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Subseries: Global Climate Change Files

OA/ID Number: 62052
Folder ID Number: 62052-011

Folder Title:
Global Warming: Greenhouse Gases (1) [1 of 3] [1990]

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Executive Summary

30 Apr 90 19h

We are certain of the following:

The greenhouse effect is real; even without man-made contributions, natural greenhouse gases already keep the earth warmer than it would otherwise be.

* Man made emissions are substantially increasing the atmospheric concentrations of the greenhouse gases: carbon dioxide, methane, the chlorofluorocarbons, nitrous oxide and tropospheric ozone. These increases will lead to a warming of the Earth's surface. The main greenhouse gas, water vapour, is also expected to increase in response to a global warming.

We calculate with confidence that:

Some gases are potentially more effective than others at changing climate, and their relative effectiveness can be estimated. Carbon dioxide has been responsible for over half the total man-made greenhouse effect in the past, and is likely to remain so in the future.

Atmospheric concentrations of the long-lived gases (carbon dioxide, nitrous oxide and the CFCs) adjust only slowly to changes in emissions. Present day emissions of these gases are committing us to increased concentrations for decades to centuries. The longer emissions continue at present day levels, the greater reductions would have to be to stabilise at a given concentration.

The long-lived gases would require reductions in man-made emissions of 60-80% to stabilise their concentrations at today's levels; methane would require only a 15-20% reduction.

* Using current models, due to IPCC Business-as-Usual emissions, the following changes are predicted:

* global mean temperatures about 2°C above pre-industrial ¶ (with a range of uncertainty of 1.4°C to 2.9°C) by the year 2035; by the year 2100 temperatures about 4°C above pre-industrial # (with a range of 2.7°C to 6°C)

* rate of change of global mean temperature 0.2 - 0.4°C per decade over the next century

* land surfaces warmed more rapidly than the oceans, and high northern latitudes warmed more than the global mean in winter.

Our confidence in prediction of regional changes is low, though in general they will be different from the global mean. For example, temperature increases in Southern Europe and continental North America are predicted to be higher than the global mean, accompanied by reduced summer precipitation and soil moisture.

- * sea level 10cm to 32cm higher than today by 2030, 33cm to 75cm higher by 2070 ; due mainly to the thermal expansion of the oceans and the melting of some land ice.

There are many uncertainties in our predictions which arise in particular from an incomplete understanding of :

- * sources and sinks of greenhouse gases, which affect predictions of future concentrations
- * clouds, which strongly influence the magnitude of climate change
- * oceans, which influence the timing and patterns of climate change

These processes are at least partially understood, and we are confident that the uncertainties can be reduced by further research

Our best judgement is that:

- * Global - mean surface air temperature has increased by 0.3 to 0.6 °C over the last 100 years, with the six global-average warmest years being in the 1980s. This rise has not been smooth with time, nor uniform over the globe.

The size of this warming is broadly consistent with predictions of climate models. If the sole cause of the warming was man-made greenhouse gases, then the implied climate response would be near the lower end of model predictions. However, the warming to date is within the range of natural variability, so that this variability and other man-made factors could have offset a larger man-made greenhouse warming.

There is no evidence that climate has become more variable over the last few decades. However, with an increase in the mean temperature, episodes of high temperatures will most likely become more frequent in the future, and cold episodes less frequent,

Ecosystems will be affected both favourably and unfavourably by a changing climate and by increasing carbon dioxide concentrations. If climate changes rapidly the composition of ecosystems will change and some species will be unable to migrate or adapt fast enough and may become extinct. However, enhanced levels of carbon dioxide may increase productivity and efficiency of water use of vegetation.

To improve our predictive capability, we need:

- * to understand better the various climate processes, particularly those associated with clouds, oceans and the carbon cycle
- * to improve our monitoring of a number of climate variables on a global basis, and further investigate changes which took place in the past

* to **develop** improved computer-based numerical models of the earth's climate system.

* to **Increase** support for national and international research activities

¶ 0.9°C to 1.8°C above 1990 temperatures

2.2°C to 4.5°C above 1990 temperatures

1. Doubling relative to pre industrial 600 Bt in atmosphere
2. Up to 750 B in 1990 — up 25%
3. In dec. & doubling — "effective" is implied
4. But in 1990 $\text{CH}_4 + \text{NO}_2$ etc have increased
"greenhouse gas" i.e. radiative forcing
by same as CO_2 i.e. another 25%
5. So we are now 50% above pre industrial.
These are hard data

6. Used in IPCC "High Emissions" — which US has
just agreed — "Business as Usual" but trying
to change.

— ~~we~~ assume that from year no
— which is 5.9 Wt we increase in Rf from
and is 7-8B by 2030-2050.

7. In all IPCC CO_2 — 50-60% of total radiative
forcing.

8. So we are looking to increase to 900B i.e.
 CO_2 up by 50% and other gases contribute other
50%

9. Do only need another 150 BT in atmosphere.
10. At 6 BT/a, and [$\frac{1}{2}$ stay in atmosphere] we need 300 BT increase i.e. 50 yrs
11. But if we assume increase in rate then < 50 yrs.
— 1 —
12. Over last 100 yrs. $(50 \pm 10)\%$ is fraction that stays in the atmosphere. We understand fossil fuel input but deforestation 0.6-2.5 (1980 estimate) - could now be larger. - Usually 1.6 rounded to 2
— 1 —
13. 3-3.5 BT measured to stay in atmosphere per year - (from Hawaii data)
14. We have fossil fuel at 5.9 and 1.6 i.e. 7.5 BT/yr. 3.5 stays in atmosphere. - so other 4 BT go?
15. Ocean uptake is 2.5 BT/a classically. (RCC says 2 BT/a on this page) Could be anywhere between 1 and 3 BT/a.

16. Specified 2 Gt/a missing!

Sequestered biopne takes this up — or the deforestation number is closer to zero. But really no understanding.

17. One just says There is a regrowth forest in mid latitudes. NO_x + some SO_2 could have stimulated biomass Carbon sequestration. More CO_2 itself could have directly fertilized biomass.

18. What entraps ocean uptake is how does CO_2 get translated from surface to depth. — not the interface air-water. We don't understand very well this factor.

19. Surface (CO_2) measurements are so variable in time and space so measurements are very spotty and we may have a very lousy average. Very few in far Southern, High latitude, data.

20 IPCC has effective doubling by 2025.

#2 scenario gives 2050

#3 scenario gives 2090.

— worked backwards from simple assumption of these dates.

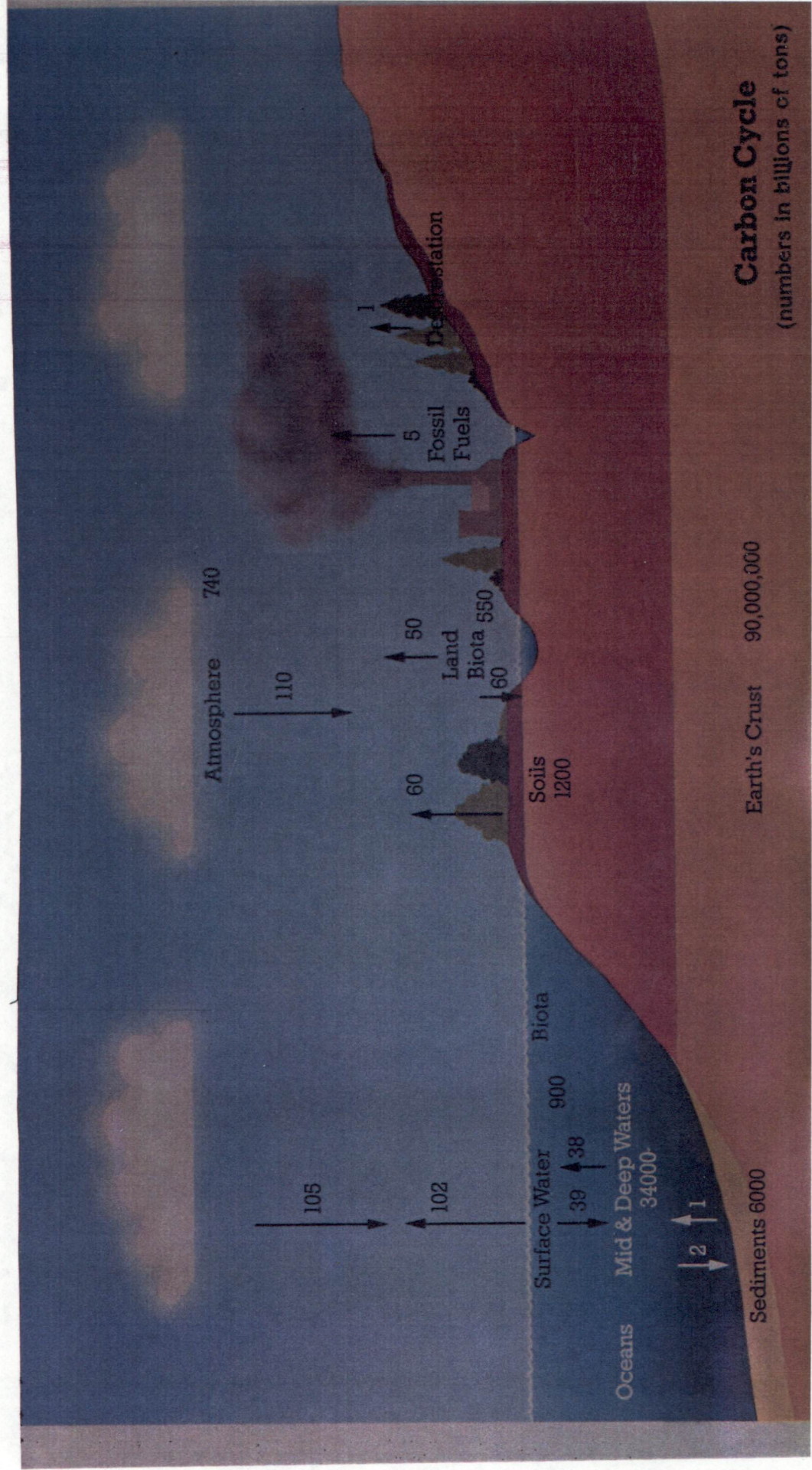
21) But bottom up study (EPA + Dutch) under IPCC Wg #3. — study CO_2 only and got ~~doubling~~ even faster growth than business as usual.

22. IPCC estimates $2^\circ C \pm (1.4 - 2.9)$ relative preindustrial $(0.9 - 1.8)$ above today.

23. Models would predict that today global temp would have ~~to~~ increased $(0.5 - 1.1)^\circ C$ by 1750.

24. If we assume banning CFC's by 2000 IPCC keeps 50%+ growth in LDC's so

overestimates. This is a fundamental flaw would increase time to "effective doubling" by a decade or so.



Carbon Cycle
(numbers in billions of tons)

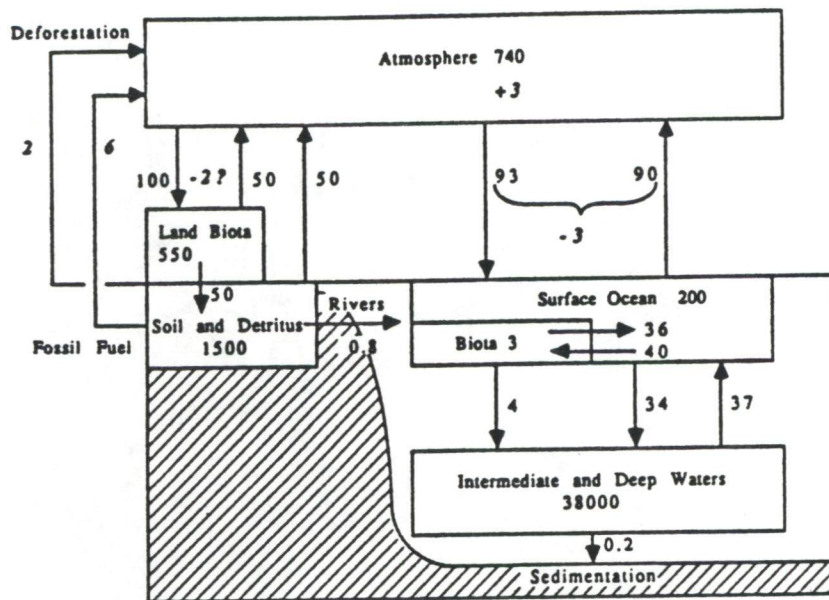


Figure 1: Global carbon reservoirs and fluxes. The numbers apply for the present-day situation and represent typical literature values. Fluxes, e.g. between atmosphere and surface ocean, are gross annual exchanges. Numbers in bold-face italics indicate net annual perturbation fluxes to the atmosphere and annual atmospheric CO_2 increase due to human action. Units are gigatons of carbon (GtC ; $1\text{Gt} = 10^9$ metric tons = 10^{12}kg) for reservoir sizes and GtC yr^{-1} for fluxes. More details and discussions are found in several reviews (Sundquist, 1985; Trabalka, 1985; Bolin, 1986; Siegenthaler, 1986).

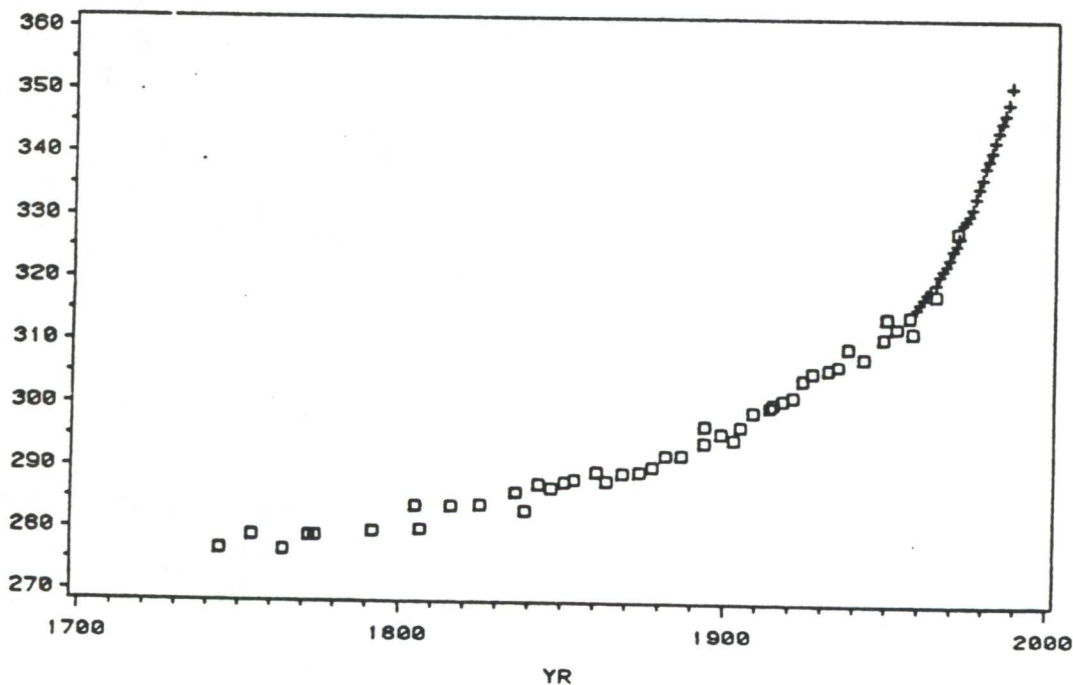


Figure 2: Atmospheric CO_2 increase in the past 200 years, as indicated by measurements on air trapped in ice from Siple Station, Antarctica (squares; Neftel et al., 1985a; Friedli et al., 1986) and by direct atmospheric measurements at Mauna Loa, Hawaii (crosses; Keeling, et al., 1989). Units: parts per million by volume (ppmv).

The term "atmospheric stabilisation" is often used to describe the limiting at present day values of the concentration of the greenhouse gases. The amount by which man-made emissions of a greenhouse gas must be reduced in order to achieve this is shown in the box below.

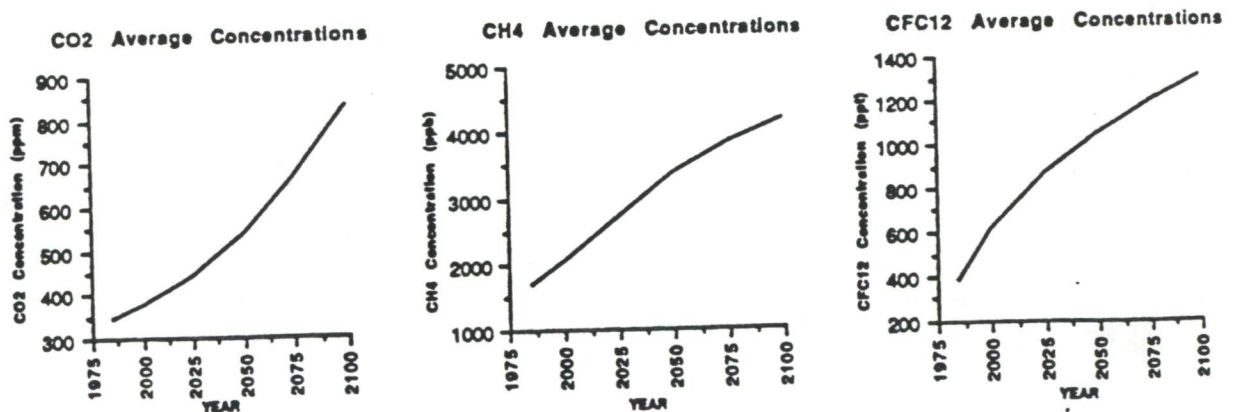
STABILISATION OF ATMOSPHERIC CONCENTRATIONS

Reductions in the man-made emissions of greenhouse gases required to stabilise concentrations at present day levels:

Carbon Dioxide	60 - 80%
Methane	15 - 20%
Nitrous Oxide	70 - 80%
CFC11	70 - 75%
CFC12	75 - 85%
HCFC22	40 - 50%

Natural sources and sinks are assumed to remain unchanged
Note that the stabilisation of each of these gases would have different effects on climate, as explained in the next section.

The same models used for these calculations are also used to determine future concentrations of the greenhouse gases which would arise from scenarios of future emissions generated by IPCC Working Group 3. Shown below are the concentration trends of some of the greenhouse gases expected to result from the High Emissions scenario.



Which gases are the most important?

We know, with certainty, that more greenhouse gases means more radiative forcing and hence a global warming. We can calculate the the radiative forcing due to the increase in concentration of each gas with much more confidence than the calculation of the resulting climate change because the former only involves laboratory measurements of the gases: how strongly, and where in the spectrum,

GLOBAL WARMING POTENTIALSThe warming effect of an emission of 1kg of each gas relative to that of CO₂

	time horizon		
	20 yr	100 yr	500 yr
Carbon dioxide	1	1	1
Methane	63	21	9
Nitrous oxide	270	290	190
CFC11	4500	3500	1500
CFC12	7100	7300	4500
HCFC22	4100	1500	510

Global Warming Potentials for a range of CFCs and potential replacements are given in the full text

Although the table shows carbon dioxide to be the least effective greenhouse gas on an equal emissions basis, the warming effect will depend on the size of emissions. In the example shown below, 1990 emissions of carbon dioxide will contribute well over half the man-made warming effect over the next 100 years.

THE RELATIVE CUMULATIVE CLIMATE EFFECT OF 1990 EMISSIONS

	GWP (100yr horizon)	1990 emissions (Tg)	relative contribution over 100yr
carbon dioxide	1	26000†	61%
methane	21	300	15%
nitrous oxide	290	6	4%
CFC11	3500	0.3	2%
CFC12	7300	0.4	7%
HCFC22	1500	0.1	0.4%
Others (including indirect effects, eg that of NO ₂ on surface ozone)			10.6%

†26000Tg of carbon dioxide = 7Gt of carbon

Patrick Buchanan is a nationally syndicated columnist.

tant secretary of state for international organization affairs and ambassador to the U.N. Economic and Social Council.

see LERNER, page D4

Max Lerner is a nationally syndicated columnist.

WARREN BROOKES

On Thursday, Aug. 2, Michael Deland, chairman of the White House Council on Environmental Quality, stripped the clothes from the European green imposters when he said:

"When the [European Community] environmental ministers come to town ... the first question out of their mouth is 'You're not going to sign on to a carbon-dioxide reduction plan, are you?'"

Even as these same ministers assailed the U.S. refusal to set a target for carbon-dioxide emissions reductions, they are privately encouraging the Bush administration to keep dragging its own collective feet. One week earlier, Britain released its environmental white paper without any reference to a CO2 tax.

Good thing, too. The next day after Mr. Deland demonstrated that

British Prime Minister Margaret Thatcher was talking tough but waving a wet green noodle, Iraqi strongman Saddam Hussein reminded us that folks who play on coal piles should stay away from greenhouse lawn parties.

In one lightning stroke, Mr. Hussein demonstrated that any country that now gets more than 58 percent of its electricity from coal (as the United States does) and for whom coal is the leading domestic energy source has no business toying with unproven scientific theories that would make that vital source obsolete — overnight.

Not only are we among the world's most coal-dependent nations, we are now the world's largest and most-efficient coal exporter. Indeed, since 1981, while U.S. oil and gas extraction have plunged by 22 percent, U.S. coal production has

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Tanks crush greenhouse

soared by 21 percent, and this year we will export more than 100 million short tons. Coal is one of the reasons we now have a trade surplus with the EC nations.

Since coal burning produces three times as much carbon dioxide per British thermal unit as oil and gas, it would have to be taxed three times as much to cut CO2 emissions. In 1989, the Environmental Protec-

man equation as well as the environment."

Sadly, that concern is not shared by his own State Department or the EPA, who will dominate the staffing of the U.S. delegation to the Intergovernmental Panel on Climate Change (IPCC) in Stockholm Aug. 29 to work for a 1992 "consensus" for actions that would endanger the United States as well as coal-dependent China and the Soviet Union.

Yet that consensus will fly in the face of growing dissent over the evidence for global warming theories and models. As Science editor Philip Abelson said of this issue on March

TAXES AND COSTS TO CONTROL CARBON-DIOXIDE EMISSIONS

	To stabilize CO2 tax rate	cost	To reduce CO2 20 % tax rate	cost
Oil	\$6/bbl*	\$30 bill.	\$25/bbl	\$90 bill.
Natural gas	\$0.50/MCF**	\$10 bill.	\$2/MCF	\$30 bill.
Coal	\$25/ton	\$25 bill.	\$200/ton	\$80 bill.
Totals		\$65 bill.		\$200 bill.

*bbl = per barrel **MCF = per 1,000 cubic feet

Total costs estimated based on reduced usage of fuels.

Source: Environmental Protection Agency (rates)

The Washington Times

tion Agency said to reduce CO2 emissions by 20 percent would require a \$200-per-ton tax on coal, \$25 a barrel on oil, \$2 per thousand cubic feet on natural gas, a probable total of \$200 billion per year! (See Table.)

No wonder President Bush was willing to take collective heat for cooling down global warming at Houston last month, warning that: "On the environmental extreme, they don't want this country to grow. I am not going to talk about no growth for the United States because I feel a deep concern about the hu-

30, "There has been more hype than solid facts."

That editorial accompanied the publication of a 10-year satellite temperature study by NASA that showed "no warming or cooling trend" in 1979-1988. Those satellite data were known to NASA's James Hansen in the drought of June 1988 when he went before Congress and launched the global-warming panic, saying he was "99 percent confident" greenhouse warming had started.

see BROOKES, page D4

BROOKES

From page D1

Yet just six months later in December 1988, the U.S. National Climate Data Center published a "de-urbanized" analysis of the U.S. temperature record since 1895, which showed "no significant warming trend." That was followed 10 months later by the Massachusetts Institute of Technology study that showed there had been no ocean surface warming since 1850.

Then in March 1990, in the Bulletin of the American Meteorological Society, MIT's Richard Lindzen pointed out that whereas all of the climate models predict a warming of 5 degrees to 14 degrees Celsius in the Arctic during the winter, from 1940-1987, the Atlantic Arctic winter temperatures have fallen 2.5 degrees C.

At his largely ignored April Interparliamentary Conference on Climate Change, Sen. Albert Gore, Tennessee Democrat, was so frustrated by this he charged any scientist who studied the Antarctic ice cores and did not consider global warming an immediate "crisis" was being "unethical."

Yet a month later, on May 29 in Baltimore, two University of Rhode Island scientists, Todd Sowers and Michael Bender, told the American Geophysical Union that in the Antarctic ice-core data the lag between changes in CO₂ over the centuries and ice-cap melting, i.e., basic climate change, ranged from 3,000 to 8,000 years.

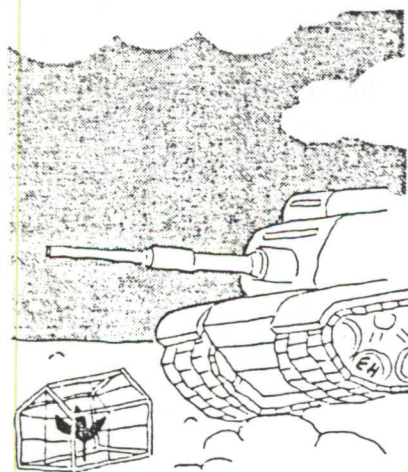
Then last week the New York Times reported that a new Swedish tree-ring study shows that "average summer temperature in far Northern Europe has fluctuated almost 7 degrees over the last 1,400 years, il-

lustrating how widely the Earth's climate varies naturally and suggesting that in the northernmost parts of the world [where all the warming is now predicted], the effects of any global warming will be hard to feel for a long time."

Then yesterday, British Independent Television (ITV-channel 4) presented a devastating one-hour documentary on its science show "Equinox" called "The Greenhouse Conspiracy." It argued bureaucratic politics, turf-building and environmental hysteria are driving this issue far beyond actual scientific evidence.

Why? Because it fits the no-growth agenda of environmentalism. As Mr. Gore's keynote speaker at his April conference — Brazil's Secretary of the Environment Jose Antonio Lutzenberger — told an enthusiastic audience that, "The worst and most dangerous aspect of our modern economic thinking is the dogma of the necessity for continuous economic growth. Even the idea of sustainable growth is a contradiction in terms."

It would seem Saddam Hussein drove a big tank through that greenhouse garden party in just the nick of time.



General Change We

Document Originally
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Following Page

Chart 1
Projected U. S. Greenhouse Gas Emissions
 (With Current Policy Commitments)
 Based on IPCC CO2 Baseline Estimate

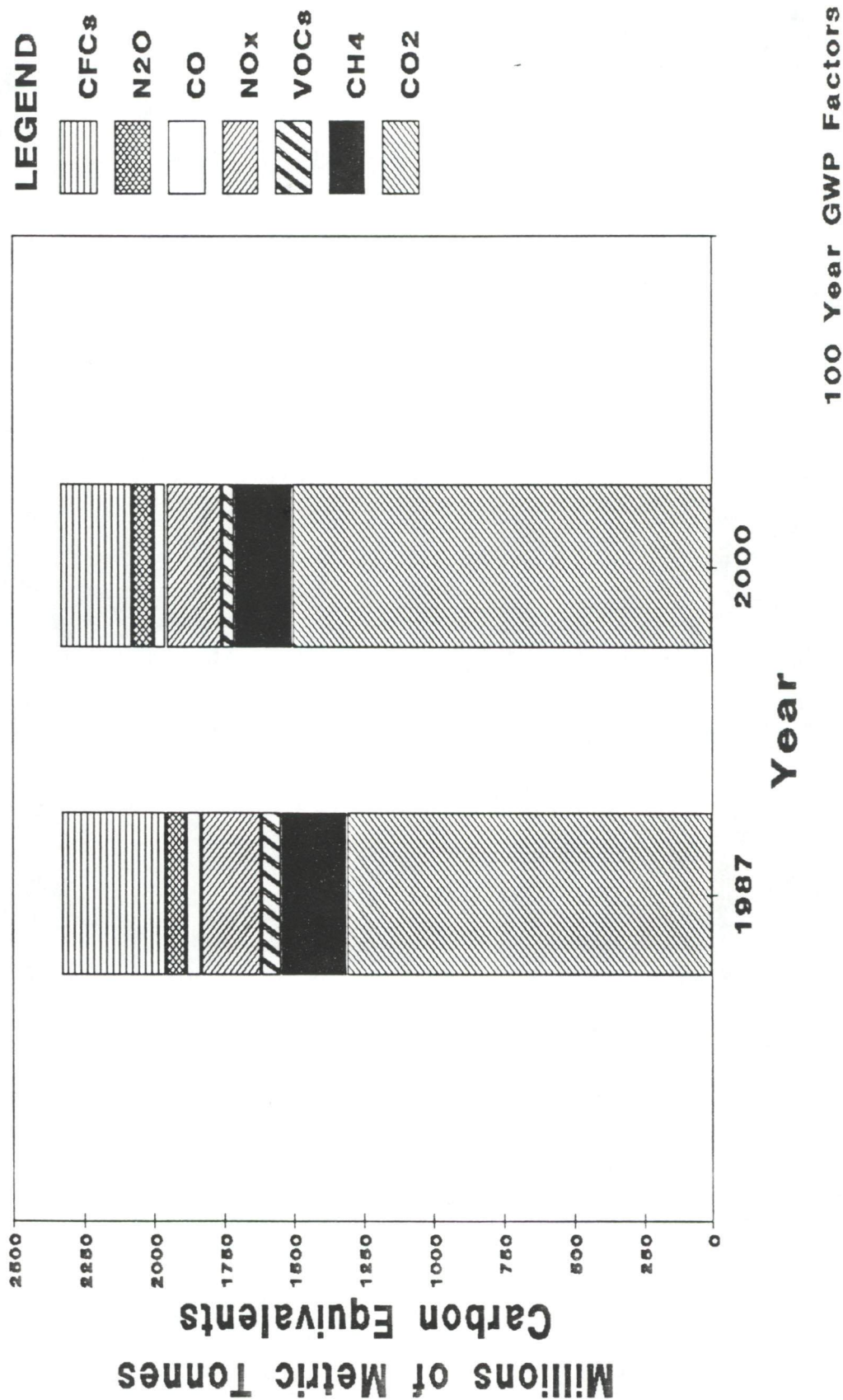
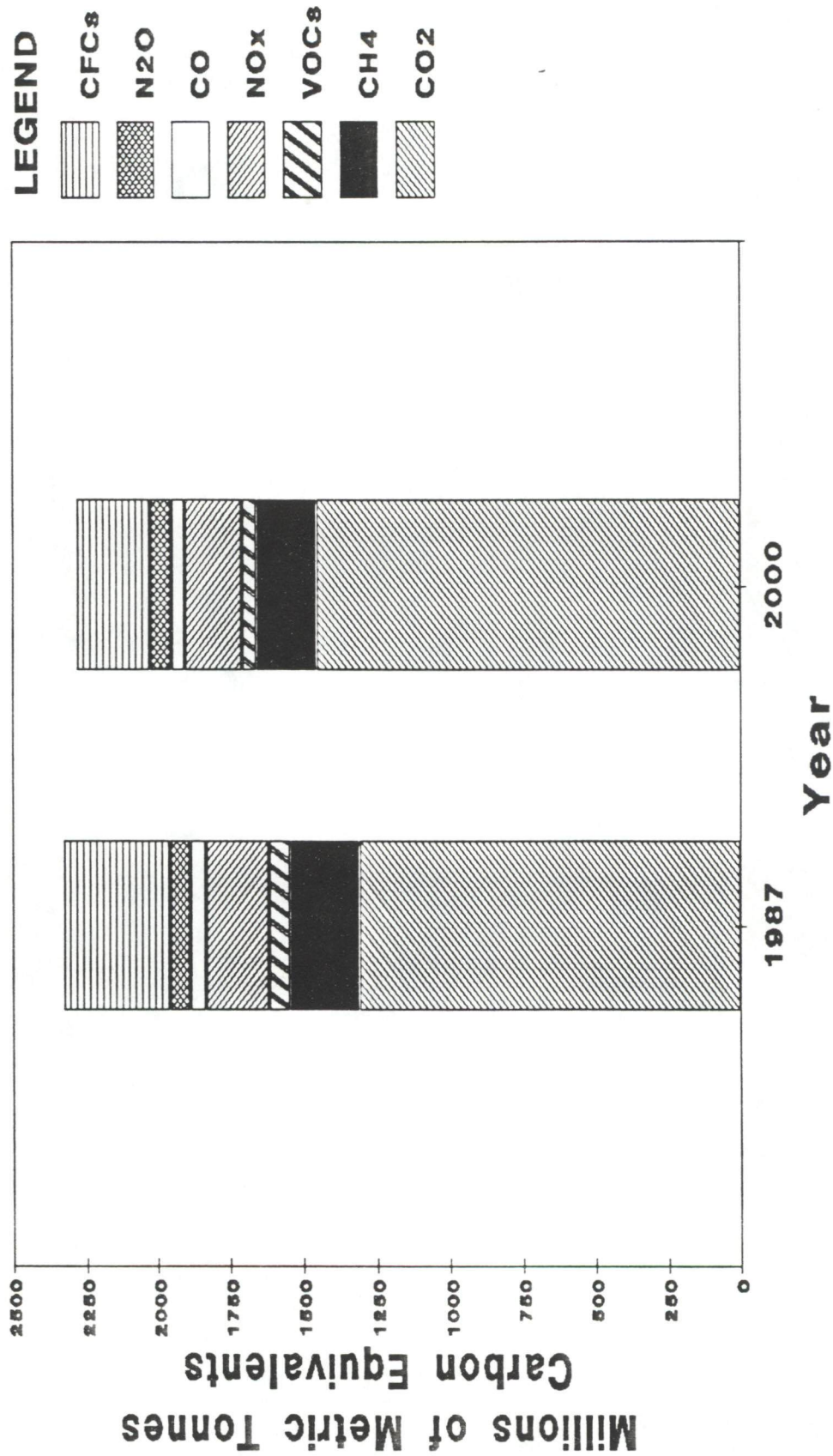


Chart 2
Projected U. S. Greenhouse Gas Emissions
 (With Current Policy Commitments)
 Based on Jorgenson/Wilcoxon (1990) Model



100 Year GWP Factors

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Chart 3
Phase-Out of Major CFCs
Montreal Protocol (P) Compared
With U. S. Clean Air Act

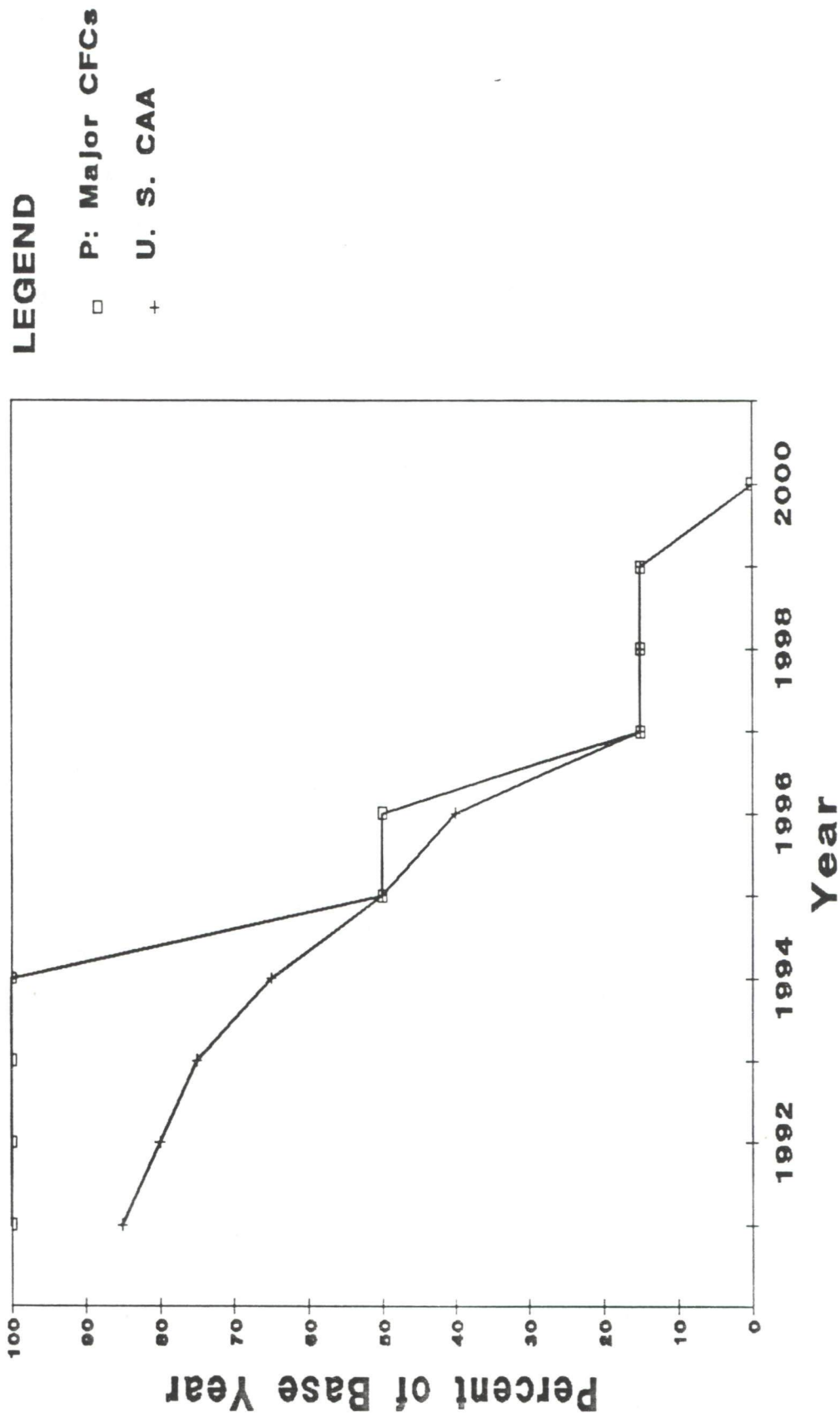


Chart 4
Phase-Out of Halons
Montreal Protocol (P) Compared
With U. S. Clean Air Act

