp(r(t_2 - t_1)) - 1} \right] \right\} . \tag{B.21}$$

If t_1 is long compared to 1/r, F calculated from Eq. B.20 will be constant and identical to F calculated from B.18.

The above derivation can be extended to show that the fraction of the input in any reservoir becomes constant after a time long compared to 1/r.

Some Considerations of Climatic Variability in the Context of Future CO_2 Effects on Global-Scale Climate

J. Murray Mitchell, Jr., Environmental Data Service,
National Oceanic and Atmospheric Administration,
Silver Spring, Maryland

1. INTRODUCTION

When contemplating the significance to climate of a long-term increase of atmospheric CO_2, we must recognize that the natural climatic system being perturbed by the CO_2 is a system that exhibits a considerable degree of inherent variability. This is important for two reasons:

1. The inherent variability of climate, which would continue in the future at the same time that the CO_2 is being injected, would tend to obscure the climatic change that is properly attributable to the CO_2 increase.
2. The many natural sources of climatic variability, including those related to "forcing" of the climatic system by systematically changing environmental conditions (such as possible solar energy variations and the sporadic injection of volcanic particles into the stratosphere) would in the future compete with the added CO_2 to produce future changes of climate.

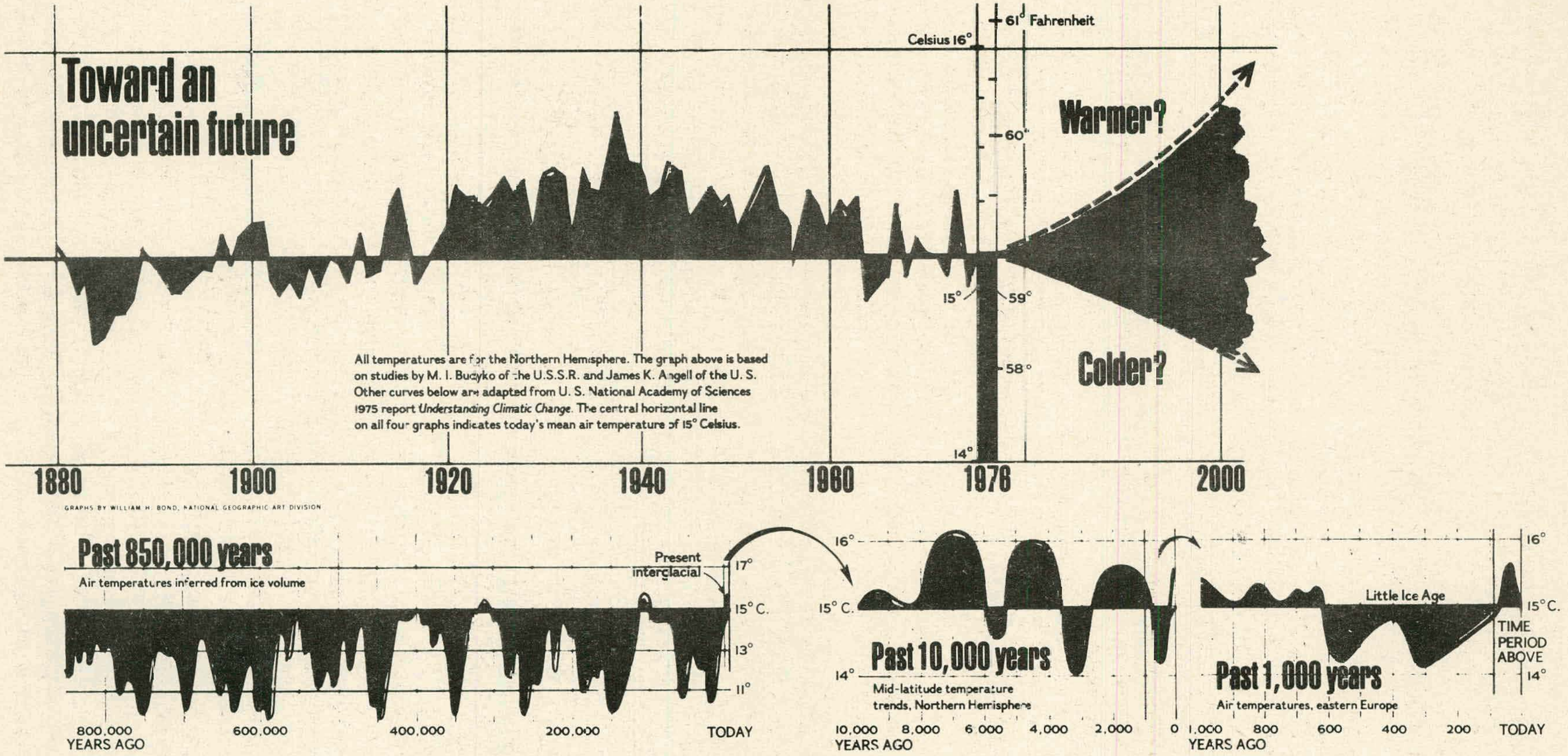
In both these respects, the task of reliably assessing the significance of future CO_2 changes to future climate is much more complicated than it would be if the natural climate were as "well behaved" as the kind of climate we are able to simulate in the present generation of general circulation and climate models. In experiments with these models everything is held constant except the atmospheric CO_2 level (the "sensitivity study" approach). In real life, on the other hand, the CO_2 level is but one of innumerable respects in which the environmental parameters governing the state of global climate are simultaneously changing.

There is general agreement that the primal climatic effect of a CO_2 enrichment of the atmosphere would be a planetary-scale warming, through the so-called "greenhouse effect." It so happens that the best documentation we have of climatic changes in past centuries, millennia, and geological epochs is in terms of temperature conditions. It therefore makes sense for us to pool information about atmospheric temperature into hemispheric- or global-average statistics, and to determine as best we can how the mean temperature over the world as a whole has changed down through history. If the theoretically expected changes of future planetary mean temperature that are associated with reasonable scenarios of future atmospheric CO_2 changes are compared with observed past changes of planetary mean temperature, we can gain at least some idea as to the importance of the CO_2 changes, relative to other (natural) controls, in altering our future climate.

With the above in mind, it is the purpose of this paper to consider aspects of past variations of planetary-scale temperature conditions on the earth that may be helpful to us in assessing the importance of future temperature changes possibly related to atmospheric CO_2 increases.

2. PAST FLUCTUATIONS OF PLANETARY-SCALE MEAN TEMPERATURE

I begin by summarizing what we know about general global-scale conditions of temperature today, compared with conditions existing in earlier periods of the earth's climatic history. A graphic summary of the situation is shown in Fig. 1 (Matthews, 1976). This consists largely of information gleaned from a 1975 National Academy of Sciences report (NAS, 1975), updated for the most recent years with data from Angell and Korshover (1977). Wherever the



All temperatures are for the Northern Hemisphere. The graph above is based on studies by M. I. Budyko of the U.S.S.R. and James K. Angell of the U. S. Other curves below are adapted from U. S. National Academy of Sciences 1975 report *Understanding Climatic Change*. The central horizontal line on all four graphs indicates today's mean air temperature of 15° Celsius.

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Fig. 1. History of global-scale mean temperature of the earth as inferred from a variety of paleoclimatic indicators, after Matthews (1976). ©National Geographic Society. Reprinted by permission of the National Geographic Society.

curves in Fig. 1 are seen to lie below the heavy horizontal lines in the different panels of the figure, general temperature levels (in the northern hemisphere, at least) are inferred from various kinds of paleoclimatic evidence to have been colder than they are now.

Referring to the top panel of Fig. 1, we see the history of the last 100 years of average temperature in the northern hemisphere, according to Budyko (1969) and updated since 1968 by Angell and Korshover (1977). With due allowance for errors of estimation that are undoubtedly present in this series, the interannual variations of temperature shown here may be indicative of slight changes in the overall global heat budget as affected by such factors as volcanic dust in the stratosphere, carbon dioxide accumulation in the atmosphere, and perhaps solar radiation changes. These are the kinds of things that might be expected to influence globally-averaged temperatures, in addition to whatever other effects they may have on climate of a regionally preferential character. The fact that temperatures were generally higher in the 1920s, 1930s, and 1940s than they had been at the end of the last century, and higher than they are now, may mean something quite significant if we are clever enough to decipher it. For example, Oliver (1976) has interpreted the relative warmth of those three decades as a consequence of the lack of major volcanic activity leading up to that period, in contrast to the much higher level of volcanic activity between the 1880s and 1910s and a latter-day resurgence of such activity in the 1950s and 1960s.

In any event, much of the world has evidently been cooling in the last 30 or 40 years. The extent to which that cooling is likely to be still in progress today, and therefore likely to continue for some period of time in the future, is a matter of considerable interest vis-à-vis the problem of fixing public attention on the CO_2-climate problem as a credible societal threat. I shall return to this matter for a closer look in the next section.

Referring next to the lower right panel of Fig. 1, we see the present situation in the perspective of the last thousand years. The temperature index in that panel happens to be derived for the area of eastern Europe, where considerable historical information is available to construct such an index. We cannot be certain that this index is representative of the hemisphere as a whole, or of the world as a whole, but it is probably indicative of global-scale changes during the past millennium. The "Little Ice Age" was the most remarkable feature here. This consisted of two particularly cool periods, one in the latter 15th century and the other in the latter 17th century, which were characterized by expanded mountain glaciation, a general increase of arctic sea ice, heightened storminess of the North Atlantic, and a general planetary-scale cooling of the order of 1°C. We have the impression that the Little Ice Age event has by now run its full course, although a few climatologists have speculated otherwise.

In the panel at the bottom center of Fig. 1, we find a rough index of how the temperature of the earth has changed over the past 10,000 years, that is, since the recovery of the earth from the last major ice age. There are many paleoclimatic indications that this postglacial period was punctuated by a whole series of "neoglacial" events, each resembling the Little Ice Age, at intervals of two or three millennia (see, for example, Denton and Karlén, 1977). There is some reason to believe that these events are related to long-term variations of solar activity (Eddy, 1977).

Referring, finally, to the lower left panel of Fig. 1, we see a still longer view of the earth's climate, based on deep-sea sediment analyses (Shackleton and Opdyke, 1973; NAS, 1975). This spans a substantial part of the Quaternary (Ice Age) period, which is the appropriate reference for the climate of today on the geological scale of time. What stands out very clearly here is the fact that, within the last million years or so, intervals of time when the earth was warmer than now are few and far between. That is to say, the prevailing condition of the earth's climate during the present ice-age epoch has been a great deal colder than the climate of today. This invites concern that, in the geological perspective at least, the warm (extreme interglacial) period we are in today is not likely to persist much longer, and indeed that warmth may already be slipping away. I concur in this general line of reasoning, but the pace at which this refrigeration of the earth will be likely to take place can scarcely be rapid enough to warrant any kind of alarm with regard to events in the next few centuries. Clearly, it is the more rapid variations of climate, including those associated with the Litle Ice Age events, that justify our attention here.

3. UPDATE ON THE RECENT GLOBAL COOLING

In the face of an expected future warming attributable to increasing atmospheric CO_2, it may seem paradoxical that our global climate has shown no indication of warming in the last few decades. On the contrary, the general trend of planetary-scale temperature since about 1940 has been in the direction of cooling, despite an estimated 8% increase of atmospheric CO_2 in that time (roughly two-thirds of the total CO_2 increase since the last century). In consideration of the fact that atmospheric temperature has shown wide variability in the past, on all time scales, and undoubtedly responds to a wide variety of natural climate forcing phenomena, it should come as no surprise that the recent trend of planetary temperature has been inconsistent with what one would expect on the basis of the response to only one forcing agent such as CO_2. In this respect, the "paradox" I have indicated is more apparent than real. Yet, present-day rates of increase of atmospheric CO_2 are accelerating to a point where, in the rather near future, it would become surprising if the present-day climatic cooling were to continue unabated, as if the atmosphere were to take no note of the CO_2 effect we anticipate in the decades ahead. By the same token, however, should the global climatic cooling soon reverse itself and become a warming (even a rather rapid warming), we could not unequivocally claim that this warming would be proof of a CO_2 effect. Every cooling cycle of global climate in the past has sooner or later been followed by a warming cycle, and who can say that a warming in the near future would be anything other than nature following a normal cycle, CO_2 or no CO_2?

In spite of the difficulties of interpretation I have indicated, it is, nevertheless, a matter of considerable interest to monitor very carefully the progress of the global-scale cooling in the last few decades, and to determine if that cooling is—or is not—showing evidence of faltering at the present time.