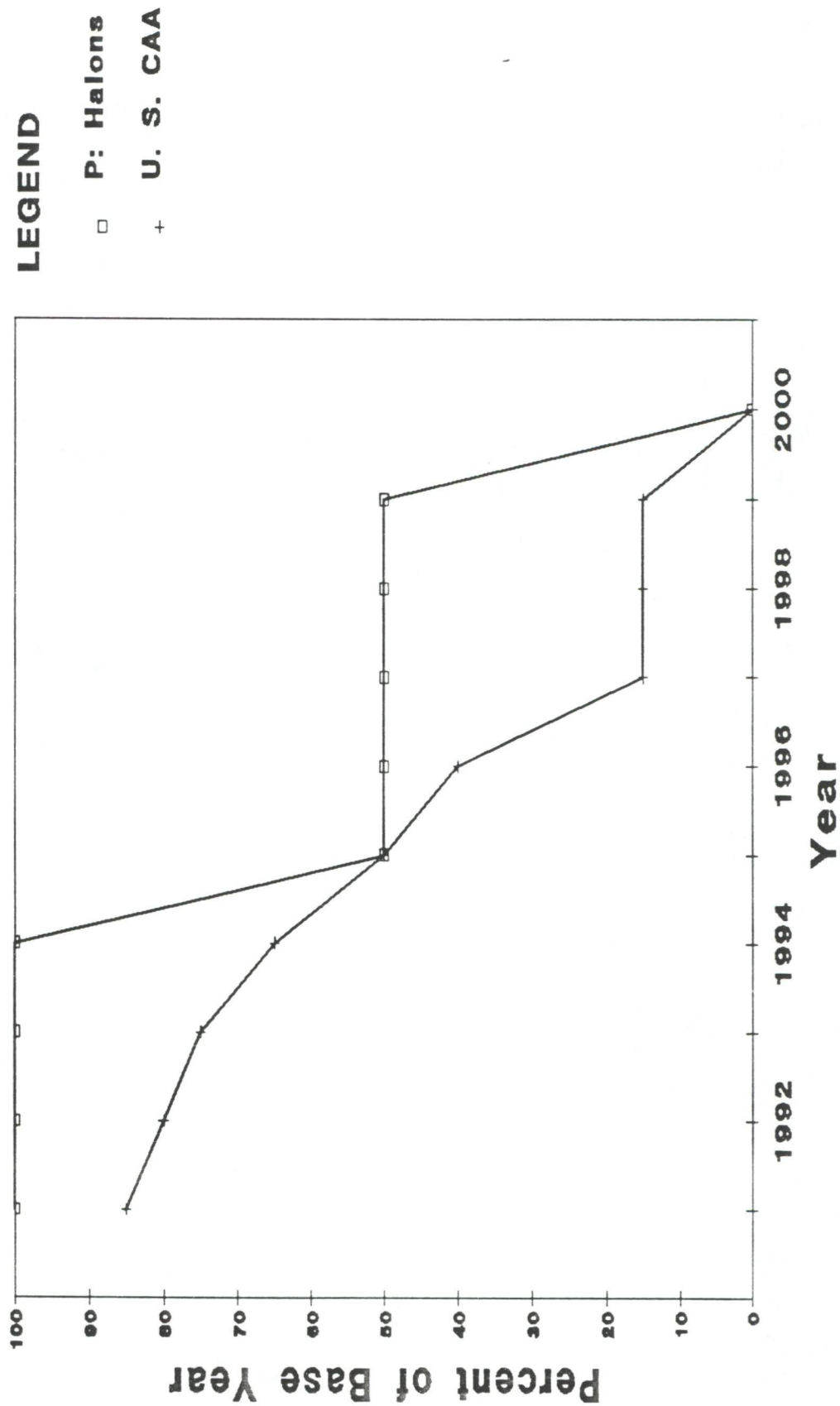


Chart 5
Phase-Out of Other CFCs
Montreal Protocol (P) Compared
With U. S. Clean Air Act

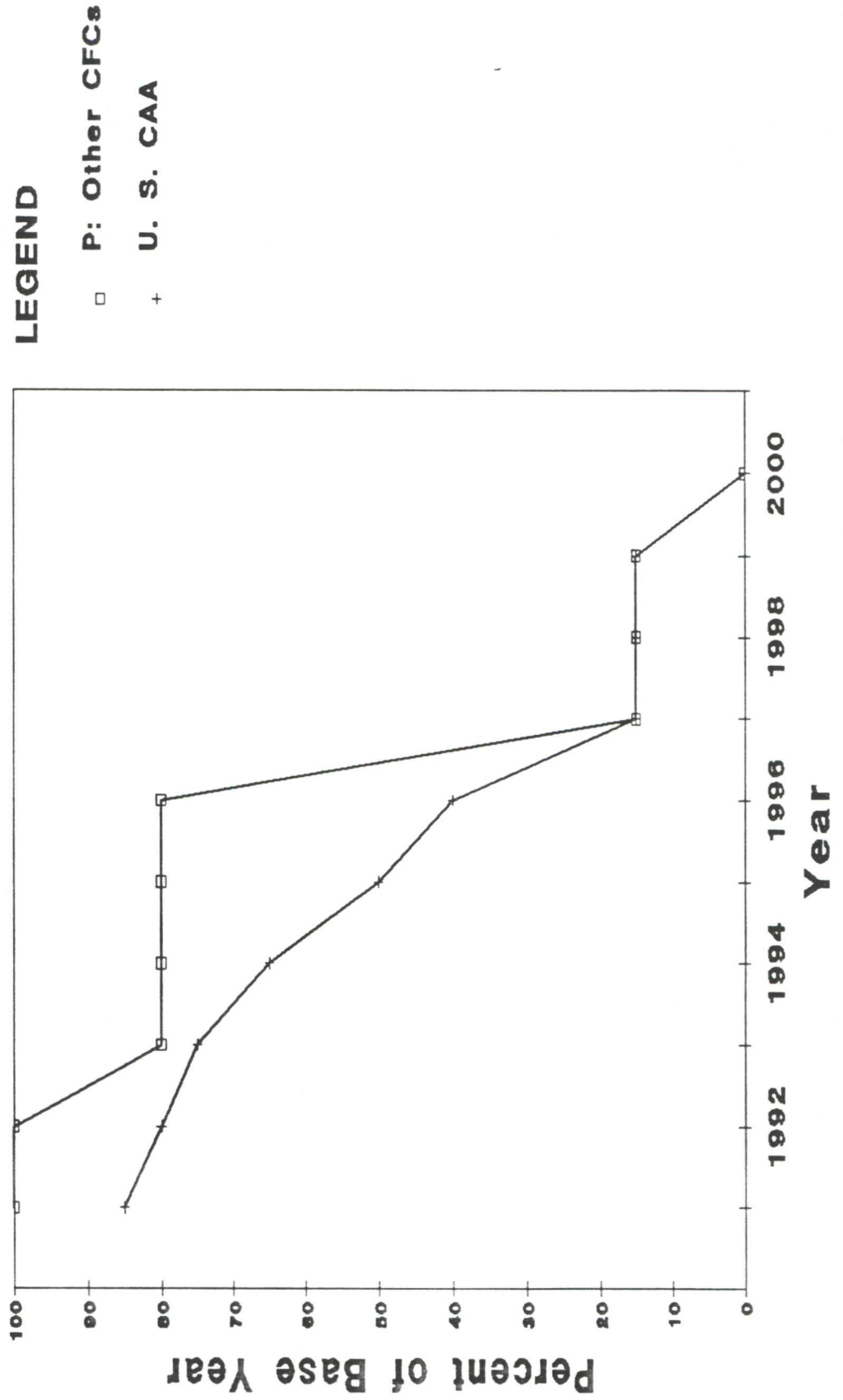


Chart 6
Phase-Out of Carbon Tetrachloride
Montreal Protocol (P) Compared
With U. S. Clean Air Act & User Fee

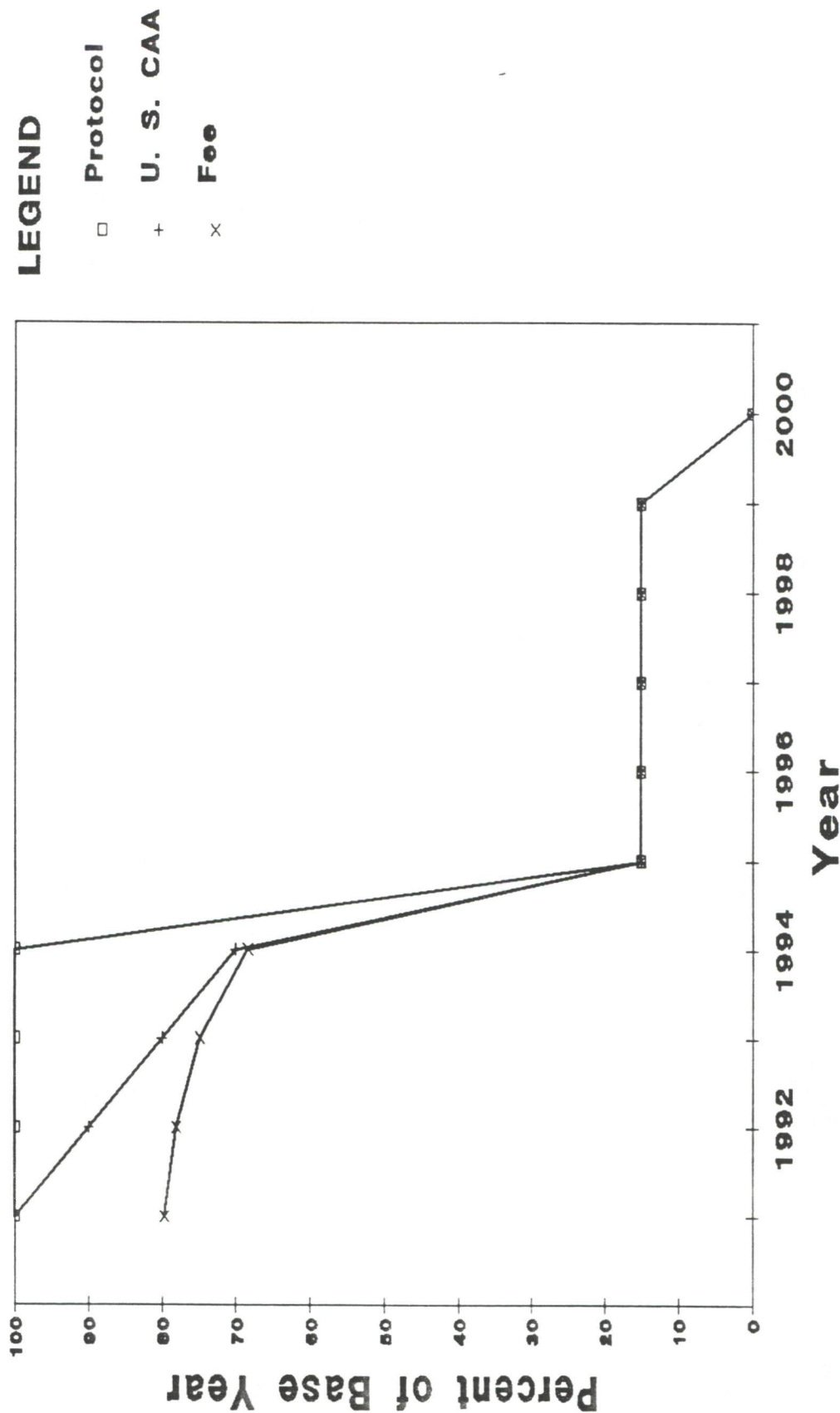
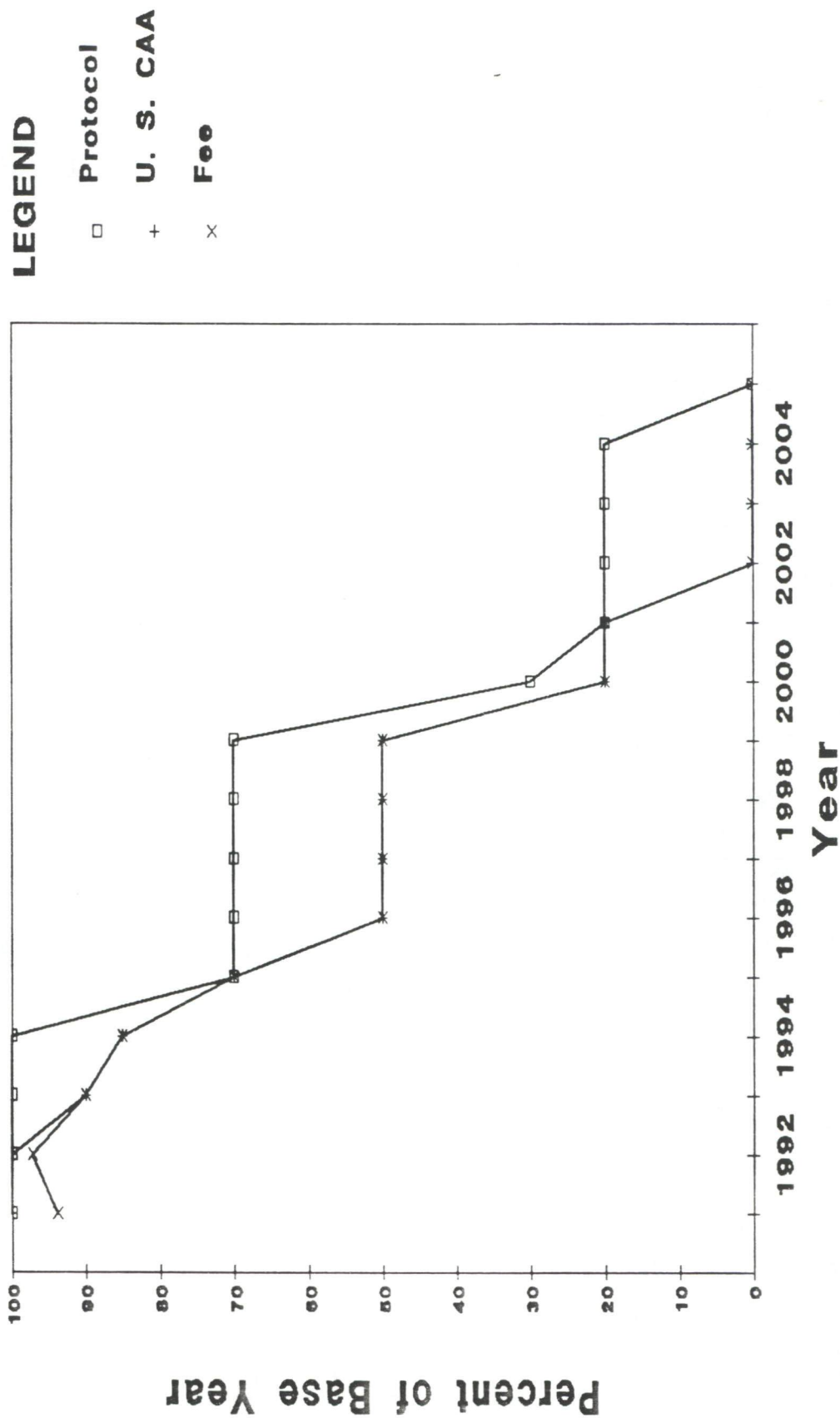


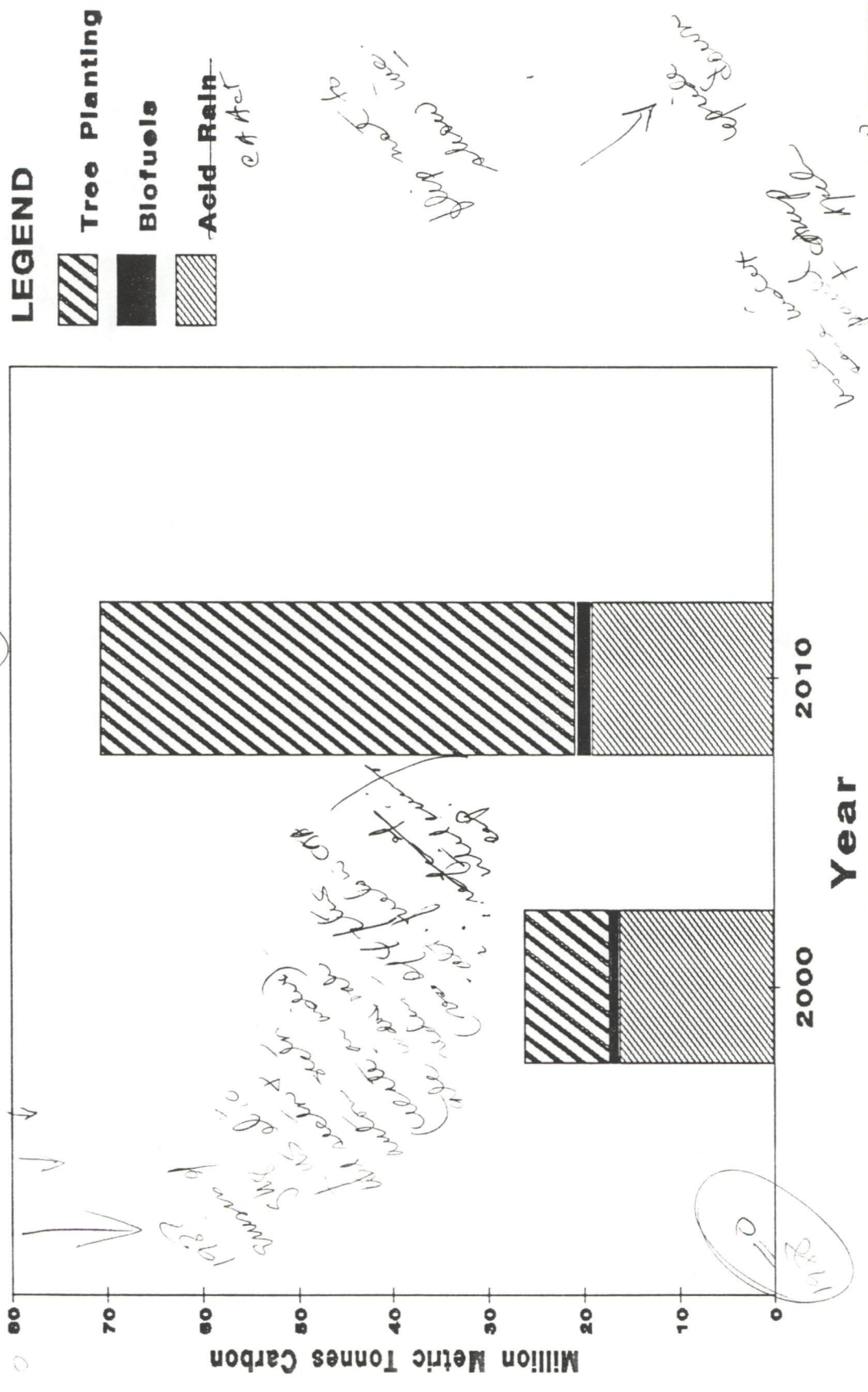
Chart 7

**Phase-Out of Methyl Chloroform
Montreal Protocol (P) Compared
With U. S. Clean Air Act & User Fee**



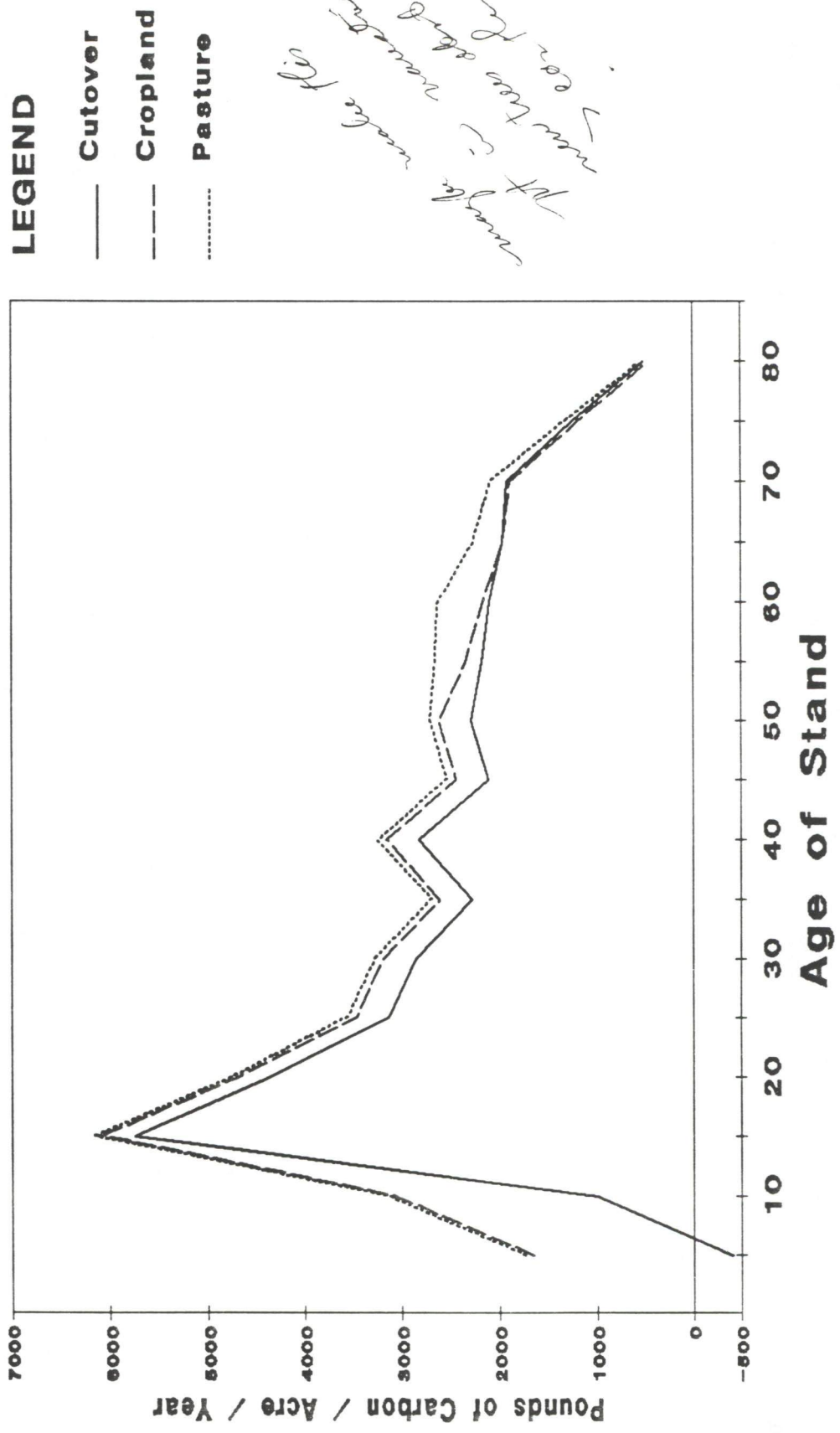
How does it work?
Col. J. J. J.
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Chart 8
Reductions in Carbon Dioxide Emissions
From Clean Air Act and Trees Initiative



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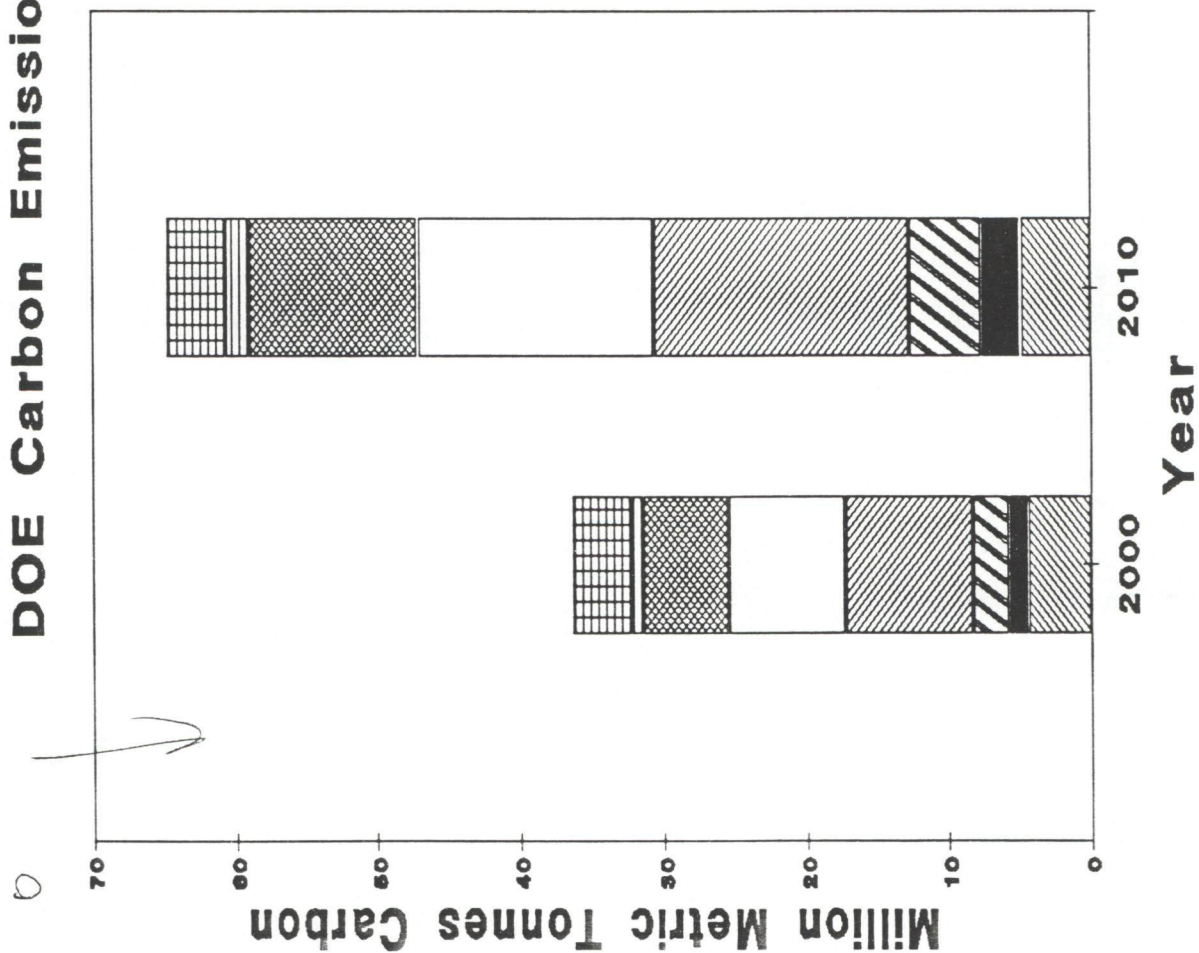
Chart 9
Tree Planting Initiative
Example Carbon Sequestration by Age of Stand *



* SE Region, S. Pine, High Quality Site, Improved Stock

Chart 10
DOE Carbon Emissions Initiatives

newt
base as



2

Chart 11
Emissions of CO₂ Per Unit of GDP, 1988

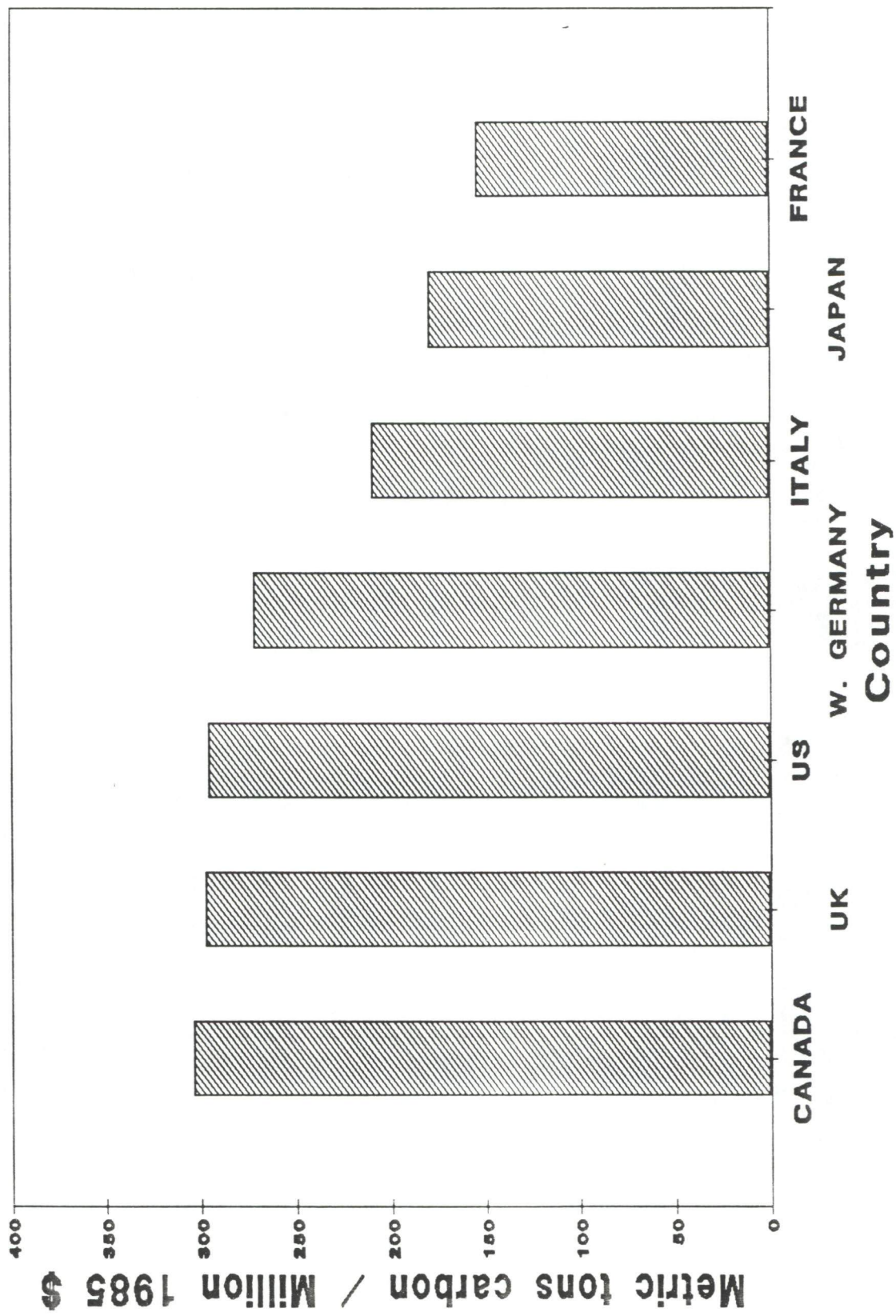
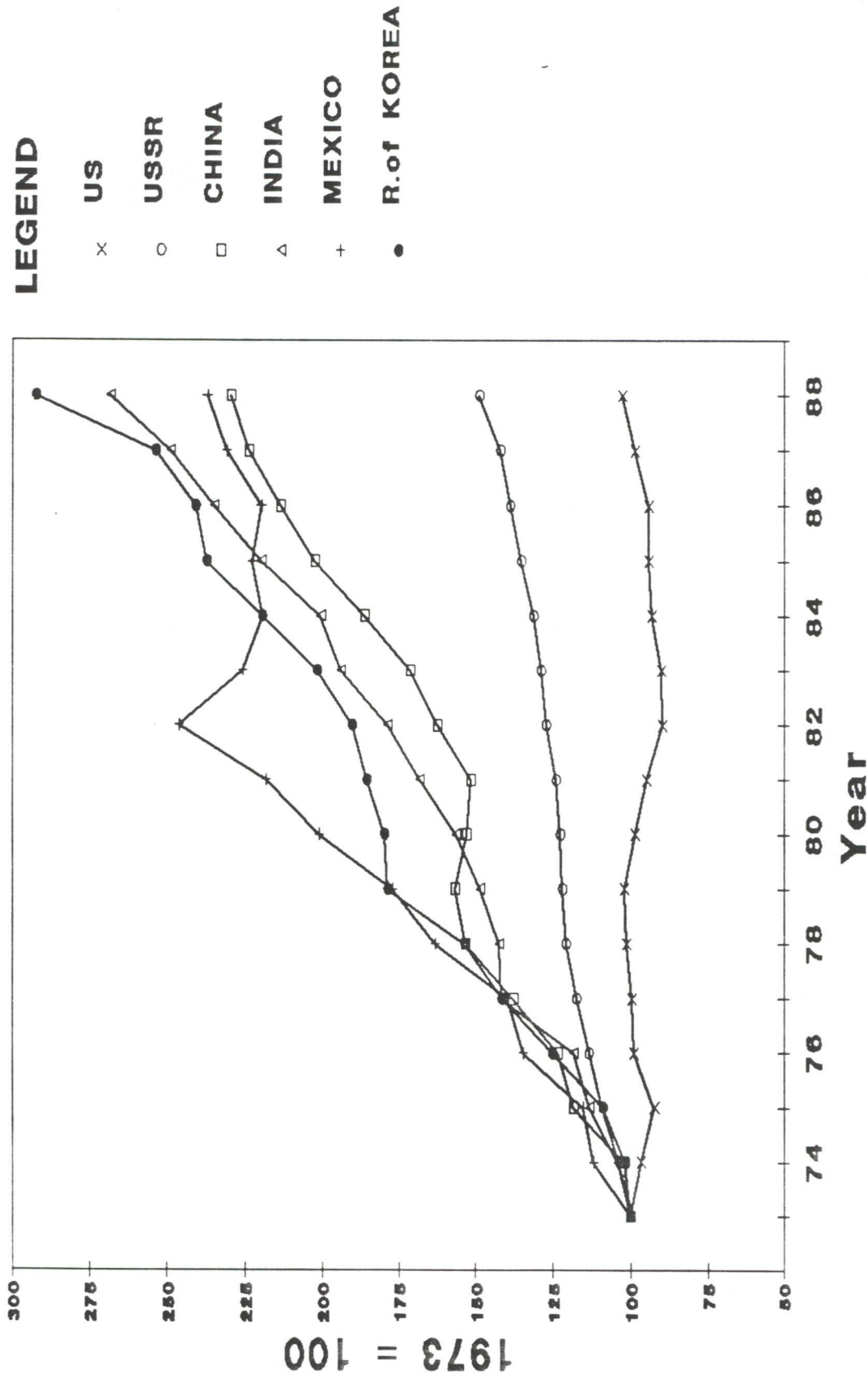


Chart 12
Trends in Carbon Dioxide Emissions
U.S. and High Growth Countries



Trends in Carbon Dioxide Emissions

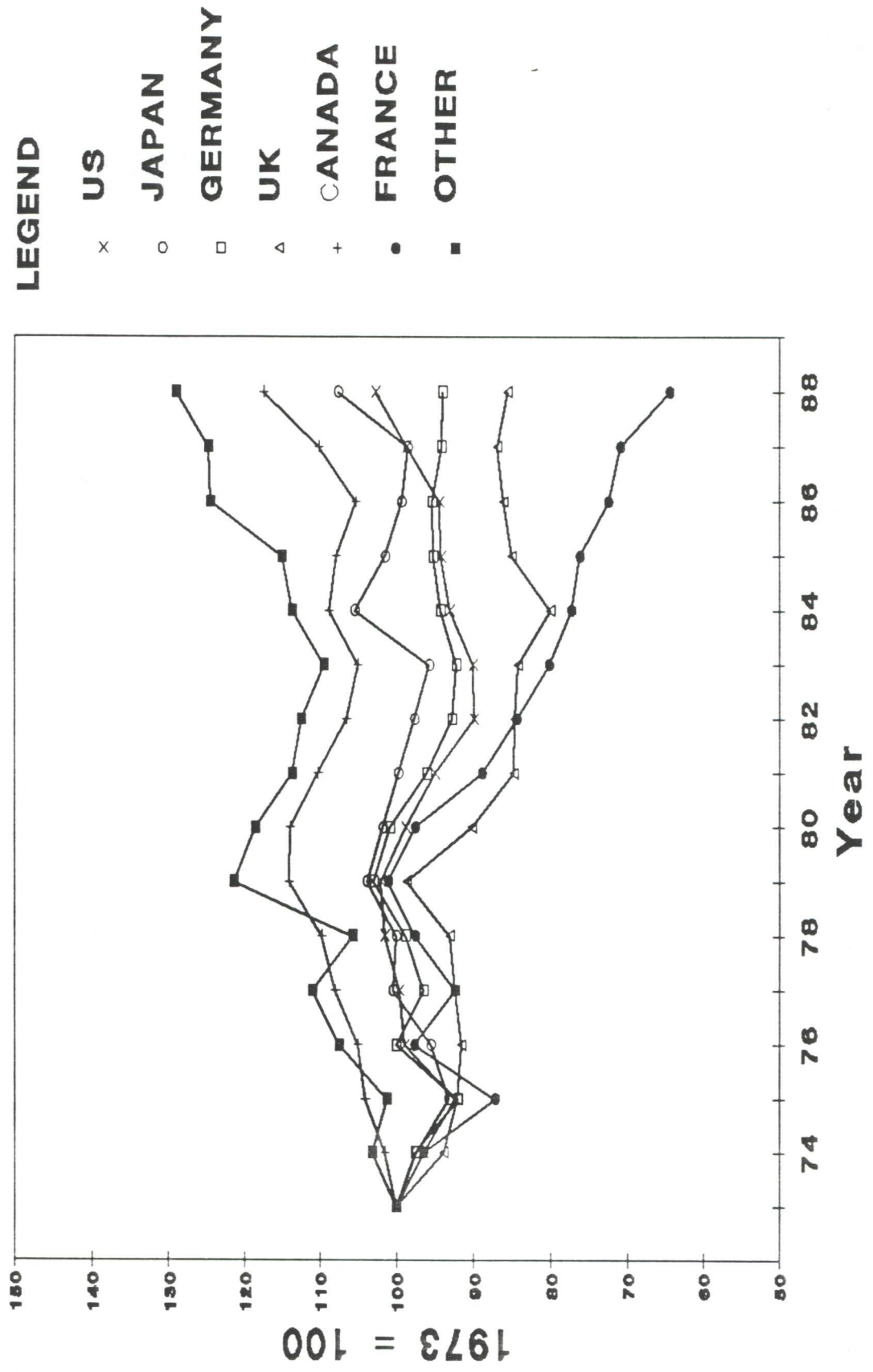


Chart 14

U. S. Production of CFCs, Methyl Chloroform (MCF) and Carbon Tetrachloride (CTC)

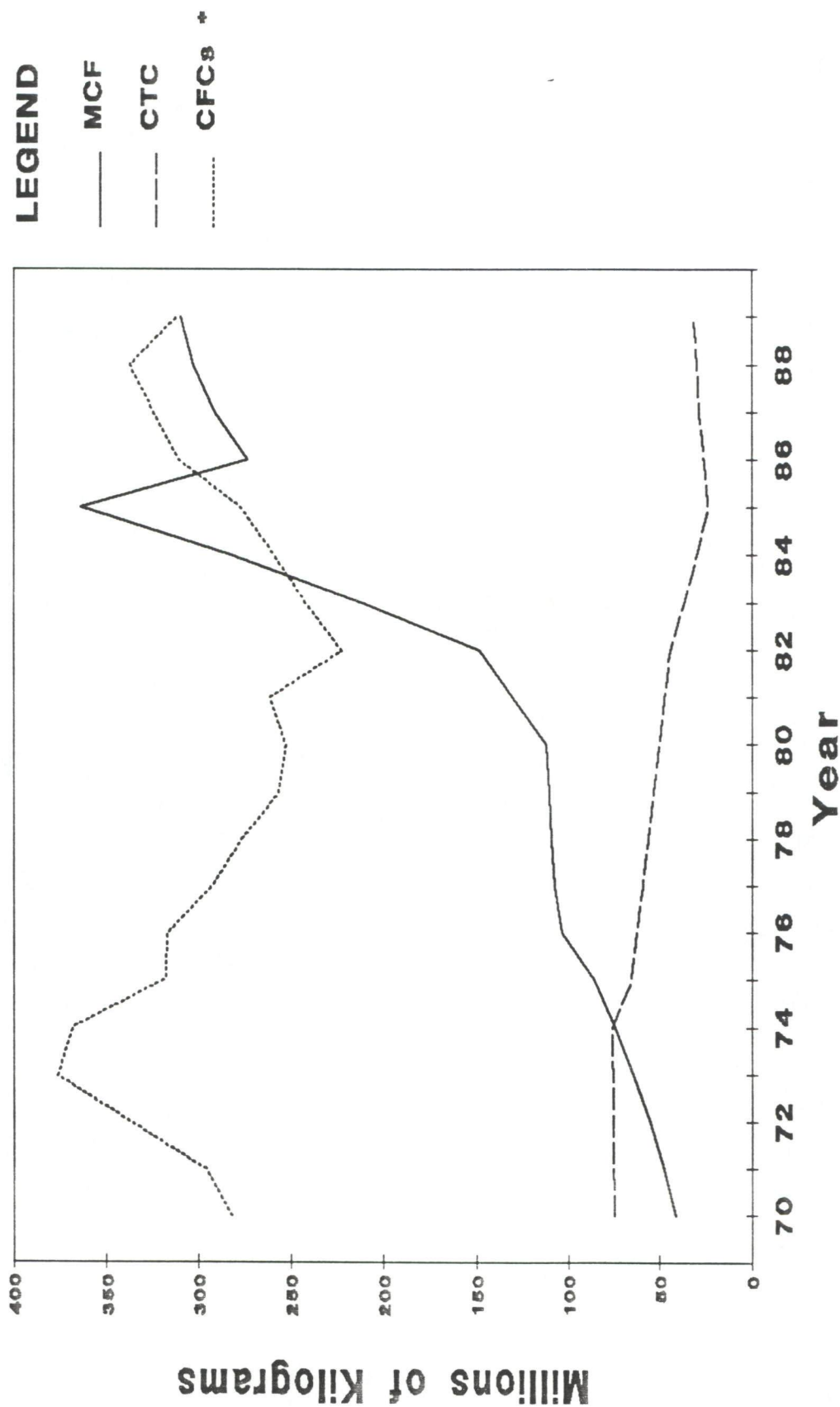


Chart 15
Trends in U. S. Emissions of Carbon Monoxide

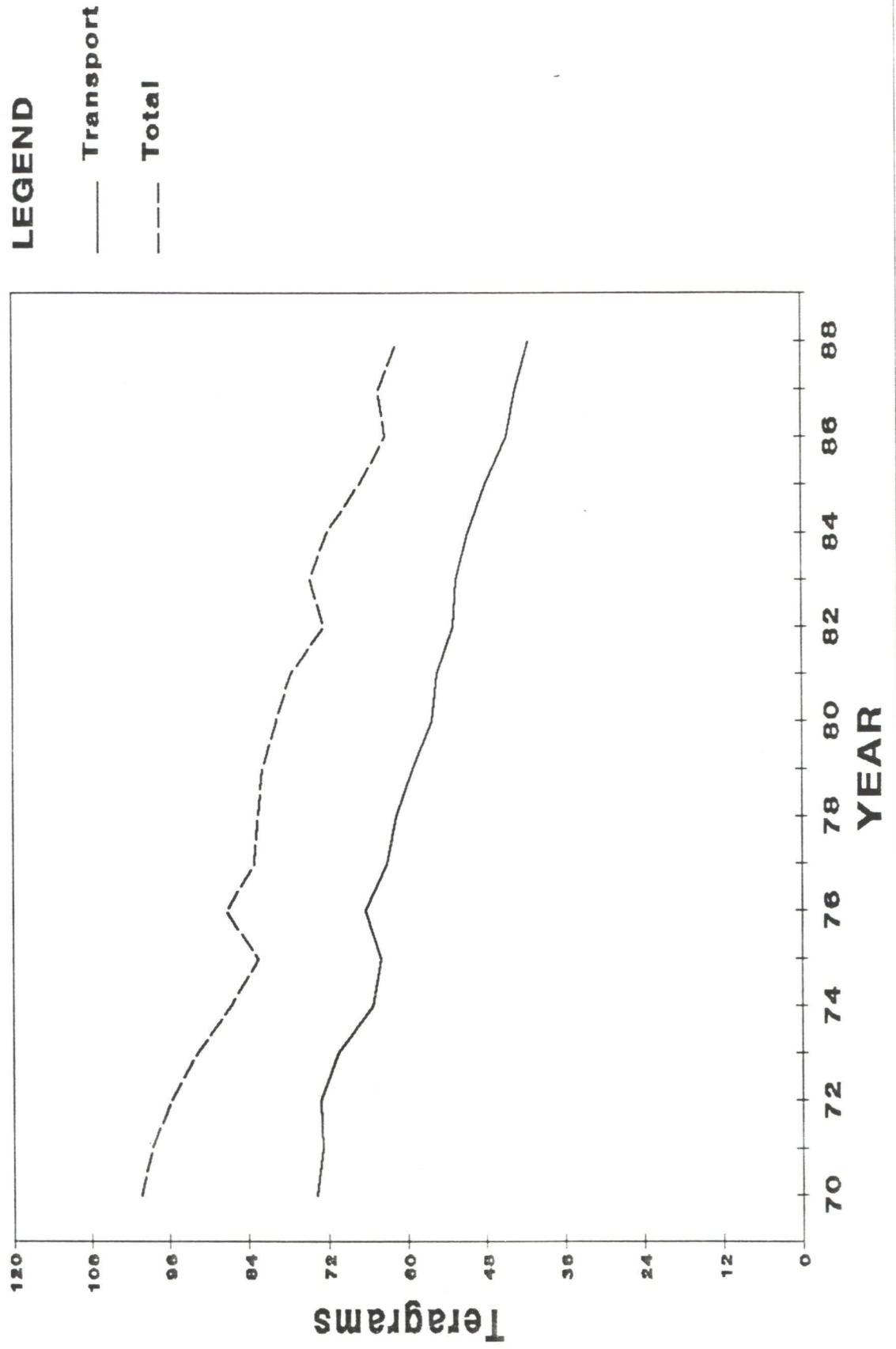


Chart 16
Trends in U. S. Emissions of Reactive
Volatile Organic Compounds

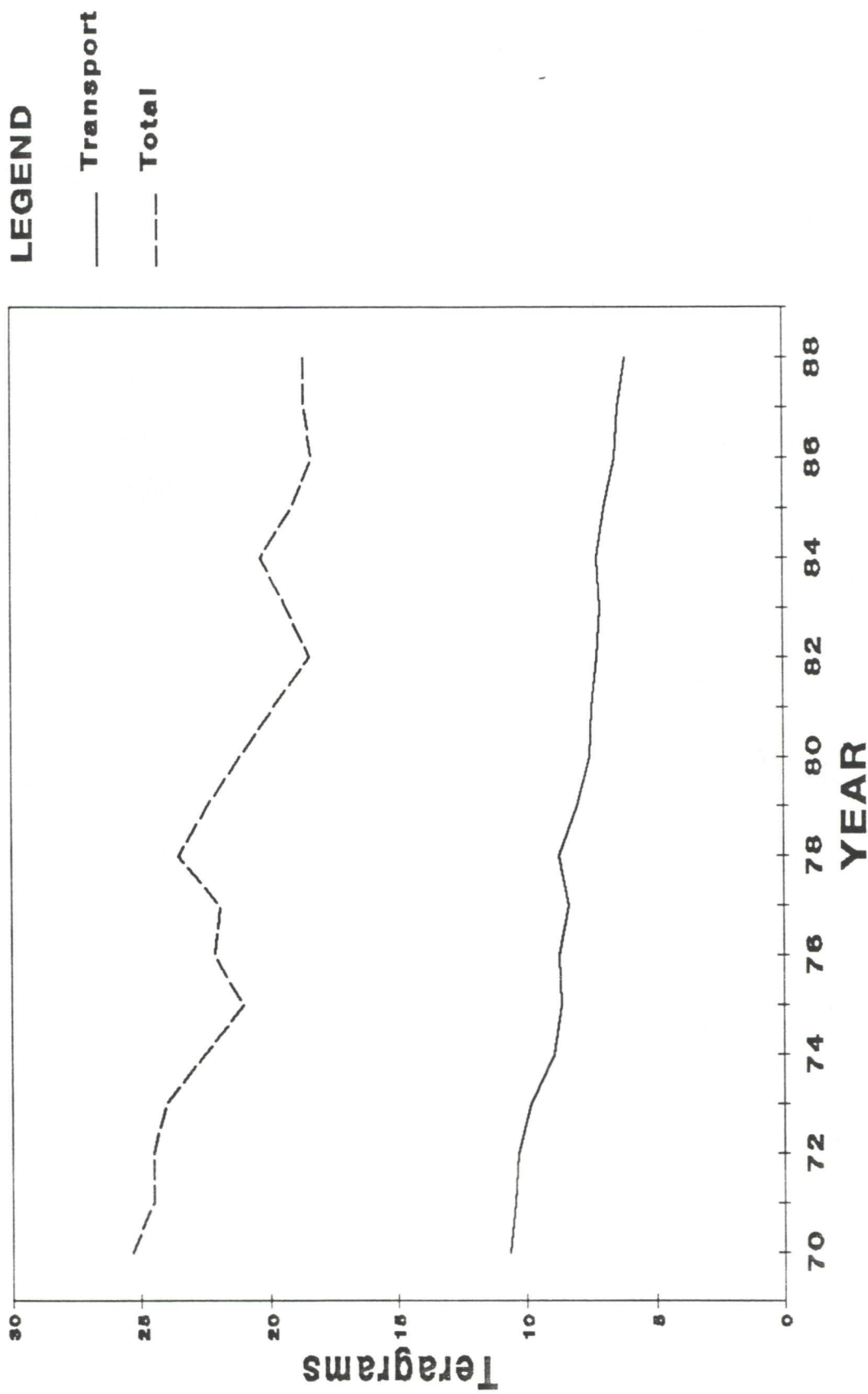
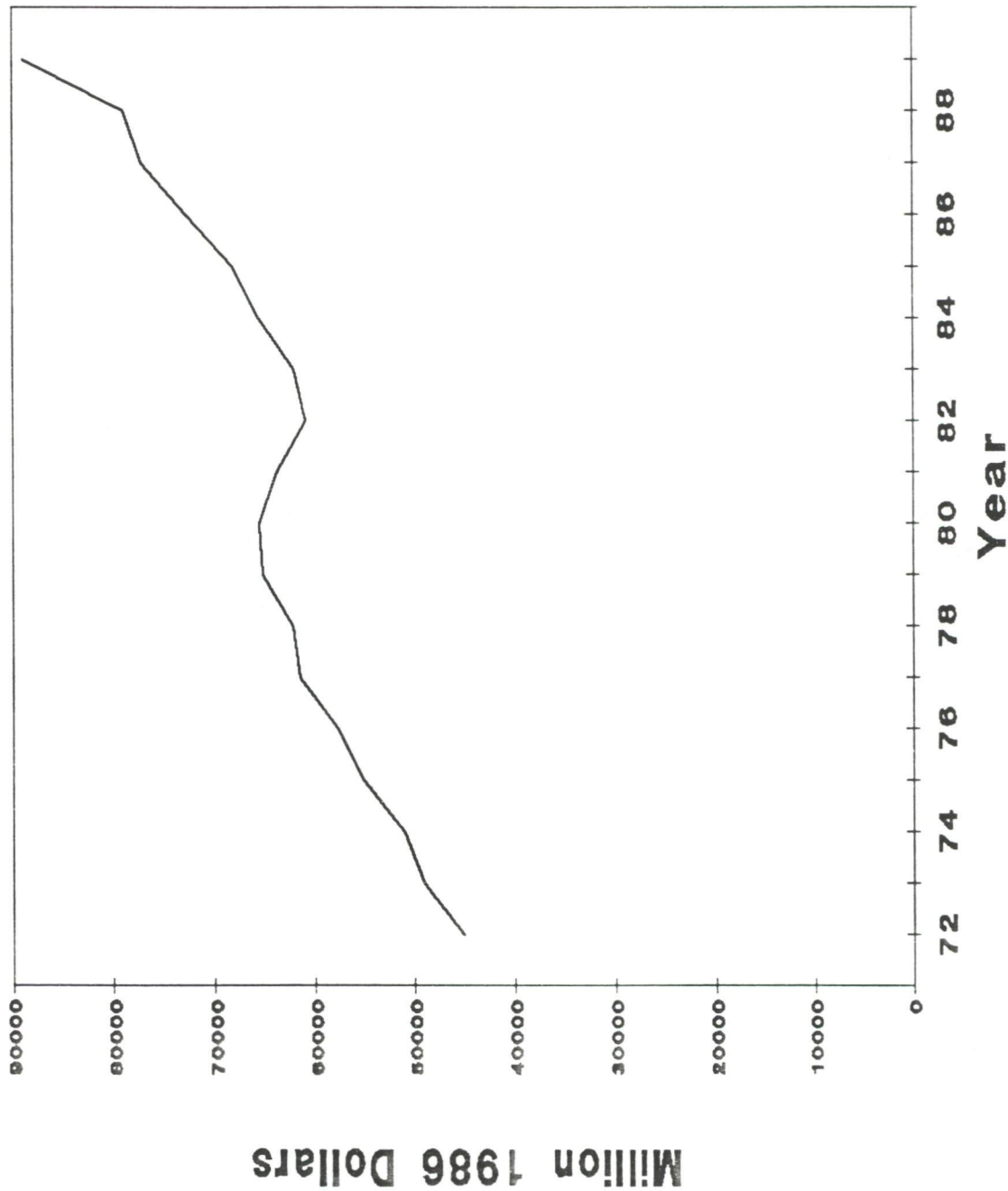


Chart 17
U. S. Pollution Control Expenditures



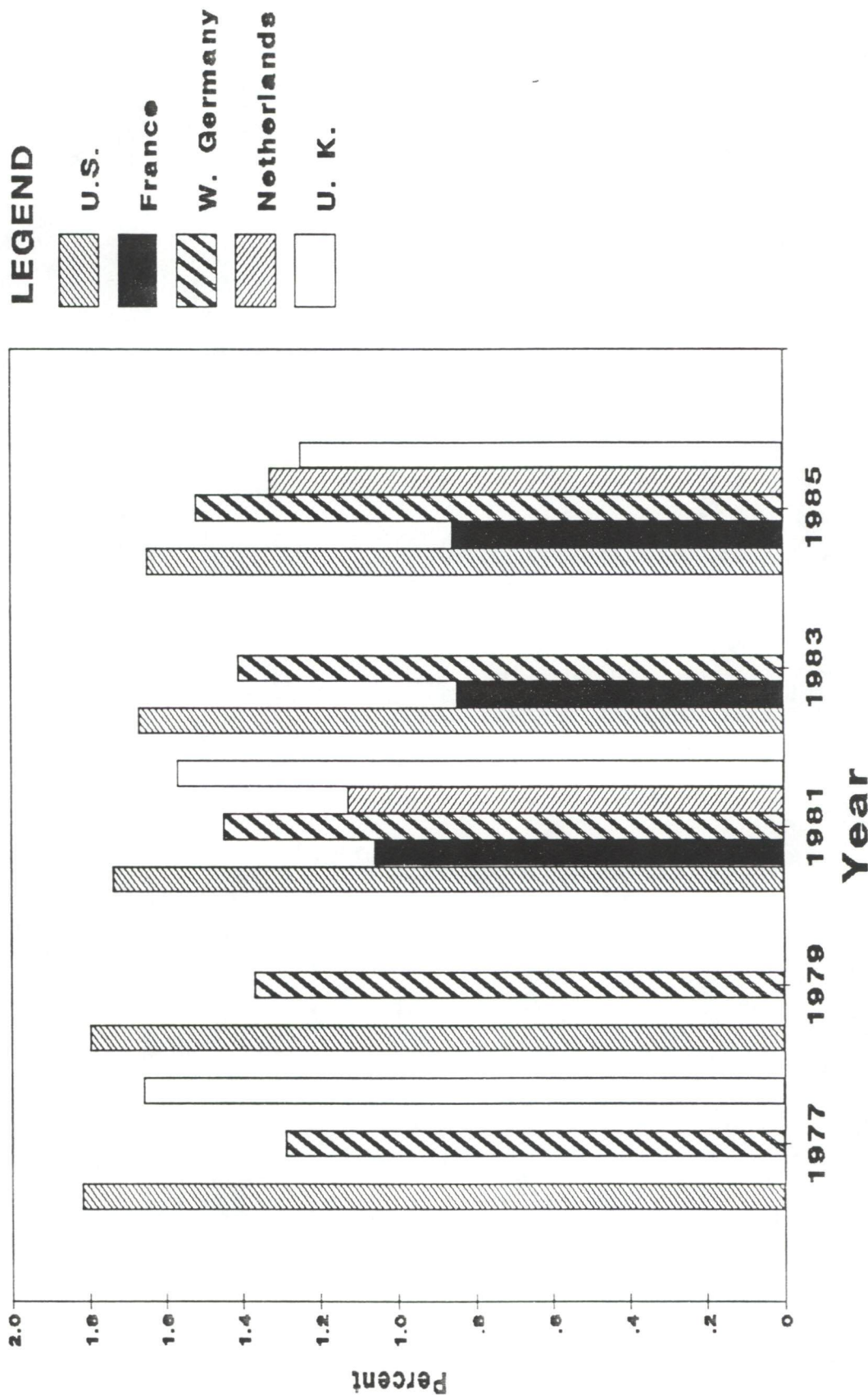
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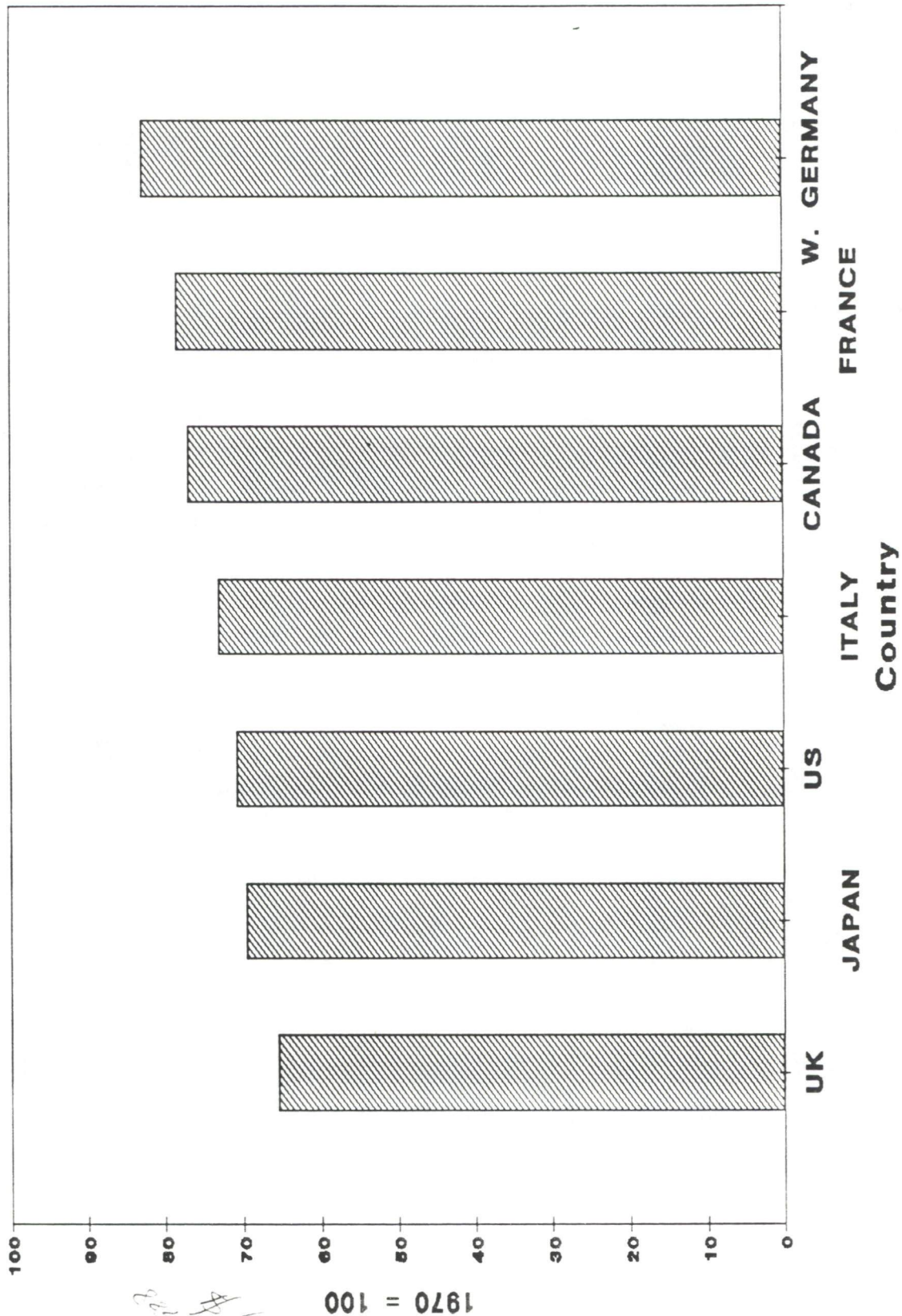
Chart 18
Pollution Control Expenditures
As a Percentage of Gross Domestic Product



*Don't know
 need to be with
 1985*

Chart 19
Energy Consumption per Unit of GDP
(Gross Domestic Product)

Chart 20
Indices of Energy Consumption / Unit of GDP, 1988



#19
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1970 = 100
 374
 1970

Chart 21
Energy Consumption/GPO in Industry
 (Gross Product Originating)

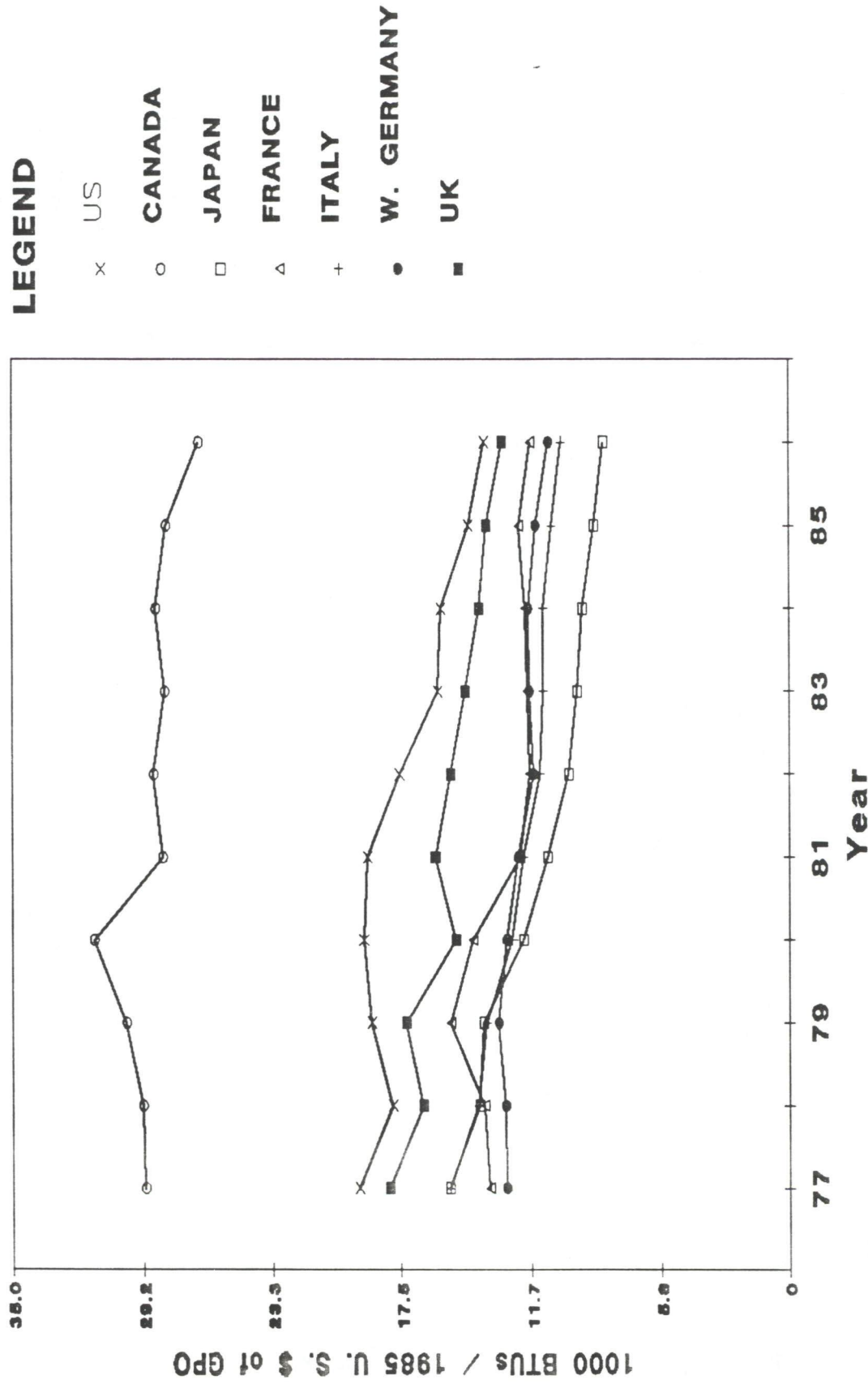


Chart 22
Indices of Energy Consumption/GPO
Industry, 1986

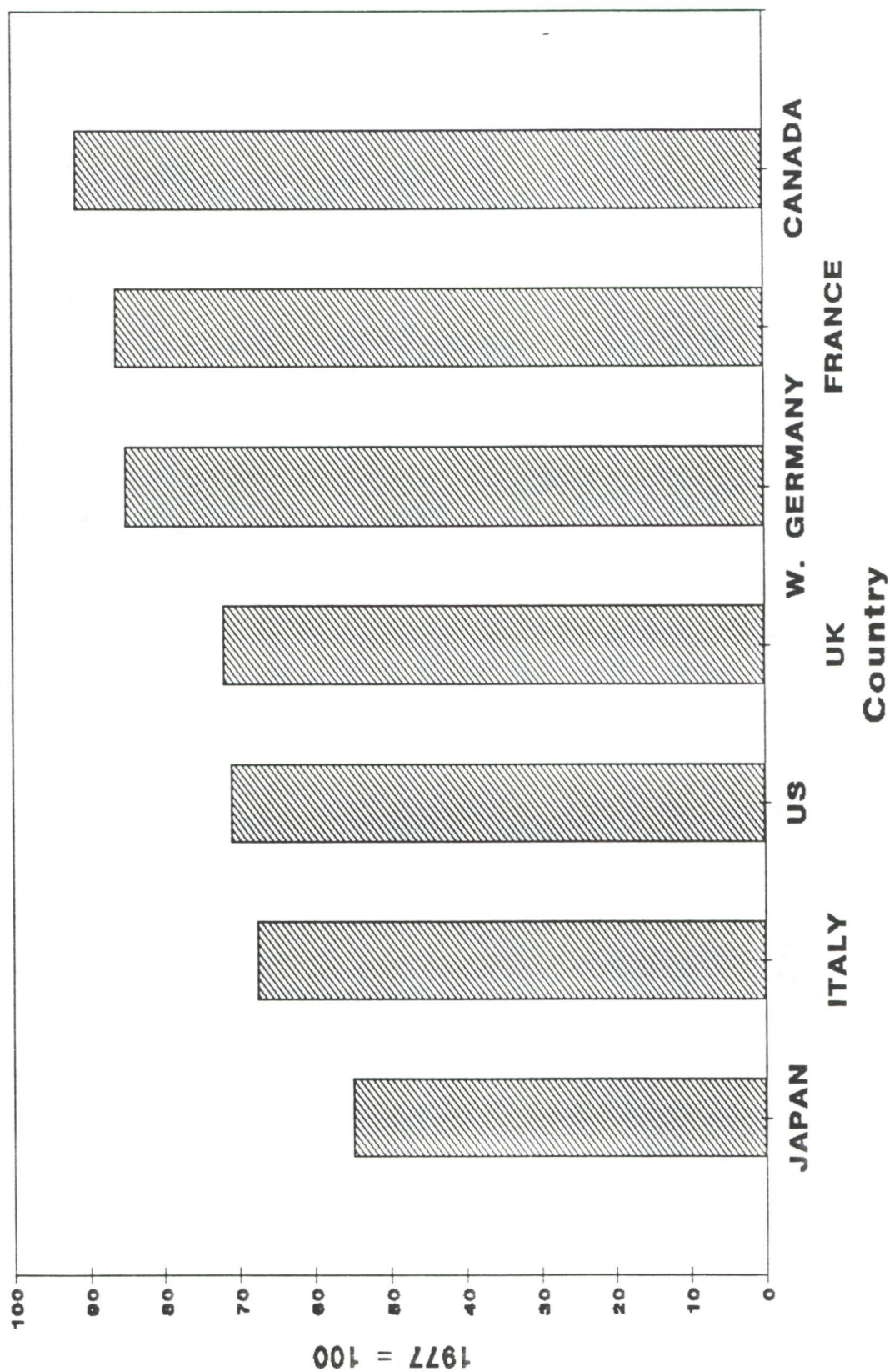


Chart 23
Population Density Per Square Mile, 1988

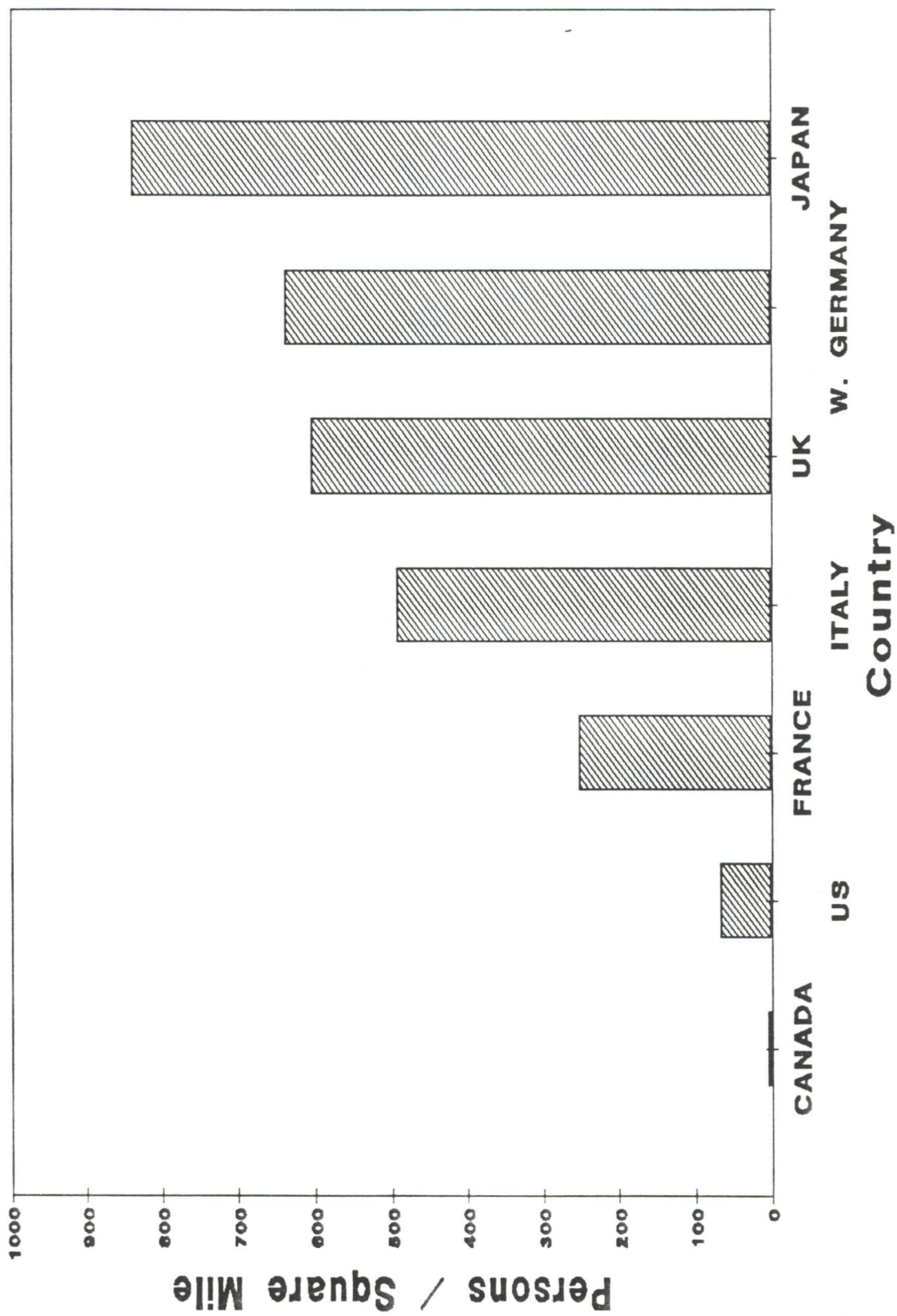


Chart 24
Transportation Energy Consumption
Per Capita, 1987

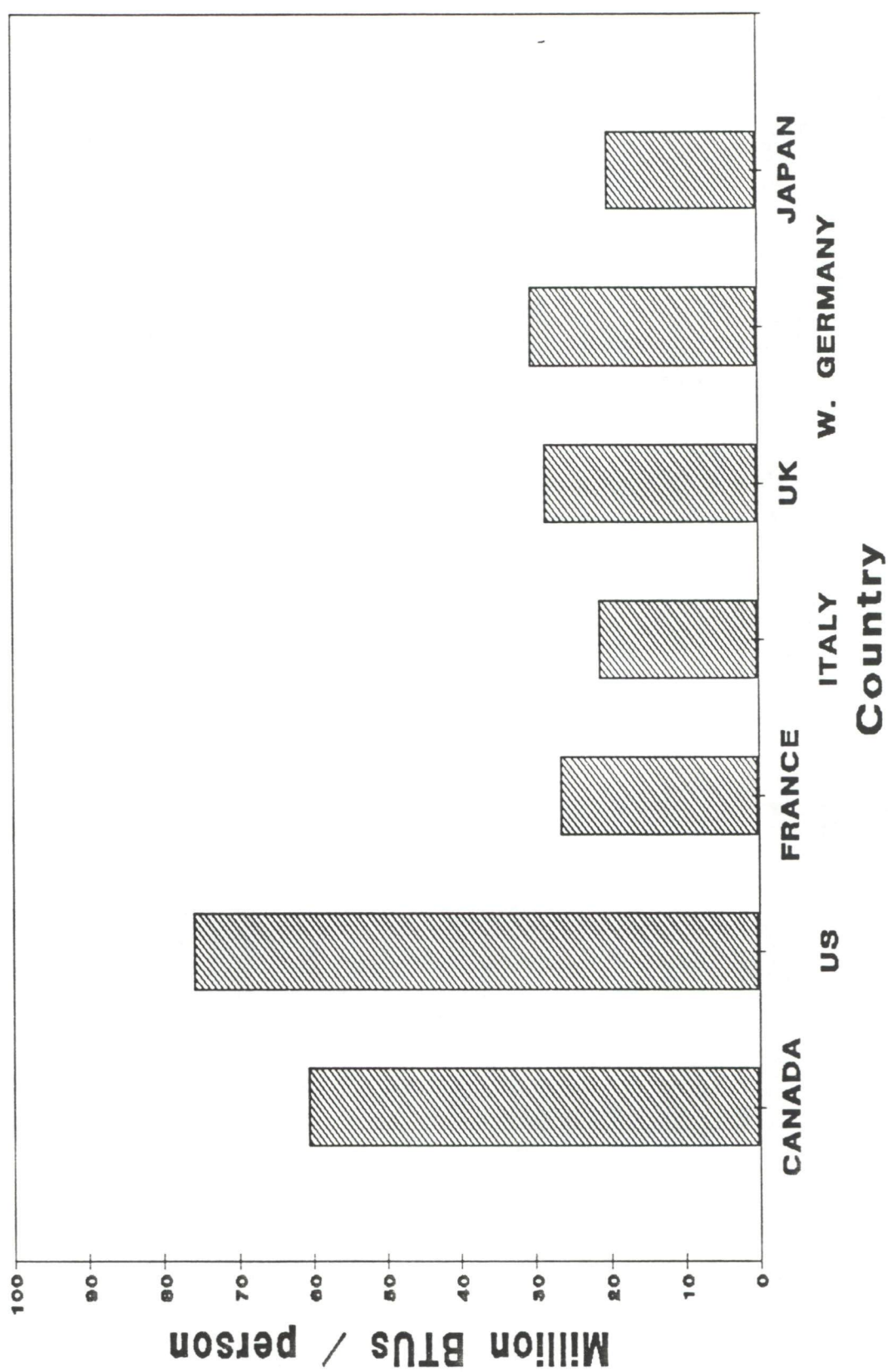
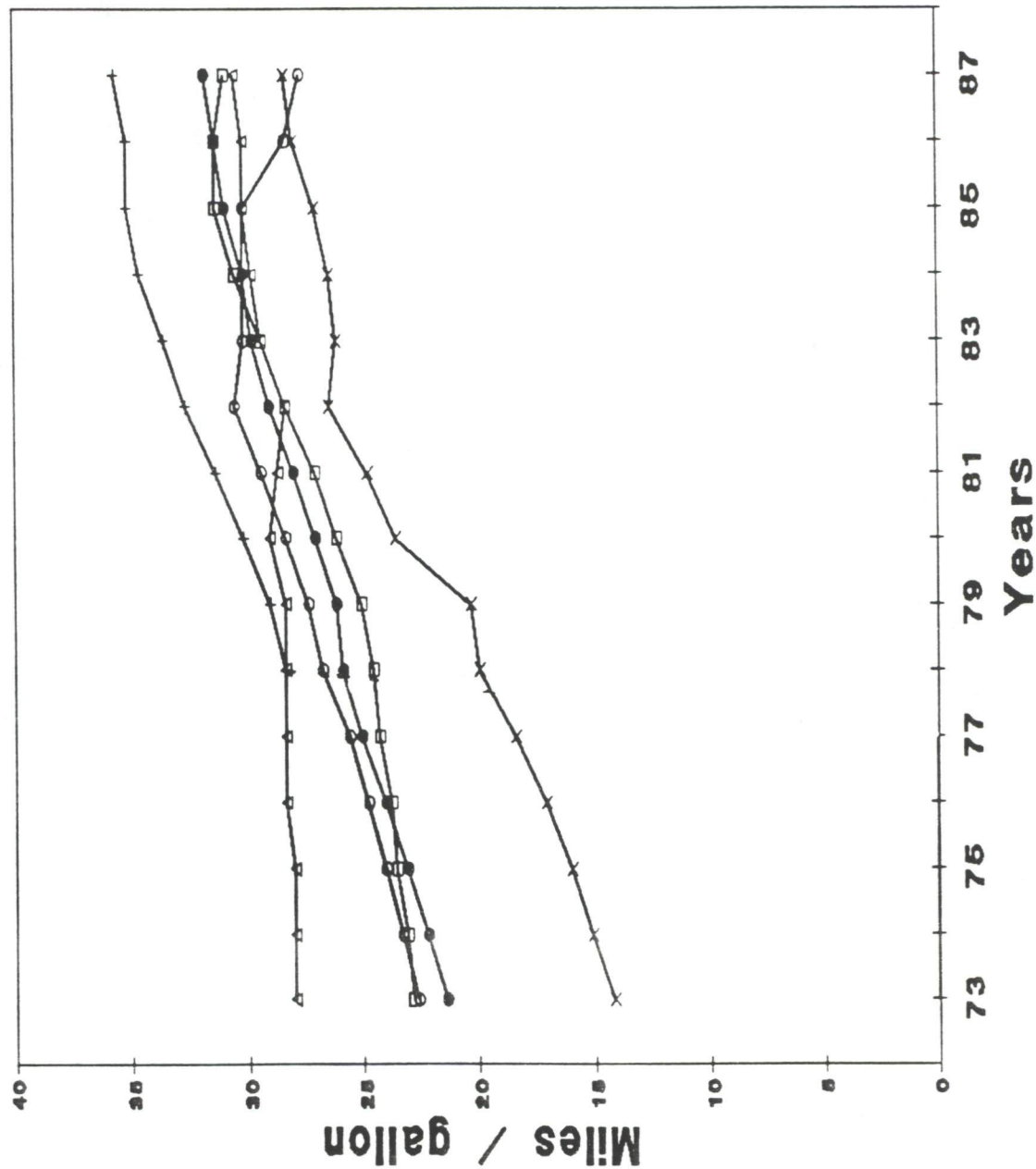