In recent years many studies have been made to document the progress of global-scale temperature changes. Most have been done since the 1950s (e.g., Dronia, 1974; Reitan, 1974; Yamamoto, Iwashima, and Hoshiai, 1975; Brinkmann, 1976; Borzenkova et al., 1976; Damon and Kunen, 1976; Angell and Korshover, 1977; and Kukla et al., 1977). Some interesting differences, as well as parallels, are found in the conclusions of these various investigations. For the most part, the studies confirm a continuation of the general cooling that set in about 1940 (top panel of Fig. 1), but they show indications of considerable variation in the year-by-year changes including an excess cooling in the years immediately following the Agung volcanic event of 1963. Angell and Korshover (1977) demonstrate that the cooling (since 1958) was common to virtually all latitude belts of the world, with the notable exception of the antarctic region which has apparently warmed in that period. The results of Damon and Kunen (1976) are anomalous in that they indicate a significant warming of the southern hemisphere as a whole in the last two decades (an evident consequence of the fact that those authors failed to properly area-weight their data when they pooled the trends for Antarctica with those for the lower southern latitudes).

It is quite clear from the study of Angell and Korshover (1977), which is geographically the most comprehensive of all the up-to-date studies, that the cooling of the last two decades has been shared not only by the upper atmospheric layers as well as the surface layer, but in virtually equal measure by the northern and southern hemispheres (notwithstanding the antarctic anomaly). This important result contradicts the suggestion by Damon and Kunen (among others) that the southern hemisphere is already responding to the CO_2 warming, unlike the northern hemisphere where (as they propose) the atmosphere may still be cooling because of the overriding effect of increasing particulate pollution concentrated in that hemisphere.

There is some indication in Angell and Korshover's data that the global-scale cooling might have reached a nadir in both hemispheres between five and ten years ago (see Fig. 2). It will, however, require a few additional years of data to verify this with confidence, especially since preliminary information for the year 1976 (not shown in Fig. 2) suggests that that year was one of the coldest since the start of the analysis in 1958 (Angell, personal communication).

4. CHANGES OF CLIMATIC VARIABILITY

An important point that comes up in connection with climatic change is the matter of possible changes of climatic *variability*. The changes in the mean temperature, for example, are relatively slow. If the variability of the climate around that mean were to increase or decrease, this would present us with a set of problems no less troublesome than the problems created by changes of the mean.

What may have first attracted the attention of most of us to the question of climatic variability is the kind of thing we can see in Fig. 3, taken from McQuigg (1973). This shows a time series of the year-to-year variations in corn crop yields in the United States which are attributable only to weather variations. This series has been calculated on the basis of constant technology for the entire period.

Figure 3 suggests, first, that the mean corn yields have been relatively high in recent years, and, second,

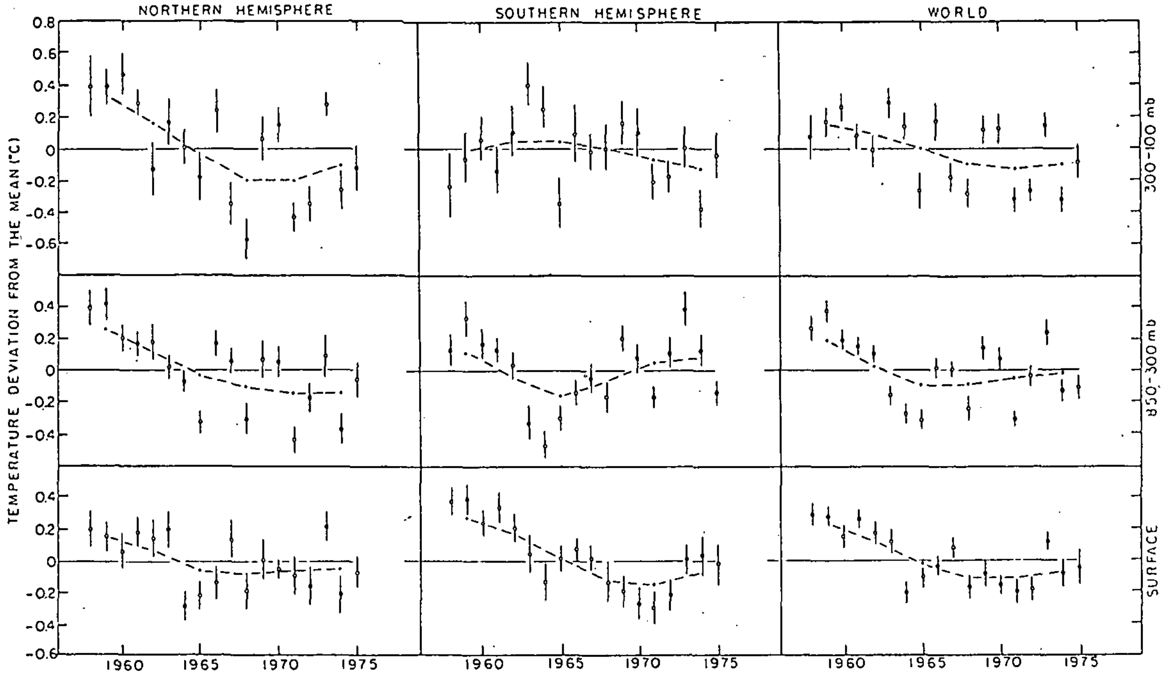

Fig. 2. Trends of annual mean temperature since 1958, for the northern hemisphere, southern hemisphere, and the world as a whole, and for various levels of the atmosphere. Circles and vertical bars indicate relative temperature and its standard error of estimate for each year; dashed curves are binomially weighted moving averages to emphasize trends. After Angell and Korshover (1977).

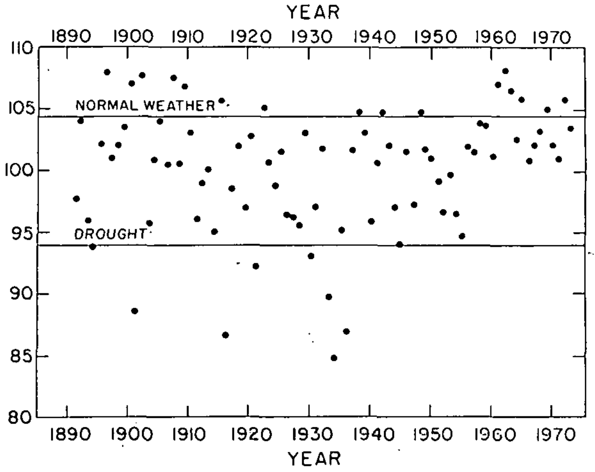

Fig. 3. Weather-related variations since 1891 of simulated average yield of corn crop in principal corn-producing areas of the United States (Ohio, Indiana, Illinois, Iowa, and Missouri), normalized to 1973 technology and harvested acreage. After McQuigg (1973).

that the variability among those recent high yields has been small—astonishingly small. Does this indicate that the climate has stabilized recently, in the sense of becoming less variable?

On the face of it, it does appear that way. However, looking at another measure of climatic variability, this time for the world and not just for an agricultural zone of the United States, I refer to Fig. 4, taken originally from a study by the Japan Meteorological Agency and reproduced in a number of other publications recently (e.g., GARP, 1975). Based on records for about 65 stations around the world, this figure shows the number of station-months in each decade since that of the 1910s in which the mean temperature ranged beyond the 5 and 95 percentile points of temperature for the entire 60-year record. There is an indication in this analysis that the total number of extreme excursions of temperature around the world has been rising slightly in the last two or three decades, and that the proportion of the total number of excursions that involves cold events has been increasing quite substantially. The proportion that involves warm events has been decreasing. It would be a matter of great practical value to determine whether this trend is physically significant and whether it can be extrapolated somehow into the future.

Still another, and in certain respects conflicting, bit of evidence with regard to changes of climatic variability is shown in Fig. 5. The upper part of this figure shows the year-by-year history of the average annual temperature of the entire United States since 1892. This series makes it appear that there was a slight warming up to around 1935 or 1940 and a slight cooling since then. What stands out more than trends is the year-to-year variability. The lower half of Fig. 5 shows an 11-year moving standard deviation of the annual average temperature for the United States. By this measure of variability, the year-to-year temperature variability for the United States as a whole has been decreasing since 1920, and has become especially small after about 1960.

This is quite typical of what one finds if one goes to any regional part of the United States and calculates the changes in the variability of temperature for each part individually. One nevertheless has the impression that, in the last few years, the variability has increased again.

Comparing Fig. 5 with Fig. 4, which suggests that the last 20 years were a period of increasing variability in the world, we see something of a contradiction, at least in the United States. Perhaps this contradiction arises because we are using dissimilar measures of variability here.

By any measure, the variability of climate has been high enough in the last few years, in different parts of the world, that it is tempting to conclude we are moving away again from a period of low variability into one of greater variability. But that is speculation,

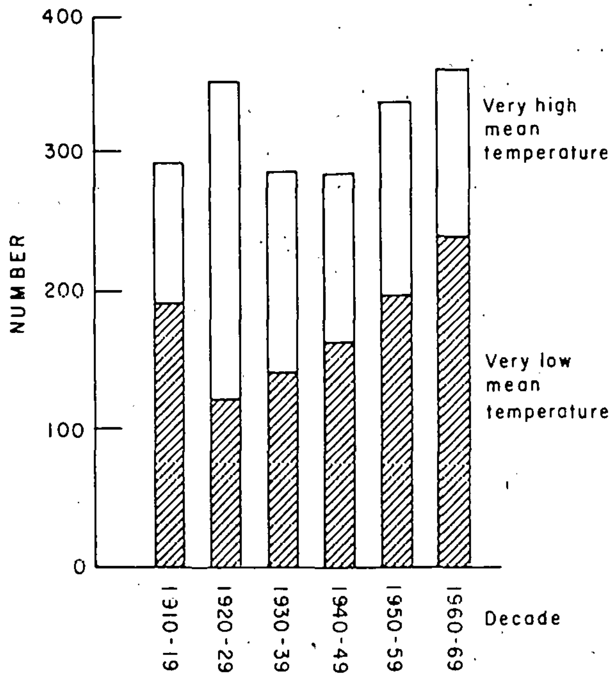

Fig. 4. Number of extremes of mean monthly temperature observed in each decade since 1910, at 65 selected stations around the world. Extreme temperatures are defined as those lying above or below the range expected 90% of the time in all 60 years of record. After Japan Meteorological Agency study; see GARP (1975).

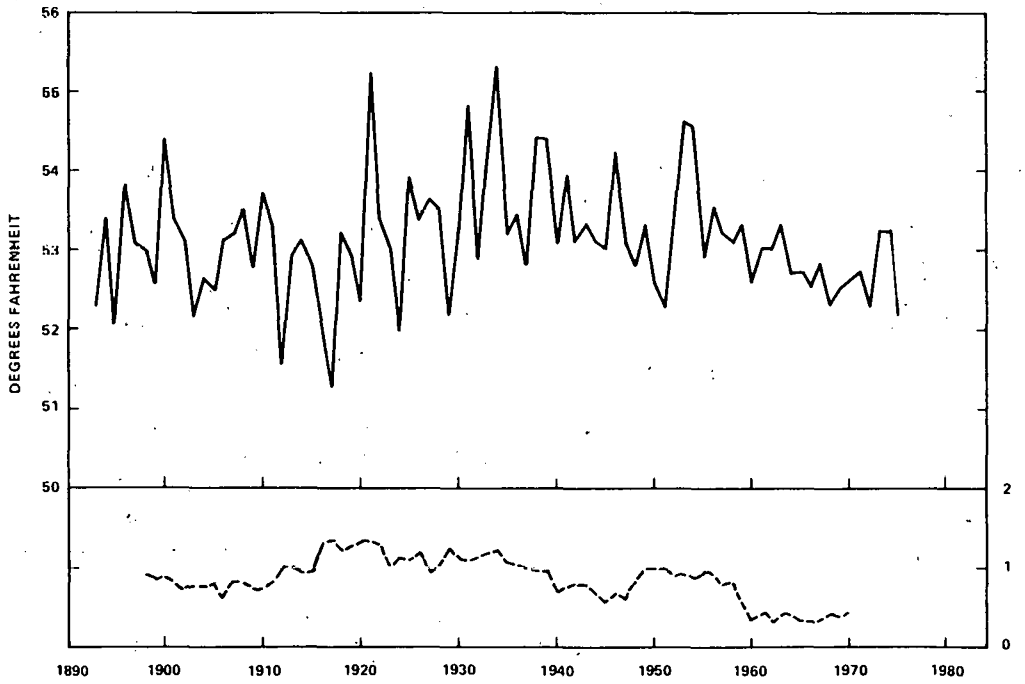

Fig. 5. Above: annual mean temperature of the contiguous United States, 1893 to 1975, derived from averages for each state weighted by their 1970 population. Below: 11-year moving standard deviation of annual mean temperature.

and we need to be very careful in looking at the data we have to verify that we can be justified in drawing any kind of conclusion about systematic changes of climatic variability.