Chart 25
New Car Fuel Efficiency

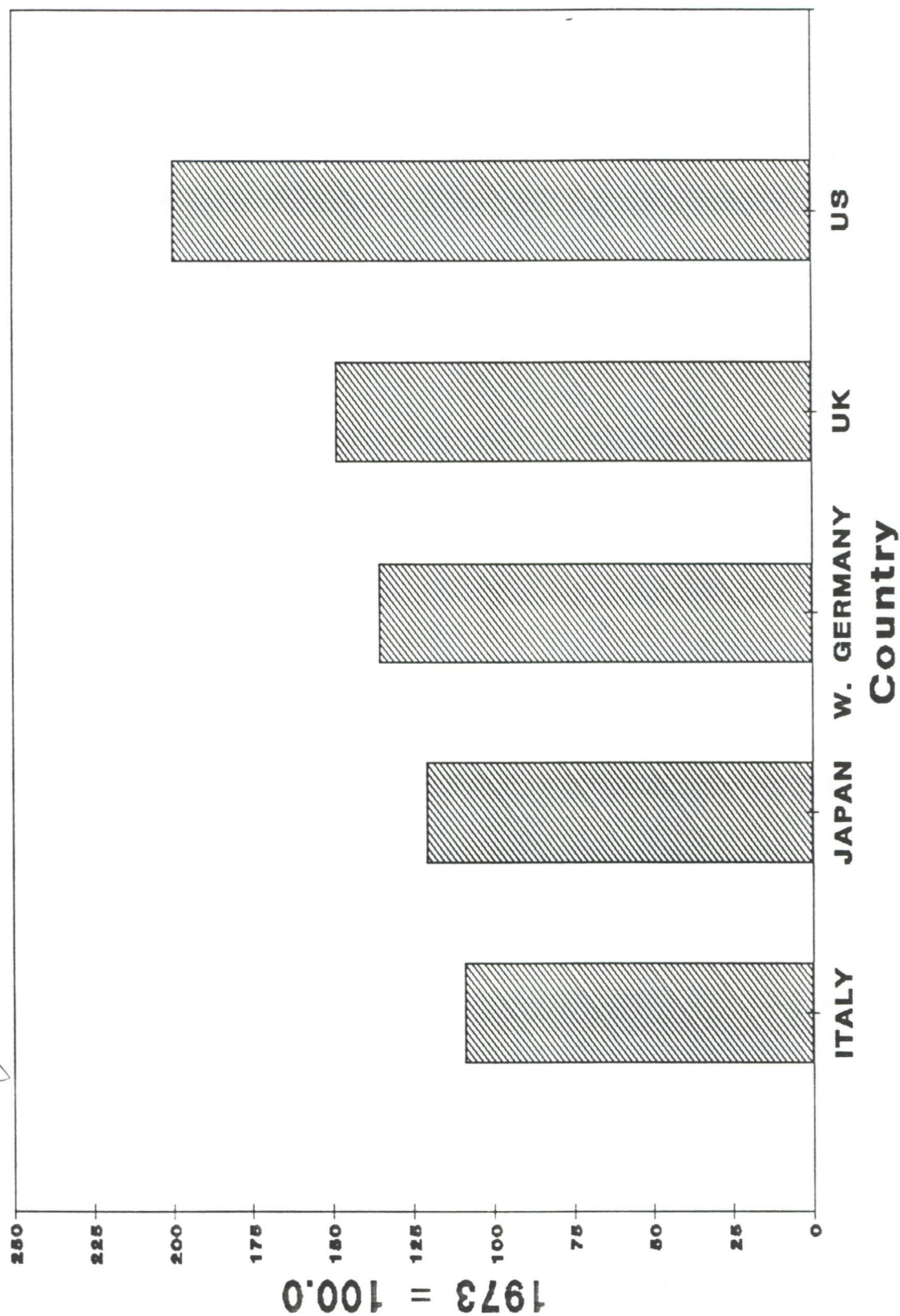


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Chart 26
Indices of New Car Fuel Economy, 1987



1973-1987
 1986-1987

Chart 27
Heating Intensities: Dwellings
(Climate Corrected)

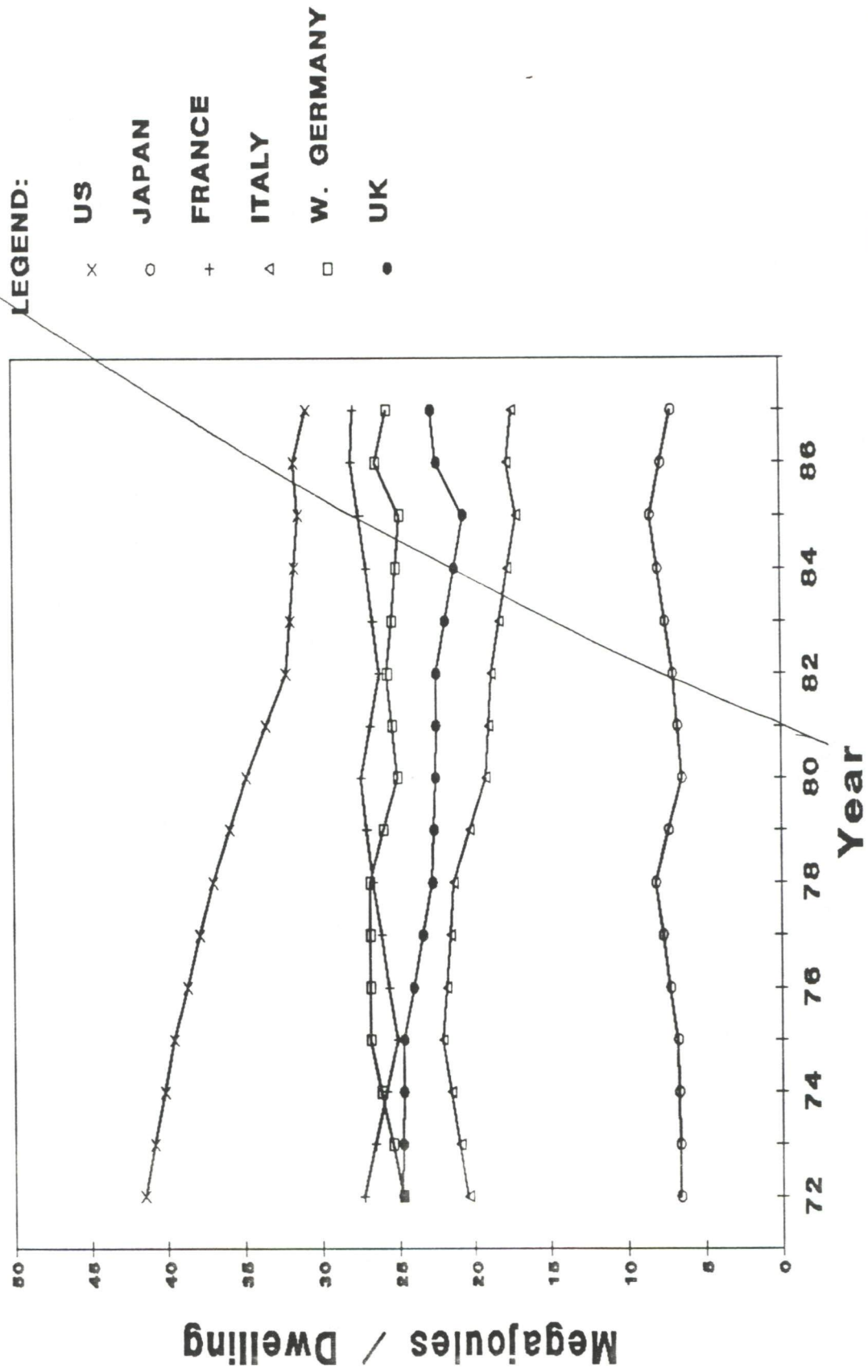


Chart 28
Indices of Heating Intensities
Dwellings, 1987

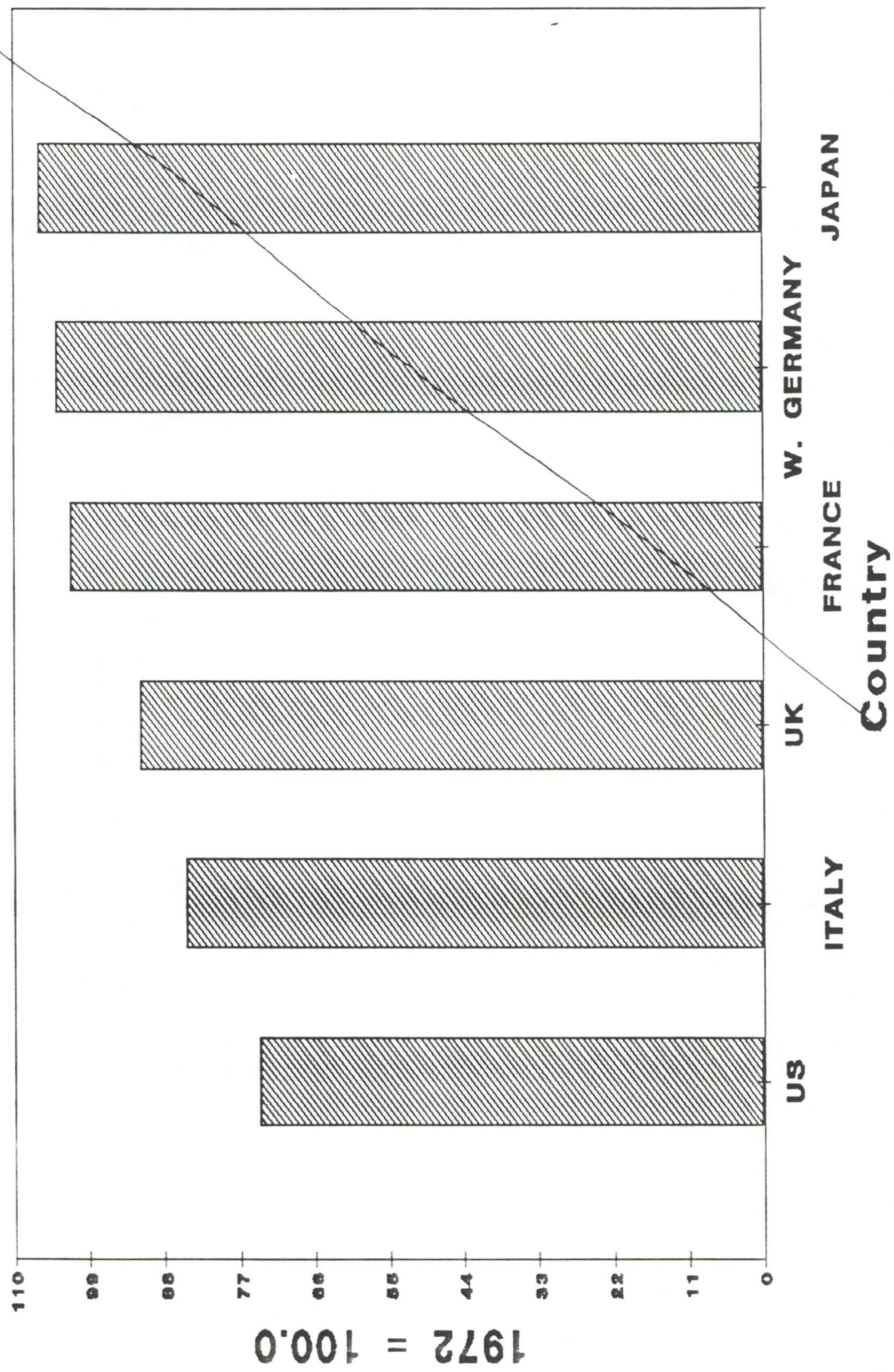


Chart 29
Heating Intensities: Floor Space
 (Climate Corrected)

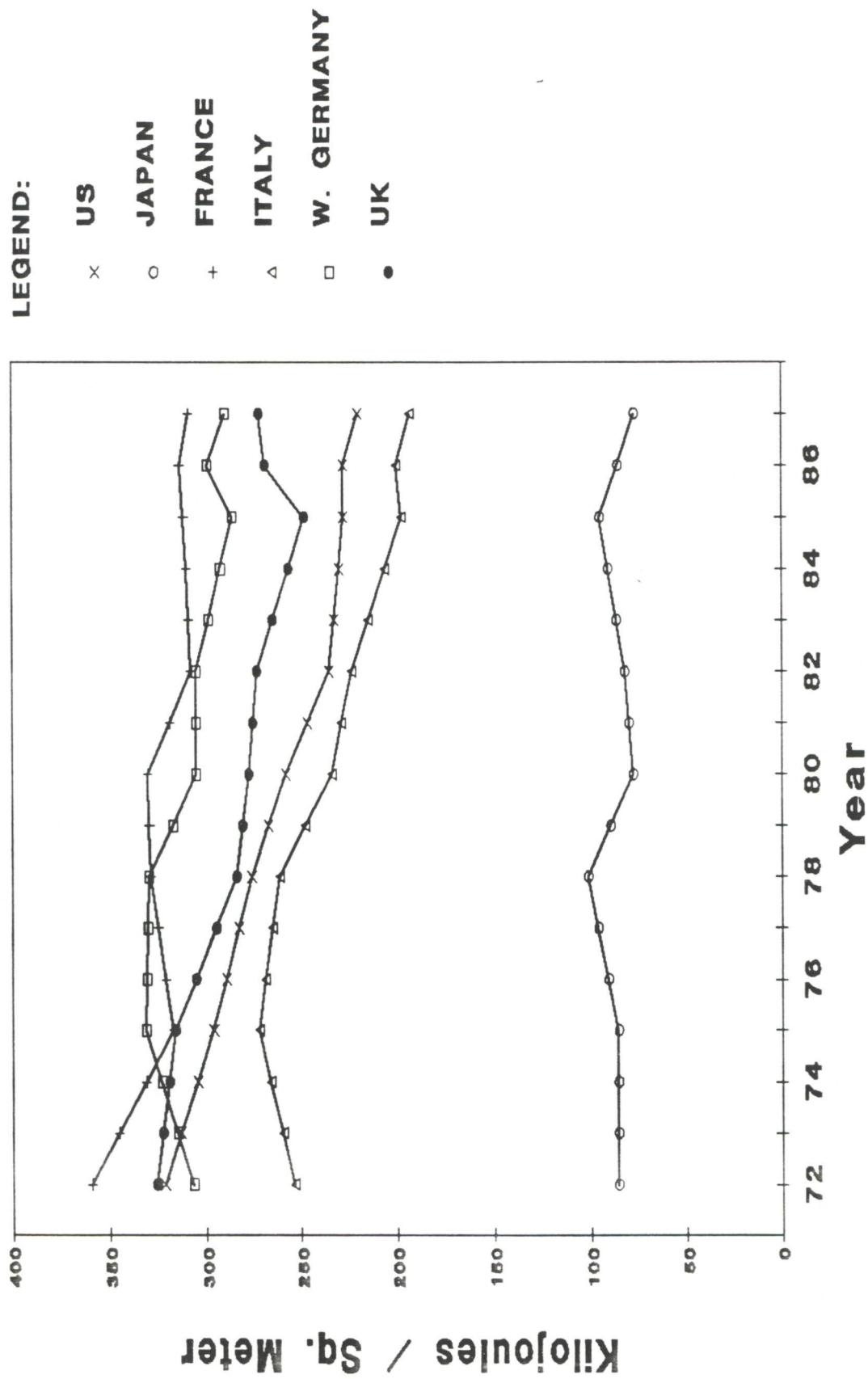
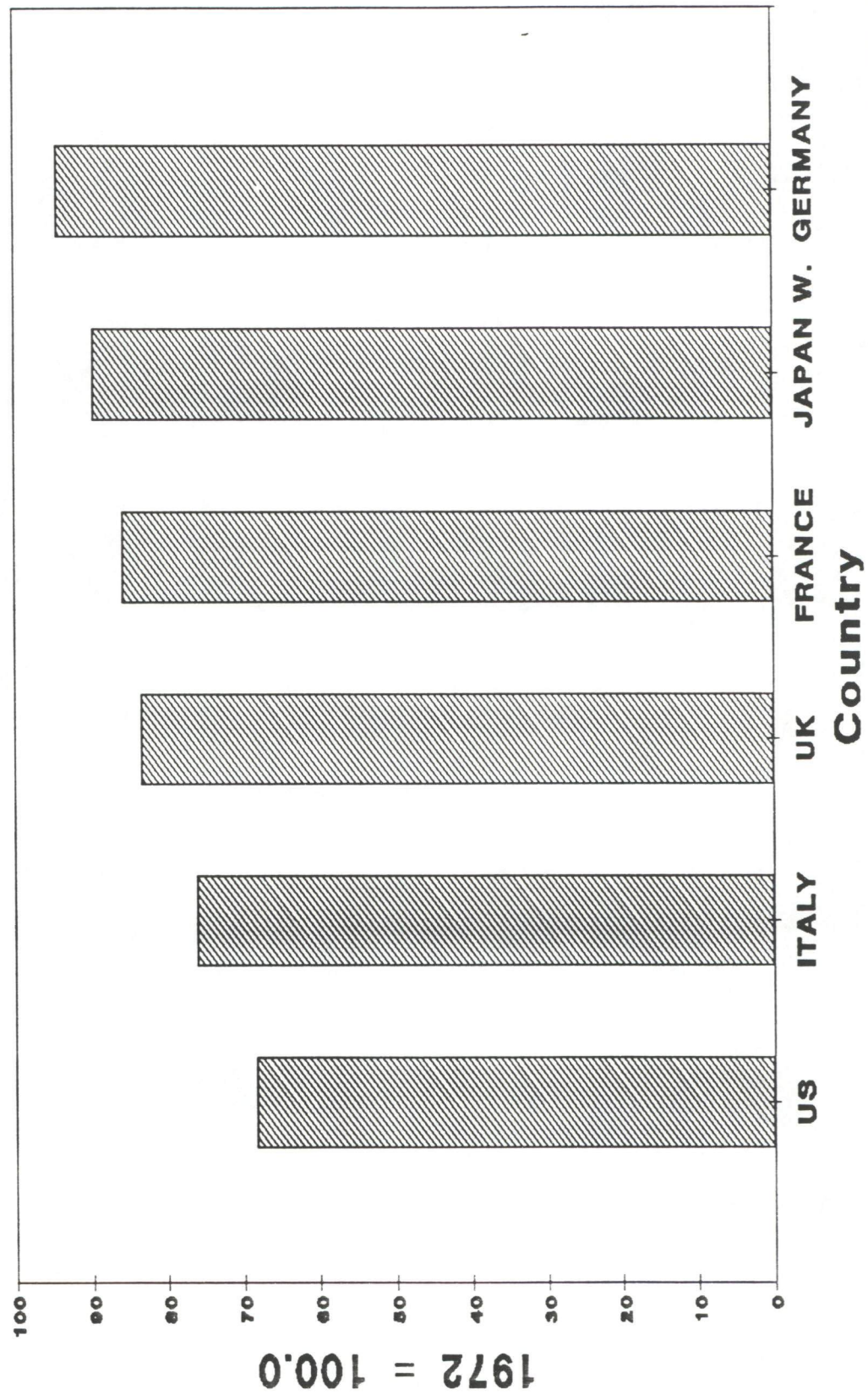


Chart 30
Indices of Heating Intensities
Floor Space, 1987
(Climate Corrected)



Global Change file

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Atmospheric Carbon Dioxide Levels Over Phanerozoic Time

ROBERT A. BERNER

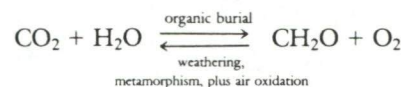
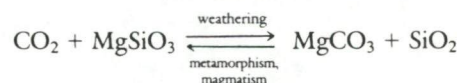
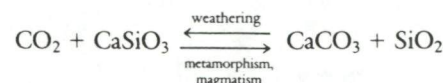
A new model has been constructed for calculating the level of atmospheric CO₂ during the past 570 million years. A series of successive steady states for CO₂ is used in order to calculate CO₂ level from a feedback function for the weathering of silicate minerals. Processes considered are: sedimentary burial of organic matter and carbonates; continental weathering of silicates, carbonates, and organic matter; and volcanic and metamorphic degassing of CO₂. Sediment burial rates are calculated with the use of an isotope mass-balance model and carbon isotopic data on ancient seawater. Weathering rates are calculated from estimates of past changes in continental land area, mean elevation, and river runoff combined with estimates of the effects of the evolution of vascular land plants. Past degassing rates are estimated from changes in the rate of generation of sea floor and the shift of carbonate deposition from platforms to the deep sea. The model results indicate that CO₂ levels were high during the Mesozoic and early Paleozoic and low during the Permo-Carboniferous and late Cenozoic. These results correspond to independently deduced Phanerozoic paleoclimates and support the notion that the atmospheric CO₂ greenhouse mechanism is a major control on climate over very long time scales.

CARBON DIOXIDE IN THE ATMOSPHERE IS A GREENHOUSE gas. In other words, because of its radiation absorbing properties, changes in CO₂ concentration over time can bring about changes in global climate. As is well known, increases in CO₂ during the past century due mainly to human activities have given rise to the environmental problem of global warming and its many and varied consequences. Also, changes in CO₂ levels, due mainly to natural perturbations of the ocean-atmosphere-biosphere system, have been documented for the past 150,000 years by the analysis of bubbles of air trapped in glacial ice (1). Thus, the atmospheric CO₂ level does change with time, but what about at longer time scales? Studies of the geochemical carbon cycle (2) have shown that the transfer of carbon from rocks to and from the surficial environment over millions of years is far greater than that present at any one time. Thus, it is quite reasonable to speculate that natural long-term processes involving carbon stored in rocks have also brought about changes in atmospheric CO₂ (but at rates that are far too slow to explain the changes observed on human time

scales). The purpose of this article is to suggest that gradual CO₂ variations have occurred during the past 570 million years (Phanerozoic time) and that they have been large.

The approach adopted is to model the geochemical carbon cycle and its effect on atmospheric CO₂. The geochemical carbon cycle (2) refers to the transfer of carbon between rocks and the ocean-atmosphere system on a multimillion-year time scale and is illustrated in Fig. 1. The ocean and atmosphere are lumped together in Fig. 1 because of their very small carbon contents and short residence times compared to that of the rock reservoirs. The major processes of the cycle are: (i) uptake of CO₂ from the atmosphere during the weathering of silicate and carbonate rocks and transformation of it to dissolved HCO₃⁻ in ground water and river water; (ii) precipitation of the HCO₃⁻ in the oceans, after transfer there by rivers, as CaCO₃ and other carbonate minerals; (iii) the thermal decomposition of carbonate minerals upon burial to great depths by way of magmatic and metamorphic activity, resulting in transfer of CO₂ back to the atmosphere plus oceans; (iv) the burial of organic matter, representing the remains of dead organisms, in sediments followed much later by the weathering of this material after uplift onto the continents; and (v) the thermal breakdown at depth via diagenesis, metamorphism, and magmatism of organic matter followed by air oxidation of reduced carbon-containing gases upon evasion to the atmosphere.

The major processes of the geochemical carbon cycle can be summarized most simply by the following reactions:



The first two reactions represent the combination of steps (i), (ii), and (iii) above and are normally referred to as the Urey reactions (3). (The formulas given for minerals are generalized to represent all calcium and magnesium silicates and carbonates and do not refer to any specific mineral.) The third reaction is of the same form as that normally written for the processes of photosynthesis and respiration but refers instead to their geological equivalents. In other words, going from left to right the reaction refers to net photosynthesis or the difference between global photosynthesis and respiration as represented by the burial of organic matter in sediments. Likewise, the reaction going from right to left represents the oxidation of rocks and volcanic gases, analogous to long-term respiration.

The weathering of carbonate minerals, which is an essential

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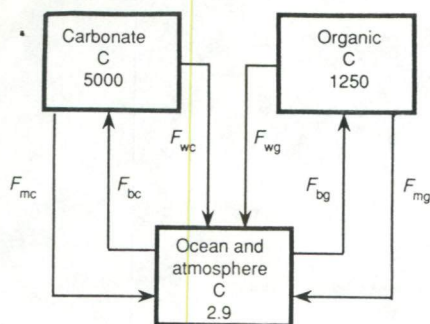


Fig. 1. Schematic representation of the geochemical carbon cycle. Masses of carbon (in 10^{18} moles) are based on the data of (6). Symbols for fluxes are defined in Table 1.

component of the geochemical carbon cycle (Fig. 1), is not represented in the above reactions. This is because the atmospheric CO_2 that is taken up by this reaction is soon returned to the atmosphere during precipitation of HCO_3^- in the ocean as carbonate minerals (2). On a multimillion-year time scale there is a negligible effect on atmospheric CO_2 . By contrast, because of the long lag between the transformation of calcium and magnesium silicates to carbonates and the thermal decarbonation of these carbonates, a large imbalance in CO_2 fluxes can occur. In other words, it is the Urey reactions (and to a lesser extent the organic matter reactions) that dominate the system.

The Model

Modeling of the geochemical carbon cycle, in whole or in part, has been done by many workers (4). The model presented here is basically a greatly expanded steady-state version of the carbon mass balance model of Garrels and Lerman (5). It is much simpler mathematically than those that the BLAG group constructed earlier (6, 7), but more complex geologically and biologically. Mathematical simplification comes about by lumping all carbonate minerals together, following only the cycle of carbon and not that of other elements, combining the oceans and atmosphere into one reservoir, and calculating CO_2 level in the atmosphere as a succession of steady states. The steady-state assumption has been shown to be justified by other studies (6, 8, 9). The procedure is to assume that during

each 1-million-year period (the time step of the numerical model) inputs of CO_2 to the ocean-atmosphere system are balanced by outputs. This assumption is reasonable because of the very rapid turnover of carbon in the ocean-atmosphere system—a result of its very small capacity for carbon storage compared to that of rocks.

Geological and biological complexity are added to the Garrels and Lerman model by introducing dimensionless parameters to represent the effects of geological and biological evolution over time. Because of the assumption of a steady-state atmosphere-plus-ocean reservoir, the actual calculation of CO_2 level is done differently than that used in the earlier models (6, 7). The method is to calculate the value of the feedback function for weathering $f_B(\text{CO}_2)$ by a combination of mass balance expressions for both total carbon and carbon-13 and expressions for weathering and metamorphic and volcanic degassing. Atmospheric CO_2 is then calculated from $f_B(\text{CO}_2)$. The appropriate expressions are:

$$F_{wc} + F_{mc} + F_{wg} + F_{mg} = F_{bc} + F_{bg} \quad (1)$$

$$\delta_c(F_{wc} + F_{mc}) + \delta_g(F_{wg} + F_{mg}) = \delta_o F_{bc} + (\delta_o - \alpha_c) F_{bg} \quad (2)$$

$$F_{wc} = f_B(\text{CO}_2) f_A(t) f_D(t) f_E(t) k_{wc} C \quad (3)$$

$$F_{mc} = f_G(t) f_C(t) k_{mc} C \quad (4)$$

$$F_{wg} = f_A(t) f_D(t) f_R(t) k_{wg} G \quad (5)$$

$$F_{mg} = f_G(t) k_{mg} G \quad (6)$$

$$F_{ws} = F_{bc} - F_{wc} = f_B(\text{CO}_2) f_A(t) f_D(t) f_R(t) f_E(t) F_{ws}(0) \quad (7)$$

where all terms are defined in Table 1.

The procedure of calculation was as follows. Values of the dimensionless parameters $f_A(t)$, $f_D(t)$, $f_R(t)$, $f_E(t)$, $f_G(t)$, and $f_C(t)$ and values for k_{wc} , k_{mc} , k_{wg} , k_{mg} , and δ_o were taken from the literature or estimated (see Table 2 and Figs. 2 and 3). Initial values for C , G , $f_B(\text{CO}_2)$, δ_c , and δ_g at 570 million years ago (Ma) were assumed, and the program was run. At each 1-million-year time step appropriate values of F_{wc} , F_{mc} , F_{wg} , and F_{mg} were calculated and substituted into Eqs. 1 and 2, and the two equations were then solved for F_{bc} and F_{bg} . From values of F_{bc} and F_{wc} , Eq. 7 was solved for $f_B(\text{CO}_2)$ and, thus, for CO_2 level (see below). The process was

Table 1. Definition of terms used in equations.

Parameter	Definition
F_{wc} , F_{wg}	= rate of release of carbon to the ocean-atmosphere system by the weathering of carbonates (c) and organic matter (g)
F_{mc} , F_{mg}	= rate of carbon release to the ocean-atmosphere by the metamorphic and volcanic breakdown of carbonates (c) and organic matter (g)
F_{bc} , F_{bg}	= burial rate of carbon as carbonates (c) and as organic matter (g) in sediments
F_{ws}	= rate of uptake of CO_2 for the weathering of Ca and Mg silicates followed by precipitation of the Ca and Mg as carbonates (Urey reactions)
$F_{ws}(0)$	= value of F_{ws} at present
$f_B(\text{CO}_2)$	= dimensionless feedback function expressing the dependence of weathering on atmospheric CO_2
RCO_2	= ratio of atmospheric $\text{CO}_2(t)$ to atmospheric $\text{CO}_2(0)$
$f_A(t)$	= ratio of land area (t)/land area (0)
$f_D(t)$	= ratio of river runoff (t) to river runoff (0) (per unit area) due to changes in paleogeography. [Dependence of runoff on atmospheric CO_2 is included in $f_B(\text{CO}_2)$; see (6)]
$f_R(t)$	= ratio of mean land elevation (t) to mean land elevation (0)
$f_E(t)$	= dimensionless parameter expressing the dependence of weathering rate on soil biological activity due to land plants [$f_E(t) = 1$ at present]
$f_G(t)$	= degassing rate (t) divided by degassing rate (0) due to tectonics; assumed to be equal to the rate of new sea-floor generation (sea-floor spreading rate)
$f_C(t)$	= dimensionless parameter expressing the dependence of degassing rate on the relative proportions of carbonates on shallow platforms and in the deep sea [$f_C(t) = 1$ at present]
δ	= $\delta^{13}\text{C}$ value (‰); δ_c for carbonates, δ_g for organic matter, and δ_o for the oceans
α_c	= isotope fractionation factor between organic matter and carbonates
k_{wc} , k_{wg}	= rate constants for weathering of carbonates and organic matter, respectively
k_{mc} , k_{mg}	= rate constants for metamorphic and volcanic breakdown of carbonates and organic matter, respectively
C , G	= masses of carbon as carbonates and as organic matter
t	= time
(0)	= present (average for past 1 million years)

Fig. 2. Plots versus time of $f_A(t)$, land area at time t divided by land area today, and $f_D(t)$ (continental runoff at time t divided by continental runoff today). Values of $f_A(t)$ are based on the paleogeographic reconstructions of Barron and others (22) and Ronov (23). Values of $f_D(t)$ are based on the work of Tardy *et al.* (24), who calculated runoff changes per unit area by keeping track of changes in precipitation minus evaporation on the continents that resulted from past continental drift beneath fixed (present-day) latitudinal climatic zones. No account of changes in orography or monsoonality was made; my, million years.

then iterated until a constant value of $f_B(\text{CO}_2)$ was obtained (rapid convergence required minimal iteration). New values of C , G , δ_c , and δ_g were then calculated with the use of four additional mass balance expressions:

$$dC/dt = F_{bc} - (F_{wc} + F_{mc}) \quad (8)$$

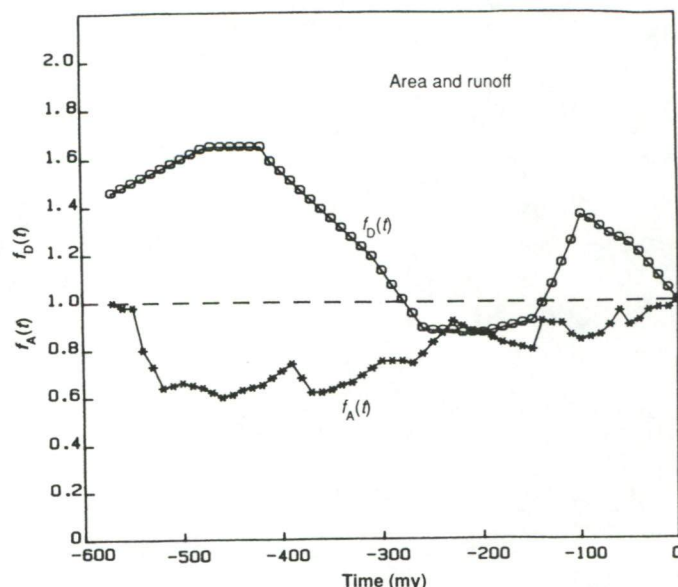
$$dG/dt = F_{bg} - (F_{wg} + F_{mg}) \quad (9)$$

$$d(\delta_c C)/dt = \delta_o F_{bc} - \delta_c (F_{wc} + F_{mc}) \quad (10)$$

$$d(\delta_g G)/dt = (\delta_o - \alpha_c) F_{bg} - \delta_g (F_{wg} + F_{mg}) \quad (11)$$

and the procedure was repeated at the next 1-million-year time step and so on. Initial values were then adjusted so that present-day values for the various fluxes were retrieved (see Table 2). Use of shorter time steps was found to be unnecessary in that the calculation procedure converged rapidly.

As one can see, the terms $f_E(t)$ and $f_B(\text{CO}_2)$ are omitted from the expression for the weathering of organic matter. This was done because organic matter is attacked by O_2 , not CO_2 , and the biological production of CO_2 or soil acids should have little effect on



organic matter oxidation. The parameter $f_R(t)$ is omitted from the weathering expression for carbonates because their weathering has been shown to be relatively unaffected by changes in elevation or relief (10).

Once values of the weathering feedback function, $f_B(\text{CO}_2)$, were obtained the value of RCO_2 , the ratio of mass of CO_2 in the atmosphere at time t to that at present, was calculated for each time period. For this purpose expressions were used that expand on the earlier BLAG formulation for $f_B(\text{CO}_2)$ (6). In the BLAG model weathering feedback appears only in terms of changes in global mean surface temperature that are attributed to changes in atmospheric CO_2 levels according to the greenhouse effect. There is no direct response of weathering to atmospheric CO_2 levels, and changes in temperature are felt as changes in river runoff and in temperature-controlled rates of CO_2 uptake (HCO_3^- production) by weathering. In the present model allowance for the direct fertilization by CO_2 of land plant productivity is included, and this change results both in greater rock weathering and a stronger negative feedback. The BLAG feedback term, which can be represented in simplified form as $f_B(\text{CO}_2) = \text{RCO}_2^{0.22}$, is multiplied by an additional CO_2 fertilization term on the basis of a similar approach by Volk (9). In other words:

$$f_B(\text{CO}_2) = [2\text{RCO}_2/(1 + \text{RCO}_2)]^{0.4} \text{RCO}_2^{0.22} \quad (12)$$

The first term on the right within brackets represents a Michaelis-Menten type formulation, and is used to express the concept that global productivity at high atmospheric CO_2 levels is ultimately limited by factors other than the CO_2 level. The 0.4 power signifies that on a global scale many natural (nonagricultural) plant ecosystems probably respond more weakly to CO_2 because of nutrient, water, or light limitation (11), and that weathering rate does not linearly track productivity (9). It is equivalent to assuming that only 35% of land vegetation responds to CO_2 fertilization as it affects weathering rate. Because the global response of plant productivity to CO_2 is not well established (11), let alone its effect on rock weathering, Eq. 12 should be considered only as a crude first-order approximation. However, regardless of exact formulation, it is necessary to introduce stronger feedback than is available in the BLAG model because over Phanerozoic time the BLAG formulation results in excessive CO_2 variation (for example, $\text{RCO}_2 > 100$ during the Cretaceous).

This new fertilization-BLAG feedback formulation is only applicable to the land biota at present. Because vascular land plants did

Table 2. Values of key parameters used in the model for CO_2 ; my, million years.

Parameter	Total range or value	Source
$f_A(t)$	See Fig. 2	(22, 23)
$f_D(t)$	See Fig. 2	(24)
$f_R(t)$	1.00 to 0.79	$= 1 + \log_{10} f_A(t)$
$f_C(t)$	1.00 to 0.50	For 570 to 150 Ma: $f_C(t) = C$ For 150 to 0 Ma: $f_C(t) = C + (1 - C) \exp[-0.03(150 - t)]$ $C = 0.50, 0.75, 1.00$ [see (25, 26)]
$f_E(t)$	1.00 to 0.15	For 570 to 350 Ma: $f_E(t) = 0.15, 0.30, 0.50, 1.00$ (16) For 300 to 130 Ma: $f_E(t) = 0.50, 0.63, 0.75, 1.00$ (17) For 80 to 0 Ma: $f_E(t) = 1.00$ For 350 to 300 Ma [rise of vascular land plants (12)] and 130 to 80 my [rise of angiosperms (13)], $f_E(t)$ calculated by linear interpolation
$f_G(t)$	See Fig. 3 (i) 1.00 to 1.36 (ii)	(i) Degassing equals sea floor generation rate from inversion of sea level data (27); (ii) One-half the variation from sea level inversion due to other effects on sea level (28)
$\delta_c(570)$	1.5 per mil;	(18)
$\delta_g(570)$	-23.5 per mil	(18)
δ_o	-0.80 to 6.20 per mil	(29) (smoothed curve fit)
α_c	21.8 to 30.0 per mil	For 570 to 140 Ma: $\alpha_c = 25$ per mil; for 140 to 0 Ma (30)
k_{wc}	0.00267 my^{-1}	Total $k_{wc} + k_{mc}$ from (18); k_{wc}/k_{mc} from (6)
k_{mc}	0.00133 my^{-1}	Total $k_{wc} + k_{mc}$ from (18); k_{wc}/k_{mc} from (6)
k_{wg}	0.0030 my^{-1}	Total $k_{wg} + k_{mg}$ from (18)
k_{mg}	0.0010 my^{-1}	Total $k_{wg} + k_{mg}$ from (18)
$C(0)$	$5000 \times 10^{18} \text{ mol}$	(6)
$G(0)$	$1250 \times 10^{18} \text{ mol}$	(6)

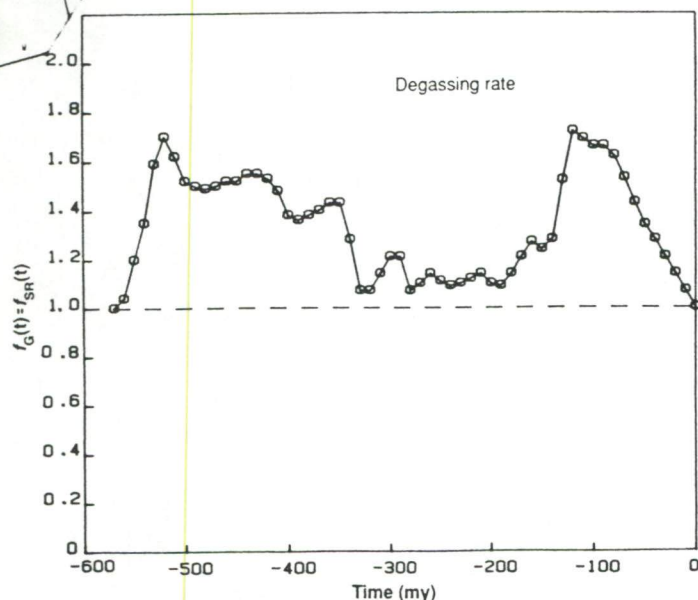


Fig. 3. Plot of $f_G(t)$, the rate of tectonically controlled volcanic and metamorphic degassing of CO_2 at time t divided by the same rate today. The parameter $f_G(t)$ is set equal to $f_{SR}(t)$, the rate of sea-floor spreading at time t divided by the rate today. Values of $f_{SR}(t)$ are based on the sea level inversion calculations of Gaffin (27). Long-term eustatic sea level change has been shown to correlate with the rate of generation of sea floor or spreading rate [see references in (27)], and thus, can be used as an indicator of crustal turnover by plate tectonics. [The problem is complicated, however, by the effects of continental breakup and collision on sea level (28)—see Table 2.] The further assumption made here [see also (6, 7)] is that the global rate of CO_2 degassing is linearly related to crustal turnover rate, which is in turn indicated by long-term changes in sea level.

not arise until the Silurian (12) and angiosperms until the Cretaceous (13), some allowance in the weathering feedback mechanism for these evolutionary changes must be made. I assumed that before vascular land plants, from 570 to 350 Ma, the weathering response was directly proportional to the level of atmospheric CO_2 such that $f_B(\text{CO}_2)$ equals $R\text{CO}_2$. Little is known of how weathering proceeded at this time in the presence of primitive nonvascular plants, but a direct response to the atmospheric CO_2 level should represent the strongest possible feedback. On the basis of laboratory experiments on mineral dissolution (14) and the potential mediation by soil microflora, it is possible that the feedback during the early Paleozoic could have been much less than linear. (Use of feedbacks based on laboratory observations results in $R\text{CO}_2$ values ranging from 30 to 400.)

The date of 350 Ma represents the time when vascular plants had become widespread on dry land and important as weathering agents. In order to represent the changeover to vascular plant-affected weathering, a linear mix of weathering functions was used between 350 and 300 Ma. Angiosperms may induce faster weathering than more primitive plants (15), but did not arrive until approximately 130 Ma (13). To account for this an intermediate gymnosperm-pteridophyte feedback function was introduced between 300 and 130 Ma. From 130 to 80 Ma a linear mix of functions was employed, and from 80 Ma to the present the functionality of $f_B(\text{CO}_2)$ was assumed to be the same as that of today, that is Eq. 12.

Accompanying changes in $f_B(\text{CO}_2)$ over time the model also considers corresponding changes in $f_E(t)$, the parameter expressing the effects of soil biological activity on actual weathering rate. There is little doubt that plants and associated soil biota enhance chemical weathering by the production of CO_2 and organic acids in the soil.

The question is how much. A study of waters draining lava flows in Iceland (16) showed a threefold enhancement of dissolved HCO_3^- concentration from vegetated to nonvegetated areas. However, much more work is needed to check this result. There is also a suggestion that angiosperms induce more rapid weathering than gymnosperms (15), and on this basis Volk (17) has constructed an appropriate global expression, which is incorporated in the present model (Table 2).

Results and Discussion

Model results showing sensitivity to some key parameters are shown in Fig. 4. One simulation is for the situation of no change in global tectonics, in other words for the same geographic, degassing, and general tectonic setting as at present. In this case $f_A(t) = 1$, $f_D(t) = 1$, $f_R(t) = 1$, $f_C(t) = 1$, and $f_G(t) = 1$ for all time. The curve expresses the residual effects of the burial rate of organic matter, as calculated from carbon isotopic data, plus the effect of the evolution of land plants. Another simulation is for no biological evolution. Here $f_E(t)$ is set equal to one for all time and the weathering feedback function $f_B(\text{CO}_2)$ is assumed equal for all time to that used for the present (Eq. 12). Here the curve reflects tectonics plus the burial rate of organic matter. (The result of setting all dimensionless parameters equal to one, such that the controlling factor is only organic burial rate, results in a curve similar to the no-tectonics situation except that CO_2 values before 350 Ma are about one-third those shown.) In either the no-tectonics or no-evolution cases the curves show a consistent trend of high atmospheric CO_2 levels during the Mesozoic (240 to 65 Ma) and early Paleozoic (570 to 350 Ma) and low levels during the late Paleozoic (330 to 260 Ma) and late Cenozoic (last 30 million years). Thus, neither geology nor biology dominate the trend in CO_2 levels for all times. It is a combination of factors that results in the calculated CO_2 values. This response comes about partly because some biological factors cancel each other and some tectonic factors cancel each other. For example,

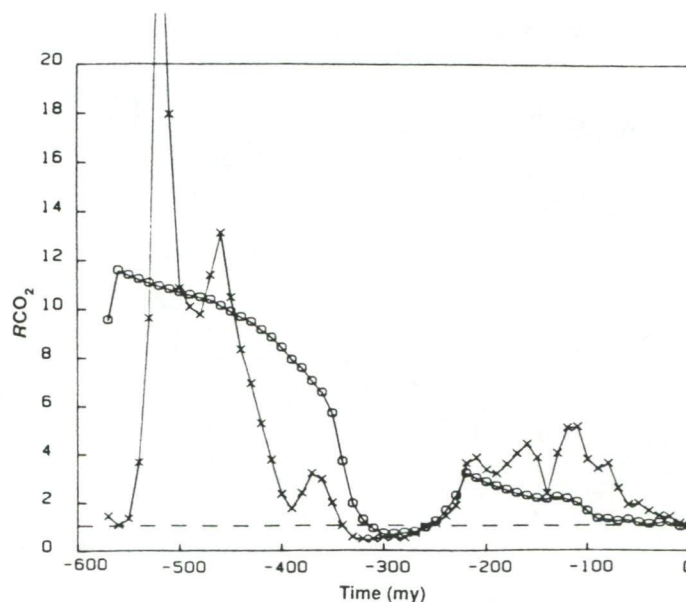


Fig. 4. Sensitivity plots versus time of $R\text{CO}_2$, the ratio of mass of CO_2 in the atmosphere at time t to that in the present atmosphere, illustrating the effects of major processes. (O) Situation for no change in tectonics over time; in other words for $f_A(t)$, $f_D(t)$, $f_R(t)$, $f_C(t)$, and $f_G(t)$ all equal to 1. (X) Situation for no biological evolution; in other words for $f_E(t) = 1$ and use for all time of the present-day formulation for the weathering feedback function $f_B(\text{CO}_2)$ (Eq. 12).

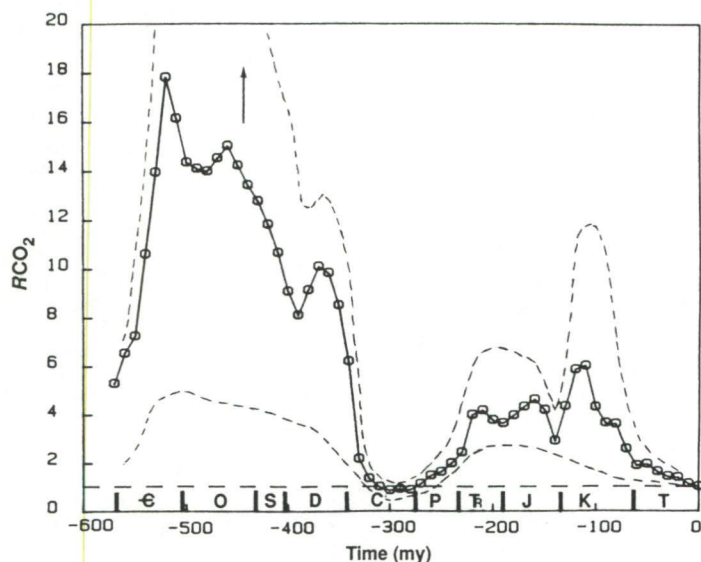


Fig. 5. Standard plot versus time of RCO_2 , the ratio of mass of CO_2 in the atmosphere at time t to that in the present atmosphere, based on the carbon mass balance model of the present study. The curve represents the best estimate of the various parameters that go into the model. Dashed lines show the envelope of approximate error based on sensitivity analysis. The vertical arrow denotes that early Paleozoic CO_2 levels may have been even higher than those shown; see text for discussion. Values used (see Table 2): $C = 0.75$; $f_E(t) = 0.15$ for 570 to 350 Ma; $f_E(t) = 0.75$ for 300 to 130 Ma; $f_A(t)$ and $f_D(t)$ from Fig. 2; $f_G(t)$ from Fig. 3.

setting $f_E(t)$ equal to one during the early Paleozoic reduces the level of CO_2 , but use of also the present-day formulation for $f_B(CO_2)$ at that time raises the level. Likewise, because changes in $f_D(t)$, representing CO_2 consumption, are generally parallel to changes in $f_G(t)$, representing CO_2 production (compare Figs. 2 and 3), holding both constant results in no great change in the trend of CO_2 levels with time.

Standard results for best estimates of all parameters are shown in Fig. 5 along with a crude estimate of the ranges in error from sensitivity analysis. (Because of a severe lack of knowledge of the role of primitive nonvascular land plants on weathering during the early Paleozoic, a wide range of error is shown for this time.) Again, the pattern of high CO_2 levels during the Mesozoic and early Paleozoic and low CO_2 levels during the late Paleozoic and late Cenozoic are apparent. Furthermore, use of the concept of rapid recycling (18) did not change the qualitative aspects of these results. Thus, although the curve shown in Fig. 5 represents only one choice of imperfectly known parameters, it can be used in a semiquantitative sense to represent the approximate levels of atmospheric CO_2 over Phanerozoic time.

If the values of CO_2 calculated here are at all correct, then because of the atmospheric greenhouse effect one would expect warm global climates on the average during the Mesozoic and early Paleozoic and cold climates during the late Paleozoic and late Cenozoic. This conclusion is in agreement with independent estimates of paleoclimate (19) and the contention of Fischer (20) that Phanerozoic climates can be characterized as a succession of greenhouses and icehouses. The two most important glaciations during Phanerozoic time occurred during the Permo-Carboniferous and the latest Cenozoic, precisely at times of minimum atmospheric CO_2 levels calculated by the model. (There is a problem with a smaller glaciation that occurred during the Ordovician, and this requires further refinement of the model.)

The model presented here is primarily an attempt to illustrate how one can use the geochemical carbon cycle as a means of exploring the various factors affecting atmospheric CO_2 over long geologic times. The results should be interpreted in the same manner as they are presented, as semi-quantitative trends. There are many problems with values of key parameters, but at least the modeling points to the importance of obtaining better estimates for them (21). For example, the parameterization of degassing rate as related to sea level is a crude assumption; other approaches are needed such as good estimates of the original abundance of igneous rocks as a function of time. Also, much more is needed on the quantitative role of land plants in chemical weathering. Nevertheless, the trend of changing CO_2 levels with time calculated with the model is consistent with paleoclimatic trends, and the greenhouse theory of climatic change, on a multimillion-year time scale, is supported by the results.

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The ice-core record: climate sensitivity and future greenhouse warming

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The prediction of future greenhouse-gas-induced warming depends critically on the sensitivity of Earth's climate to increasing atmospheric concentrations of these gases. Data from cores drilled in polar ice sheets show a remarkable correlation between past glacial-interglacial temperature changes and the inferred atmospheric concentration of gases such as carbon dioxide and methane. These and other palaeoclimate data are used to assess the role of greenhouse gases in explaining past global climate change, and the validity of models predicting the effect of increasing concentrations of such gases in the atmosphere.

INFRARED-ABSORBING gases in the Earth's atmosphere raise the mean global surface temperature from -18°C , which it would be in the absence of an atmosphere, to $\sim 15^{\circ}\text{C}$. This trapping of infrared energy emitted by the Earth's surface is known as the greenhouse effect¹⁻⁴. The largest effect is from water vapour, with a significant contribution from carbon dioxide and smaller contributions from ozone, methane, nitrous oxide and, in the past few decades, anthropogenic chlorofluorocarbons (CFCs). The naturally occurring greenhouse effect makes our planet habitable, but there is growing concern about the future of the Earth's climate because of increasing atmospheric concentrations of CO_2 , CH_4 , N_2O and CFCs, which are largely a result of man's activities. This radiative forcing (the change of the planetary radiation balance) is just over 4 W m^{-2} for a doubling of the CO_2 concentration (from, say, 300 to 600 p.p.m.v.). The increase in atmospheric CO_2 , CH_4 , N_2O and CFCs that have occurred since the beginning of the industrial revolution results in a forcing of just over 2 W m^{-2} (ref. 5); thus the anthropogenic greenhouse forcing has already reached a level of about half of the doubled- CO_2 radiative forcing, the canonical large forcing used in theoretical climate studies.

The radiative forcing resulting from doubled atmospheric CO_2 would increase the surface and tropospheric temperature by $\Delta T_e \approx 1.2^{\circ}\text{C}$ if there were no feedbacks in the climate system (see Box 1). But there are many possible feedbacks which may occur in response to an initial forcing. For example, a warming climate is likely to alter the amount of water vapour in the atmosphere, the cloud properties, snow cover and sea ice^{1-4,6}. Doubled- CO_2 sensitivity experiments using the most comprehensive climate models, the three-dimensional general circulation models (GCMs), yield equilibrium warmings between 1.9 and 5.2°C ⁶⁻¹¹. The uncertainty in climate sensitivity obtained from GCMs may, however, be even greater than this. A recent comparison of 14 GCMs showed about a threefold variation in climate sensitivity, attributable mainly to differences in the cloud feedbacks¹². Some simple one-dimensional climate models suggest essentially no net feedback³. Thus the net feedback factor f (the amplification of the no-feedback radiative-equilibrium temperature change) is uncertain in the large range $f \approx 1$ – 4 (Box 1).

One implication of the uncertainty in climate sensitivity is an even greater uncertainty in the climate response time. The time required for the climate system to approach a new equilibrium temperature after a change of radiative forcing varies at least linearly with f ^{6,13} because the feedbacks come into play in response to the change of temperature, not in response to the climate forcing. A consequence of the large uncertainty in the response time, which may be anywhere from a decade to more than a century, is that the observed global temperature trend of

the past century does not provide a very stringent empirical limit on climate sensitivity; indeed, allowing for the possibility of small unknown climate forcings, such as changes of tropospheric aerosols and solar irradiance, and for unforced climate variability, the temperature trend of the past century does not provide a useful bound on climate sensitivity^{6,14,15}.

A more promising source of empirical information on global climate sensitivity is provided by the changes in Earth's climate during the past few hundred thousand years because the global temperature change was very large, ~ 4 – 5°C ^{16,17}, the climate change was linked with changes of greenhouse-gas concentration (which by themselves resulted in a direct radiative forcing of $\sim 2\text{ W m}^{-2}$) and the changes were maintained for a sufficiently long time for equilibrium to be achieved, unlike the present greenhouse-gas changes. Based on these considerations there have already been attempts^{6,19} to estimate climate sensitivity from conditions during the Last Glacial Maximum, 18,000 years before present (18 kyr BP).

Box 1 Equilibrium temperature, feedback processes and $2\times\text{CO}_2$ experiments

Earth's atmosphere is heated by short-wavelength solar radiation and cooled by the emission of long-wavelength radiation to space. The planetary radiative energy budget per unit area, N , can be written as the difference of these terms

$$N = \frac{1-\alpha}{4} S - \epsilon \sigma T_s^4 \quad (1)$$

where S is the solar constant ($1,370\text{ W m}^{-2}$), α the planetary reflectivity or albedo (0.3), ϵ the effective emissivity of the atmosphere, σ the Stefan-Boltzmann constant and T_s the surface temperature⁶⁵. At equilibrium ($N=0$) and with no atmosphere this yields $T_e = 255\text{ K}$. Earth's surface temperature is actually 288 K , primarily because of the presence of the atmosphere and the present greenhouse warming is therefore $\sim 33^{\circ}\text{C}$.

If the atmospheric structure and all other factors remain fixed, the response of the system to an increase of radiative forcing ΔQ would be a change in T_s necessary to restore radiative equilibrium, $\Delta T_s = T_s \Delta Q / (1-\alpha)S$ leading to $\Delta T_s = 0.3 \Delta Q$. The radiative forcing for a doubling of the CO_2 concentration is $\Delta Q \approx 4\text{ W m}^{-2}$ corresponding to $\Delta T_s \approx 1.2^{\circ}\text{C}$.

The surface temperature changes, ΔT_s , calculated in recent GCM studies for doubled CO_2 ⁶⁻¹¹ are shown in the table. Here we have adopted the feedback terminology of Hansen *et al.*⁶ in which a net feedback factor f is defined by assuming that the radiative forcing for doubled CO_2 is $\sim 4\text{ W m}^{-2}$ and hence $\Delta T_s = 1.2 f$. The feedbacks that have been found to be significant in these models (water vapour, clouds, sea ice and snow cover) can be described as fast feedbacks, because they respond rapidly to climate change and are thus important on decadal timescales.

Recently, ice cores (see Box 2) from the Greenland^{20,21} and Antarctic^{22,23} ice sheets and, in particular, the Vostok record²⁴⁻³⁰, which spans a full glacial-interglacial cycle²⁶, have allowed documentation of the relationship between greenhouse radiative forcing (CO_2 and CH_4) and climate over a period of large climate changes. These new data support the orbital theory of ice ages, but they are also directly relevant to the problem of greenhouse gases and climate sensitivity. Here we compare information about climate sensitivity derived from ice cores with the GCM results. Although a full understanding of the causes and sequences of climate fluctuations remains elusive, much information is revealed from comparison of glacial-interglacial palaeotemperature data with concurrent changes in the planet's energy balance that are linked with atmospheric CO_2 and CH_4 concentrations.

The ice-core data suggest that greenhouse gases have had a significant role in explaining the magnitude of past global temperature changes and indicate a net feedback for fast processes of $f \approx 3$, consistent with a climate sensitivity of $\sim 3-4^\circ\text{C}$ for a doubled atmospheric CO_2 . Studies of palaeoclimate changes therefore provide a clue to help validate estimations of the climate impact of increasing greenhouse gases.

Ice-core data

Ice-core data are unique in providing, in the same samples,

Box 2 Ice cores, climate and atmospheric data

Although we cannot expect to find an ideal record of climate, ice is a rather close approximation to it as summarized in the table below⁶⁸.

CLIMATE FACTORS AND QUANTITIES MEASURED IN ICE CORES

Atmosphere	Ice core
Temperature	D/H, $^{18}\text{O}/^{16}\text{O}$
Precipitation	D/H, $^{18}\text{O}/^{16}\text{O}$, ^{10}Be
Humidity	D/H, $^{18}\text{O}/^{16}\text{O}$
Aerosols	
natural (continents, sea, volcanoes, biosphere)	chemicals (Al, Ca, Na, H, SO_4 , NO_3)
man-made	SO_4 , NO_3 , Pb, radioactive fallout particles
Circulation	
Gases: natural and man-made	O_2 , N_2 , CO_2 , CH_4 , N_2O

Palaeotemperature reconstruction is based on the present correlations between the ratios of ^2H (D) and ^1H and of ^{18}O and ^{16}O in the snow and the temperature conditions just above the inversion layer, where the precipitation is formed, and at the surface. These are correlated because of the fractionation processes that take place during the atmospheric water cycle and, although they depend on several parameters, lead to linear isotope-temperature relationships in both polar ice sheets^{69,70}. The validity of using the present observed relationship for palaeotemperature reconstruction is supported by various evidence including atmospheric isotope models⁷¹. Although there is no doubt that the concentration of a species in the air is reflected in snow deposits at the site, quantitative estimation of atmospheric concentrations suffers from an incomplete understanding of the deposition processes, such as the relative contribution of 'wet' and 'dry' processes. In contrast, relating gas concentrations obtained from ice-core bubbles to the atmospheric value is quite straightforward. Owing to the air-enclosure process in the ice, however, the air is isolated from the atmosphere well after the precipitation has been deposited; in a site of low accumulation like Vostok the correction ranges from 3 to 6 kyr with an uncertainty of up to ~ 1 kyr^{24,25}.

Adequate site selection is required to avoid disturbance of the atmospheric record by ice flow. Current deep-drilling projects to obtain long-term records include US and European efforts in central Greenland; hopefully this will supplement the few ice cores available now and provide the awaited counterpart of the Vostok Antarctic record needed for a better understanding of glacial-interglacial forcings and feedbacks.

access to both climate and climate forcings (Box 2). The climate information includes an estimate of temperature changes at the atmospheric level where the snow formed and at the surface, and the information about climate forcing includes data on the amount of aerosols in the atmosphere and the atmospheric chemical composition. So far, there are only two deep ice cores from Greenland (Camp Century²⁰ and Dye 3²¹) and three from Antarctica (Byrd²², Dome C²³ and Vostok²⁴⁻³⁰) which provide long-term climate and environmental records extending over the last ice age. Only Vostok provides a complete undisturbed series over the last climate cycle. The principle findings obtained from these deep ice cores have been recently reviewed³¹.

The Vostok temperature record is reconstructed from the continuous deuterium profile measured along the core²⁷. Here we refer to the atmospheric temperature change, ΔT_a , just above the inversion layer (Box 2) because this is the parameter that is most directly accessible from the snow deuterium content, which depends on the temperature of formation of the precipitation (such an interpretation is well supported by a theoretical approach) and because this temperature is certainly more relevant for characterizing the global atmosphere as the existence of a strong inversion is restricted to central Antarctica. The record (Fig. 1b) shows the last two glacial-interglacial transitions with atmospheric temperature changes of $\sim 6^\circ\text{C}$. The last ice age is characterized by three minima, around 20, 60 and 110 kyr BP, separated by slightly warmer episodes (interstadials).

The three Antarctic ice cores yield remarkably similar isotope-derived temperature records over the last ~ 60 kyr (ref. 32). The surface warmings associated with the last deglaciation are quite comparable with values slightly below 10°C and the Vostok record can be considered to be representative of a large area in Antarctica. The deglaciation warming for Dye 3 is similar to the one observed in Antarctica (whereas the Camp Century core suggests a somewhat larger change) and in fact, the main features of the Vostok climate record are of global significance, at least qualitatively speaking. This significance is suggested by the comparison with the global ice volume changes derived from isotope measurements in deep-sea sediments (during glacial periods the accumulation of isotopically depleted ice over the continents leads to a measurable isotope enrichment of the ocean). In particular the SPECMAP record of marine ^{18}O content³³, which essentially represents global ice volume (Fig. 1c), and the Vostok temperature record corresponds very closely (correlation coefficient, $r^2 = 0.87$) down to 110 kyr BP²⁷. Further back there is a discrepancy in the start and in the duration of the previous interglacial which can probably be explained by dating uncertainties for this part of the record (see discussion below).

The close association between greenhouse gases and glacial-interglacial climate changes was first revealed from the analysis of air trapped in Greenland and Antarctic cores showing that the atmospheric CO_2 content was $\sim 190-200$ p.p.m.v. during the Last Glacial Maximum compared with an average close to 270-280 p.p.m.v. during the Holocene³⁴⁻³⁷. The detailed reconstruction from Byrd³⁸ for the period between 50 and 15 kyr fully confirmed the association between cold climate and low atmospheric CO_2 levels (Fig. 2a). The Vostok data confirmed the existence of a strong CO_2 -climate relationship over a full climate cycle²⁴. This CO_2 record exhibits large changes between two levels that are centred near 190-200 and 260-280 p.p.m.v., with the low and high levels associated with full glacial and interglacial conditions, respectively (Fig. 1a). Moreover, despite some noticeable differences, there is a remarkable correlation ($r^2 = 0.79$) between atmospheric CO_2 and temperature change throughout the record.

Such a correlation between climate change and radiatively active gases also exists for methane. Analysis of Dye 3 and Byrd samples showed an increase in the CH_4 concentration from 0.35 p.p.m.v. for the Last Glacial Maximum to Holocene values of 0.65 p.p.m.v.³⁹, and a similar increase paralleling the tem-

perature change was revealed from the Vostok ice for the transition between the penultimate ice age and the last interglacial²⁹. These results are now fully confirmed³⁰ by the detailed and continuous CH₄ profile obtained along the Vostok core (Fig. 1c), which also exhibits four well marked maxima during the glacial period. These maxima occur at the time of relatively warm interstadials and although there are marked differences between the CH₄ and CO₂ record, the correlation with the temperature record is quite similar ($r^2=0.78$ in the case of CH₄).

Superimposed on these long-term trends, rapid and large changes have been documented both for CO₂ and CH₄. In the time interval between 40 and 30 kyr a series of rapid CO₂ variations (Fig. 2b), occurring in a century or less between low values in the range 180–200 p.p.m.v. and higher values in the range 240–260 p.p.m.v., were found first in Dye 3⁴⁰. The existence of these rapid events, associated with large changes in surface temperature (5–6 °C) and in the dust content, has been confirmed in the Camp Century core but has not been identified in the Antarctic Byrd core, a difference not yet satisfactorily explained. There is also a clear indication of rapid climate change at the end of the last deglaciation extensively documented from isotope and dust studies on Greenland cores⁴¹. This so-called Younger Dryas cooling dated at ~11–10 kyr BP is well marked in North Atlantic and adjacent regions. Although there is also a weak cooling in Antarctica the dating is not yet sufficiently accurate to allow it to be firmly related to the Younger Dryas. Recent Vostok measurements³⁰ show that the Antarctic cooling was associated with a large decrease in CH₄ concentration (Fig. 1c), as is also suggested in the data from the Dye 3 core³⁹.

Reliable N₂O data can also be obtained from ice cores. Existing results suggest that N₂O concentrations during the Last

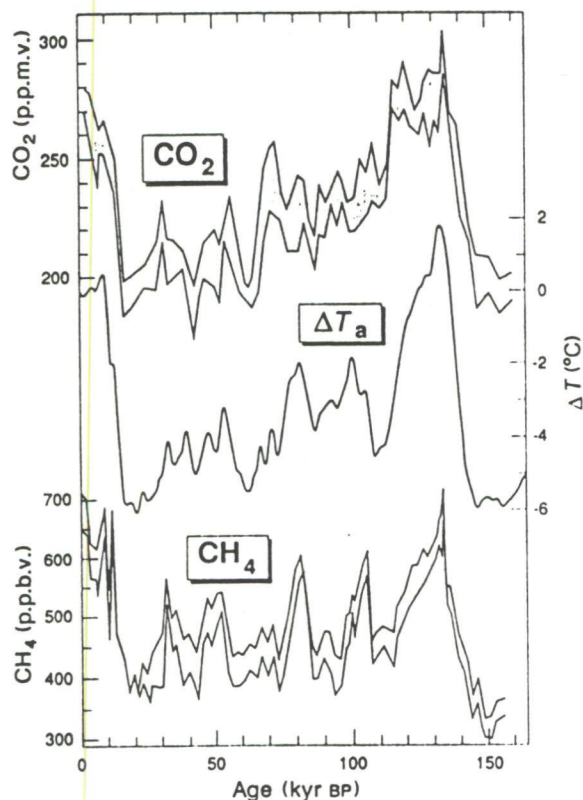


FIG. 1 Variation during the last climate cycle, as derived from measurements along the 2,083-m Vostok ice core, of the CO₂ atmospheric concentration²⁴, the atmospheric temperature change over Antarctica²⁷, and the CH₄ atmospheric concentration³⁰. For CO₂ and CH₄ the envelope shown has been plotted taking into account the different sources of uncertainty whereas the temperature record is a smoothed curve.

Glacial Maximum were possibly smaller but not drastically different from those of the recent period⁴². We can therefore estimate the direct radiative forcing exerted by all naturally occurring greenhouse gases (except changes in water vapour and possibly in ozone). In terms of radiative forcing such changes in the concentration of greenhouse gases are important because, for example, of the logarithmic increase with CO₂ and CH₄ concentrations⁴³. From glacial to interglacial this forcing increases by ~2 W m⁻² (ΔT_e of ~0.7 °C), half of that corresponding to a CO₂ doubling from 300 to 600 p.p.m.v.

Ice-core data and the theory of ice ages

The long Vostok ice core offered a unique opportunity to examine a continental record of the link between a time series for atmospheric temperature and the astronomical theory of the ice ages. Among the orbital parameters the obliquity of the Earth's axis (period of ~41 kyr) and the precession of the equinoxes (periods of 23 and 19 kyr) are the most important as they strongly affect the distribution of available energy between latitudes and seasons⁴⁴; for example, the variation of the summer insolation at 65 °N, a key latitude in the astronomical theory of the ice ages, is ~60 W m⁻² over the last climate cycle. Spectral analysis shows that the Vostok temperature record is indeed dominated by a strong obliquity component and influenced to a lesser degree by the precession²⁷. Aside from the interval of ~100 kyr between interglacials, the main features of the Vostok temperature record show that the minima are in good agreement with those of the annual insolation received at the Vostok latitude (78° S) and that there are also similarities with the 65° N summer insolation. These results further support the role of orbital forcing in Pleistocene glacial-interglacial cycles demonstrated by deep-sea sediment records⁴⁵. The orbital forcing is, however, relatively weak when considered on an annual globally averaged basis (the total insolation received by Earth has varied by <0.7 W m⁻² over the past 160 kyr). The amplification of this forcing, the observed dominant 100-kyr cycle and the synchronized termination of the main glaciations and their similar amplitude in the Northern and Southern Hemispheres cannot easily be explained despite developments including the nonlinear response of ice sheets to orbital forcing⁴⁶. The discovery of significant changes in climate forcing linked with the composition of the atmosphere has led to the idea that changes in the CO₂^{24,28,47–49} and CH₄^{29,30} content have played a significant part in the glacial-interglacial climate changes by amplifying, together with the growth and decay of the Northern Hemisphere ice sheets, the relatively weak orbital forcing and by constituting a link between the Northern and Southern Hemisphere climates.

The observed changes in CO₂ and CH₄ imply modifications in their sources and sinks that probably involve very different processes such as ocean circulation and marine production for CO₂ (see ref. 50 for a review) and fluxes of emission from natural wetlands for CH₄^{29,30}. The mechanisms behind these modifications are not yet fully understood, especially those involving CO₂, but we note that the orbital frequencies are present in both of these Vostok series with, in particular, a strong precessional signal^{24,30}. This suggests that the changes are in some way orbitally driven and, even if the orbital forcing is not the only important effect, this supports the idea that orbital changes are one initial cause of the ice ages.

Further, the astronomical theory cannot easily explain the rapid events recorded in ice cores. Rather, although the mechanisms are still unknown, rapid changes could be connected to a flip-flop mechanism in the North Atlantic ocean, perhaps a turning on and off of the North Atlantic current⁵¹. We note, however, that the correspondence between cold episodes and low CO₂ concentration (Fig. 2b) and conversely between warmest periods and high concentrations seems to hold true during the last ice age and possibly during the Younger Dryas

event, as suggested by the CH₄ data (Fig. 1c). There is therefore some indication that greenhouse forcing also participated in the amplitude of these temperature changes.

Ice-core data and the greenhouse effect

The glacial-interglacial changes include not only the fast feedback processes associated with water vapour, cloud properties, snow cover and sea ice but also longer-term processes such as those linked with slow changes in several boundary conditions and in the atmospheric composition, which help maintain the ice age cold relative to the interglacial (Box 1). We now derive information about the role of these slow changes from the Vostok temperature series and compare it with other palaeodata and climate forcings. Using the data on the direct radiative forcing associated with changes in the concentration of greenhouse gases, we derive information about the role of fast feedback processes. This does not require a solution of the 'chicken and egg' problem, that is, we do not have to address fully the question of causes of the glacial-interglacial cycles and of the sequence of possible forcing factors^{28,52}. For example, whether the temperature changes lead or lag the changes in CO₂ or CH₄ concentrations is not relevant for the study of fast feedbacks. There is, however, the important constraint that the planet must be in near radiation balance with space. This may require several thousand years and we therefore cannot use this approach to address the question of transients or rapid events, which may occur in a few decades⁴¹. Within these limits, we may assume that greenhouse gases have contributed to the glacial-interglacial temperature change through their direct radiative forcing associated with fast feedback processes. As discussed below, such an assumption is well supported by the GCM experiments described in refs 49 and 53 and we can take advantage of the long records of climate forcings and climate outputs, which reflect equilibrium conditions, to evaluate (or at least reasonably bracket) the role of greenhouse-gas radiative forcing in determining past climate.

Genthon *et al.*²⁸ followed such an approach in performing a multivariate analysis of the Vostok temperature record and three climate inputs. This approach decomposes the temperature variance into a part that can be accounted for by the inputs plus a residual, which is due to noise not related to these inputs. The relative contribution of each input is calculated so that the noise

is minimized in a least-squares sense⁵⁴. To be able to partition properly the output series as a combination of the input series, the input series must be independent. For climate forcing, Genthon *et al.*²⁸ used CO₂, a Southern Hemisphere component represented by the local (78°S) annual insolation (Fig. 3d) and a Northern Hemisphere component derived either from the 65°N July insolation or from the $\delta^{18}\text{O}$ marine record, a proxy of the Northern Hemisphere ice volume (Fig. 3c). We have used this approach and extended it to account for the following points.

First, such a multivariate analysis implicitly assumes that the system can be described by its steady-state behaviour. This is not the case for a climate system affected by nonlinearities associated with the growth and decay of the ice sheets. To account for this, we consider cases in which the ice volume reconstructed from the $\delta^{18}\text{O}$ record is taken as a proxy of the Northern Hemisphere forcing. This proxy combines the orbital insolation forcing and the nonlinearities mentioned above. We can therefore assume that it represents a large part of the slow feedbacks associated with the changes in boundary conditions, all the more because the increased area of ice sheets provides a large forcing ($\sim 3 \text{ W m}^{-2}$ mainly owing to albedo effects but also to topographic changes).

Second, we account for changes in both CO₂ and CH₄ to evaluate the combined greenhouse radiative forcing. For this purpose we use the CO₂ and CH₄ Vostok records (including data from other cores would not result in significant changes, except for rapid events not considered here) and calculate the associated direct radiative forcing through the formulae given by Hansen *et al.*⁴³. For CH₄, this direct radiative forcing was augmented by a factor of ~ 2 because of chemical feedbacks linked with the increase of stratospheric water vapour³⁰. The total greenhouse radiative forcing expressed in terms of equilibrium temperature change (with no climate feedback) given in Fig. 3b shows a glacial-interglacial amplitude of $\sim 0.7^\circ\text{C}$.

Third, we found it useful to explore a wider range of possible atmospheric radiative forcings such as those resulting from changes in the aerosol loading^{18,55} and in the amount of cloud condensation nuclei (CCN). A strong increase (up to a factor of 30) in the dust fallout has been recorded for the Last Glacial Maximum both in Greenland and Antarctic cores, as illustrated in Fig. 2c for Dome C⁵⁶. It has been claimed that the resulting increase in the optical depth of the atmospheric aerosols could make a significant contribution to the cooling during this period¹⁸. This is still uncertain because it is difficult to evaluate the global aerosol loading using data coming mostly from polar ice cores. Also, a change in the optical depth and thus in direct radiative forcing depends strongly on the dust optical properties, which are not known. Analysis of the Vostok core⁵⁷ showed a strong dust increase at the end of the penultimate ice age and during the cold interstadial centred around 60 kyr BP, which also indicates a relation between cold periods and high aerosol loading. The Vostok dust record is very spiky, however, and not simply related with the smoother temperature record. There is no dust peak during the cold period around 110 kyr and the correlation ($r^2 = 0.36$) with the temperature record is much lower than the corresponding one for greenhouse gases. It has recently been proposed that the number concentration of CCN, mainly sulphate particles produced by the oxidation of dimethylsulphide emitted from the ocean, influences the albedo of marine stratus cloud and hence global climate⁵⁵. In this hypothesis, a change in marine productivity may eventually act as a climate forcing through the induced change in CCN concentration. Using the Vostok non-sea-salt sulphate as a proxy for CCN concentration, Legrand *et al.*⁵⁸ suggested that this forcing may have participated in the glacial cooling. The correlation between non-sea-salt sulphate and the Vostok temperature is relatively high ($r^2 = 0.63$) but it must be noted that the role of CCN and the CCN-sulphate link are not yet firmly established.

TABLE 1 Results from recent doubled-CO₂ models

Study	Source	ΔT_s	f
Goddard Institute for Space Studies	ref. 6	4.2	3.5
National Center for Atmospheric Research, Geophysical Fluid Dynamics Laboratory	ref. 7	4.0	3.3
UK Meteorological Office	ref. 8	4.0	3.3
Oregon State University	ref. 9	5.2	4.3
UK Meteorological Office	ref. 10	2.8	2.3
	ref. 11	1.9	1.6

Most results are adapted from ref. 1; an extreme experiment¹¹ in which cloud cover and cloud radiative properties depend strongly on temperature cloud water content is also included. The values of ΔT_s obtained by the GCMs, 1.9–5.2 °C, are substantially larger than the values of ΔT_s , indicating that the modelled processes significantly amplify the direct greenhouse forcing. More general discussions of the equilibrium climate sensitivity relevant to short timescales fall in the range $\Delta T_s = 1.5\text{--}5.5^\circ\text{C}^{1-3,14,15}$. Thus the uncertainty in f seems to be $f \approx 1\text{--}4$. There are other potentially important feedbacks, which are not investigated by the GCM studies because they are treated as either fixed boundary conditions or as specified climate forcings. Many of these could be classified as slow feedbacks which may be significant on long timescales, such as glacial-interglacial climate change. Examples include changes of land ice, floating ice shelves⁶⁶ vegetation cover⁶, and ocean circulation⁶⁷. Possible atmospheric changes include the abundance of greenhouse gases, dust¹⁸ and cloud condensation nuclei⁵⁵, including the impact of these on optical properties of clouds¹⁸.

Finally, we reexamined the previously noted influence of the dating²⁸. Indeed, the multivariate analysis must be done with all time series having a common timescale. This is not a problem for the temperature, non-sea-salt sulphate and dust Vostok series and, within the uncertainty associated with the correction due to the air-ice difference (Box 2), for the greenhouse forcing. There is, however, a noticeable dating discrepancy between the Vostok and ice volume records before 110 kyr, which has been shown to affect significantly the results of the multivariate analysis performed by Genthon *et al.*²⁸. We tested numerous redatings of the oldest part of the Vostok series accounting for the recent interpretation of the Vostok dust record, which suggests that the marine and Vostok records are roughly in phase before 110 kyr²⁷. A new multivariate analysis was then performed using five forcing factors: greenhouse gases, dust and non-sea-salt sulphate concentrations, ice volume and local insolation. The results may be summarized as follows:

There is no case for which the contribution of the greenhouse effect is below 40%. The lowest calculated value is 42% when the Vostok record is put exactly in phase with the marine $\delta^{18}\text{O}$ record, with a similar contribution from the ice volume (45%) and with contributions of $\leq 5\%$ for each of the three other climate inputs. With this redating, Genthon *et al.*²⁸ obtained a CO_2 contribution of 35%. Indeed, for all analyses performed, adding CH_4 leads to a $\sim 10\%$ increase over the effect derived using only CO_2 . This may be related⁵⁹ to the higher correlation of the Vostok temperature with the combined greenhouse ($r^2 = 0.84$) than with either CO_2 ($r^2 = 0.79$) or CH_4 ($r^2 = 0.78$).