5. A LOOK TO THE FUTURE

Our very superficial understanding of climatic variability and change translates to a very superficial capability for prediction of climate. One of the principal motivations for climate research is to gain enough insight into climate system behavior to enable the skillful prediction of future climate. Yet it has to be recognized that our understanding is not yet adequate to assure us beyond a reasonable doubt that climate *can* be predicted, with practical skill, for an extended period of time into the future.

In the face of the many urgent societal needs for information about future climates, it would seem useful to indicate the essence of our present situation with regard to climate prediction.

On the one hand, it can be observed that the pioneers in the field of modern climate dynamics research have been hesitant to engage in any kind of speculation about future climatic developments. This is the case despite the fact that those pioneers are probably in the best position of anyone to speculate intelligently about the future. This hesitancy is not difficult to understand. It is born of a healthy respect for the limitations of present generations of climatic models, which necessarily omit too much of the physics that they eventually have to include before real-time predictions are likely to bear much relationship to reality.

On the other hand, one can point to the fact that in the last 20 years or so, a number of deterministic long-range projections of climate have found their way into the scientific literature. The authors of such projections have based them on a variety of simplistic reasoning, sometimes involving "blind" statistical extrapolation of what are essentially unverified periodicities in past climate, and other times involving certain external physical processes (e.g., solar activity) which they have assumed to have an overriding influence on the course of global climate. It cannot be categorically asserted that all of these projections are without merit. Unfortunately, where merit exists it will be likely to remain obscure until climate dynamics research finally progresses to a point where it can affirm that merit. By then, however, that research will doubtless provide us with a much more comprehensive assessment of reality.

If society is to embrace any deterministic climate projection, as a basis for sound future planning, it should be clear that two criteria have to be met. First, the scientific basis of the projection must be demonstrably rational. Second, the projection must be accompanied by a suitably objective and unbiased measure of its own reliability. No deterministic projections of future climate now available can be said to meet both of these criteria.

Reliable projections of future climate, of any kind, will likely remain an elusive goal until our knowledge of climatic dynamics is adequate to provide the answers to two sets of fundamental questions.

The first set of fundamental questions to be answered concerns the natural climate system: (1) If man's various impacts on environment were to be held in the future at their present-day levels, how would the earth's climate evolve from its present state in the decades and centuries ahead? (2) Does nature herself know the answer to (1)? Or is the natural evolution of climate a probabilistic process in time, such that, after a relatively short future interval, the state of global climate will be essentially unrelated to its present state? (3) To the extent that nature does in fact know the answer to (1), to what further extent is it within man's power to wrest that answer from her?

The second set of fundamental questions to be answered concerns man himself in relation to his potential capacity for interfering in the workings of the natural climate system: (4) On the basis of each of a variety of credible alternative scenarios as to future increases and proliferation of his impacts on environment, can man assess the consequences of those changing impacts on the natural evolution of climate in the decades and centuries ahead? (5) Can man establish which, if any, alternative scenarios would lead to "unacceptable" climatic consequences, and are therefore to be avoided? (6) If the answer to (5) is yes, will man come into possession of such knowledge in time to avert possible calamity?

The very essential matter of climate *predictability* is addressed by question (2). Should climate happen to be inherently unpredictable after a disappointingly short interval of future time, then the capacity of science ever to provide answers to the other questions posed here will be limited, perhaps severely so.

Viewing past climates in the perspective of events of the past million years, we can at least venture the conclusion that, in the long run, natural forces will be more likely to relax our global climate toward a glacial condition than to maintain climate indefinitely in its present, unusually warm interglacial condition. As to when the return to a glacial condition may begin, and at what rate global climate will be altered when it does begin, these are by no means clear from the paleoclimatic record.

Viewing past climates in the somewhat shorter perspective of the postglacial period, the sequence of "neoglacial" events appears to be a dominant feature. It is tempting to suggest that the last neoglacial event, the "Little Ice Age" of the 16th to 19th centuries, is behind us and that the next such event will not arrive for another thousand years or more.

Beyond such vague statements as these, disturbingly little can be said about the probable cause of natural climate in the decades and centuries ahead. Yet it would appear that by some time within the next few decades the impact of CO_2 accumulation in the atmosphere will begin to dominate over natural processes in terms of governing the general direction of global mean temperature. In Fig. 6, I illustrate this from two points of view: (1) the course of events over the next century, and (2) the CO_2 disturbance in geologic perspective. Two main conclusions follow from this figure, which is discussed in further detail by Mitchell (1977). First, it may not be until the end

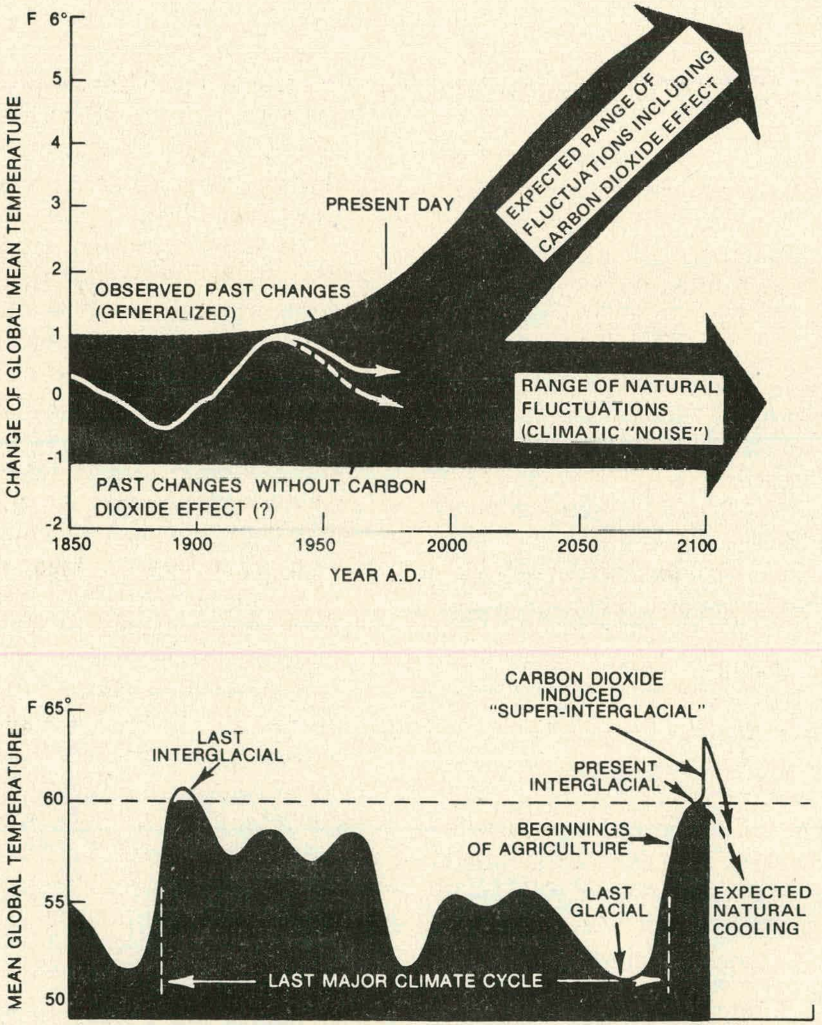

Fig. 6. Above: range of global-scale mean temperature, 1850 to 2100, with and without the projected CO_2 effect. Below: projected CO_2 temperature disturbance in the context of ice-age chronology of the past 150,000 years. After Mitchell (1977).

of the century that the warming effect of future CO_2 increases will suffice to emerge as a clearly recognizable trend against the background of natural climatic variability. By then, however, the trend would be climbing rather steeply toward temperature levels exceeding anything hitherto experienced by mankind. Second, by the time the CO_2 disturbance will have run its full course, more than a millennium into the future, temperature levels in the global atmosphere may possibly exceed the highest levels attained in the past million years of the earth's history.

These projections are, of course, contingent upon two crucial assumptions: first, that our present theoretical assessment of the climatic impacts of atmospheric CO_2 changes is at least in first approximation correct, and, second, that the world reserves of fossil fuels will in fact be extensively raided in future centuries and that a substantial fraction of the resulting addition of fossil CO_2 to the atmosphere will remain airborne as now expected.

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The Effect of Increasing the CO_2 Concentration on the Climate of a General Circulation Model

Syukuro Manabe, Geophysical Fluid Dynamics Laboratory, NOAA,
Princeton University, Princeton, New Jersey

The climatic influence of the change of CO_2 concentration in air is discussed based upon the results from a simplified general circulation model of the atmosphere (Manabe and Wetherald, 1975). The model consists of the equation of motion, thermodynamical equation, equation of radiative transfer, and the prognostic equation of water vapor. The temperature of the earth's surface is determined by the balance condition that no heat is stored there. The depth of snow cover and the amount of soil moisture is determined from the budget of snow and liquid water respectively. One of the main characteristics of the model is the estimation of heat transport by large-scale eddies through explicit computation rather than parameterization. On the other hand, the model contains the following simplifications: a limited computational domain, an idealized topography, no heat transport by ocean currents, and fixed cloudiness.

The climatic influence of the change in CO_2 concentration is estimated by comparing two quasi-equilibrium model climates with normal and twice the normal concentration of carbon dioxide. A quasi-equilibrium climate is computed from the long-term integration of the model described above. Figure 1 illustrates the difference between the zonal mean temperature of the two model atmospheres as a function of latitude and altitude. This figure reveals that the increase in CO_2 concentration raises the temperature of the model troposphere, whereas it lowers the temperature of the model stratosphere. The tropospheric warming is somewhat larger than that obtained from a one-dimensional radiative-convective model in which the positive feedback mechanism between snow albedo and temperature is not taken into consideration (see, for example, Manabe and Wetherald, 1967; Augustsson and Ramanathan, 1977). In this connection, it is important to note that the surface temperature increase in high latitudes is computed to be relatively large due to the decrease of the highly reflective snow-covered region and the thermal stability of the lower troposphere, which limits the convective heating to the near-surface layer.

In addition, the increase in CO_2 concentration causes a more active hydrologic cycle in the model atmosphere as indicated by greater rates of global mean precipitation and evaporation. The following factors are identified to be responsible for increasing the evaporation rate and thus enhancing the intensity of the hydrologic cycle:

1. Downward flux of atmospheric radiation increases not only due to the increase of CO_2 content but also due to the increase of water vapor content in response to the general warming of air temperature.
2. The derivative of saturation vapor pressure with respect to temperature increases nonlinearly with increasing temperature. Therefore, in the boundary layer, the vertical gradient of vapor pressure increases more than the vertical temperature gradient responding to the increase of surface temperature. Accordingly, the larger proportion of available radiative energy at the earth's surface is used for evaporation rather than for sensible heat flux.

The general increase in the intensity of the hydrologic cycle mentioned above does not necessarily imply the increase of wetness everywhere. In other words, the sign of wetness (aridity) change depends upon the geographical location of the model.

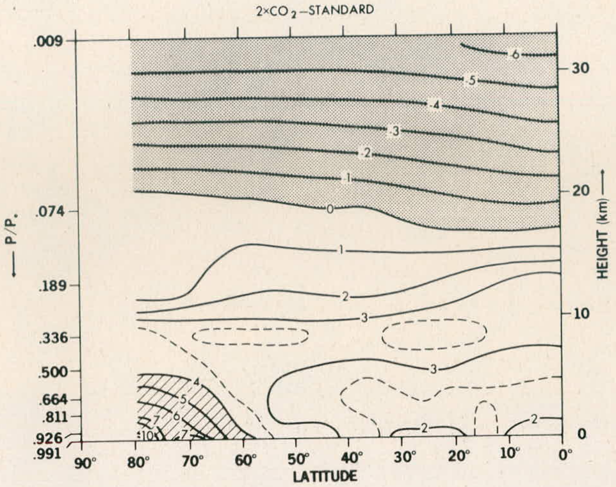

Fig. 1. Latitude-height distribution of the difference in zonal mean temperature (° K) of the model atmosphere resulting from the doubling of CO_2 concentration. Stippling indicates the decrease in temperature.

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Cryospheric Responses to a Global Temperature Increase

Roger G. Barry, Institute of Arctic and Alpine Research and Department of Geography,
University of Colorado, Boulder, Colorado

INTRODUCTION

The term *cryosphere* embraces all terrestrial manifestations of ice in snow cover, glaciers, ice sheets, seawater and freshwater ice, and ground ice or permafrost. (The minute fraction occurring as ice crystals in the atmosphere is not discussed in the present context.) Since snow and ice are persistent only with subzero temperatures, global warming induced by an increase in the atmospheric content of carbon dioxide will lead to changes in the cryosphere. The likely nature and possible magnitude of such effects are reviewed here in the context of existing literature studies.