Within the limits of dating by comparison of the Vostok dust record with marine records, the greenhouse contribution never exceeds 65%, with the sum of greenhouse and Northern Hemisphere forcing being generally over 80%. Generally $>90\%$ of the variance of the Vostok temperature is explained by the five climate inputs.

The greenhouse contribution falls in the top of this 40–65% range (between 55 and 65%) when the ice volume record of Labeyrie *et al.*⁶⁰ is used, instead of the SPECMAP curve, to represent the Northern Hemisphere forcing. This alternative approach was followed because the Labeyrie *et al.* curve probably represents the ice volume change better than the more generally used SPECMAP record, because it is corrected for the part of the isotope signal that comes from the change in ocean temperature. Also, there may be, at least for CO_2 ²⁴, some link between sea level (or ice volume) and the change in the atmospheric concentration and using different ice-volume curves

is one way to estimate the bias introduced in the multivariate analysis by two of the input series (greenhouse and ice volume) not being fully independent.

To summarize, the contribution of greenhouse gases to the Vostok temperature change can be bracketed between a lower estimate of 40% and a higher estimate of 65%. This range is somewhat uncertain, however, owing to the nonlinearities linked with the growth and decay of the Northern Hemisphere ice sheet, all the series not having a common timescale and the possibility that there may have been other important forcings that we have not taken into account. We have attempted to account for the first two factors. In addition, the introduction of a new climate input, which must correspond to a process having a potential climate input, will significantly alter the result of the multivariate analysis only if this new series is well correlated with the Vostok temperature and independent of the other inputs. One example of such a forcing is that linked with vegetation change through its influence on the albedo, but model results^{43,49} show that the climate role of the vegetation is relatively minor in maintaining the glacial climate. Although we cannot rule out another forcing, it is difficult to imagine another meeting these three requirements. So, within the limits of our multivariate analysis, $\sim 50 \pm 10\%$ is a reasonable estimate for the overall contribution of the greenhouse gases to the Vostok temperature change over the last climate cycle. This means that $\sim 3^\circ\text{C}$ of the 6°C in the glacial-interglacial change at Vostok could be attributed to the greenhouse effect. This result agrees well with the GCM simulations for the Last Glacial Maximum performed by Broccoli and Manabe⁴⁹. Using a model with prescribed cloud cover, they reduced the CO_2 concentration from 300 to 200 p.p.m.v. in experiments with and without changes in other boundary conditions. The CO_2 change alone, which in terms of radiative forcing roughly corresponds to the combined CO_2 and CH_4 glacial-interglacial change measured in ice cores, results in a temperature drop of 2–2.5 $^\circ\text{C}$ in central east Antarctica. With variable cloud, however, the GCM sensitivity is increased by 31%¹⁹ implying a change of $\sim 3^\circ\text{C}$ over this part of Antarctica. Although based on one particular GCM and experiment, this agreement between two independent approaches is worth noting.

Ice-core data and climate sensitivity

Despite the large geographical significance of the Vostok temperature record, it can be argued that looking at one particular site or area is too restrictive. Other palaeoclimate data and climate simulations are indeed helpful in examining the problem

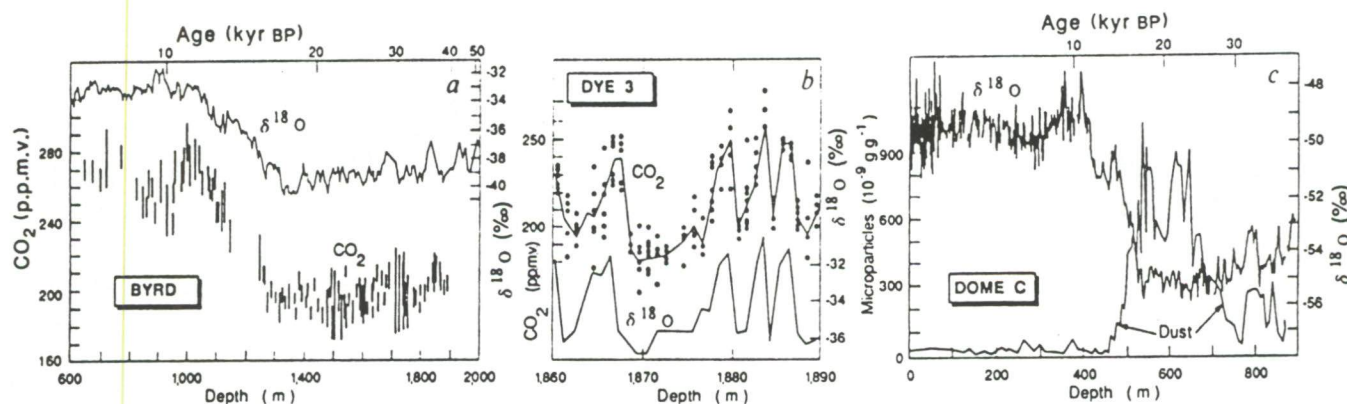


FIG. 2 This panel shows key data obtained from the Byrd (West Antarctica), Dye 3 (Greenland) and Dome C (East Antarctica) cores. The results are given as a function of depth. The corresponding timescale (upper abscissa) is also given, except for Dye 3, for which the 30-m ice increment corresponds to approximately 10 kyr between 40 and 30 kyr. The $\delta^{18}\text{O}$ record has been reported for each of these three cores ($\delta^{18}\text{O}(\text{‰}) = \{[(^{18}\text{O}/^{16}\text{O})_{\text{sample}} / (^{18}\text{O}/^{16}\text{O})_{\text{standard}}] - 1\} \times 1,000$ where the standard is

standard mean ocean water. a, Byrd: A detailed CO_2 record covering the last glacial period and part of the Holocene³⁸. b, Dye 3: Rapid CO_2 changes (circles indicate the results of single measurements and the upper solid line connects the mean value for each depth), adapted from ref. 40. c, Dome C: The Dome C dust record showing a strong increase of dust during the Last Glacial Maximum⁵⁶.

of greenhouse gases and climate sensitivity in a global perspective.

First, we note that owing to the general polar amplification of climate changes, the Vostok inferred glacial-interglacial warming of $\sim 6^\circ\text{C}$ over Antarctica is consistent with a globally averaged value of $\sim 4\text{--}5^\circ\text{C}$; for example, in doubled- CO_2 GCM experiments this amplification is, on average, $\sim 50\%$ for high-latitude continental regions³. This indicates that we can extend the conclusion that the greenhouse effect resulted in $\sim 50\%$ of the temperature change in the last climatic cycle to be a global estimate. This is supported by the similarities observed between the Vostok temperature and palaeorecords of global character; the multivariate analysis shows that generally $>80\%$ of the explained Vostok temperature variance is due to climate forcings of global significance (ice volume and greenhouse gases). Similarities between the Vostok and other palaeotemperature records further support this conclusion. For example, the Vostok temperature is well correlated with the surface temperature at the Indian Ocean site RC11-120³⁴ (with $r^2 = 0.66$ for the period before 110 kyr for which redating uncertainties are not critical). Replacing the Vostok temperature record by the RC11-120 record as climate output in the multivariate analysis left a contribution of the greenhouse gases to the explained variance of $\geq 50\%$.

The approach used here suggests that greenhouse gases contributed $\sim 2^\circ\text{C}$ to the global warming between glacial and interglacial periods. Estimating climate sensitivity then simply consists in dividing the estimated global warming associated with the greenhouse forcing by the corresponding ΔT_e (0.7°C from glacial to interglacial) to get the net feedback factor. We stress

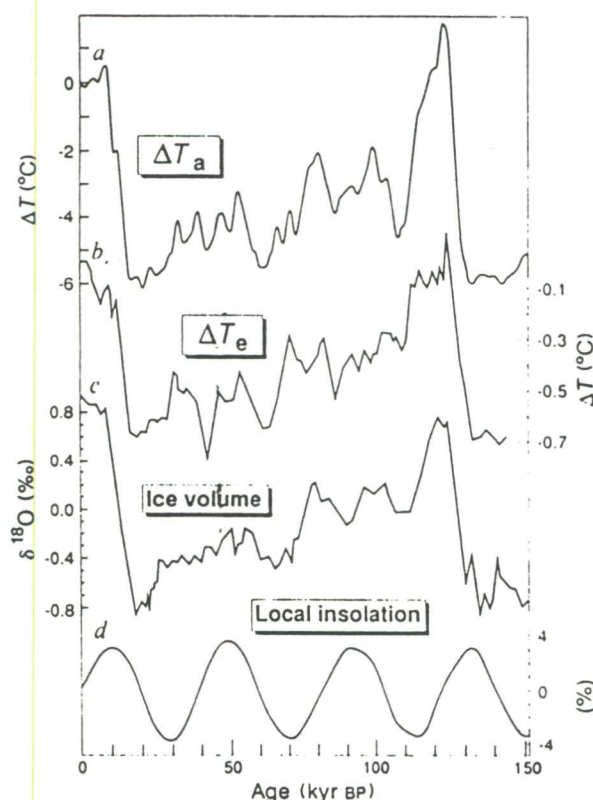


FIG. 3 Time series of the Vostok climate record and of climate forcings used in the multivariate analysis: *a*, atmospheric temperature change over Antarctica (Vostok record), ΔT_a ; *b*, the direct greenhouse radiative forcing accounting for CO_2 and CH_4 variations, ΔT_e ; *c*, $\delta^{18}\text{O}$ SPECMAP record taken as a proxy of the change in ice volume (normalized value from ref. 33); *d*, percentage change in total insolation at the Vostok latitude (78°S) during the entire year. Vostok curves (*a* and *b*) have been redated for the period before 110 kyr in such a way to put the Vostok temperature and the marine records in phase.

the independence of this method with respect to that based on GCM results. Although GCM simulations are useful in providing evidence that the regional results have a global significance, this use does not make our reasoning circular.

Second, it is of interest to compare the relative contribution of the greenhouse forcing as independently derived from the multivariate analysis and from GCMs. These results, only available for the GFDL and GISS models, indicate that a significant part of the glacial-interglacial warming may, on a global scale, be attributed to the greenhouse effect. In the Broccoli and Manabe experiments⁴⁹ the CO_2 increase from 200 to 300 p.p.m.v. and associated fast feedbacks, accounts for $\sim 40\%$ of the global glacial-interglacial warming. These experiments also showed that most of the remaining warming is due to the retreat of the Northern Hemisphere ice sheet in agreement with our multivariate analysis: Using the GISS model, Rind *et al.*⁵³ recently mentioned that an average warming of 2.5°C would result from a CO_2 increase from 230 to 300 p.p.m.v. This corresponds to at least $\sim 50\%$ of the glacial-interglacial warming predicted by the GISS model for various simulations of the Last Glacial Maximum⁶. Such agreement supports the belief that the net effect of the fast feedback processes which amplify the direct greenhouse radiative forcing, or at least their combined effect, are realistically evaluated in these two models (or in models having similar climate sensitivity).

To the extent that atmospheric feedback processes are independent of the detailed causes of climate changes but governed by the overall climate response, the same feedback processes would operate now as during the Last Glacial Maximum¹⁸. Applying the net feedback factor of three deduced from ice-core data to the present climate may be done without further assumptions (at least within the limits of uncertainty of our method for estimating f). For greenhouse model experiments, climate sensitivity depends primarily on fast feedback processes. Analysis of ice-core results and palaeodata over a full glacial-interglacial cycle suggests that a warming induced by doubled CO_2 concentrations of $3\text{--}4^\circ\text{C}$ ($f \approx 3$) may be a realistic value which, although being in the middle of the range of values inferred from the GCM experiments (Box 1), corresponds to a relatively high climate sensitivity.

Concluding remarks

Ice-core records have been important in determining that climate and greenhouse-gas concentrations were intensely interactive. Palaeoclimate reconstructions are now recognized as an important element in climate studies because they: (1) allow us to assess the degree of natural variability and put observed changes in present climate in a broader perspective; (2) assist in understanding the causes and mechanisms of climate change (by relating climate and forcing factors, for example); (3) contribute to validating models through the comparison of outputs with empirical data sets; (4) may potentially be used in predicting geographical patterns of future climate based on past analogues.

We have approached the evaluation of climate sensitivity to greenhouse gases from the point of view of both the modellers and palaeoclimatologists. Despite the relative agreement of the two approaches there is still much research to be done and our assertion about the irrelevance of fully addressing the causes of glacial-interglacial cycles for estimating climate sensitivity must not be misunderstood. We attach the utmost importance to a full understanding of the physical, chemical and biological processes by which subtle changes in insolation are amplified to induce long-term changes in global climate. For this, a knowledge of the sequence of events and the exact timing of forcings and of the climate responses in various parts of the Earth system is essential. This includes acquisition of new well dated series from land, ocean and ice records to document fully the changes that have affected our planet over the last climate cycle; obtaining a globally averaged temperature record would improve our multivariate approach.

As far as ice cores are concerned, there is still a lot of information directly relevant to the interaction of greenhouse gases and climate buried in both Antarctic and Greenland polar ice caps. Obtaining Northern Hemisphere ice records of climate, greenhouse gases and other climate forcings over a full glacial-interglacial cycle is one important objective for a better understanding of the past complex interactions between Northern and Southern Hemispheres on this timescale. This objective is part of the GRIP (Greenland Ice Core Project) and GISP II (Greenland Ice Sheet Project) now being conducted in north central Greenland by European and American scientists. These drillings are expected to reach the bedrock (the ice thickness is 3.2 km) in 1992 and to cover the last climate cycle and hopefully more. These cores will allow further documentation of the rapid climate changes discussed here. With a snow accumulation higher than at Vostok they should also allow a better determination of the relative timing (phase lag) of climate and greenhouse forcing. Plans are also being developed for further drilling in Antarctica. Drilling at Vostok was interrupted at a depth of 2,546 m but Soviet drillers are starting a new hole with the aim of reaching the bedrock, that is, obtaining a 3.5-km ice core. Several other nations are also planning deep drilling in Antarctica (Australia, France, Japan, USA, for example). Scientists and operating agencies should develop a plan to organize an array of shallow, intermediate and deep ice cores ensuring both proper geographical coverage and coverage of relevant timescales at appropriate resolution. These plans should aim to obtain records of climate and forcing factors but also address other problems such as the interaction between climate and ice sheets and the

role of the ice-sheet mass balance in sea level changes.

A strong interaction between data acquisition and interpretation and modelling of palaeoclimate and palaeoenvironment changes including the atmosphere, ocean, cryosphere, hydrosphere and land processes and various biogeochemical cycles is also needed. As well as GCMs, which produce steady-state simulations, a hierarchy of time-dependent models ranging from simple or even conceptual^{46-48,61,62} to more complex two-dimensional models⁶³ can certainly provide information on how the various processes combine to produce observed climate changes. For this, understanding mechanisms that cause and are associated with rapid climate change is crucial. The effort in these directions must be well supported and the fact that the study of glacial-interglacial cycles constitutes one of the streams of the IGBP (International Geosphere Biosphere Program) project on Global Changes of the Past (PAGES) is very positive. Further studies of climate change will narrow the uncertainties in our present approach of estimating climate sensitivity. A modelling effort coupled with interpretation of palaeodata provides, along with the satellite data on radiative forcing⁶⁴, perhaps our best hope in the near future for predicting the impact of increasing greenhouse gases on climate. □

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*Global Change
files.*

A SURVEY OF INFORMED OPINION REGARDING THE NATURE AND REALITY OF A 'GLOBAL GREENHOUSE WARMING'

An Editorial

Last year I conducted a small survey of environmental scientists* actively involved in studying global climate change or planning activities in this area. The purpose of the survey was to obtain a sense of the scientific community regarding the reality of an anthropogenically-induced global climate change. The survey consisted of 16 questions, 13 dealing with technical aspects of the issue and 3 with institutional concerns. Of the 43 scientists contacted, 22 responded to the survey.

The device of a relatively short questionnaire to solicit scientific opinion on a topic as complex as climate change may appear of limited utility. However, there is an authority inherent in the massed, yet *independent* responses of a sizable number of active and reputable scientists that perhaps cannot be obtained in any other way.

Virtually all methods for obtaining a sense of the opinion on a scientific issue have potential problems. As an example, Federal pronouncements may be self-serving. An individual scientist's sentiment must be viewed with the suspicion of unintended prejudice. The joint statement of a group can be either skewed by the more vocal members of that group or presented as a bland commentary, heavily caveated and from which all of the meaningful controversy has been removed. Facts gathered by the news media may mislead by the media's tendency to prefer the dramatic over the substantive.

The results of this poll should also be viewed with a well-founded skepticism for another reason: timing. Were this poll conducted in the early spring of 1988, prior to the drought of that summer, rather than a year later, it is likely that at least some of the sentiments would have been different. Indeed, during that summer, the New York Times and the Washington Post were running a daily average of 10 column inches on the drought each day as well as editorials, in-depth heuristic articles and even a page or two of paid advertisement. It seems likely that such publicity had some effect, even on scientists' perceptions. Were this survey to have been conducted two decades ago, prevailing scientific sentiment might have divided between advocates of climate warming, climate stability or even climate cooling.

The following is a selection from the questions that appeared on the survey form. Most of the survey questions were deliberately framed to require a numerical score, generally from 1 to 10. However, non-quantitative comments were solicited and profusely proffered by the respondents adding to the substance of the survey.

* I wish to thank the following for participating in this survey: James Angell, Charles Baes, Jr., D. James Baker, Ralph J. Cicerone, Gordon MacDonald, Robert Dickinson, John Eddy, Hugh Ellsaesser, John Firor, Hermann Flohn, W. Lawrence Getes, James Hansen, Martin Hoffert, Michael MacCracken, Gerry Mahlman, Reginald Newell, Eugene Rasmusson, Stephen Schneider, Kevin Trenberth, George Woodwell and two equally distinguished scientists who preferred not to be identified.

Question on the Reality of the Greenhouse Effect

On a scale of zero to 10 (where zero indicates profound disbelief) what is your rating of the ultimate reality of a significant climate change, resembling current conceptions, that can be confidently ascribed to anthropogenically-produced, radiatively-active gases?

Answer

'Reality' range	Number in each range
0- 3	2
4- 6	2
7-10	18

Question on Unexpected Climate Occurrences

Are 'surprises' likely in some future features of climate (significant deviations from currently conceived types, rates, patterns or intensity of changes)?

Answer

Category	# of responses in category
Yes	22
No	0
No opinion	0

In addition to the unexpected unanimity, this question received the largest number of comments in the survey. The comments identified the interactions of clouds and the oceans in the climate system as the most important among the likely causes of the surprises. Research on these topics is given the highest priority within the most heavily funded scientific element in the U.S. Global Change Research Program plan, 'Our Changing Planet: The FY 1990 Research Plan' (1989).

Question on the Expected Magnitude of Climate Change

What is your opinion of the magnitude of the changes predicted for a doubling of radiatively-active gas concentration (or for any other bench mark that you use). Use a scale of zero to 10 where zero is much less than predicted, 5 is about as predicted and 10 is much greater than predicted. 'No basis for an opinion' is a valid answer as is, 'other'.

Answer

Category

0- 3
4- 6
7-10
No opinion

Question on Whether a Greenhouse Effect Has Already Begun

On a scale of zero to 10 (where zero indicates profound disbelief) what is your rating of the ultimate reality of a significant climate change, resembling current conceptions, that can be confidently ascribed to anthropogenically-produced, radiatively-active gases?

Answer

Value

0- 3
4- 6
7-10

Question on General Circulation Model (GCM) Simulations

On a scale of zero to 10 (where zero indicates profound disbelief) what is your rating of the ability of GCMs to simulate climate and climate after a doubling of CO₂ (mark you yourself subscribe)?

Answer

0- 3
4- 6
7-10

An allied question sought to determine the degree to which the changes. While about half of the comments that supplied the 'surprises' case, the writer actually of

The three questions on adm

Answer

Category	# of responses in each category
0- 3	3
4- 6	14
7-10	2
No opinion	3

Question on Whether a Greenhouse Climate Change Effect Has Actually Been Observed

On a scale of zero to 10 (where zero indicates complete disbelief) how confident are you with the proposition that there is some evidence that (an anthropogenically-induced) greenhouse effect has already received at least minimal observational verification?

Answer

Value	Number of responses for each value
0- 3	4
4- 6	10
7-10	8

Question on General Circulation Models (GCMs) Predictive Ability

On a scale of zero to 10 (where zero means useless and 10 is perfect) what is your rating of the ability of GCMs to simulate the general characteristics of past climate, current climate and climate after a doubling of radiatively-active gases (or to whatever benchmark you yourself subscribe)?

Answer

	Past	Present	Future
0- 3	4	1	4
4- 6	16	8	11
7-10	2	13	7

An allied question sought to identify alternatives to GCMs for simulating climate changes. While about half of the respondents answered that there *were* alternatives, it was the comments that supplied the information that in almost every 'yes, there are alternatives' case, the writer actually offered supplements to GCMs but not alternatives.

The three questions on administrative aspects of a lead institution for planning and

managing the U.S. effort suggested that there was no clear preference for the structure of the lead institution although there was some preference for a scientific leadership (as opposed to leadership by agency officials). Additionally, there was a preference for funding from a number of agencies rather than from just one source. Further, about half of the respondents thought that social sciences studies were appropriate to a climate change research program.

The results of this survey show:

- a general agreement that the issue is real,
- a unanimous accord that we may look forward to climatic 'surprises' although the gross magnitude of the climate change may be about as predicted,
- that there is evidence suggesting that some of the predicted environmental dislocations may already have been observed,
- that GCMs at least adequately describe the general features of a future climate and that there are no serious alternatives for simulating climate and, finally
- that there is no clear consensus regarding the structure of the institution or organization that scientists wish to see lead the United States' effort.

It is felt that the results of this survey, and what may be inferred from these results, constitute a strong and independent endorsement of the U.S Global Climate Change Research Program's position on the scale and breadth of the research program it is developing.

However, the unanimous agreement of the surveyed scientists that future surprises are likely argues for some additional research program mechanism that is flexible and agile enough to respond rapidly when a 'surprise' is recognized. Such an agility is not, but can be made, a concomitant of the large U.S. Program. This may be particularly necessary if the surprise is startling and not among the Program's long-term commitments for substantial research funding.

Reference

Committee on Earth Sciences: 1989, *Our Changing Planet: The FY 1990 Research Plan*, Washington, D.C.

Environmental Sciences Consultation and Services
859 Loxford Terrace
Silver Spring, MD 20901, U.S.A.

DAVID H. SLADE

GREENHOUSE GUESSING

An Editorial

For the past few thousand years we have heated the atmosphere by burning fuel. Since the Industrial Revolution the heating has increased enormously and other greenhouse gases, such as nitrogen, have been produced. The atmosphere is now disturbing a natural balance on prolonged (geological) time scales. Oceans and rocks. Many of the effects are the greenhouse effect.

The greenhouse effect is a prediction disputed. The greenhouse effect is more than most domestic ovens and planetary warming that may be. Without any radiatively active gases a global average. In fact, the greenhouse effect is +15°C. It is the greenhouse effect well below zero to comfortable habitable.

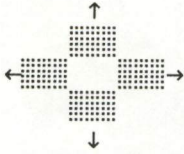
The rate of buildup of the greenhouse effect is fast that it can be confidently predicted sufficient quantities of all greenhouse gases equivalent to an effective doubling of the greenhouse effect.

Scientists have measured the greenhouse effect for over thirty years and we have measurements of the carbon dioxide concentration. It is no doubt that mankind's activities are the predictions of global climate change also well agreed. It is anticipated temperature increase of about 1°C (evaporation, cloud formation).

Observations of the buildup of the greenhouse effect potential impact on climate, but the predictions are not.

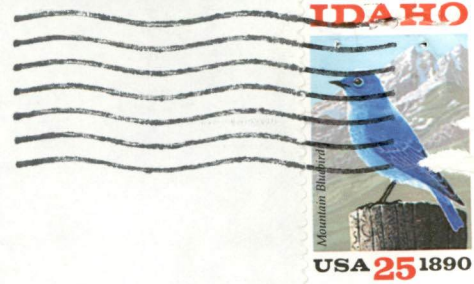
Observations of air and sea surface temperatures around the globe show an increase of about 1°C since 1900, globally averaged temperature degree Celsius. These observations are made by climate modellers and predictions. There are other factors which are also considered.

The dilemma which faces



DAVID H. SLADE
ENVIRONMENTAL SCIENCES
CONSULTATION AND SERVICES

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Silver Spring, Maryland 20901 USA



Dr. D. Allan Bromley
Office of Science and Technology Policy
Old Executive Office Building
Washington, DC 20506

"CORRESPONDENCE TRACKING"

TYPE: Information

DOCUMENT NUMBER: 9021259

FROM: Knauss, John A., US Depart. of Commerce

TO: Bromley

From Global Warming files

DATE OF
CORRESPONDENCE: 03/14/90

SUBJECT: 50 year business plan to make money from greenhouse
gas mitigation.

ASSIGNED TO:

ACTION REQUIRED: NONE

SENDER'S DUE DATE:

OSTP DUE DATE:

DATE COMPLETED:

COPIES TO: James Wyngaarden

WHITE HOUSE TRACKING #:

CONTACT PERSON:

REMARKS:

DATE RECEIVED: 03/19/90

FILE: NEOB/LS/GREENHOUSE GAS

MAR 19 1990



The Under Secretary of Commerce
for Oceans and Atmosphere
U.S. Department of Commerce
Washington, D.C. 20230

MAR 14 1990

The Honorable D. Allan Bromley
Assistant to the President for
Science & Technology
Old Executive Office Building
17th & Pennsylvania Avenue N.W.
Suite 358
Washington, D.C. 20506

Dear Allan,

Tom Murrin suggested I forward the enclosed. The first is a summary of a March 2 meeting I had with Director General Okamatsu, the most important part of which is the latter half in which he outlines MITI's "fifty year business plan" to make money from greenhouse gas mitigation. The second is a document he left with me that expands on that outline.

How does one get a "free market economy" to make fifty year business plans?

Should we?

Sincerely,

A handwritten signature in dark ink, appearing to be "J. Knauss", is written above the printed name.

John A. Knauss

cc : Tom Murrin

Withdrawal/Redaction Sheet

(George Bush Library)

Document No. and Type	Subject/Title of Document	Date	Restriction	Class.
01. Notes	Notes on meeting with Sozaburo Okamatsu, Director General for Environmental Policy at MITI (3 pp.)	3/5/90	(b)(1)	

Collection:

Record Group: Bush Presidential Records
Office: Science and Technology Policy, Office of (OSTP)
Series: Bromley, D. Allan, Files
Subseries: Global Climate Change Files
WHORM Cat.:
File Location: Global Warming: Greenhouse Gases (1) [1 of 3] [1990]

Date Closed: 4/6/2010	OA/ID Number: 62052-011
FOIA/SYS Case #: 2005-0336-F	Appeal Case #:
Re-review Case #:	Appeal Disposition:
P-2/P-5 Review Case #:	Disposition Date:
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 (b)(8) Release would disclose information concerning the regulation of financial institutions [(b)(8) of the FOIA]
 (b)(9) Release would disclose geological or geophysical information

The Global Environmental Challenge:
Japanese Initiative for Technological Breakthrough

Ministry of International Trade and Industry
The Government of Japan

March 1990

1. Initiative

1) The Need for Technological Breakthrough

Unprecedented levels of economic activity have raised such environmental issues as global warming, depletion of the ozone layer, deforestation and others which pose a major threat to the continued existence and well-being of mankind. Current and future efforts to ensure energy efficiency, and introduce alternative energy sources, reforestation and other measures will not be enough to combat global environmental problems. "Sustainable Development" as advocated by the World Commission on Environment and Development (Report: Our Common Future) and by the UNEP Governing Council will require unprecedented technological breakthroughs in a wide array of fields including energy and biotechnology, as recommended by the "Tokyo Conference on the Global Environment and Human Response toward Sustainable Development" in September 1989.

2) The Development of a New Industrial Technology

As part of the global response to this technological challenge, the Ministry of International Trade and Industry of Japan (MITI) intends to launch a series of research and development projects to push forward the frontiers of technology, and create a new environmentally friendly industrial technology. This new industrial technology will reconcile the apparent dichotomy between economic growth and environmental protection by integrating the production and consumption processes into the ageless cycle of energy and matter in the geo-biosphere. Such efforts must be undertaken now if the 21st century is to be a century of sustainable development.

3) The Role of the New Institute

In order to provide a strong and effective institutional base for these research projects, MITI intends to establish a new research and development center called "The Institute of Industrial Technology for the Global Environment." The Institute will work closely with major research and development institutions, both foreign and domestic.

Scientists and engineers from abroad are more than welcome to participate in its activities.

4) Promoting Environmental Technology

Grants will be available for research and development for innovative environmental technologies. Here again, we encourage inquiries from scientists and engineers abroad. The grants will be awarded to projects on their own merits, regardless of country of origin.

2. New Projects for "Environmentally Friendly" Technology

1) Development of Full Carbon Cycle Technology

(CO₂ fixation and reutilization)

CO₂ emissions from industrial activities is currently the major source of greenhouse gases. The objective of this project is to develop industrial technology for converting CO₂ from industrial sources into useful products such as methanol and other hydrocarbon compounds. Both chemical and biological methods will be studied.

2) Development of Environmentally Benign Materials

a) Development of advanced substitutes for CFCs

In this project, advanced CFC substitutes will be developed, not only to protect the stratospheric ozone layer, but also to reduce the potential greenhouse effect of CFC substitutes.

b) Development of biodegradable plastics

Various biodegradable plastics will be developed in order to reduce the accumulation of environmentally burdensome conventional plastics.

3) Development of Environmentally Benign Production Processes

Biological reactions, which occur under low-temperature and low-pressure, will be developed to replace energy intensive chemical reactions in production processes. This will result in enormous energy savings.

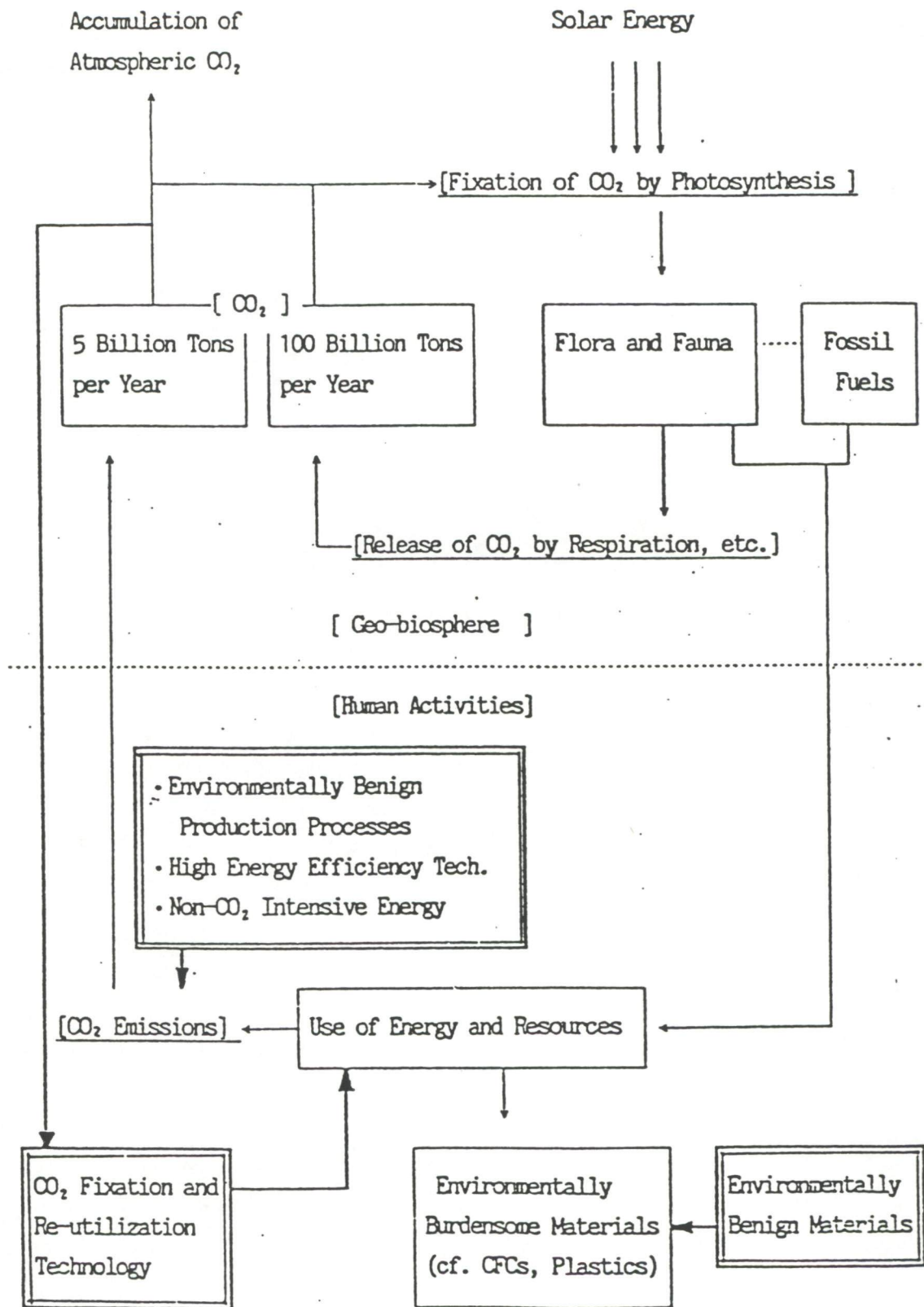


Figure 1. The Global Carbon Cycle and the Technological Response

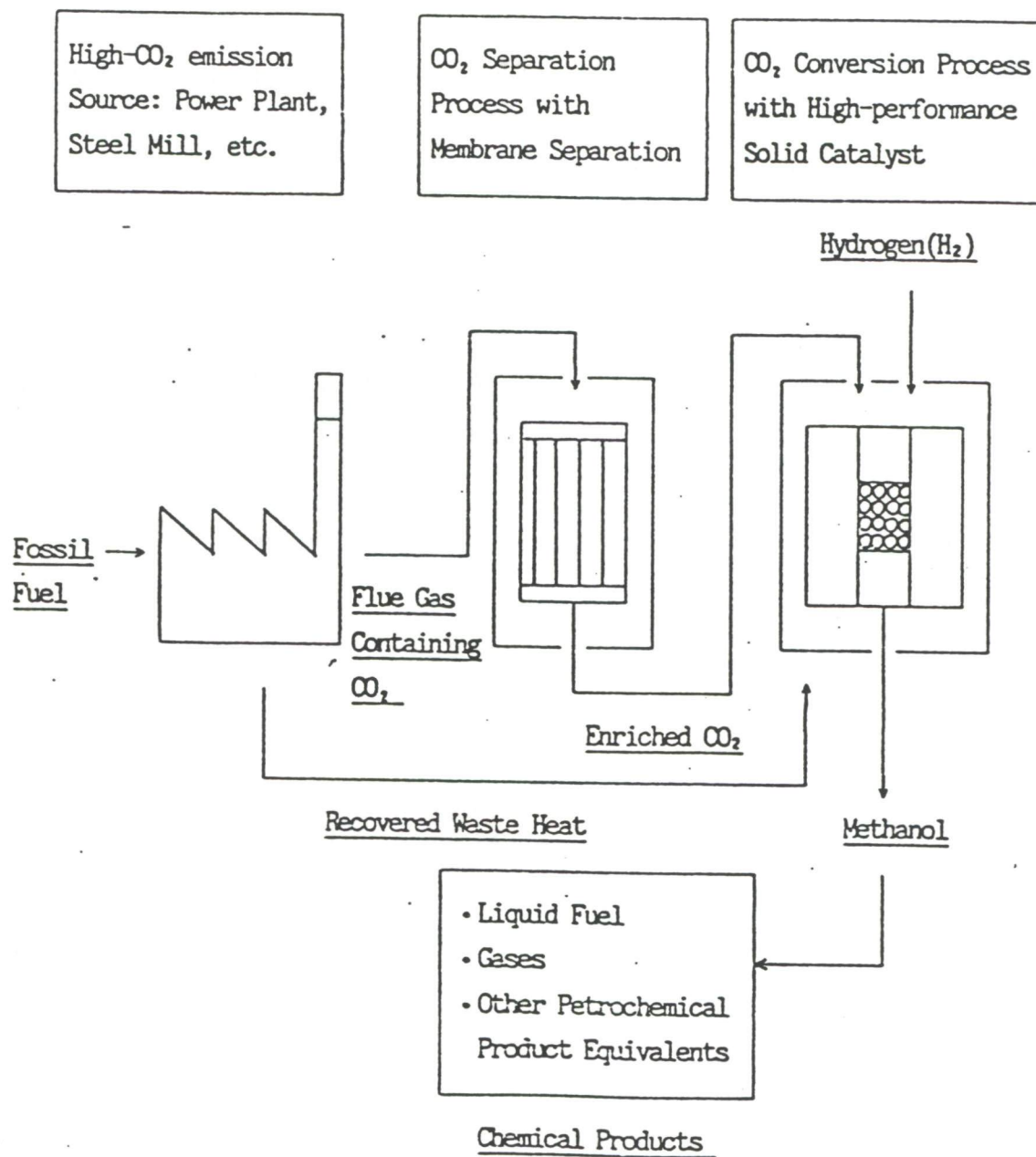


Figure 2. CO₂ Fixation and Re-utilization Technology
- Catalytic Hydrogenation -

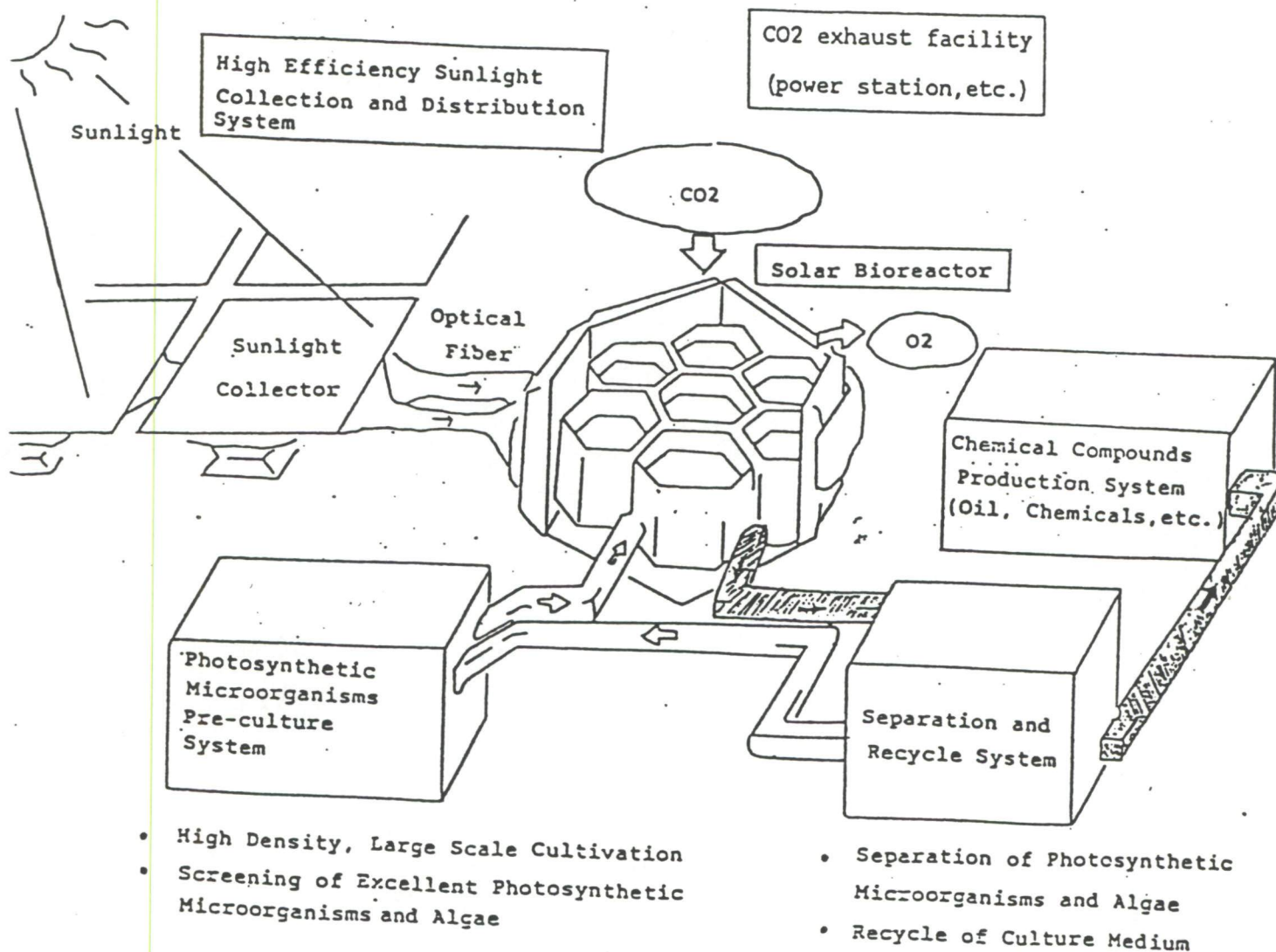


Figure 3. CO2 Fixation and Re-utilization Technology
- Photosynthetic Microorganisms and Algae -

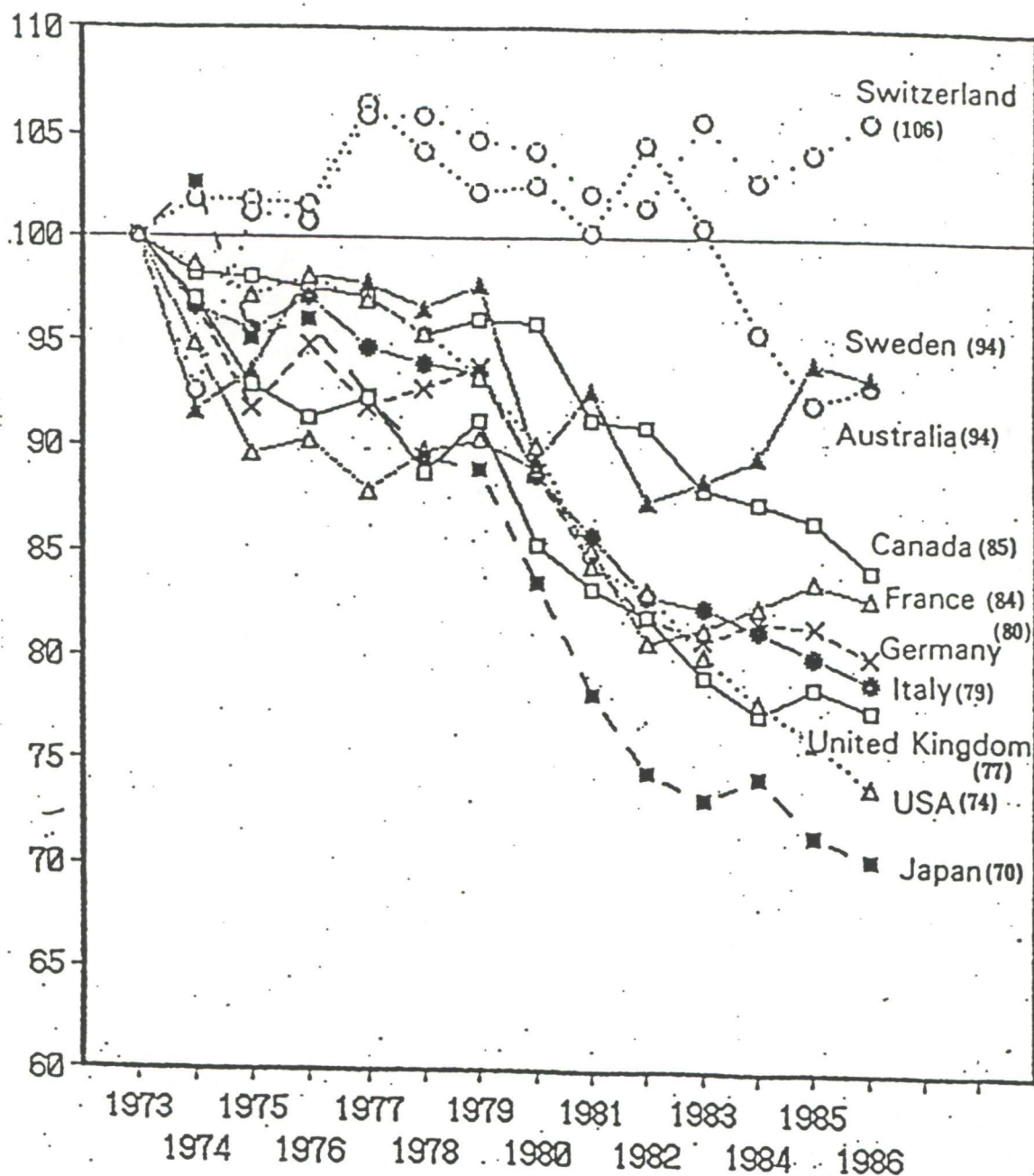


Figure 4. Changes in Index of Energy Consumption/GDP Rate
in Principal OECD Countries

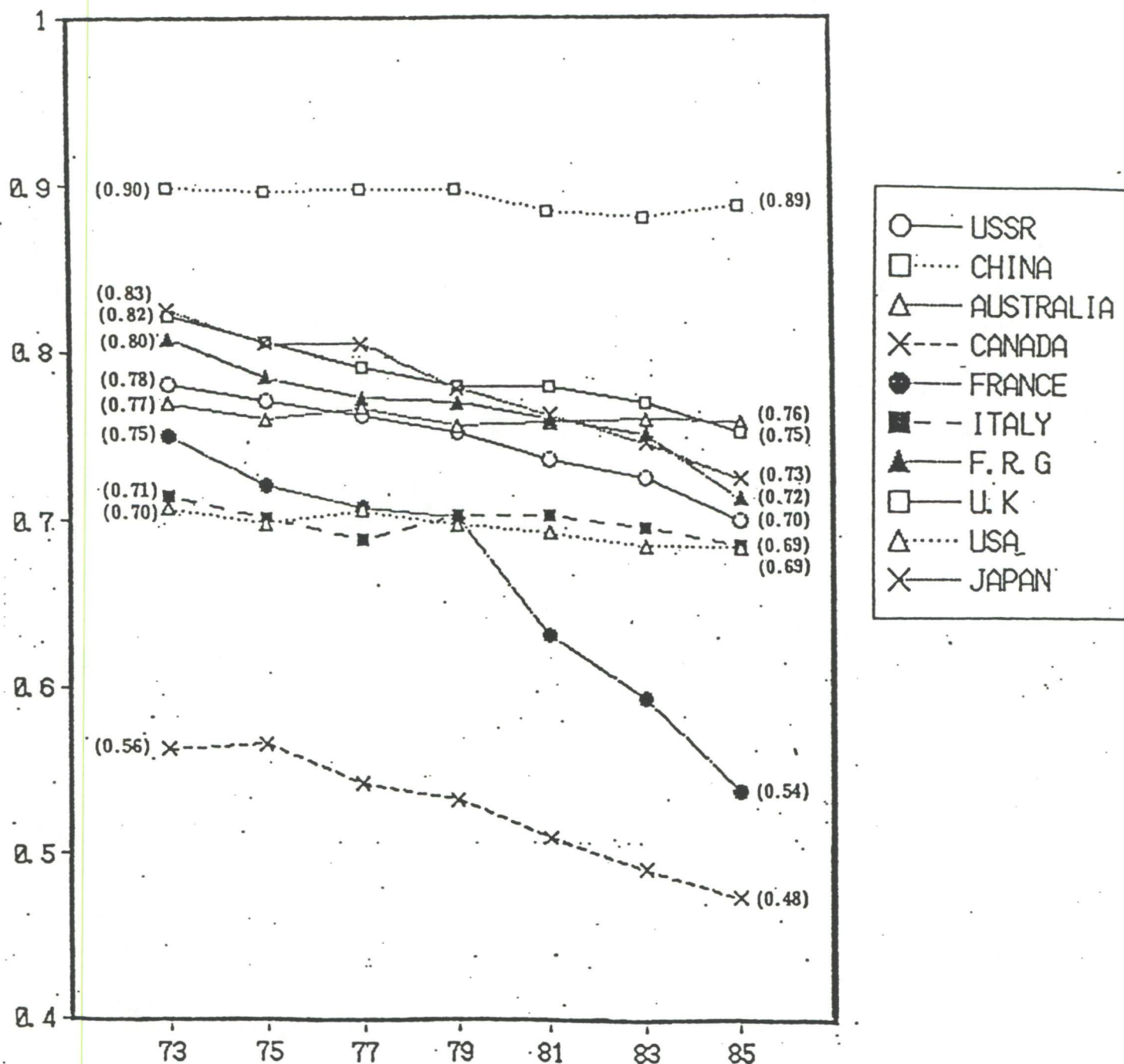


Figure 5. Changes in the CO₂ Emission/Energy Rate
in Major Countries (tonnes of carbon/ toe)

Table. MITI's Budget Plan for Fiscal Year 1990

	(Million Yen)
1) Development of Full Carbon Cycle Technology (CO ₂ fixation and reutilization) and Related Research	2,210
2) Development of Environmentally Benign Materials	541
a) Development of advanced substitutes for CFCs	361
b) Development of biodegradable plastics	180
3) Development of Environmentally Benign Production Processes	72
4) Environment Protection Technology Development in the Private Sector	2,000
5) Other Research Programs, etc.	1,187
Total	6,010

SECTION 1

**GREENHOUSE GASES
AND AEROSOLS**

**Peer Reviewed Assessment
for WG1 Plenary**

April 25 1990

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SECTION 1: GREENHOUSE GASES AND AEROSOLS

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EXECUTIVE SUMMARY

The Earth's climate is dependent upon the radiative balance of the atmosphere, which in turn depends upon the input of solar radiation and the atmospheric abundances of radiatively active trace gases (i.e., greenhouse gases), clouds and aerosols.

Since the industrial revolution the atmospheric concentrations of several greenhouse gases, i.e., carbon dioxide (CO₂), methane (CH₄), chlorofluorocarbons (CFCs), nitrous oxide (N₂O), and tropospheric ozone (O₃), have been increasing, primarily due to human activities. Several of these greenhouse gases have long atmospheric lifetimes, decades to centuries, which means that their atmospheric concentrations respond slowly to changes in emission rates. In addition, there is evidence that the concentrations of tropospheric aerosols have increased at least regionally.

Carbon Dioxide: The atmospheric CO₂ concentration, at 353 ppmv in 1990, is now about 25% greater than the pre-industrial (1750-1800) value of about 280 ppmv, and higher than at any time in at least the last 160,000 years. Carbon dioxide is currently rising at about 1.8 ppmv (0.5%) per year due to anthropogenic emissions. Anthropogenic emissions of CO₂ are estimated to be 5.7±0.5 Gt C (in 1987) due to fossil fuel burning, plus 0.6 - 2.5 Gt C (in 1980) due to deforestation. The atmospheric increase during the past decade corresponds to (48±8)% of the total emissions during the same period with the remainder being taken up by the oceans and land. Indirect evidence suggests that the land and oceans sequester CO₂ in roughly equal proportions, though the mechanisms are not all well understood. The time taken for atmospheric CO₂ to adjust to changes in sources or sinks is of order 50-200 years, determined mainly by the slow exchange of carbon between surface waters and deeper layers of the ocean. Consequently, CO₂ emitted into the atmosphere today will influence the atmospheric concentration of CO₂ for centuries into the future. Three models have been used to estimate that even if anthropogenic emissions of CO₂ could be kept constant at present day rates, atmospheric CO₂ would increase to 415 - 480 ppmv by the year 2050, and to 460 - 560 ppmv by the year 2100. In order to stabilize concentrations at present day levels, an immediate reduction in global anthropogenic emissions by 60-80 percent would be necessary.

Methane: Current atmospheric CH₄ concentration, at 1.72 ppmv, is now more than double the pre-industrial (1750-1800) value of about 0.8 ppmv, and is increasing at a rate of about 0.015 ppmv (0.9%) per year. The major sink for CH₄, reaction with hydroxyl (OH) radicals in the troposphere, results in a relatively short atmospheric lifetime of about 10 years. Human activities such as rice cultivation, domestic ruminant rearing, biomass burning, coal mining, and natural gas venting have increased the input of CH₄ into the atmosphere, which combined with an possible decrease in the concentration of tropospheric OH, yields the observed rise in global CH₄. However, the quantitative importance of each of the factors contributing to the observed increase is not well known at present. In order to stabilize concentrations at present day levels, an immediate reduction in global anthropogenic emissions by 15-20 percent would be necessary.

Chlorofluorocarbons: The current atmospheric concentrations of the anthropogenically produced halocarbons, CCl₃F (CFC-11), CCl₂F₂ (CFC-12), C₂Cl₃F₃ (CFC-113) and CCl₄ (carbon tetrachloride) are about 280 pptv, 484 pptv, 60 pptv, and 146 pptv, respectively. Over the past few decades their concentrations, except for CCl₄, have increased more rapidly (on a percentage basis) than the other greenhouse gases, currently at rates of at least 4% per year. The fully halogenated CFCs and CCl₄ are primarily removed by photolysis in the stratosphere, and have atmospheric lifetimes in excess of 50 years. Future emissions will, most likely, be eliminated or significantly lower than today's because of current international negotiations to strengthen regulations on chlorofluorocarbons. However, the atmospheric concentrations of CFCs 11, 12 and 113 will still be significant (30 - 40% of current) for at least the next century because of their long atmospheric lifetimes.

Nitrous Oxide: The current atmospheric N_2O concentration, at 310 ppbv, is now about 8% greater than in the pre-industrial era, and is increasing at a rate of about 0.8 ppbv (0.25%) per year. The major sink for N_2O , photolysis in the stratosphere, results in a relatively long atmospheric lifetime of about 150 years. It is difficult to quantitatively account for the source of the current increase in the atmospheric concentration of N_2O but it is thought to be due to human activities. Recent data suggest that the total annual flux of N_2O from combustion and biomass burning is much less than previously believed. Agricultural practices may stimulate emissions of N_2O from soils and play a major role. In order to stabilize concentrations at present day levels, an immediate reduction of 70 - 80% of the additional flux of N_2O that has occurred since the pre-industrial era would be necessary.

Ozone: Ozone is an effective greenhouse gas especially in the middle and upper troposphere and lower stratosphere. Its concentration in the troposphere is highly variable because of its short lifetime. It is photochemically produced in-situ through a series of complex reactions involving carbon monoxide (CO), CH_4 , non-methane hydrocarbons (NMHC), and nitrogen oxide radicals (NO_x), and also transported downward from the stratosphere. The limited observational data support positive trends of about 1% per year for O_3 below 8 km in the northern hemisphere (consistent with positive trends in several of the precursor gases, especially NO_x , CH_4 , and CO) but probably close to zero trend in the southern hemisphere. There is also evidence that O_3 has decreased by a few percent globally in the lower stratosphere (below 25 km) within the last decade. Unfortunately, there are no reliable long-term data near the tropopause.

Aerosol particles: Aerosol particles have a lifetime of at most a few weeks in the troposphere and occur in highly variable concentrations. A large proportion of the particles that influence cloud processes and the radiative balance is derived from gaseous sulphur emissions. Due to fossil fuel combustion, these emissions have more than doubled globally, causing a large increase in the concentration of aerosol sulphate especially over and around the industrialized regions of Europe and North America. Future concentrations of aerosol sulphate will vary in proportion to changes in anthropogenic emissions. Aerosol particles derived from natural (biological) emissions may contribute to climate feedback processes. During a few years following major volcanic eruptions the concentrations of natural aerosol particles in the stratosphere can be greatly enhanced.

1.1 INTRODUCTION

The Earth's climate is dependent upon the radiative balance of the atmosphere, which in turn depends upon the input of solar radiation and the atmospheric abundances of radiatively active trace gases (i.e., greenhouse gases), clouds and aerosols. Consequently, it is essential to gain an understanding of how each of these climate "forcing agents" varies naturally, and how some of them might be influenced by human activities.

The chemical composition of the Earth's atmosphere is changing, largely due to human activities (Table 1). Air trapped in Antarctic and Greenland ice shows that there have been major increases in the concentrations of radiatively active gases such as carbon dioxide (CO₂), methane (CH₄), and nitrous oxide (N₂O) since the beginning of the industrial revolution. In addition, industrially produced chlorofluorocarbons (CFCs) are now present in the atmosphere in significant concentrations, and there is evidence that the concentrations of tropospheric O₃ and aerosols have increased at least regionally. Atmospheric measurements indicate that in many cases the rates of change have increased in recent decades. Many of the greenhouse gases have long atmospheric life-times, decades to centuries, which implies that their atmospheric concentrations respond slowly to changes in emission rates.

TABLE 1
Summary of Key Greenhouse Gases Influenced by Human Activities ¹

Parameter	CO ₂	CH ₄	CFC-11	CFC-12	N ₂ O
Pre-industrial atmospheric concentration (1750-1800)	280 ppmv ²	0.8 ppmv	0	0	288 ppbv ²
Current atmospheric concentration (1990) ³	353 ppmv	1.72 ppmv	280 pptv ²	484 pptv	310 ppbv
Current rate of annual atmospheric accumulation	1.8 ppmv (0.5%)	0.015 ppmv (0.9%)	9.5 pptv (4%)	17 pptv (4%)	0.8 ppbv (0.25%)
Atmospheric lifetime ⁴ (years)	(50-200)	10	65	130	150

¹ Ozone has not been included in the table because of lack of precise data.

² ppmv = parts per million by volume;

ppbv = parts per billion by volume;

pptv = parts per trillion by volume.

³ The current (1990) concentrations have been estimated based upon an extrapolation of measurements reported for earlier years, assuming that the recent trends remained approximately constant.

⁴ For each gas in the table, except CO₂, the "lifetime" is defined here as the ratio of the atmospheric content to the total rate of removal. This time scale also characterizes the rate of adjustment of the atmospheric concentrations if the emission rates are changed abruptly. CO₂ is a special case since it has no real sinks, but is merely circulated between various reservoirs (atmosphere, ocean, biota). The "lifetime" of CO₂ given in the table is a rough indication of the time it would take for the CO₂ concentration to adjust to changes in the emissions (see section 1.2.1 for further details).

The effectiveness of a greenhouse gas in influencing the Earth's radiative budget is dependent upon its atmospheric concentration and its ability to absorb outgoing long-wave terrestrial radiation. Tropospheric water vapour is the single most important greenhouse gas, but its atmospheric concentration is not significantly influenced by direct anthropogenic emissions. Of the greenhouse gases that are affected by human activities, CO₂ has the largest radiative effect, followed by the CFCs, CH₄, tropospheric O₃, and N₂O. Although the present rate of increase in the atmospheric concentration of CO₂ is about a factor of 70,000 times greater than that of CCl₃F (CFC-11) and CCl₂F₂ (CFC-12) combined, and a factor of about 120 times greater than that of CH₄, its contribution to changes in the radiative forcing during the decade of the 1980s was about 55%, compared to 17% for CFCs (11 and 12), and 15% for CH₄ (see Section 2). Other CFCs and N₂O accounted for about 8%, and 5%, respectively, of the changes in the radiative forcing. While the contribution from tropospheric O₃ may be important, it has not been quantified because the observational data is inadequate to determine its trend. This pattern arises because of differences in the efficiencies of the gases to absorb terrestrial radiation.

Aerosol particles play an important role in the climate system because of their direct interaction (absorption and scattering) with solar and terrestrial radiation, as well as through their influence on cloud processes and thereby, indirectly, on radiative fluxes.

There is a clear need to document the historical record of the atmospheric concentrations of greenhouse gases and aerosols, as well as to understand the physical, chemical, geological, biological and social processes responsible for the observed changes. A quantitative understanding of the atmospheric concentrations of these gases requires knowledge of: the cycling and distribution of carbon, nitrogen and other key nutrients within and between the atmosphere, terrestrial ecosystems, oceans and sediments; and the influence of human actions on these cycles. Without knowledge of the processes responsible for the observed past and present changes in the atmospheric concentrations of greenhouse gases and aerosols it will not be possible to predict with confidence future changes in atmospheric composition, nor therefore the resulting changes in the radiative forcing of the atmosphere.

1.2 CARBON DIOXIDE

1.2.1 The Cycle of Carbon in Nature

Carbon in the form of CO₂, carbonates, organic compounds, etc. is cycled between various reservoirs, atmosphere, oceans, land biota and marine biota, and on geological time scales, also sediments and rocks (Figure 1.1; for more detailed reviews see Sundquist, 1985; Bolin, 1981, 1986; Trabalka, 1985; Siegenthaler, 1986). The largest natural exchange fluxes occur between the atmosphere and the terrestrial biota and between the atmosphere and the surface water of the oceans. By comparison, the net inputs into the atmosphere from fossil fuel combustion and deforestation are much smaller, but are large enough to modify the natural balance.

The turnover time of CO₂ in the atmosphere, measured as the ratio of the content to the fluxes through it, is about 4 years. This means that on average it takes only a few years before a CO₂ molecule in the atmosphere is taken up by plants or dissolved in the ocean. This short time scale must not be confused with the time it takes for the atmospheric CO₂ level to adjust to a new equilibrium if sources or sinks change. This adjustment time, corresponding to the lifetime in Table 1, is of the order of 50 - 200 years, determined mainly by the slow exchange of carbon between surface waters and the deep ocean. The adjustment time is important for the discussions on global warming potential, cf. Section 2.2.7.

Because of its complex cycle, the decay of excess CO₂ in the atmosphere does not follow a simple exponential curve, and therefore a single time scale cannot be given to characterize the whole adjustment process toward a new equilibrium. The two curves in Figure 1.2, which represent simulations of a pulsed input of CO₂ into the atmosphere using atmosphere-ocean models (one box model and one General Circulation Model (GCM)), clearly show that the

initial response (governed mainly by the uptake of CO₂ by ocean surface waters) is much more rapid than the later response (influenced by the slow exchange between surface waters and deeper layers of the oceans). For example, the first reduction by 50 percent occurs within some 50 years, whereas the reduction by another 50 percent (to 25 percent of the initial value) requires approximately another 200 years. The concentration will actually never return to its original value, but reach a new equilibrium level; about 15 percent of the total amount of CO₂ emitted will remain in the atmosphere.

1.2.1.1 The role of the atmosphere

The mean annual concentration of CO₂ is relatively homogeneous throughout the troposphere because the troposphere is mixed on a time scale of about 1 year. The pre-industrial atmospheric CO₂ concentration was about 280 ppmv, as reconstructed from ice core analyses (c.f. Section 1.2.4.1), corresponding to an atmospheric amount of 594 Gigatonnes of carbon (GtC: 1 Gt = 10⁹t = 10¹⁵g; 1 ppmv CO₂ of the global atmosphere equals 2.12 GtC and 7.8 Gt CO₂); today, the level is about 353 ppmv (Figures 1.3 and 1.4). The atmospheric increase has been monitored since 1958 at a growing number of stations (Keeling and Heimann, 1986; Keeling et al., 1989a; Beardsmore and Pearman, 1987; Conway et al., 1988).

1.2.1.2 The role of the ocean

On time scales of decades or more, the CO₂ concentration of the unperturbed atmosphere is mainly controlled by the exchange with the oceans, since this is the largest of the carbon reservoirs. There is a continuous exchange of CO₂ in both directions between the atmosphere and oceans. The net flux into (or out of) the ocean is driven by the difference between the atmospheric partial pressure of CO₂ and the equilibrium partial pressure of CO₂ (pCO₂) in surface waters.

The exchange of carbon between the surface and deeper layers is accomplished mainly through transport by water motions. Ventilation of the thermocline (approximately the uppermost km of the ocean) is particularly important for the downward transport of anthropogenic CO₂. The deep circulation is effective on time scales of 100-1000 years.

The natural carbon cycle in the ocean and in particular pCO₂ in surface ocean water are strongly influenced also by biological processes. The marine biota serve as a "biological pump", transporting organic carbon from surface waters to deeper layers as a rain of detritus at a rate of about 4 GtC per year (Eppley and Peterson, 1979), which is balanced by an equal upward transport of carbon by deeper water richer in CO₂ than surface water. This "biological pump" has the effect of reducing surface pCO₂ very substantially: without the biological pump ("dead ocean") the pre-industrial CO₂ level would have been higher than the observed value of 280 ppmv, at perhaps 450 ppmv (Wenk, 1985; Bacastow and Maier-Reimer, 1990). Alterations in the marine biota due to climatic change could therefore have a substantial effect on CO₂ levels in the future. Note, however, that the "biological pump" does not help to sequester anthropogenic CO₂ (see Section 1.2.4.2).

1.2.1.3 The role of terrestrial vegetation and soils

The most important processes in the exchange of carbon are those of photosynthesis, autotrophic respiration (i.e., CO₂ production by the plants) and heterotrophic (i.e., essentially microbial) respiration converting the organic material back into CO₂ mainly in soils (c.f. Section 10 for a detailed discussion). Net primary production (NPP) is the net annual uptake of CO₂ by the vegetation; NPP is equal to the gross uptake (gross primary production, GPP) minus autotrophic respiration. In an unperturbed world, NPP and decomposition by heterotrophic respiration are approximately balanced on an annual basis; formation of soils and peat corresponds to a (relatively small) excess of NPP.

The carbon balance can be changed considerably by the direct impact of man (land use changes, particularly deforestation), by climate changes, and by other changes in the environment, e.g., atmospheric composition. Since the pools and fluxes are large (NPP 50-60 GtC per year, GPP 90-120 GtC per year; Houghton et al, 1985b), any perturbations can have a significant effect on the atmospheric concentration of CO₂.

1.2.2 Anthropogenic Perturbations

The concentrations of CO₂ in the atmosphere are primarily affected by two anthropogenic processes: release of CO₂ from fossil fuel combustion, and changes in land use such as deforestation.

1.2.2.1 Historical fossil fuel input

The global input of CO₂ to the atmosphere from fossil fuel combustion, plus minor industrial sources like cement production, has shown an exponential increase since 1860 (about 4% per year), with major interruptions during the two world wars and the economic crisis in the thirties (Figure 1.5). Following the "oil crisis" of 1973, the rate of increase of the CO₂ emissions first decreased to approximately 2% per year, and after 1979 the global emissions remained almost constant at a level of 5.3 GtC per year until 1985, when they started to rise again, reaching 5.7 GtC per year in 1987 (Figure 1.5). The cumulative release of CO₂ from fossil fuel use and cement manufacturing from 1850 to 1987 is estimated at 200 GtC $\pm$ 10% (Marland, 1989).

Ninety five percent of the industrial CO₂ emissions are from the northern hemisphere, dominated by industrial countries, where annual releases reach up to about 5 tC per capita (Rotty and Marland, 1986). In contrast, CO₂ emission rates in most developing countries lie between 0.2 and 0.6 tC per capita per year. However, the relative rate of increase of the CO₂ emissions is much larger in the developing countries (ca 6% per year), showing almost no slowing down after 1973 in contrast to Western Europe and North America where the rate of increase decreased from about 3% per year (1945-72) to less than 1% per year (1973-84).

1.2.2.2 Historical land use changes

The vegetation and soils of unmanaged forests hold 20 to 100 times more carbon per unit area than agricultural systems. The amount of carbon released to the atmosphere compared to that accumulated on land as a result of land use change depends on the amounts of carbon held in biomass and soils, rates of oxidation of wood products (either rapidly through burning or more slowly through decay), rates of decay of organic matter in soils, and rates of regrowth of forests following harvest or abandonment of agricultural land. The heterogeneity of terrestrial ecosystems makes estimation of global inventories and fluxes difficult.

The total release of carbon to the atmosphere from changes in land use, primarily deforestation, between 1850 and 1985 has been estimated to be about 115 GtC (Houghton and Skole, 1990), with an error limit of about $\pm$ 35 GtC. The components of the flux to the atmosphere are: (1) burning associated with land use change; (2) decay of biomass on site (roots, stumps, slash, twigs etc.); (3) oxidation of wood products removed from site (paper, lumber, waste etc.); (4) oxidation of soil carbon; minus (5) regrowth of trees and redevelopment of soil organic matter following harvest. Although the greatest releases of carbon in the nineteenth and early twentieth centuries were from lands in the temperate zone (maximum 0.5 GtC per year), the major source of carbon during the past several decades has been from deforestation in the tropics, with a significant increase occurring since 1950. Over the entire 135 yr period, the release from tropical regions is estimated to have been 2-3 times greater than the release from middle and high latitudes. Estimates of the flux in 1980 range from 0.6 to 2.5 GtC (Houghton et al. 1985a, 1987, 1988; Derwiler and Hall, 1988): virtually all of this flux is from the tropics. The few regions for which data exist suggest that the annual flux is higher now than it was in 1980.

1.2.3 Long-Term Atmospheric Carbon Dioxide Variations

The most reliable information on past atmospheric CO₂ concentrations is obtained by the analysis of polar ice cores. The process of air occlusion lasts from about 10 up to 1000 years, depending on local conditions (e.g., precipitation rate), so that an air sample in old ice reflects the atmospheric composition averaged over a corresponding time interval.

Measurements on samples representing the last glacial maximum (18,000 yr before present) from ice cores from Greenland and Antarctica (Neftel et al, 1982, 1988; Delmas et al., 1980) showed CO₂ concentrations of 180-200 ppmv, i.e., about 70 percent of the pre-industrial value. Analyses on the ice cores from Vostok, Antarctica, have provided new data on natural variations of CO₂, covering a full glacial-interglacial cycle (Figure 1.6; Barnola et al., 1987). Over the whole period there is a remarkable correlation between polar temperature, as deduced from deuterium data, and the CO₂ profile. The glacial-interglacial shifts of CO₂ concentrations must have been linked to large-scale changes in the circulation of the ocean, and in the whole interplay of biological, chemical and physical processes, but the detailed mechanisms are not yet very clear. The CO₂ variations were large enough to potentially contribute, via the greenhouse effect, to a substantial (although not the major) part of the glacial-interglacial climate change (Hansen et al., 1984; Broccoli and Manabe, 1987).

Ice core studies on Greenland ice indicate that during the last glaciation, CO₂ concentration shifts of the order of 50 ppmv may have occurred within less than 100 years (Stauffer et al., 1984), parallel to abrupt, drastic climatic events (temperature changes of the order of 5°C). These rapid CO₂ changes have not yet been identified in ice cores from Antarctica (possibly due to long occlusion times; Neftel et al., 1988), therefore, it is not yet clear if they are real or represent artefacts in the ice record.

1.2.4 The Contemporary Record of Carbon Dioxide - Observations and Interpretation

1.2.4.1 The carbon dioxide increase from pre-industrial period

Relatively detailed CO₂ data have been obtained for the last millennium from Antarctic ice cores (Neftel et al, 1985a; Friedli et al, 1986; Siegenthaler et al., 1988; Raynaud and Barnola, 1985, Pearman et al., 1986). They indicate that during the period 1000 to 1800, the atmospheric concentration was between 270 and 290 ppmv. The relative constancy seems surprising in view of the fact that the atmosphere exchanges about 30 percent of its CO₂ with the oceans and biota each year. This indicates that the sensitivity of atmospheric CO₂ levels to minor climatic changes such as the Little Ice Age (lasting from the end of the 16th to the middle of the 19th century), where global mean temperatures probably decreased by about 1°C, is small.

A precise reconstruction of the CO₂ increase during the past two centuries has been obtained from an ice core from Siple Station, Antarctica (Figure 1.3; Neftel et al., 1985a, Friedli et al., 1986). These results indicate that CO₂ started to rise around 1800 and had already increased by about 15 ppmv by 1900. Precise direct atmospheric measurements started in 1958, when the level was about 315 ppmv and the rate of increase 0.6 ppmv per year. The present atmospheric CO₂ level has reached 353 ppmv, and the mean growth rate has now reached about 1.8 ppmv per year (Figure 1.4; Keeling et al, 1989a).

1.2.4.2 Uptake by the ocean

The ocean is an important reservoir for taking up anthropogenic CO₂. The relative increase of dissolved inorganic carbon (total CO₂) in ocean water is smaller than in the atmosphere (only 2-3 percent until now, see below). Precise measurements of dissolved inorganic carbon can be made with present analytical tools. However, an accurate determination of the trend in dissolved inorganic carbon is difficult because of its variability in time and space. Hence, repeated transects and time series will be required to assess the total oceanic CO₂ uptake with good precision.

The net flux of CO₂ into (or out of) the ocean is given by the product of a gas transfer coefficient and $\Delta p\text{CO}_2$ (the CO₂ partial pressure difference between ocean and atmosphere). The gas transfer coefficient increases with increasing wind speed and also depends on water temperature. Therefore, the net flux into the ocean can be estimated from a knowledge of the atmospheric CO₂ concentration, pCO₂ in surface water (for which the data are still sparse), the global distribution of wind speeds over the ocean as well as the relation between wind speed and gas transfer coefficient (which is known to $\pm 30\%$ only). There have been several

estimates of the global net uptake of CO₂ by the oceans using observations (e.g., Enting and Pearman, 1982, 1987). The most recent estimate yields 1.6 GtC per year (Tans et al., 1990); the error of this estimate is, according to the authors, not easy to estimate.

Estimates of oceanic CO₂ uptake, in the past and in the future, require models of the global carbon cycle that take into account air-sea gas exchange, aqueous carbonate chemistry and the transport from the surface to deep ocean layers. The aqueous carbonate chemistry in sea water operates in a mode that if the atmospheric CO₂ concentration increases by e.g., 10%, then the concentration of dissolved inorganic carbon in sea water increases by only about 1% at equilibrium. Therefore, the ocean is not such a powerful sink for anthropogenic CO₂ as might seem first when comparing the relative sizes of the reservoirs (Figure 1.1).

The rate at which anthropogenic CO₂ is transported from the surface to deeper ocean layers is determined by the rate of water exchange in the vertical. It is known from measurements of the radioactive isotope ¹⁴C that on average it takes hundreds to about one thousand years for water at the surface to penetrate to well below the mixed layer of the major oceans (e.g., Broecker and Peng, 1982). Thus, in most oceanic regions only the top several hundred metres of the oceans have at present taken up significant amounts of anthropogenic CO₂. An exception is the North Atlantic Ocean where bomb-produced tritium has been observed even near the bottom of the sea, indicating the active formation of new deep water.

The rain of biogenic detrital particles, which is important for the natural carbon cycle, does not significantly contribute to a sequestering of excess CO₂, since the marine biota do not directly respond to the CO₂ increase. Their activity is controlled by other factors, such as light, temperature and limiting nutrients (e.g., nitrogen, phosphorus, silicon). Thus, only the input of fertilizers (phosphate, nitrate) into the ocean through human activities may lead to an additional sedimentation of organic carbon in the ocean; different authors have estimated the size of this additional sink between 0.04 and 0.3 GtC per year (see Baes, 1985). It seems thus justified to estimate the fossil fuel CO₂ uptake to date considering the biological flux to be constant, as long as climatic changes due to increasing greenhouse gases, or natural causes, do not modify the marine biotic processes. Although this appears a reasonable assumption for the past and present situation, it may well not be so in the future.

The carbon cycle models used to date to simulate the atmosphere-ocean system have often been highly simplified, consisting of a few well-mixed or diffusive reservoirs (boxes) (e.g., Oeschger et al. 1975; Broecker et al., 1980; Bolin, 1981; Enting and Pearman, 1987; Siegenthaler, 1983). Even though these box models are highly simplified they are a powerful means for identifying the importance of the different processes that determine the flux of CO₂ into the ocean (e.g., Broecker and Peng, 1982; Peng and Broecker, 1985). The results of these models are considered to be reasonable because, as long as the ocean circulation is not changing, the models need only simulate the transport of excess CO₂ from the atmosphere into the ocean, but not the actual dynamics of the ocean. In the simple models, the oceanic transport mechanisms, eg., formation of deep water, are parameterized. The transport parameters (e.g., eddy diffusivity) are determined from observations of transient tracers that are analogs to the flux of anthropogenic CO₂ into the ocean. If a model reproduces correctly the observed distribution of, e.g., bomb-produced ¹⁴C, then it might be expected to simulate reasonably the flux of CO₂ into the ocean. A 1-D box-diffusion model yields an oceanic uptake of 2.4 GtC per year on average for the decade 1980 - 1989, and an outcrop-diffusion model (both described by Siegenthaler, 1983) 3.6 GtC per year. The latter model most probably overpredicts the flux into the ocean, because it includes an infinitely fast exchange between high-latitude surface waters and the deep ocean.

However, it is obviously desirable to use 3-dimensional (3-D) general circulation models of the oceans for this purpose. At this time, only a few modeling groups have started to do this. One 3-D model (Maier-Reimer and Hasselmann, 1987) gives a similar CO₂ uptake as a 1-D box-diffusion model of Siegenthaler (1983), as illustrated by the model response to a pulse input of CO₂ (Figure 1.2). In a recent revised version of this model (Maier-Reimer et al., personal communication) the ocean takes up less CO₂, about 1.2 GtC per year on average for the decade 1980 - 1989. The GFDL 3-D ocean model (Sarmiento et al, 1990) has an oceanic uptake of

1.9 GtC per year for the same period. 3-D ocean models and especially coupled atmosphere-ocean models are the only means to study in a realistic way the feedback effects that climate change may have on atmospheric CO₂ via alteration of the ocean circulation (cf. Section 1.2.7.1). However, models need to be constrained by more data than are presently available.

The oceanic uptake of CO₂ for the decade 1980 - 1989, as estimated based on carbon models (e.g., Siegenthaler and Oeschger, 1987; Maier-Reimer et al., personal communication, 1990; Goudriaan, 1989; Sarmiento et al., 1990) is in the range 2.0±0.8 GtC per year.