At present, perennial ice covers 7% of the world's ocean and 11% of the land surface, primarily in high latitudes. Data on the extent of snow and ice are summarized in Table 1 (Unstersteiner, 1975). Latitudinal distributions are tabulated by Shumskiy, Krenke, and Zotikov (1964). Land ice, which represents 75% of all terrestrial fresh water, has by far the largest volume, with the antarctic and Greenland ice sheets accounting for nearly all of this. The snow cover is distributed about 2/3 on land and 1/3 on sea ice, and is about twice as extensive in the northern hemisphere as in the southern. It is the most seasonally variable component of the cryosphere. Figure 1 shows the duration of snow cover and the range of variation of its extent (except on the polar sea ice). The average depth of maximum late-winter snow cover ranges from about 40 cm on the high arctic tundra to 150 cm in the northern taiga of central Siberia (Mellor, 1964; Potter, 1965). Fresh snow has a typical density in the range of 0.05 to 0.20 g/cm^3, whereas in late-winter snow packs it averages 0.25 to 0.40 g/cm^3 (McKay and Thompson, 1968). The snow line rises toward the equator from sea level at approximately 65°S and 90°N. Maximum snowline elevations exceed 6 km in the continental subtropics of central Asia and Bolivia-Peru as a result of dryness and high solar radiation intensities (Hermes, 1965). On equatorial high mountains, snowfall is characteristically a diurnal, rather than a seasonal, phenomenon.

CRYOSPHERIC SENSITIVITY

The response of snow and ice covers to climatic factors varies widely in terms of time scale. Typical residence times of solid precipitation in the various reservoirs have the following orders of magnitude: 10^{-1} to 1 year for seasonal snow cover, 1 to 10 years for sea ice, and 10^3 to 10^5 years for ground ice and ice sheets. In all instances, however, certain phases of the seasonal regime are particularly critical for the occurrence of snow or ice and their response to climatic variations. Of primary importance are the times of seasonal temperature transition across the 0°C threshold (or −1.8°C in the case of seawater). Other threshold effects, which influence radiative and turbulent energy exchanges, arise as a result of the large albedo differences between snow cover (about 0.80) and snow-free ground (0.10 to 0.25) or between ice (0.65) and water (0.05 to 0.10).

In the case of snow cover, maps prepared by Dickson and Posey (1967) indicate that, in all months of the year, extensive areas of North America and Eurasia have a probability of snow cover (exceeding 1 in.) ranging from 0 to 50%. This wide range is apparent in the large year-to-year variability in northern hemisphere snow cover, especially in Eurasia, reported by Wiesnet and Matson (1976). Global warming could eliminate the occurrence of

Table 1. Components of the cryosphere[a]

	Area (10^6 km^2)	Volume (10^6 km^3 water)
Land ice:	16	31
Antarctica	13.9	28
Greenland	1.8	2.7
Ground ice (excluding Antarctica):		
continuous permafrost	7.5	≤2
discontinuous permafrost	17.5	≤0.5
Sea ice:		
Antarctic, September	20	0.05
March	2.5	
Arctic, September	8.4	
March	15	
Snow cover,[b] March (North America, Eurasia)	38	
Total snow, land ice, sea ice:[c]		
January: northern hemisphere	58	
southern hemisphere	18	
July: northern hemisphere	14	
southern hemisphere	25	
Global mean annual	59	

[a] After Untersteiner, N. 1975. Sea ice and ice sheets and their role in climatic variations. Pages 206–26 in *The physical basis of climate and climate modelling*, GARP Publ. Ser. 16, World Meteorological Organization Geneva, Switzerland.

[b] From Wiesnet, D. R. and M. Matson. 1976. A possible forecasting technique for winter snow cover in the northern hemisphere and Eurasia. *Mon. Weather Rev.* 104:828–35.

[c] From Kukla, G. J. and H. J. Kukla. 1974. Increased surface albedo in the northern hemisphere. *Science* 183:709–14.

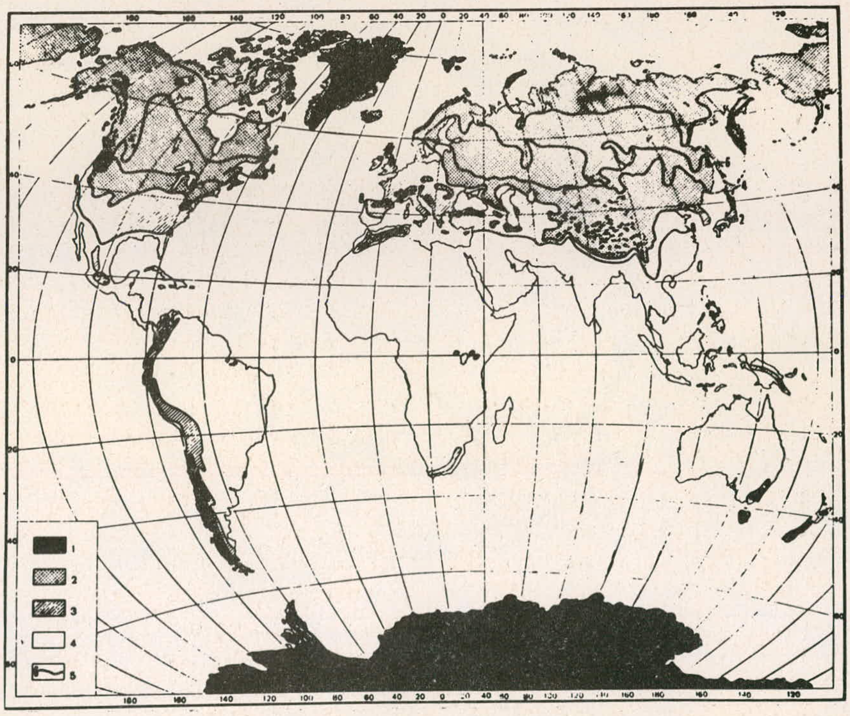

Fig. 1. World distribution of snow cover. 1: permanent cover of snow and ice; 2: a stable snow cover of varying duration forms every year; 3: a snow cover forms almost every year, but is not stable; 4: no snow cover; 5: duration of snow cover (months). (From Mellor, 1964, after Rikhter.)

snow covers in the broad areas of low frequency depicted by Dickson and Posey.

Sea ice in the southern hemisphere displays a wide seasonal variation in extent (Fig. 2A). This is a result of the well-mixed structure of the waters on the shelf around Antarctica; the entire column down to about 200 m must be cooled to −1.8° C before freezing can take place. In contrast, in the largely closed Arctic Ocean (Fig. 2B), a shallow layer of low salinity is maintained by continental runoff and ice melt in summer and by haline convection below refreezing leads in winter. There is a pronounced pycnocline at 25 to 50 m depth, and hence freezing occurs readily and there is rather limited summer breakup.

In the case of ground ice or ice sheets, the long residence times greatly reduce their sensitivity to any climatic change. Nevertheless, change does occur on long time scales, as illustrated by the geological and oceanographic records of glacial history. Land ice increased to cover approximately 25% of the land surface at the Last Glacial Maximum, which ended about 16,000 years ago.

CRYOSPHERIC RESPONSES TO WARMING

Snow

The effects of a temperature increase on snowfall and snow cover must be considered on a seasonal and a geographical basis. In middle latitudes, where the occurrence of snow rather than rain is frequently marginal (the 50% probability limit for snow/rain occurrence generally occurs at a surface air temperature of 1.5° C or a freezing level height of 250 m), warming will decrease the frequency of snowfall and the duration of snow cover on the ground. For southeast England, Manley (1969) shows a shorter snow season in 1931 to 1960 compared with 1901 to 1930 or the nineteenth century. The period with snow lying in Britain was a minimum in the 1920s and

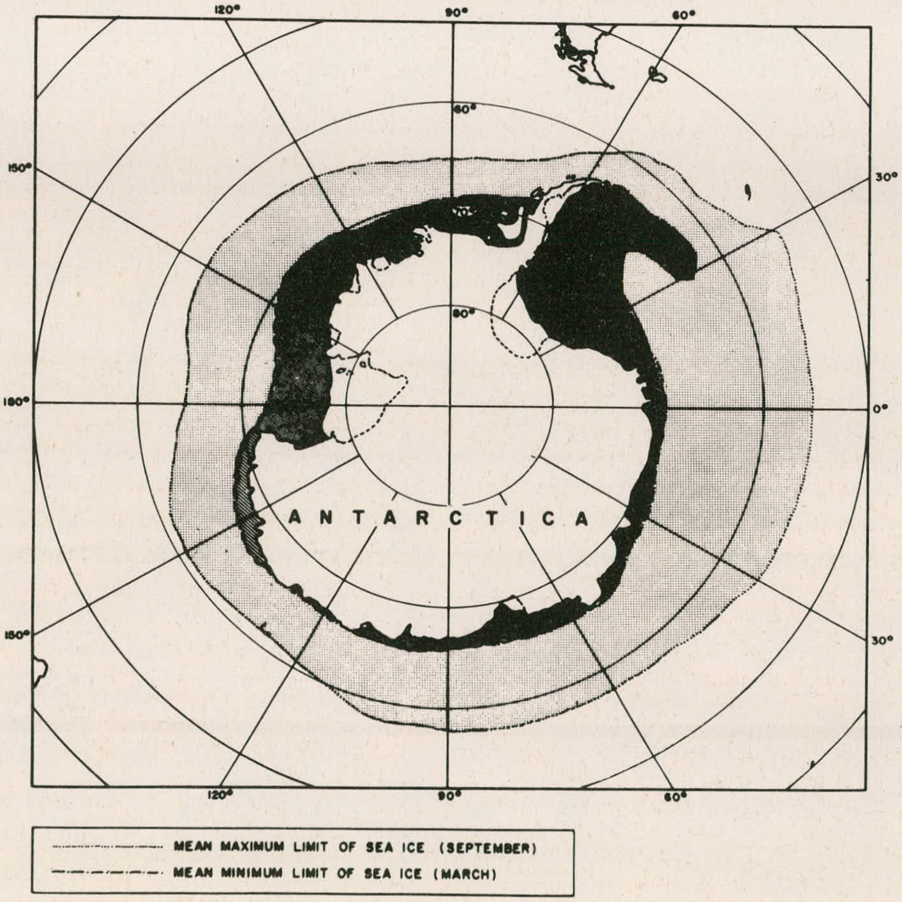

Fig. 2A. Mean maximum and minimum limits of sea ice in the Antarctic. (From Mellor, 1964.)

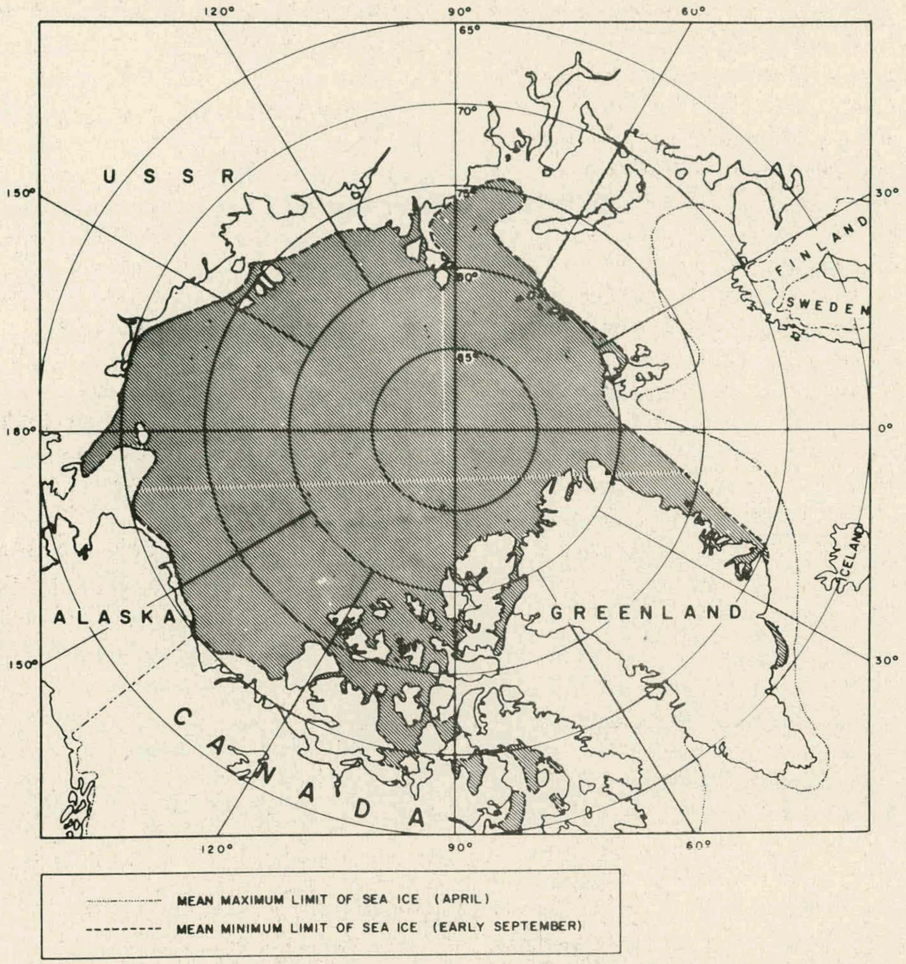

Fig. 2B. Mean maximum and minimum limits of sea ice in the Arctic. (From Mellor, 1964.)

1930s, corresponding to the period of maximum warmth. In response to the recent cooling trend, for example, Green (1973) reports an increase in days with snow lying at Balmoral, Scotland (283 m) from 51 days for 1951-52 to 1960-61 to 70 days for 1961-62 to 1970-71, with the major increases occurring in spring and autumn. Investigations in the equatorial mountains of New Guinea show that the rising snow line associated with an inferred warming of 0.6°C over the last century has caused a retreat of the glaciers on Mt. Jaya and disappearance of several small ice bodies on other mountains in Irian Jaya (western New Guinea) (Peterson, Hope, and Mitton, 1973; Allison and Kruss, 1977).

In high latitudes, where snowfall is limited by the low vapor content of the air due to low temperatures, warmer winters are generally more snowy. At Barrow, Alaska, for example, for the 30 winters between 1946 and 1976 there is a −0.57 correlation between freezing degree-days and snowfall during December to February. In Labrador-Ungava and Keewatin, snowy winter months tend to be associated with an average temperature departure of +1° to 3°C (Brinkmann and Barry, 1972). Occasionally, incursions of very warm air in winter may cause rain to fall even in arctic latitudes, and this forms ice layers in the snowpack which may be disastrous for populations of lemmings and voles and even for caribou and reindeer herds.

While increased snowfall will provide a deeper snowpack, the duration of snowcover is likely to be only marginally affected, at least in the arctic. Weller and Holmgren (1976) show that the snowmelt period is typically five to seven days and requires only 2 to 3 kJ/cm^2 for the melt process. There is also some interaction between snowpack depth and the development of the active (unfrozen) layer in the ground in summer. Simulation studies, carried out by S. I. Outcalt and C. Goodwin under the United States International Biological Programme (IBP) Tundra Biome Program, indicate that with snow depths increased by three to five times there is decreased cold penetration into the ground and some delay in the melt. However, the total active layer thickness developed during the summer is primarily determined by soil moisture and is unlikely to be affected significantly by variations in spring snow depth.

Sea Ice

Temperature and any related snowfall changes will affect sea ice thickness and extent. Some estimates of the likely magnitude of these changes can be made from empirical and modeling studies.

Observational evidence suggests the importance of snow cover persisting on the arctic pack ice through early summer until the solar radiation has passed its maximum and incipient melt has resulted in an increase in stratus cloud cover. Chernigovskiy (1967) shows that the net radiation available to a bare ice surface in the arctic is five times that of ice with an only partially-melted snow cover. Nevertheless, the model of Maykut and Untersteiner (1971) demonstrates that the thickness of sea ice remains essentially unchanged for snow covers of between 0 and 70 cm. Bare ice allows the ice to thicken by bottom accretion, although losses by ablation in summer are large. The ice only thickens significantly in the model for snow depths in excess of 100 cm.

Simple empirical estimates of changes in ice thickness due to warming can be based on the Zubov-Bilello relationships to freezing degree-days (Bilello, 1961). Under present winter conditions at Barrow, Alaska, with a freezing degree-day total of 4833°C, Zubov's equation gives an ice thickness of 173 cm and Bilello's gives a value of 247 cm. The two expressions indicate that a 5°C rise would lead to a reduction of 15 to 17% in ice thickness, and a 10°C rise would produce a reduction of 35 to 40%.

Analogous empirical relationships have been developed for late summer retreat of the pack ice margin off Barrow by J. C. Rogers. As shown in Fig. 3, there is a correlation of 0.81 between distance to the limit of 4/8 ice concentration north of Point Barrow on September 15 and the thawing degree-day total (see Barry, 1977). The mean distance for 1953 to 1975 was 155 km. A rise in mean summer temperature of 5°C would increase this to 600 km and a rise of 10°C to 1000 km. For the antaractic, Budd (1975a) calculates that a 1°C change in annual mean temperature corresponds to a 70-day variation in the duration of sea ice and a 2.5° latitude variation in maximum extent at the Antarctic Circle. It is likely that nonlinearities would act to augment the simple temperature effect. Retreat of the antarctic ice two weeks earlier than usual would cause a 7% increase in ocean heating, for example (Fletcher, 1969), which could amplify the initial effect by ocean-atmosphere feedbacks. However, Budd's results indicate that the feedback may *not* be of the unstable type.

Sedimentary records from the arctic basin indicate that the pack ice cover has been present for at least the last 700,000 years (Clark, 1971). There is, however, evidence from the distribution of pumice and driftwood in the Canadian arctic archipelago of more open conditions during the Thermal Maximum (Hypsithermal) about 5000 years before the present (B.P.) (Blake, 1970).

Several model experiments have been performed to assess the possible atmospheric effects of an ice-free Arctic Ocean (Donn and Shaw, 1966; MacCracken, 1970; Fletcher et al., 1973; Newson, 1973;

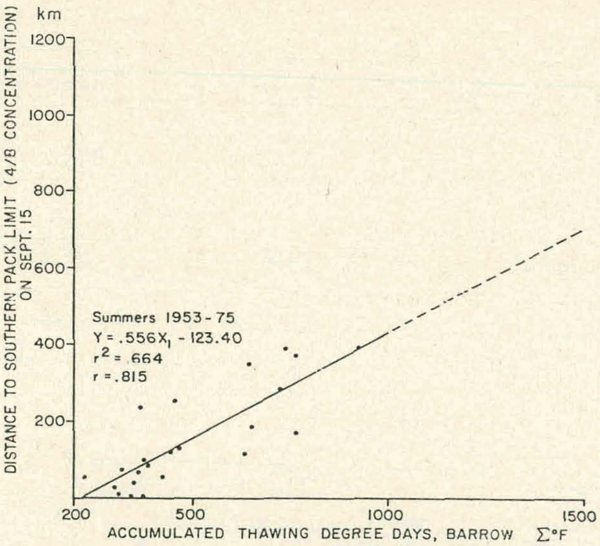

Fig. 3. Distance to 4/8 ice concentration September 15 off Point Barrow versus accumulated thawing degree-days. Data for 1953 to 1975. Projections illustrate the possible implications of higher summer temperatures, such as might occur in association with increasing CO_2 levels, assuming that a linear relationship obtains. (Prepared by J. C. Rogers, Institute of Arctic and Alpine Research, Boulder, Colo.)

Warshaw and Rapp, 1973). The results all show major warming of the lower troposphere over the arctic basin and cooling over continental areas in middle latitudes. However, these simulations do not take account of the likely ocean-atmosphere couplings in terms of oceanic advection. Empirical estimates by Drozdov (cited by Lamb, 1971), based on +5° C temperature anomalies in the arctic, suggest large precipitation increases over most of the arctic and in low latitudes but decreases over middle-latitude continents, in both warm and cold seasons of the year.

Permafrost

Quantitative assessments of the effects of a warming on temperatures in permanently frozen ground are provided by the work of Lachenbruch, Greene, and Marshall (1966) in Alaska. The zero isotherm occurs at 395 m (1200 ft) on Cape Thompson. Lachenbruch and colleagues show that, with an equilibrium profile at $\partial T/\partial z$ for the present surface temperature of −5° C, 25% of the present thickness is unstable (Fig. 4). A warming of 2° C over 75 to 100 years is inferred. However, he notes that an imbalance of 335 J $\mathrm{cm^{-2} yr^{-1}}$ (80 cal $\mathrm{cm^{-2}\ yr^{-1}}$) over 60 years would melt only about 0.6 m of ice at the surface.

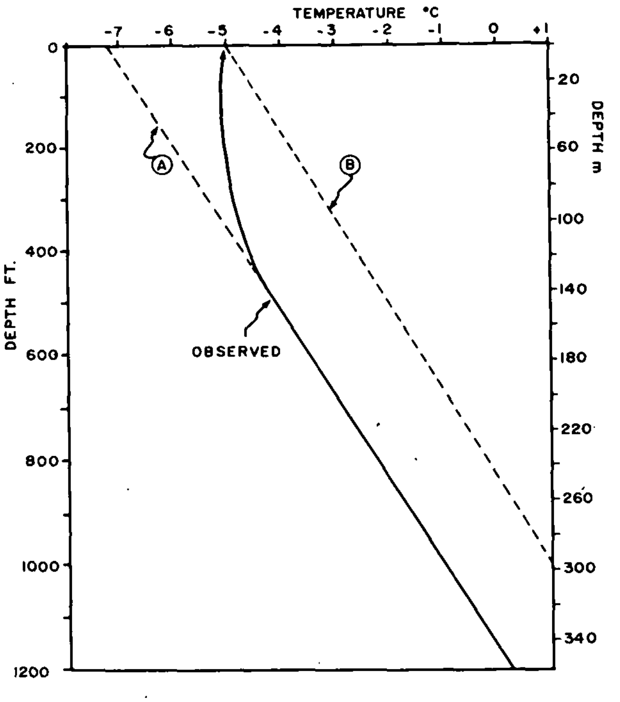

Fig. 4. Ground temperature profile at Cape Thompson, Alaska (after Lachenbruch, Greene, and Marshall, 1966). (A) shows the hypothetical original profile prior to the recent warming. (B) shows the equilibrium profile for the present conditions.

A warming trend would be most significant, in terms of effects on soil moisture and vegetation, in the areas of discontinuous permafrost along the southern margins. Thie (1974) finds general melting and retreat of the permafrost over the last 120 years in Manitoba, for example.

Ice Sheets

The effects of a warming on large ice sheets are likely to be complex. As outlined above, warming may augment precipitation in polar latitudes, giving more accumulation; but, if the net effect were to cause a negative mass balance, world sea levels would begin to rise. Mercer (1968) has suggested that higher summer temperatures during the Last Interglacial (125,000 B.P.) caused destabilization and disintegration of most of the west antarctic ice sheet and a 5 to 6 m rise in sea level. The temperature increase during this period was 1° C in the western North Atlantic (Sancetta et al., 1972), although regional increases in mean annual temperature ranged up to 6° to 7° C in Russia and central Siberia according to pollen evidence (Frenzel, 1973, p. 123). Hughes (1975) has examined possible mechanisms for such disintegration. He suggests that instability of ice streams entering the Ross Sea initiated a similar event prior to 10,000 B.P. Disintegration could be triggered either by the warming itself, by interaction of the rising sea level with the grounded ice, or by internal ice processes. Hollin (1970) postulates that such major ice sheet "surges" would lead to 10 to 30 m rises of sea level in less than 100 years, although unequivocal geological evidence of such events has not yet been identified.

The rate of propagation of a surface warming to the base of a 2000-m-thick ice sheet has been calculated using a thermodynamic model by M. D. Jenssen (personal communication, 1977). In response to a 10° C surface warming, the basal temperature rises 0.4° C after 5000 years and 5.9° C after 50,000 years (Fig. 5). This suggests that there should be a considerable lag in any direct response by the ice to a climatic warming, which does not appear to have been the case. Surge behavior in glaciers does not appear to be climatically induced (Budd, 1975b), although decoupling of the ice from its bed by basal

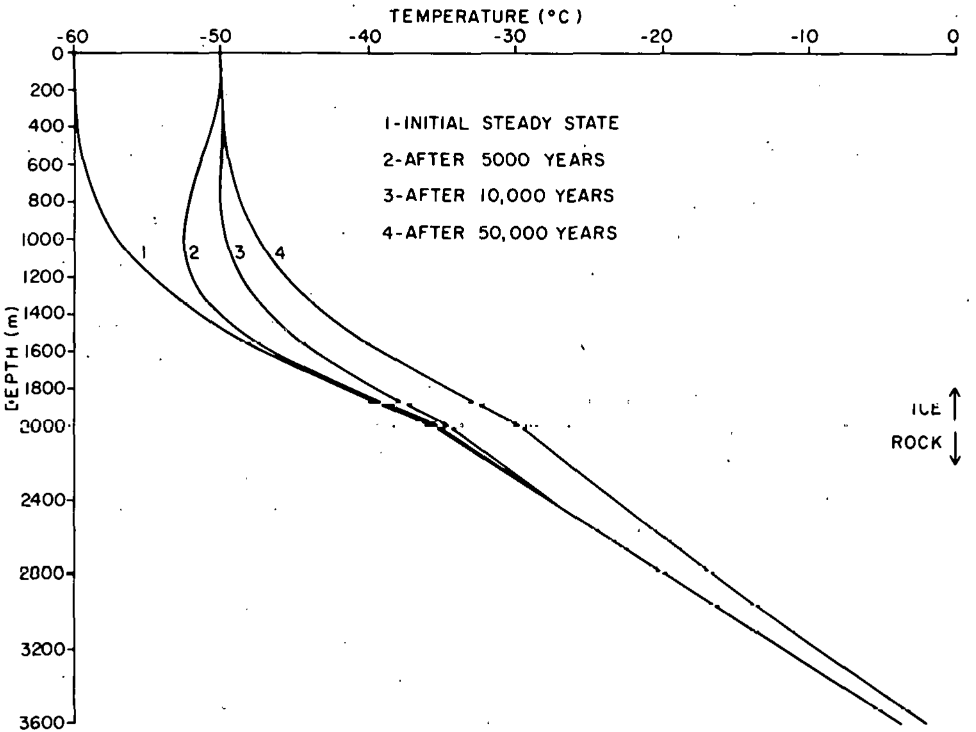

Fig. 5. Temperature profiles in ice computed by a thermodynamic model (M. D. Jenssen). The assumed conditions are 0.2 m/year of accumulation, frictional heating of 0.01°C/m at the base, geothermal heat flux of 0.02°C/m: thermal diffusivities are 42 m^2/year for ice and 32 m^2/year for rock. (Prepared by M. D. Jenssen, University of Melbourne.)

meltwater may be a primary causal factor (Weertman, 1969). Indeed, the magnitude of surges may be inversely related to the general temperature level. Denton and Karlén (1977) report that surges of glaciers in the St. Elias Mountains have been successively smaller in this century than in the last, matching the general glacier retreat.

OUTSTANDING PROBLEMS

There are several areas where our data or understanding of climate-cryosphere interactions are at present inadequate to evaluate the effects of increased atmospheric carbon dioxide. Major deficiencies exist in data on snow cover extent and depth. Satellite records of snow cover are available only since 1967, and it is not yet possible to assess the typical range of year-to-year variability under the present climate (Kukla and Kukla, 1974; Wiesnet and Matson, 1976). The age structure, thickness, and mass balance of the Arctic Ocean ice are still poorly known (Koerner, 1973). Comparison of ice thickness data from the *Sedov* (1938 to 1940) with those of Nansen from the *Fram* (1893 to 1895) suggest a significant thinning (Dunbar, 1976), but this conclusion may be invalid in view of the complexity of the ice composition from year-to-year. (A decrease of 360 cm to 215 cm is cited by Dunbar, but tables in Nansen's account indicate 260 cm in 1894.) There are few published data on the thickness of permafrost or on changes in ground ice temperature and thickness during this century. The decay and disappearance of areas of ground ice is perhaps less known than the occurrence of its formation in new sediments. Likewise, the minimum extent of land ice during previous warmer episodes is generally much less well-documented than are periods of maximum extent. The long-term mass balance of the major ice sheets is also known with less certainty than that of many smaller glacier systems.

Many research problems remain. A major question concerns the effects of a temperature rise on the melting of glaciers and ice sheets, and therefore the effects on world sea level. Sea levels rose about 5 to 6 m during the Last Interglacial (120,000 B.P.) and fell to about −120 ± 40 m during the Last Glacial Maximum (20,000 to 18,000 B.P.). Since then, sea level has risen to its present position, although there is dispute as to the precise character of this trend. Moreover, the eustatic change related to changes in ice volume is greatly complicated by isostatic responses to ice unloading, by meltwater loading of the ocean floor (Walcott, 1972), and by the associated redistribution of matter in the crust, which act together to change the geoid (Clark, 1976). Recent calculations by Farrell and Clark (1976) indicate that sea-level changes during the last 5000 years (since the Hypsithermal) can be explained without invoking eustatic effects. Changes in eustatic sea level were probably less than 1 m during this time period.

A major climatological question that needs to be resolved before cryospheric effects can be properly evaluated is the temperature response to increased carbon dioxide to be anticipated in high latitudes. Simulations with the Geophysical Fluid Dynamics Laboratory (GFDL) general circulation model (Manabe and Wetherald, 1975) have not assessed seasonal effects, but the temperature structure in the arctic changes markedly with season. In winter, there is an 85% frequency of a surface-based temperature inversion with an average intensity of 8° to 10°C. In summer, there is a weaker upper inversion with a 40% frequency. The winter inversion is determined by radiation, advection, and subsidence; while the summer one is primarily an advective feature (Vowinckel and Orvig, 1967). A CO_2-induced global warming is likely to be associated with strong warming in the surface layer and cooling at the inversion level. This would weaken the inversion in winter, but the effect in summer is less certain.

Finally, it is worth noting that certain types of ice mass in critical locations can provide valuable integrated indicators of climatic trends. Shallow plateau ice caps in high latitudes would be highly sensitive to warming trends as a result of changes in equilibrium line altitude, for instance (see Bradley, 1975). Ice patches in equatorial mountains should also be monitored.

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Technical Fixes for the Climatic Effects of CO_2

Freeman J. Dyson, Institute for Advanced Study, Princeton, New Jersey
and Gregg Marland, Institute for Energy Analysis, Oak Ridge
Associated Universities, Oak Ridge, Tennessee

INTRODUCTION

There already exists a sizable literature on this subject, and we are attempting here this week to further assess and to predict the effects of increasing atmospheric carbon dioxide on the climate of the earth. Such attempts are useful and necessary but, as brought out in the preceding papers, they run into formidable technical difficulties. The mean global temperature change caused by a given quantity of CO_2 is subject to great uncertainty; and the effects of CO_2 on local and time-variable phenomena (which may be crucially important to agriculture and other human activities) are more uncertain still. It is even possible that the rise in CO_2 will be, on balance, beneficial to mankind, especially in reducing climatic extremes in very cold regions. The prevailing opinion, however, is that the dangers greatly outweigh the potential benefits. Nonetheless, it seems inevitable that we shall continue for decades to burn fossil fuels and to increase the level of atmospheric CO_2, without knowing with any degree of certainty the consequences of our actions.

It seems the time is now ripe to ask another question. Suppose that the level of CO_2 does continue to rise and we do indeed run into an acute ecological disaster. Will it then be possible for us to halt or reverse the rise in CO_2, within a few years, by means less drastic than the shutdown of industrial civilization? Here it is important to distinguish between the long-term and short-term responses to a threat of catastrophe caused by CO_2. The long-term response, if such a catastrophe becomes imminent, must be to stop burning fossil fuels and convert our civilization to nuclear or solar-based fuels. However, a worldwide shift from fossil to nonfossil fuels could not be carried out in a few years. A time scale of decades is more likely. Such a response might well not be rapid enough to avoid a climatic or economic disaster.

In a paper presented at the 1976 Dahlem Workshop on Global Chemical Cycles and Their Alterations by Man, Harvey Brooks (1976) noted the problems of societal response to long-term environmental threats. Brooks envisioned the limitations of dealing with such distant threats as fivefold:

1. The predicted threats are subject to a high degree of uncertainty, raising problems in establishing a consensus for action.
2. Biological and ecological impacts are even more uncertain than the physical and chemical effects.
3. When the effects are long-term and cumulative, the costs of delaying action appear small compared with the immediate costs of action.
4. There are real trade-offs between the interests of current and future generations.
5. The threats can seldom be dealt with by a single discrete action but require sustained effort on a worldwide basis.

One can easily conclude that if a CO_2 disaster really looms on the horizon, we are likely to proceed well up to it before a "consensus for action" is achieved and then ask, "What do we do now?" The problem at that point will be one of arresting further atmospheric CO_2 growth while the energy delivery system is switched to alternate fuels, or perhaps even of trying to bring about a reduction in the atmospheric CO_2 level. The purpose of this paper is to present order-of-magnitude estimates of the feasibility and size of emergency plans which might be proposed. It is highly unlikely that the particular programs which we discuss now will ever be implemented. Our discussion is necessarily unrealistic and enormously oversimplified. From our present perspective, some of the possibilities assume science-fiction-like aspects. Our intent is to begin a process of mental preparation so that we can realistically visualize our options if ever the danger of catastrophe from CO_2 accumulation becomes acute. To have plans ready is a form of life insurance, valuable even if the threatened catastrophe never happens.

A quick review of the carbon cycle emphasizes that our concern is focused on the accelerated flux of carbon from fossil organic matter to the relatively small reservoir of the atmosphere. Since natural exchange times are frequently large in comparison to the time scale of man's fossil fuel burning, this results in a buildup in the atmosphere, which, though of short duration in a geologic sense, may have important consequences for man. A solution to this dilemma must focus on ways to either accelerate carbon transfer from the atmosphere to other reservoirs of carbon or to bypass the atmosphere and place the CO_2 directly into other reservoirs.

We wish to be prepared to sequester a quantity of carbon comparable with the current rate of growth of atmospheric CO_2. If the withdrawal were to begin now (1977), the quantity would be 2.5×10^9 metric tons of carbon per year. Current projections (Baes et al., 1976) indicate that in the year 2000 the rate of fossil fuel use is likely to be between two and four times the present rate. The required rate of CO_2 withdrawal between now and the year 2000 will therefore lie in the range from 2.5×10^9 to 1×10^{10} tons of carbon per year. We take the value 5×10^9 tons per year as illustrative of the magnitude of the problem that would arise if the withdrawal plan had to be implemented at some unspecified time within the next 25 years. Beyond the year 2000, the problem may or may not become more severe, depending on a multitude of unpredictable contingencies. We assume here that a rational decision on the global climatic effect of increasing CO_2 can be made long before a sizable fraction of the total world fossil fuel resource is consumed, and we are thus committed to a significant and essentially permanent displacement of the equilibrium atmospheric concentration.

Obvious candidates for the disposal of CO_2 are the deep ocean, the long-term biosphere, and the lithosphere (solid carbonates). No rational scheme for transfer to the lithosphere is presently apparent. The Solvay process does react common salt, in the form of a saturated brine, with ammonia and carbon dioxide to precipitate sodium bicarbonate, but one must burn $CaCO_3$ to recover the ammonia. The remaining options are the deep ocean and the biosphere.

PHYSICAL TRANSFER OF CO_2 TO THE DEEP OCEAN

It has been demonstrated that the rate of CO_2 accumulation in the atmosphere is on the order of half the rate at which fossil fuel burning releases CO_2 to the atmosphere (e.g., Bacastow and Keeling, 1973). At least a large fraction of the remainder seems to be taken up by the surface ocean. The oceanic carbon reservoir is immense with respect to fossil fuel carbon release, and the released carbon could be accommodated in the water mass with little change (see, for example, Broecker, Li, and Peng, 1971; Siegenthaler and Oeschger, 1977). The problem is that the mixing rate for the deep ocean is on the order of 500 to 2,000 years. What we need is a scheme for accelerating the downward mixing of carbon into the deep ocean.

Marchetti (1976) claims that "the basis for a technologically and economically feasible operation does exist" for putting CO_2 directly into the deep ocean. He envisions that CO_2 would be collected at large installations such as power stations, blast furnaces, and, ultimately, hydrogen production facilities. He further suggests that, with slight management of coal and oil, as much as 50% of the carbon could be removed by CO_2 scrubbers similar to those in use for CO_2 manufacture. The CO_2 would be compressed to a liquid at 60 to 70 atm at room temperature, transported by pipeline, and injected to the deeper ocean via natural circulations such as the Mediterranean Intermediate Water.

Marchetti claims that, in fact, perhaps ten times the entire current CO_2 production could be injected into the Mediterranean Outflow water with an acceptable increase in CO_2 concentration. We are not prepared to speculate as to whether this is environmentally acceptable or even technically feasible. However, we would point out that—accepting

Marchetti's number for the volume of water flowing out through the Gibraltar Straits (10^{14} tons/year)—the current emission of fossil fuel carbon is already approximately twice the carbon flux in the Mediterranean Outflow water.

For this operation, the cost to the consumer, Marchetti reports, would be 10% of the fuel value for recovery of 50% of the CO_2. The *Kirk-Othmer encyclopedia of chemical technology* (Standen, 1964) suggests that the energy demands for CO_2 recovery may be much greater. Mr. Donald Hess of the C. I. Girdler Corporation (personal communication, 1977), a manufacturer of CO_2 scrubbing equipment, speculated that the majority, if not all, of the power plant output could be required for CO_2 processing. The discrepancy may be one of scale, since current CO_2 production facilities are small by comparison; but this important issue should be addressed more carefully.

Marchetti cites Nordhaus as proposing a similar collection scheme in which a high-density liquid CO_2 slug would then be allowed to sink to and rest on the sea bottom. The same energy requirement question is obviously encountered.

TRANSFER OF CO_2 TO THE LONG-TERM BIOSPHERE

The magnitude and rate of biosphere response for CO_2 uptake is dramatically evident in the large amplitude and short lag time of the CO_2 annual cycle observed at, for example, Mauna Loa (Pales and Keeling, 1965). There are those (Bacastow and Keeling, 1973) who maintain that the rate of growth of atmospheric CO_2 is presently being limited by an increase in uptake by the biosphere; and it has, indeed, been demonstrated that many plant species are CO_2-limited. On the other hand, it has been argued (Degens, 1977) that forest clearing, especially in the tropics, is currently leading to a net decrease in the mass of the long-term biosphere and that this has contributed to the atmospheric growth of CO_2. Other human activities such as heavy use of fertilizer and plowing back of crop residues into the ground, the eutrophication of lakes and estuaries, and the harvesting of lumber and pulp may be significantly altering the carbon cycle. In summary, there is insufficient evidence to decide whether the carbon content of the biosphere has increased, decreased, or remained stationary in response to the manifold human activities of recent decades.

Thus, while the present carbon content of the biosphere is only roughly known, the carbon in living plants is believed to be roughly equal to the total atmospheric carbon. Moreover, the net quantity of carbon photosynthetically transferred (gross photosynthesis minus respiration) from the atmosphere to the biosphere is approximately 5×10^{10} tons carbon/year (SCEP, 1970, p. 161), about 20 times the annual increase of atmospheric carbon. It may be possible, in case of a worldwide emergency, to plant enough trees and other fast-growing plants to incorporate the excess CO_2 and retard the annual atmospheric increase. This would have to be viewed as an emergency scheme to provide a short-term response to hold the CO_2 at bay while the shift to nonfossil fuels was implemented. The newly grown plants would have to be regarded as a "carbon bank" and should not be used as fuel except in places where they could directly replace fossil fuels.

There are two chief ways in which carbon can be transferred from the atmosphere into long-term storage in the biosphere: (1) by growing long-lived trees which are left unharvested or (2) by growing short-lived plants which are converted to humus or allowed to accumulate in artificial peat bogs. Either method requires large areas of land. Some of the planting may be done on land which is at present unproductive, but much of it may have to be done on existing agricultural or productive forest land with some changes in the management of crops and fertilizers. There is no necessary incompatibility between increased yields of food and forest products and increased inventories of carbon in trees and humus.

A great variety of trees could be grown on marginal land in various parts of the world. As an example of a tree crop suited to the southeastern United States, we consider American sycamore as grown experimentally by the School of Forest Resources of the University of Georgia (Steinbeck and Brown, 1976). This yields about 750 metric tons carbon km^{-2} $year^{-1}$. We do not know how much vegetation was destroyed in the clearing of the land before the trees were planted, nor do we know the effect of the plantation on the inventory of carbon in humus in the soil. If we assume that we can achieve a net increase of fixed carbon of 750 tons km^{-2} $year^{-1}$, then we should need about 7×10^6 km^2 of land to counter the atmospheric CO_2 growth with this particular crop—5% of the world's land area.

Of course, the area suitable for American sycamore plantation is much less than 7×10^6 km^2. The area of the United States suitable for some kind of tree plantation is estimated to be about 7×10^5 km^2 (Fraser, Henry, and Vail, 1976). This is the area that has at least 50 cm of annual rainfall and is at present unproductive. Much of this area is already covered with low-grade forest, and detailed surveys would be required to determine what fraction of it should be replanted and what fraction should be merely fertilized and managed so as to promote rapid growth. The United States could thus provide about one-tenth of the required growth of trees. The total area of forest in the world is about 5×10^7 km^2. To absorb 5×10^9 tons carbon/year from the atmosphere, it would be necessary to achieve an average worldwide increase in long-term forest growth of 100 tons carbon km^{-2} year^{-1}, or a growth of 750 tons km^{-2} year^{-1} on one-seventh of the area. The additional growth should be in species that live 50 years or longer, so that their carbon is not returned to the atmosphere by decay before it can be safely harvested. Most commercial species are sufficiently long-lived for this purpose.

An alternative to tree planting is the intensive cultivation of aquatic plants. Water hyacinths are especially interesting because of their rapid growth and hardiness (Lecuyer and Marten, 1976; Poole and Williams, 1976). They yield about 6000 tons carbon km^{-2} year^{-1}, about eight times as much as American sycamore. They can only be grown on flooded land with abundant fresh water, or in canals or ponds. To fix 5×10^9 tons carbon/year we would need an area of 8×10^5 km^2. The total area of tropical swampland that might, with some adaptation, be suitable for water hyacinths is about 25×10^5 km^2 (Poole, 1974). Most of this land is in South America, Africa, and Southeast Asia. The United States has only 2×10^5 km^2 of such land, and the greater part of this is the Mississippi floodplain which has long ago been drained and used for farming. Still, there is enough wetland left in the United States to make a nonnegligible contribution to the amelioration of the CO_2 problem.

The intensive cultivation of water hyacinths has been proposed as a source of pipeline-quality gas (essentially pure methane) obtained by digesting the plants anaerobically in large fermentation tanks (Lecuyer and Marten, 1976). If the plants are to be used as a sink for atmospheric CO_2, it is essential to store them after harvesting in a semipermanent fashion so that digestion to methane does not occur. We would have to duplicate artificially the conditions of acidity and sterility that cause vegetation in natural peat bogs to form peat rather than methane. Experimentation is required to find out whether artificial peat bogs on the required scale are either feasible or environmentally acceptable.

There are many types of soft plants other than water hyacinth which could be used to build up carbon reserves in the form of humus, either in cropland or in natural swamps and wetlands. Some of these plants, like sugarcane, would have commercial value as food crops or energy crops (Alich and Inman, 1976) in addition to their value as carbon-fixers. We do not attempt here any assessment of the various possible alternatives.

One of the major problems in building up a "carbon bank," either with trees or with soft plants, is that the vegetation does not consist of pure cellulose but also incorporates substantial quantities of the elements nitrogen, phosphorus, and potassium. In most parts of the world, natural forests grow without artificial fertilization, but any plan for forced, rapid growth of trees or plants will require the use of fertilizers. To estimate the magnitude of this problem, we use the input-output data from the experimental planting of American sycamore seedlings in Georgia (Steinbeck and Brown, 1976). For a two-year cycle yielding a crop of 3000 tons dry weight/km^2 (this means about 1350 tons of fixed carbon), the quantities of fertilizer input were 24 tons nitrogen, 2.2 tons phosphorus, and 4.5 tons potassium; and the quantities of these elements in the harvested crop were 4.60 tons nitrogen, 0.56 tons phosphorus, and 2.14 tons potassium. Roughly speaking, these input quantities set an upper bound and the output quantities set a lower bound to the quantities of fertilizer required for an intensive rapid growth program. The larger the area planted, the less need there would be for heavy fertilization. If we use the output quantities to estimate the requirements, then for fixing 5×10^9 tons carbon/year we would need to consume fertilizers at an annual rate of 1.7×10^7 tons nitrogen, 5×10^6 tons P_2O_5, and 10^7 tons K_2O. These numbers represent roughly 40% of the world production of nitrogen and K_2O and 20% for P_2O_5 in fertilizers for 1974 (UN, 1975). The availability of fertilizers will probably be the critical factor limiting the scale and speed of carbon fixation. Using numbers from Leach (1976), some 1.4×10^{12} MJ—on the order of 1% of current world energy consumption—would be required for manufacture of the fertilizer.

In fact, we may already be doing just this experiment on a limited scale. At Ralph Rotty's suggestion, we have looked at the rate of lumber harvest in the world. The assumption is, of course, that the lumber harvested is replaced by a new growth of trees and that the lumber is (or could be) sequestered out of the carbon cycle for some significant length of time. The United Nations (1975) reports that 2.5×10^9 m^3 of roundwood was harvested in 1973. This is solid volume, without bark, although admittedly some of it is used for fuel. Seventeen percent of it shows up in their statistics for sawn wood. If we take 0.60 as an average density for dry wood and 0.45 as the carbon content of dry wood, this comes to 0.68×10^9 tons carbon/year, about one-seventh of the current fossil fuel carbon flux.

BIOLOGICAL TRANSFER OF CO_2 TO THE DEEP OCEAN

In the world ocean, a sizable fraction of the downward transport of carbon takes place as biologically created detritus. We consider now a scenario in which the surface waters are supplied with whatever happens to be the biologically limiting nutrient (phosphate and/or nitrate), and hence the growth of organisms and settling of carbon is much accelerated. Schindler and his co-workers (1972) have shown that for small Canadian shield lakes, the invasion of atmospheric CO_2 supplies sufficient carbon to produce an algal bloom in proportion to the amount of phosphorus and nitrogen available, and invasion rates as high as 400 tons carbon km^{-2} $year^{-1}$ have been measured for the ocean (Schindler, 1975).

Just such a scheme has been described by Broecker (1977) in a draft paper circulated to conference participants. Broecker envisions spreading phosphorus (as Na_3PO_4) and fixed nitrogen (as $NaNO_3$) on the ocean. If carbon, nitrogen, and phosphorus eventually sink and accumulate in the ratio 120:15:1, then the fertilizer requirement for incorporating 5×10^9 tons carbon/year comes to 7.2×10^8 tons of nitrogen and 1.1×10^8 tons of phosphorus per year. The fertilizer requirement is on the order of 50 times greater than suggested for growing American sycamore in Georgia, apparently because of the much higher nitrogen and phosphorus content of algae as opposed to that of woody trees. If we again use Leach's (1976) number of 80 MJ/kg of nitrogen as fertilizer, we will consume 58.0×10^{12} MJ of energy for nitrogen production (54 quads, or three-fourths of current U.S. energy consumption). On the other hand, if the oceans can, like some lakes (Schindler, 1977), correct nitrogen deficiencies by selecting for nitrogen-fixing algae, it may be possible to get by supplying only phosphorus. At 14 MJ/kg phosphorus (Leach, 1976), the energy requirement for fertilizer would be 1.5×10^{12} MJ, on the same order as previously calculated for fertilizing trees. This is still ten times the 1974 world production rate of phosphate fertilizers, but it begins to look more manageable. Larry Small (personal communication, 1977) has pointed out that in such a program it might be necessary to supply silica as well.

This experiment is also already being done on a small scale. The phosphorus runoff from the United States in 1968 amounted to 4.4×10^5 tons—60% from municipal sewage (SCEP, 1970). Presumably much of this ends up in lakes and estuaries. If the current world phosphorus runoff is on the order of five times the 1968 U.S. level, we may be already sequestering 2% of the fossil fuel carbon in eutrophic lakes and estuaries.

Broecker goes on to calculate that phosphorus is currently being delivered to the surface ocean by upwelling at a rate of 8.4×10^7 tons phosphorus/year—about equal to what he proposes be added. While the excess phosphorus would produce a considerable change in ocean ecology, especially if we envision large populations of undesirable nitrogen-fixing species, Broecker does not see an ecologic disaster. The most serious effect might be the sizable depletion of deep-sea oxygen.

OTHER TECHNICAL FIXES

There are at least two additional procedures that one might consider for controlling atmospheric CO_2. We describe first a scheme for physically extracting CO_2 from the air, and second an alteration outside the carbon cycle that has been proposed for balancing the CO_2 effect.

Originally conceived as a means of providing liquid motor fuel for a mobile army, a technically feasible scheme exists for using nuclear-generated electricity to extract CO_2 from the atmosphere and convert it into a liquid hydrocarbon (Beller and Steinberg, 1965; Baron and Steinberg, 1975). To be of benefit in controlling CO_2, the product methanol would have to be stored or used directly to displace fossil fuels. In simplest terms, CO_2 is absorbed from the atmosphere on a molecular sieve, hydrogen is generated by the electrolytic dissociation of water, and the two are reacted at 280 kg/cm^2 (4000 psi) and 300°C over a zinc oxide catalyst to produce methanol.

The energy requirement for this process is 11 kWhr(e) per kilogram of methanol, or 8 kWhr(e) per kilogram of CO_2 removed. To remove 5×10^9 tons carbon/year thus requires 40×10^9 MWhr(e)/year or 5.7×10^6 MW (5700 1000-MW plants) of dedicated nuclear capacity operating at 80% capacity factor. The process may have merit as a storage method for off-peak nuclear, solar, or wind electric energy or as an ultimate source of liquid fuels, but it cannot meet our criteria for short-term amelioration of a CO_2 crisis. Over a long period of time, however, off-peak electricity might be used in this mode to help slowly bring down an excessive CO_2 concentration.

As a final possibility for a technical fix, we recognize the suggestion of balancing the CO_2-induced heat retention with artificial alteration of the earth's albedo. Such schemes are fraught with difficulty in that both the primary effects and the potential side effects are poorly understood (see, for example, SMIC, 1971). There has been much debate on the impact of anthropogenic dust and aerosols, and our intent here is only to introduce the issue and make a couple of brief observations.

One can talk about painting all roads and parking lots white or floating styrofoam plates on the oceans; since the majority of the earth's reflectivity now occurs in the atmosphere, however, it seems most useful to direct our attention there first. The earth reflects about 28% of the incoming solar radiation: 19% from clouds, 6% from air, dust, and haze, and 3% from the surface (Rotty, 1975). Another 22% of the incoming radiation is absorbed by water, dust, haze, and clouds in the atmosphere, and a fraction of this is reradiated to space without ever interacting at the surface. A small increase in the atmospheric backscatter could have a sizable impact on the atmospheric energy balance.

There is also evidence to support the contention that it is inherently possible to decrease the tropospheric temperature through the injection of dust into the lower stratosphere. Mount Agung on the island of Bali erupted on March 17, 1963 and injected a large quantity of dust into the stratosphere at a height of 22 to 23 km (Dyer and Hicks, 1968; Schneider and Mass, 1975). The transmission of direct solar radiation dropped nearly 2% at Mauna Loa Observatory, and the effective decrease in the solar "constant" was probably on the order of 0.5%. While other minor eruptions may have had an effect, the Mauna Loa measurements took seven years to return to normal (Ellis and Pueschel, 1971). Newell and Weare (1976) show that this eruption can be associated with a 0.5° C drop in the mean temperature of the tropical troposphere.

In 1968 the direct solar radiation at Mauna Loa was still 1% below pre-Agung levels. The SCEP (1970) study group concluded that a fleet of 500 supersonic transports using 0.05% sulfur fuel would produce nearly comparable particle concentrations in the geographic regions of their maximum activity.

Flohn (1976) envisions that major earth surface changes, such as a 12% change in snow cover or a 20% change in arctic sea ice, would yield energy contributions somewhat smaller than a large-scale volcanic eruption.

SUMMARY

At this point we have provided order-of-magnitude estimates for a variety of potential technical fixes that have been proposed, seriously or otherwise, for dealing with a rising atmospheric CO_2 concentration, should that concentration rise ever be viewed as an imminent ecological emergency. While all of the possibilities are massive in scale, the biologically based scenarios seem to have the greatest possibility for utilization, and there is no law of nature which prohibits their application and success.

On the other hand, the massive scale of all technical fixes raises the question of whether practical application is really feasible. Rotty (1977) has demonstrated the extent to which CO_2 emission is distributed among many nations and the likely increased importance of more participants as time progresses. Recall that in our tree planting scenario only one-tenth of the required planting can be achieved by the United States alone.

Harrison Brown has proposed a series of syndromes for world disaster, one dealing with CO_2. He raises the very important question of whether developing nations, for example, would be willing, or able, to curtail fossil fuel burning or to implement expensive CO_2 control measures.

We raise a further issue. It seems probable that as the climatic effects of increased carbon dioxide become well defined, the changes will impact on various nations quite differently (Manabe, 1977). Some nations may perceive the change as being to their benefit—perhaps more humid conditions in the Middle East or warmer climates in Siberia; others will perceive the changes as being to their detriment—perhaps drier conditions in the American Midwest (Brown, 1977). Regardless of the total impact on humanity, it seems likely that global agreement will be difficult to achieve if the costs and benefits are unequally distributed on this political earth.

In the meantime, we continue to conduct what Revelle and Suess (1957, p. 19) described as a "large-scale geophysical experiment"; and we monitor its progress.

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Workshop Participants

Dr. L. Allen, Jr.
USDA-ARS-SWAP Project
304 Newell Hall
Agronomy Department
University of Florida
Gainesville, Florida 32611

Dr. Robert Bacastow
Scripps Institution of Oceanography
2314 Ritter Hall, A-020
La Jolla, California 92093

Dr. C. F. Baes
Chemistry Division
Oak Ridge National Laboratory
P.O. Box X
Oak Ridge, Tennessee 37830

Dr. David S. Ballantine
Division of Biomedical and Environmental Research
U.S. Energy Research and Development
Administration (U.S. Department of Energy)
Washington, D.C. 20545

Dr. Roger G. Barry
INSTAAR
University of Colorado
Boulder, Colorado 80302

Dr. W. Dwight Billings
Department of Botany
Duke University
Durham, North Carolina 27706

Dr. O. Bjorkman
Department of Plant Biology
Carnegie Institute of Washington
Stanford, California 94305

Dr. A. Björkström
Department of Meteorology
University of Stockholm
Fack, S-104 05 Stockholm
SWEDEN

Dr. T. Bramryd
University of Lund
Department of Plant Biology
Ecology Bldg., Helgonavagen 5
S-223 62 Lund
SWEDEN

Dr. Wallace S. Broecker
Geochemistry Laboratory
Lamont-Doherty Geological Observatory
Columbia University
Palisades, New York 10964

Dr. R. Burling
Institute of Oceanography
University of British Columbia
V6T 1W5 British Columbia
CANADA

Professor Melvin Calvin
Laboratory of Chemical Biodynamics
University of California
Berkeley, California 94720

Dr. Robert Cess
Laboratory for Planetary Atmospheric Research
State University of New York
Stony Brook, New York 11794

Dr. C. C. Delwiche
Soils and Plant Nutrition
University of California
Davis, California 95616

Dr. M. Drosdoff
Agronomy Department
Cornell University
Ithaca, New York 14853

Dr. William P. Elliott
NOAA—Air Resources Laboratories
8060 13th Street
Silver Spring, Maryland 20910

Dr. Arthur Fairhall
Department of Chemistry
University of Washington
Seattle, Washington 98195

Dr. Manuel E. Fiadeiro
Department of Geology and Geophysics
Yale University
New Haven, Connecticut 06520

Dr. Tzvi Gal-Chen
Department of Physics
University of Toronto
Toronto, CANADA M5S 1A4

Dr. Robert Garrels
Department of Geology
Northwestern University
Evanston, Illinois 60201

Dr. David M. Gates
Department of Botany
University of Michigan
Ann Arbor, Michigan 48105

Dr. Joris Gieskes
Scripps Institution of Oceanography
P.O. Box 1529
La Jolla, California 92093

Dr. Charles A. S. Hall
Ecology and Systematics
127 Langmuir Hall
Cornell University
Ithaca, New York 14853

Dr. Kirby Hanson
National Oceanic and Atmospheric Administration
Environmental Research Laboratory, R329
Boulder, Colorado 80302

Dr. Wilmot Hess
Environmental Research Laboratories
National Oceanic and Atmospheric Administration
Boulder, Colorado 80302

Dr. Martin I. Hoffert
Division of Applied Science
New York University
26 Stuyvesant Street
New York, New York 10003

Mr. Ernest Hughes
Analytical Chemistry Division
National Bureau of Standards
Room B 326
Washington, D.C. 20234

Dr. William J. Jenkins
Department of Chemistry
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

Dr. William Kellogg
National Center for Atmospheric Research
P.O. Box 3000
Boulder, Colorado 80302

Dr. Stephan Kempe
Scope Representative
Universität Hamburg
Geolog.-Palaontolog. Institut
2 Hamburg 13, Bundesetrasse 55
WEST GERMANY

Dr. K. K. Knoerr
School of Forestry and Environmental Studies
Duke University
Durham, North Carolina 27706

Dr. John Laurman
Division of Applied Mechanics
Durand Building
Stanford University
Palo Alto, California 94305

Dr. E. R. Lemon
Agricultural Research Service
North Atlantic Area—Northeastern Region
Bradfield Hall
Cornell University
Ithaca, New York 14853

Dr. R. S. Loomis
Agronomy and Range Science
University of California
Davis, California 95616

Dr. Michael C. MacCracken
Deputy Division Leader
Atmospheric and Geophysical Sciences Division
Physics Department
Lawrence Livermore Laboratory
P.O. Box 808
Livermore, California 94550

Dr. Lester Machta, Chairman
Air Resources Laboratories
National Oceanic and Atmospheric Administration
8060 13th Street
Silver Spring, Maryland 20910

Dr. Thomas Malone
Holcomb Research Institute
Butler University
Indianapolis, Indiana 46208

Dr. Syukuro Manabe
Geophysical Fluid Dynamics Laboratory
Princeton University
P.O. Box 308
Princeton, New Jersey 08540

Dr. Gregg Marland
Institute for Energy Analysis
Oak Ridge Associated Universities
P.O. Box 117
Oak Ridge, Tennessee 37830

Dr. J. Murray Mitchell
Environmental Data Service, R32
National Oceanic and Atmospheric Administration
8060 13th Street
Silver Spring, Maryland 20910

Dr. Reginald Newell
Department of Meteorology
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

Dr. H. Oeschger
Physikalisches Institut
Universität Bern
Sidlerstr. 5
CH-3012 Bern
SWITZERLAND

Dr. Jerry S. Olson
Environmental Sciences Division
Oak Ridge National Laboratory
Oak Ridge, Tennessee 37830

Dr. H. Göte Ostlund
School of Marine and Atmospheric Science
University of Miami
4600 Rickenbacker Causeway
Miami, Florida 33149

Dr. Graeme I. Pearman
CSIRO
Division of Atmospheric Physics
P.O. Box 77
Mordialloc, Victoria 3195
AUSTRALIA

Dr. S. S. Penner
Energy Center
University of California
P.O. Box 109
La Jolla, California 92037

Dr. V. Ramanathan
National Center for Atmospheric Research
P.O. Box 3000
Boulder, Colorado 80302

Dr. Elmar Reiter
Department of Atmospheric Science
Colorado State University
Fort Collins, Colorado 80521

Dr. Roger Revelle
Scripps Institution of Oceanography
La Jolla, California 92093

Dr. Wolfgang Roether
Zweites Physikalisches Institut der Universität
Heidelberg
Philosophenweg 12
Heidelberg 1
WEST GERMANY

Dr. N. J. Rosenberg
Department of Horticulture and Forestry
University of Nebraska
Lincoln, Nebraska 68503

Dr. Ralph Rotty
Institute for Energy Analysis
Oak Ridge Associated Universities
P.O. Box 117
Oak Ridge, Tennessee 37830

Dr. Bert Rust
Institute for Energy Analysis
Oak Ridge Associated Universities
P.O. Box 117
Oak Ridge, Tennessee 37830

Dr. George Saunders
Division of Biomedical and Environmental Research
U.S. Energy Research and Development
Administration (U.S. Department of Energy)
Washington, D.C. 20545

Dr. David Schink
Department of Oceanography
Texas A&M University
College Station, Texas 77843

Dr. M. E. Schlesinger
Department of Atmospheric Science
Oregon State University
Corvallis, Oregon 97331

Dr. Steven Schneider
National Center for Atmospheric Research
P.O. Box 3000
Boulder, Colorado 80302

Mr. Jack Siegel
U.S. Energy Research and Development
Administration (U.S. Department of Energy)
Washington, D.C. 20545

Mr. David H. Slade
Deputy Manager
Environmental Programs
Division of Biomedical and Environmental Research
U.S. Energy Research and Development
Administration (U.S. Department of Energy)
Washington, D.C. 20545

Dr. L. F. Small
School of Oceanography
Oregon State University
Corvallis, Oregon 97331

Dr. Minze Stuiver
Department of Geological Sciences and Zoology
University of Washington
Seattle, Washington 98195

Dr. Hans Suess
Department of Chemistry
University of California at San Diego
La Jolla, California 92093

Mr. Eric Sundquist
Hoffman Laboratories
Harvard University
20 Oxford Street
Cambridge, Massachusetts 02138

Dr. Taro Takahashi
Geochemistry Section
Lamont-Doherty Geological Observatory
Columbia University
Palisades, New York 19604

Dr. Richard Temkin
National Center for Atmospheric Research
P.O. Box 3000
Boulder, Colorado 80302

Dr. Robert L. Watters
Division of Biomedical and Environmental Research
U.S. Energy Research and Development
Administration (U.S. Department of Energy)
Washington, D.C. 20545

Dr. Kenneth Weaver
Staff Representative for the House Committee
on Atmosphere and Environment
House Office Building
Washington, D.C. 20515

Dr. Alvin M. Weinberg
Institute for Energy Analysis
Oak Ridge Associated Universities
P.O. Box 117
Oak Ridge, Tennessee 37830

Dr. C. S. Wong
Ocean Chemistry Division
Pacific Region
Department of the Environment
211 Harbour Road
Victoria, British Columbia
CANADA

Dr. George M. Woodwell
The Ecosystems Center
Marine Biology Laboratory
Woods Hole, Massachusetts 02542

Dr. P. J. Zinke
Department of Forestry
University of California
Berkeley, California 97420

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