1.2.4.3 Redistribution of anthropogenic carbon dioxide

During the period 1850 to 1986, 195±20 GtC were released by fossil fuel burning and 117±35 GtC by deforestation and changes in land use, adding up to a cumulative input of 312±40 GtC.

Atmospheric CO₂ increased from about 288 ppmv to 348 ppmv during this period, corresponding to (41±6)% of the cumulative input. This percentage is sometimes called the "airborne fraction", but that term should not be misunderstood: all CO₂, anthropogenic and non-anthropogenic, is continuously being exchanged between atmosphere, ocean and biosphere. Conventionally, an "airborne fraction" referring to the fossil input only has often been quoted, because only the emissions due to fossil fuel burning are known with good precision. However, this may be misleading, since the atmospheric increase is a response to the total emissions. We therefore prefer the definition based on the latter. The airborne fraction for the period 1980 - 1989 (see calculation below) corresponds to (48±8)% of the cumulative input.

In model simulations of the past CO₂ increase, using estimated emissions from fossil fuels and deforestation, it has generally been found that the simulated increase is larger than that actually observed. An estimate for the decade 1980-1989 is:

Emissions from fossil fuels into the atmosphere (Figure 1.5)	5.4±0.5 GtC per yr
Emissions from deforestation and land use	1.6±1.0
Accumulation in the atmosphere	3.4±0.2
Uptake by the ocean	<u>2.0±0.8</u>
Net imbalance	1.6±1.4

The result from this budget and from other studies is that the estimated emissions exceed the sum of atmospheric increase plus model-calculated oceanic uptake by a significant amount. The question therefore arises whether an important mechanism has been overlooked. All attempts to identify such a "missing sink" in the ocean have, however, failed so far. A possible exception is that a natural fluctuation in the oceanic carbon system could have caused a decreasing atmospheric baseline concentration in the past few decades; this does not appear likely in view of the relative constancy of the pre-industrial CO₂ concentration. There are possible processes on land, which could account for the missing CO₂ (but it has not been possible to verify them). They include the stimulation of vegetative growth by increasing CO₂ levels (the CO₂ fertilization effect), the possible enhanced productivity of vegetation under warmer conditions, and the direct effect of fertilization from agricultural fertilizers and from nitrogenous releases into the atmosphere. It has been estimated that increased fertilization by nitrogenous releases could account for a sequestering of up to a maximum of 1 GtC per year in terrestrial ecosystems (Melillo, private communication, 1990). In addition, changed forest management practices may also result in an increase in the amount of carbon stored in northern mid-latitude forests. The extent to which mid-latitude terrestrial systems can sequester carbon before becoming saturated and ineffective is unknown. As mid-latitude terrestrial systems become close to becoming saturated, and hence ineffective in sequestering carbon, this would allow more of the CO₂ to remain in the atmosphere.

A technique for establishing the global distribution of surface sources and sinks has been to take global observations of atmospheric CO₂ concentration and isotopic composition and to invert these by means of atmospheric transport models to deduce spatial and temporal patterns of surface fluxes (Pearman et al., 1983; Pearman and Hyson, 1986; Keeling and Heimann,

1986). The observed interhemispheric CO₂ concentration difference (currently about 3 ppmv) is smaller than one would expect given that nearly all fossil releases occur in the northern hemisphere. The results of this approach suggest that there is an unexpectedly large sink in the northern hemisphere, equivalent to more than half of the fossil fuel CO₂ release (Enting and Mansbridge, 1989; Tans et al., 1990; Keeling et al., 1989b). Furthermore, it has been concluded that the oceanic uptake compatible with oceanic and atmospheric CO₂ data and a 3-dimensional atmospheric transport model is at most 1 GtC per year (Tans et al., 1990). Thus, a significant terrestrial sink, possibly larger than the oceanic uptake, is suggested by these model analyses.

1.2.4.4 Seasonal variations

Atmospheric CO₂ exhibits a seasonal cycle, dominated by the seasonal uptake and release of atmospheric CO₂ by land plants. Its amplitude is small (1.2 ppmv peak-to-peak) in the southern hemisphere and increases northward to a maximum of order 15 ppmv peak-to-peak in the boreal forest zone (55-65° N).

The amplitude of the seasonal cycle has been observed to be increasing (e.g., Pearman and Hyson, 1981; Bacastow et al., 1985; Thompson et al., 1986). For example, at Mauna Loa, Hawaii, the seasonal amplitude has increased by nearly 20% since 1958. The increase has however, not been monotonic, and different evaluation methods yield somewhat different values; still, it is statistically significant. This increasing amplitude could point to a growing productivity (NPP) of the terrestrial ecosystems, and to a sequestering of carbon by a growing biomass, provided the increase in biomass is not fully compensated by respiration. It is important to note that such a change does not necessarily indicate increased productivity or increased storage of carbon (Pearman and Hyson, 1981; Kohlmaier et al., 1989; Houghton, 1987); it could also be due to, e.g., accelerated soil respiration in winter.

1.2.4.5 Interannual variations

Small imbalances in natural exchange fluxes are reflected in interannual CO₂ concentration fluctuations (± 1 ppmv over 1-2 years). They are correlated with the El Niño-Southern Oscillation (ENSO) phenomenon (Thompson et al., 1986; Keeling et al., 1989a), which suggests a relation to changes in the equatorial Pacific Ocean, where normally the upwelling causes a high pCO₂ peak and outgassing of CO₂ into the atmosphere. However, a closer inspection shows that this cannot be the dominating mechanism, since during El Niño, the equatorial pCO₂ peak disappears (Feely et al., 1987), while atmospheric CO₂ grows more strongly than normally. Alternatively, processes in the land biosphere, perhaps in response to climatic events connected with ENSO events, may be responsible. This explanation is supported by one set of stable carbon isotope data on atmospheric CO₂ (Keeling et al., 1989a); but not supported by a second set (Goodman and Francey, 1988).

1.2.4.6 Temporal variations of carbon isotopes

The release of CO₂ from biospheric carbon and fossil fuels, both having lower ¹³C/¹²C ratios than atmospheric CO₂, has led to a decrease of the isotope ratio ¹³C/¹²C in the atmosphere by about 1‰. The man-made emissions of ¹⁴C-free fossil fuel CO₂ have likewise caused a decrease of the atmospheric ¹⁴C concentration (measured on tree-rings) of the order of 2% from 1800 to 1950. Both isotopic perturbations can be used to constrain the history of the anthropogenic release of CO₂. The observed decreases of ¹³C, as observed in air trapped in ice cores (Friedli et al., 1986) and ¹⁴C, observed in tree rings, agree, within experimental uncertainty, with those expected from model calculations with the same carbon cycle models as used for studying the CO₂ increase (Stuiver and Quay, 1981; Siegenthaler and Oeschger, 1987). The interpretation of ¹³C trends in tree rings has proven to be difficult because of plant physiological effects on isotope fractionation (Francey and Farquhar, 1982).

1.2.5 Evidence that the Contemporary Carbon Dioxide Increase is Anthropogenic

How do we know that in fact human activity has been responsible for the well documented 25% increase in atmospheric CO₂ since the early 19th century? Couldn't this rise instead be the result of some long-term natural fluctuation in the natural carbon cycle? Simple arguments allow us to dismiss this possibility.

First, the observational CO₂ records from ice cores with good time resolution clearly show that the maximum range of natural variability about the mean of 280 ppmv during the past 1000 years was small (10 ppmv over a 100 year timescale), that is an order of magnitude less than the observed rise over the last 150 years. A value as high as the current level of 353 ppmv is not observed anywhere in the measured ice core record for the atmospheric history during the past 160,000 years; the maximum value is 300 ppmv during the previous interglacial, 120,000 years ago.

Second, the observed rate of CO₂ increase closely parallels the accumulated emission trends from fossil fuel combustion and from land use changes (c.f. Section 1.2.2). Since the start of atmospheric monitoring in 1958, the annual atmospheric increase has been smaller each year than the fossil CO₂ input. Thus, oceans and biota together must have been a global sink rather than a source during all these years. Further evidence is provided by the fact that the north-to-south CO₂ concentration difference has been observed to increase from 1 ppmv in 1960 to 3 ppmv in 1985, parallel to the growth of the (northern hemisphere) fossil fuel combustion sources (Keeling et al., 1989a).

Third, the observed isotopic trends of ¹³C and ¹⁴C agree qualitatively with those expected due to the CO₂ emissions from fossil fuels and the biosphere, and they are quantitatively consistent with results from carbon cycle modelling.

1.2.6 Sensitivity Analyses for Future Carbon Dioxide Concentrations

Future atmospheric CO₂ concentrations depend primarily on emission rates from energy use and deforestation, and on the effectiveness of the ocean and land biota as CO₂ sinks. For the sake of illustration, several schematic scenarios are shown in Figures 1.7 and 1.8. Those of Figure 1.7 are based on prescribed total CO₂ emission rates after 1990, for those in Figure 1.8 atmospheric concentrations after 1990 were prescribed and the corresponding emission rates were calculated to fit these concentrations. A box-diffusion model of the global cycle was used for these simulations (Enting and Pearman, 1982, 1987), with an oceanic eddy diffusivity of 5350 m²year⁻¹ and an air-sea gas exchange rate corresponding to an exchange coefficient of 0.12 year⁻¹. The calculations assume no biospheric-climate feedbacks, and also assume that after 1990 the net biospheric input of CO₂ is zero, i.e., the input of CO₂ from tropical deforestation is balanced by uptake of CO₂ by terrestrial ecosystems.

In case a (all emissions stopped; Figure 1.7), the atmospheric concentration declines, but only slowly (from 351 ppmv in 1990 to 331 ppmv in 2050 and 324 ppmv in 2100), because the penetration of man-made CO₂ to deeper ocean layers takes a long time. Even if the emissions were reduced by 2% per year from 1990 on (case b), atmospheric CO₂ would continue to increase for several decades. Case c (constant emission rate after 1990) gives CO₂ levels of about 450 ppmv in 2050 and 520 ppmv in 2100. A constant relative growth rate of 2% per year (case d) would yield 575 ppmv in 2050 and 1330 ppmv in 2100. Comparison of cases b, c and d clearly shows that measures to reduce emissions will result in slowing down the rate of atmospheric CO₂ growth.

Cases b' and c', in comparison to b and c, schematically illustrate the effect of reducing emissions in 2010 instead of in 1990.

If an (arbitrary) threshold of 420 ppmv, i.e., 50% above pre-industrial, is not to be exceeded (case e', Figure 1.8), then CO₂ production rates should slowly decline, reaching about 50% of their present value by 2050 and 30% by 2100. In order to keep the concentration at the present

level (case f), emissions would have to be reduced drastically to 30% of present immediately and to less than 20% by 2050.

The results of scenario calculations with a 3-D ocean-atmosphere model (Maier-Reimer and Hasselmann, 1987; Maier-Reimer et al., personal communication, 1990--revised model) give higher concentrations than those obtained with a box-diffusion model shown in Figure 1.7. For instance, about 480 ppmv in the year 2050 and about 560 ppmv in the year 2100 for scenario c, compared to about 450 ppmv and 520 ppmv in Figure 1.7. On the other hand, calculations with a box model that includes a biospheric CO₂ sink (Goudriaan, 1989) yields somewhat lower concentrations than shown in Figure 1.7, for instance about 415 ppmv in the year 2050 and 460 ppmv in the year 2100 for scenario c.

1.2.7 Feedbacks from Climate Change into the Carbon Dioxide Cycle

As increasing greenhouse gas concentrations alter the Earth's climate, changing climate and environmental conditions in their turn act back on the carbon cycle and atmospheric CO₂. The climate change Earth has experienced in the recent past is still within the range of natural short-term variability, and so are probably, therefore, the feedback effects of anthropogenic climate change. However, as the changes in the climate become larger than natural climatic variation the magnitude of the feedback effects should begin to have a significant effect. These feedbacks could in general be either positive (amplifying the initial changes) or negative (attenuating them).

1.2.7.1 Oceanic feedback effects

The following are possible feedback effects on the ocean-atmosphere carbon system:

1.2.7.1.1 Ocean temperature: Ocean temperature changes can affect sea water CO₂ chemistry. Surface-water pCO₂ will increase with increasing temperature, tending to decrease the net uptake by the oceans. The future atmospheric CO₂ increase may be amplified by something like 5 percent due to this effect (Lashof, 1989).

1.2.7.1.2 Ocean circulation: The ocean circulation may change in response to climatic change. As a consequence of increasing surface water temperatures, the thermocline may become more resistant to vertical mixing and slow down the uptake of anthropogenic CO₂. Modified wind stress may affect ocean circulation. However, the overall change in ocean dynamics, and consequently in CO₂ uptake, due to a climatic change cannot be estimated from simple considerations; a proper evaluation of such an effect can only be done using dynamical ocean models. Studies on Greenland ice cores indicates that during the last glaciation, significant CO₂ concentration shifts may have occurred within less than 100 years (c.f. Section 1.2.3), probably caused by strong changes of large-scale ocean circulation. Therefore, the possibility that, due to climatic changes, unexpected abrupt events may take place in the natural carbon system cannot be excluded.

1.2.7.1.3 Gas exchange rates: A change in the global wind pattern could influence the gas transfer from the atmosphere to the sea surface. Carbon cycle models show that the net CO₂ uptake by the global ocean is not sensitive to the gas transfer coefficients (because it is controlled mainly by vertical mixing, not by gas exchange; Oeschger et al., 1975; Broecker et al., 1980; Sarmiento et al., 1990), so this effect would probably be of minor influence.

1.2.7.1.4 Modification of oceanic biogeochemical cycling: The rain of dead organic particles corresponds to a continuous export flux of carbon (and nutrients) out of the ocean surface, which under non-perturbed conditions is balanced by an equal upward transport of dissolved carbon (and dissolved nutrients) by water motion. In polar regions and strong upwelling zones, where productivity is not limited by nitrogen or phosphorus, the balance could become disturbed consequent on variations in ocean dynamics (c.f. Section 1.2.7.1.2), so as to influence atmospheric CO₂. As a result of climate change, the distribution of marine ecosystems and species composition could change, which could affect pCO₂ in surface waters. It is not possible at present to predict the direction and magnitude of such effects.

Warming of the oceans might lead to accelerated decomposition of dissolved organic carbon, converting it into CO₂, and thus amplify the atmospheric increase (Brewer, personal communication, 1990).

1.2.7.1.5 UV-B radiation: A reduction in stratospheric O₃ would increase the intensity of UV-B radiation at the Earth's surface. This might have negative effects on the marine biota due to a decrease of marine productivity and thus on the biological carbon pump. This could lead to an increase in the concentration of CO₂ in surface waters and consequently in the atmosphere.

1.2.7.2 Terrestrial biospheric feedbacks

The following are probable feedback effects on the terrestrial biosphere-atmospheric carbon system:

1.2.7.2.1 Carbon dioxide fertilization: Short-term experiments under controlled conditions with crops and other annuals, as well as with few perennials, show an increase in the rates of photosynthesis and growth in most plants under elevated levels of CO₂ (Strain and Cure, 1985). If elevated levels of CO₂ increase the productivity of natural ecosystems, more carbon may be stored in woody tissue or soil organic matter. Such a storage of carbon will withdraw carbon from the atmosphere and serve as a negative feedback on the CO₂ increase. Of particular importance is the response of forests (Luxmoore et al., 1986), given that forests conduct about 2/3 of global photosynthesis (50% of this cycles annually through leaves, while 50% is stored in woody tissue). However, it is not clear whether the increases in photosynthesis and growth will persist for more than a few growing seasons, whether they will occur at all in natural ecosystems and to what degree they will result in an increased storage of carbon in terrestrial ecosystems.

1.2.7.2.2 Eutrophication and toxification: The increased availability of nutrients such as nitrate and phosphate from agricultural fertilizers and from combustion of fossil fuel may stimulate the growth of plants. It has been estimated that the effect of eutrophication, both on land and in the oceans, could be as large as 1 GtC per year (Melillo, private communication, 1990). However, it should be noted that the greater availability of nutrients has often been associated with increasing levels of acid precipitation and air pollution, which have been associated with a reduction in the growth of terrestrial biota.

1.2.7.2.3 Temperature: Under non-tropical conditions, photosynthesis and respiration by plants and by microbes both tend to increase with increasing temperature; but respiration is the more sensitive process, so that a warming of global air temperature is likely to result in an initially increased release of carbon to the atmosphere. Estimates indicate that the additional flux might be significant, perhaps as large as one or a few GtC per year (Woodwell, 1983; Kohlmaier, 1988; Lashof, 1989; Houghton and Woodwell, 1989). This temperature-enhanced respiration would be a positive feedback on global warming.

1.2.7.2.4 Water: Changes in soil water may affect carbon fixation and storage. Increased moisture can be expected to stimulate plant growth in dry ecosystems and to increase the storage of carbon in tundra peat. There is a possibility that stresses brought about by climatic change may be alleviated by increased levels of atmospheric CO₂. At present however, it is not possible to predict reliably either the geographical distribution of changes in soil water or the net effect of these changes on carbon fluxes and storage in different ecosystems. Changes in climate are generally believed to be more important than changes in the atmospheric concentration of CO₂ in affecting ecosystem processes (c.f. Section 10.)

1.2.7.2.5 Change in geographical distribution of vegetation types: In response to environmental change, structure and location of vegetation types may change. If the rate of change is slow, plant distributions may adjust. If, however, the rate of change is fast, large areas of forests might not be able to adapt rapidly enough, hence be negatively affected with a subsequent release of CO₂ to the atmosphere.

1.2.7.2.6 *UV-B radiation*: A reduction in stratospheric O₃ would increase the intensity of UV-B radiation at the Earth's surface. Increased UV-B may have a detrimental effect on many land biota, including crops (Teramura, 1983), thus affecting the strength of the biospheric sink of CO₂ over land.

1.2.8 Conclusions

The atmospheric CO₂ concentration is now about 353ppmv, 25% higher than the pre-industrial (1750-1800) value and higher than at any time in at least the last 160,000 years. This rise, currently amounting to about 1.8 ppmv per year, is beyond any doubt due to human activities. Anthropogenic emissions of CO₂ were 5.7 ± 0.5 GtC due to fossil fuel burning in 1987, plus 0.6 to 2.5 GtC due to deforestation (estimate for 1980). During the last decade (1980 - 1989) about 48% of the anthropogenic emissions have stayed in the atmosphere, the remainder has been taken up by the oceans and possibly by land ecosystems. Our qualitative knowledge of the global carbon cycle is, in view of the complexity of this cycle, relatively good. However, the current quantitative estimates of sources and of sinks of CO₂ do not balance; the atmospheric increase is less rapid than expected from carbon cycle models (in which CO₂ fertilization or environmental responses of the biosphere are not included). This, and model analyses of the interhemispheric CO₂ gradient, indicate that the northern hemisphere terrestrial ecosystems may act as a significant sink of carbon. Such a sink has, however, not been directly identified. To summarize: the total annual input of anthropogenic CO₂ is currently (1980-1989) about 7.0 ± 1.1 GtC, assuming a central value for the input of CO₂ from tropical deforestation; the annual uptake by the oceans is estimated (based on the box models, GCMs and Tans et al, 1990) to be about 2.0 ± 1.0 GtC; and the annual atmospheric accumulation is about 3.4 ± 0.2 GtC. Thus, the annual sequestering by the terrestrial biosphere should be about 1.6 ± 1.5 GtC. While several mechanisms have been suggested that could sequester carbon in terrestrial ecosystems, it is difficult to account for the total required sink. Therefore, it appears likely that, (i) the uptake of CO₂ by the oceans is underestimated, (ii) there are important unidentified processes in terrestrial ecosystems that can sequester CO₂, and/or (iii) the amount of CO₂ released from tropical deforestation is at the low end of current estimates.

If the land biota presently act as a sink of carbon due to a fertilization effect, then they might get saturated with respect to this fertilization some time in the future. This means that we cannot assume, that the terrestrial sink, which may be active currently, will continue to exist unchanged through the next century.

In order to avoid a continued rapid growth of CO₂ in the atmosphere, severe reductions on emissions will be necessary. The time taken for atmospheric CO₂ to adjust to changes in sources or sinks is of the order of 50-200 years, determined mainly by the slow exchange of carbon between surface waters and deeper layers of the ocean. Even if all anthropogenic emissions of CO₂ were halted, the atmospheric concentration would decline only slowly, and it would not approach its pre-industrial level for many hundreds of years. Thus, any reductions in emissions will only become fully effective after a time of the order of a century or more. Based on some model estimates, which neglect the feedbacks discussed earlier, the atmospheric concentration in the year 2050 would be between 530 - 600 ppmv for a constant relative growth of the annual anthropogenic emissions by 2% per year, and between 415 - 480 ppmv (increasing to 460 - 560 ppmv by the year 2100) for a constant anthropogenic emission rate at the 1990 level. In order not to exceed 420 ppmv (50% above pre-industrial), annual anthropogenic emissions would have to be reduced continuously to about 50% of their present value by the year 2050. In order to stabilize concentrations at present day concentrations (353 ppmv), an immediate reduction in global anthropogenic emissions by 60-80 percent would be necessary. The size of the estimated reduction depends on the carbon cycle model used.

During the millennium preceding the anthropogenic CO₂ growth, the concentration was relatively constant near 280 ppmv, with a variability of less than ± 10 ppmv. This indicates that the sensitivity of atmospheric CO₂ levels to minor climatic changes such as the Little Ice Age, where global mean temperatures probably decreased by about 1°C, is within this range.

However, the anticipated climatic and environmental changes may soon become large enough to act back on the oceanic and terrestrial carbon cycle in a more substantial way. A close interaction between climate variations and the carbon cycle is evidenced by the glacial-interglacial CO₂ variations. The ice-core record shows that CO₂ concentrations during the coldest part of the last glaciation were about 30% lower than during the past 10,000 years. The glacial-interglacial CO₂ variations were probably due to changes in ocean circulation and marine biological activity, and were correlated to variations in global climate. There is some (not fully clear) evidence from ice cores that rapid changes of CO₂, ca. 50 ppmv within about a century, occurred during and at the end of the ice age.

If global temperatures increase, this could change the natural fluxes of carbon, thus having feedback effects on atmospheric CO₂. Some of the identified feedbacks are potentially large and could significantly influence future CO₂ levels. They are difficult to quantify, but it seems likely that there would be a net positive feedback, i.e., they will enhance the man-made increase. On the longer term, the possibility of unexpected large changes in the mechanisms of the carbon cycle due to a human-induced change in climate cannot be excluded.

1.3 METHANE

Methane is a chemically and radiatively active trace gas that is produced from a wide variety of anaerobic (i.e., oxygen deficient) processes and is primarily removed by reaction with hydroxyl radicals (OH) in the troposphere. Oxidation of CH₄ by OH in the stratosphere is a significant source of stratospheric water (H₂O) where it is an important greenhouse gas.

1.3.1 Atmospheric Distribution of Methane

1.3.1.1 Paleo atmospheric record of methane

There are good data on the atmospheric concentration of CH₄ (Figure 9) from Antarctic and Greenland ice cores for the period between 10,000 and 160,000 years ago (Raynaud et al. 1988; Stauffer et al. 1988; Craig and Chou, 1982; Chappellaz et al. 1989). The minimum concentration during the last glacial periods (about 20,000 and 150,000 years ago) was around 0.35 ppmv, and rose rapidly, in phase with the observed temperature increases, to about 0.65 ppmv during the glacial-interglacial transitions (about 15,000 and 130,000 years ago). The atmospheric concentrations of CH₄ decreased rapidly, prior to, and during the last deglaciation period about 10,000 - 11,000 years ago (the Younger Dryas period when there were abrupt temperature decreases in Greenland and northern Europe), and increased rapidly thereafter.

Because of the brittle nature of the ice cores, data on the atmospheric concentrations of CH₄ are reliable only during the last 2,000 years of the Holocene period (last 10,000 years)

1.3.1.2 Contemporary record of methane

Ice core data (Figure 1.10) indicate that the atmospheric concentrations of CH₄ averaged around 0.8 ppmv between two hundred and two thousand years ago, increasing to 0.9 ppmv one hundred years ago (Craig and Chou, 1982; Rasmussen and Khalil, 1984; Stauffer et al. 1985; Pearman and Fraser, 1988; Pearman et al., 1986; Etheridge et al., 1988). Since then, the atmospheric concentration of CH₄ has increased smoothly to present levels, highly correlated with global human population. Analysis of infrared solar spectra has shown that the atmospheric concentration of CH₄ has increased by about 30% over the last 40 years (Rinsland et al. 1985; Zander et al. 1990).

Atmospheric concentrations of CH₄ have been measured directly since 1978 when the globally averaged value was 1.51 ppmv (e.g., Rasmussen and Khalil, 1981; Blake and Rowland, 1988). Currently the value is 1.72 ppmv, corresponding to an atmospheric reservoir of about 4900 Tg (1 Tg = 10¹² g) and it is increasing at a rate of 14 to 17 ppbv per year (40 to 48 Tg per year), i.e., 0.8 to 1.0% per year (Blake and Rowland, 1988; Steele et al. 1987). The atmospheric concentration of CH₄ in the Northern Hemisphere is 1.76 ppmv, compared to 1.68 ppmv in the Southern Hemisphere (Figure 1.11). The magnitude of the seasonal

variability varies with latitude (Steele et al. 1987; Fraser et al. 1984), being controlled by the temporal variability in source strengths and atmospheric concentration of OH radicals,

1.3.1.3 Isotopic composition of methane

Methane is produced from different sources with distinctive proportions of carbon ^{12}C , ^{13}C , and ^{14}C , and hydrogen isotopes H, D (^2H), and T (^3H). Similarly, the rates of processes that destroy CH_4 depend upon its isotopic composition. Consequently, the CH_4 budget can be constrained by knowledge of the isotopic composition of atmospheric CH_4 , the extent of isotopic fractionation during removal, and the isotopic signatures of CH_4 from different sources. Recent work to elucidate the sources of CH_4 has proceeded through an analysis of carbon isotopic signatures (Cicerone and Oremland, 1988; Wahlen et al., 1989; Lowe et al., 1988; and references therein). One example of this is an analysis of ^{14}C data which suggests that about 100 Tg CH_4 per year may arise from fossil sources (Cicerone and Oremland, 1988; Wahlen et al., 1989). Such a distinction is possible because CH_4 from fossil sources is ^{14}C -free, while that from other sources has essentially the ^{14}C concentration of modern carbon.

1.3.2 Sinks of Methane

The major sink for atmospheric CH_4 is reaction with OH in the troposphere, the OH concentration being controlled by a complex set of reactions involving CH_4 , CO, NMHC, NO_x , and tropospheric O_3 (discussed in Section 1.7; Sze, 1977; Crutzen, 1987). Based on the reaction rate coefficient between CH_4 and OH, and the estimated tropospheric distribution of OH, an atmospheric life-time for CH_4 of between 8 and 11.8 years has been estimated (Prinn et al. 1987). This estimate is supported by the fact that models of global OH are tested by analyses of the budgets for CH_3CCl_3 (Logan et al., 1981; Fraser et al., 1986a; Prinn et al., 1987) and ^{14}CO (Appendix to WMO, 1989b). The reaction between CH_4 and OH currently represents a sink of 400 to 600 Tg of CH_4 per year. The efficiency of this sink may, however, have decreased during the last century because the atmospheric concentration of OH in the troposphere may have decreased, hence the lifetime of CH_4 would have increased, in response to increasing concentrations of CO, NMHC, and CH_4 (Sze, 1977).

Soils may represent a removal mechanism for CH_4 . The magnitude of this sink has been estimated (this assessment) to be 30 ± 15 Tg CH_4 per year from the work of Harris et al., 1982 and Seiler and Conrad, 1987.

1.3.3 Sources of Methane

Methane is produced from a wide variety of anaerobic sources (Cicerone and Oremland, 1988). Two main pathways for CH_4 production have been identified: (i) reduction of CO_2 with hydrogen, fatty acids or alcohols as hydrogen donors, or (ii) transmethylolation of acetic acid or methyl alcohol by CH_4 -producing bacteria. Table 2 summarizes identified sources of CH_4 with ranges of likely annual emissions. The total annual CH_4 source must equal the atmospheric sink of about 500 (400 to 600) Tg CH_4 per year, the possible soil sink of about 30 (15 to 45) Tg CH_4 per year, and the annual growth of 40 to 48 Tg CH_4 in the atmosphere. The sum of the present best estimates of the sizes of the individual sources identified in Table 2 equal 525 Tg CH_4 per year. It should be noted that the newest data for rice paddies, biomass burning, and coal mining sources suggest that the values may be even less than those of Table 2, possibly indicating a missing source of CH_4 , or an overestimate of the sink for CH_4 .

Table 2		
ESTIMATED SOURCES AND SINKS OF METHANE		
	Annual Release (Tg CH ₄)	Range (Tg CH ₄)
Source		
Natural Wetlands (bogs, swamps, tundra, etc)	115	100 - 200
Rice Paddies	110	25 - 170
Enteric Fermentation (animals)	80	65 - 100
Gas Drilling, venting, transmission	45	25 - 50
Biomass Burning	40	20 - 80
Termites	40	10 - 100
Landfills	40	20 - 70
Coal Mining	35	17 - 50
Oceans	10	5 - 20
Freshwaters	5	1 - 25
CH ₄ Hydrate Destabilization	5	0 - 100
Sink		
Removal by soils	30	15-45
Reaction with OH in the atmosphere	500	400 - 600
Atmospheric Increase	44	40 - 48

1.3.3.1 Natural wetlands

Significant progress has been made in quantifying the magnitude of the source of CH₄ from natural wetlands (Svensson and Rosswall, 1984; Sebacher et al. 1986; Whalen and Reeburgh, 1988; Moore and Knowles, 1987; Mathews and Fung, 1987; Harris et al. 1985; Crill et al. 1988; Andronova, 1989; Harris and Sebacher, 1981; Burke et al, 1988; Harris et al, 1988; Aselmann and Crutzen, 1989). Recent data support earlier estimates of a global flux of 110 - 115 Tg CH₄ per year, but reverses the relative importance of tropical and high latitude systems. The data base, which is still quite limited (no data from Asia), suggests 55 Tg CH₄ per year (previously 32 Tg CH₄ per year) from tropical wetlands, and 39 Tg CH₄ per year (previously 63 Tg CH₄ per year) from high latitude wetlands. Since CH₄ is produced through biological processes under anaerobic conditions, any factors affecting the physical, chemical or biological characteristics of soils could affect CH₄ emission rates.

1.3.3.2 Rice paddies

Rice paddies are an important source of CH₄ with estimates of the globally averaged flux ranging from 25 - 170 Tg CH₄ per year (Neue and Scharpenseel, 1988; Yagi and Minami, 1989; Holzapfel-Pschorn and Seiler, 1986; Cicerone and Shetter, 1981; Cicerone et al. 1983). The flux of CH₄ from rice paddies is critically dependent upon several factors including: (i) agricultural practices (e.g., fertilization, water management, density of rice plants, double cropping systems, application of manure or rice straw), (ii) soil / paddy characteristics (soil type, acidity, redox potential, temperature, nutrient availability, substrate, profile of anaerobic environment), and (iii) time of season. One difficulty in obtaining accurate estimates is that almost 90% of the world's harvested area of rice paddies is in Asia, and of this about 60% are in China and India from which no detailed data is available. The annual production of rice since 1940 has approximately doubled as a result of double cropping practices and an increased area of cultivation. It is likely that CH₄ emissions have increased proportionally as well.

1.3.3.3 Biomass burning

Biomass burning in tropical and sub-tropical regions is thought to be a significant source of atmospheric CH₄, with estimates of global emission rates ranging from 20 to 80 Tg CH₄ per year (Andreae et al. 1988; Bingemer and Crutzen, 1987; Crutzen et al. 1979; Crutzen et al. 1985; Crutzen, 1989; Greenberg et al. 1984; Stevens et al. 1990; Quay et al. 1990). Improved estimates require an enhanced understanding of: (i) CH₄ emission factors, (ii) the amount, by type, of vegetation burnt each year on an area basis, and (iii) type of burning (smouldering vs flaming). Current estimates indicate that over the last century the rate of forest clearing by burning has increased (cf. Section 1.2.2.2).

1.3.3.4 Enteric fermentation (animals)

Methane emissions from enteric fermentation in ruminant animals, including all cattle, sheep and wild animals, is estimated to provide an atmospheric source of 65 - 100 Tg CH₄ per year (Crutzen et al. 1986; Lerner et al., 1988). Methane emissions depend upon animal populations, as well as the amount and type of food. It is difficult to accurately estimate the change in this source over the last century because the significant increase in the number of cattle and sheep has been partially offset by decreases in the populations of elephants and North American bison. One estimate suggests that the magnitude of this source has increased from 21 Tg CH₄ per year in 1890 to 78 Tg CH₄ per year in 1983 (Crutzen et al. 1986).

1.3.3.5 Termites

There is a large range in the magnitude of the estimated fluxes of CH₄ from termites; 10 - 100 Tg CH₄ per year (Cicerone and Oremland, 1988; Zimmerman et al. 1982; Rasmussen and Khalil, 1983; Seiler et al. 1984; Fraser et al. 1986b). The values are based on the results of laboratory experiments, applied to estimates of global termite populations and the amount of biomass consumed by termites, both of which are uncertain, and field experiments. It is important to determine whether the global termite population is currently increasing, and whether it is likely to respond to changes in climate.

1.3.3.6 Landfills

The anaerobic decay of organic wastes in landfills may be a significant anthropogenic source of atmospheric CH₄, 20 - 70 Tg CH₄ per year. However, several factors need to be studied in order to quantify the magnitude of this source more precisely, including amounts, trends, and types of waste materials, and landfill practices (Bingemer and Crutzen, 1987).

1.3.3.7 Oceans and freshwaters

Oceans and freshwaters are thought to be a minor source of atmospheric CH₄. The estimated flux of CH₄ from the oceans is based on a limited data set, taken in the late 1960's / early 1970's when the atmospheric concentration of CH₄ was about 20% lower. They showed that the open oceans were only slightly supersaturated in CH₄ with respect to its partial pressure in the atmosphere. There are inadequate recent data from either the open oceans or coastal waters to reduce the uncertainty in these estimates (Cicerone and Oremland, 1988).

1.3.3.8 Coal mining

Methane is released to the atmosphere from coal mine ventilation, and degassing from coal during transport to an end-use site. A recent unpublished study estimated the flux of CH₄ from coal mining, on a country basis, for the ten major coal producing countries, and deduced a global minimum emission of 17 Tg CH₄ per year. Global CH₄ fluxes from coal mining have been estimated to range from 10 - 50 Tg CH₄ per year (Cicerone and Oremland, 1988, and recent unpublished studies by others).

1.3.3.9 Gas drilling, venting and transmission

CH₄ is the major component of natural gas, hence leakage from pipelines and venting from oil and gas wells could represent a significant source of atmospheric CH₄ (Cicerone and Oremland, 1988). The global flux from these sources is estimated, based on limited data of questionable reliability, to range from 25 - 50 Tg CH₄ per year.

1.3.4 Feedbacks from Climate Change into the Methane Cycle

Future atmospheric concentrations of CH₄ will depend on changes in the strengths of either the sources or sinks, which are dependent upon social, economic, and political, and also environmental factors and in particular changes in climate. Methane emissions from wetlands are particularly sensitive to temperature and soil moisture, and hence future climatic changes could significantly change the fluxes of CH₄ from both natural wetlands and rice paddies. Tropospheric OH, which provides the atmospheric sink for CH₄, is dependent upon a number of factors, including the intensity of UV-B radiation, and the ambient concentrations of H₂O, CO, CH₄, reactive nitrogen oxides, and tropospheric O₃ (See Section 1.7) (Crutzen, 1987; Isaksen and Hov, 1987; Thompson and Cicerone, 1986).

1.3.4.1 Tropical methane sources

The major sources of CH₄ in tropical regions (natural wetlands and rice paddies) are quite sensitive to variations in soil moisture. Consequently, changes in soil moisture, which would result from changes in temperature and precipitation, could significantly alter the magnitude of these large sources of atmospheric CH₄. Increased soil moisture would result in larger fluxes, whereas a decrease in soil moisture would result in smaller fluxes.

1.3.4.2 High latitude methane sources

Methane fluxes from the relatively flat tundra regions would be sensitive to changes of only a few centimetres in the level of the water table, with flooded soils producing a factor of 100 more CH₄ than dry soils (Whalen and Reeburgh, 1988; Crill et al. 1988). Similarly, emissions of CH₄ are significantly larger at warmer temperatures, due to accelerated microbiological decomposition of organic material in the near-surface soils. Consequently, an increase in soil moisture and temperatures in high latitude wetlands would result in enhanced CH₄ emissions, whereas warmer dryer soils might have decreased CH₄ emissions.

Higher temperatures could also increase the fluxes of CH₄ at high northern latitudes from: (i) CH₄ trapped in permafrost, (ii) decomposable organic matter frozen in the permafrost, and (iii) decomposition of CH₄ hydrates (Cicerone and Oremland, 1988; Kvenvolden, 1988). Quantifying the magnitudes of these positive feedbacks is difficult. Timescales for thawing the permafrost, located between a few centimetres to metres below the surface, could be decades to centuries, while the time for warming the CH₄ hydrates could be even longer, although one study (Kvenvolden, 1988) estimated that the flux of CH₄ from hydrate decomposition could reach 100 Tg CH₄ per year within a century.

1.3.5 Conclusions

Current atmospheric CH₄ concentrations, at 1.72 ppmv, are now more than double the pre-industrial value (1750-1800) of about 0.8 ppmv, and are increasing at a rate of 0.9% per year. The ice core record shows that CH₄ concentrations were about 0.35 ppmv during glacial periods, and increased in phase with temperature during glacial-interglacial transitions. The current atmospheric concentration of CH₄ is greater than at any time during the last 160,000 years.

Reaction with OH in the troposphere, the major sink for CH₄, results in a relatively short atmospheric lifetime of 10±2 years. The short lifetime of CH₄ implies that atmospheric concentrations will respond quite rapidly, in comparison to the longer lived gases such as CO₂, N₂O, and CFCs, to changes in emissions. In order to stabilize concentrations at present day levels, an immediate reduction in global man-made emissions by 15-20 percent would be necessary (this and other scientific sensitivity analyses are discussed in the Annex). Global concentrations of OH are dependent upon the intensity of UV-B radiation, and the concentrations of gases such as H₂O, CO, CH₄, NO_x, NMHC, and O₃, and may have declined during the 20th century due to changes in the atmospheric concentrations of these gases.

The individual sources of atmospheric CH₄ have been qualitatively identified, but there are significant uncertainties in the magnitude of their strengths. Human activities such as rice cultivation, rearing of domestic ruminants, biomass burning, coal mining, and natural gas venting have increased the input of CH₄ into the atmosphere, and these combined with an apparent decrease in the concentration of tropospheric OH, yields the observed rise in global CH₄. However, the quantitative importance of each of the factors contributing to the observed increase is not well known at present.

Several potential feedbacks exist between climate change and CH₄ emissions, in both tropical and high latitude wetland sources. In particular, an increase in high latitude temperatures could result in a significant release of CH₄ from the melting of permafrost and decomposition of CH₄ hydrates.

1.4 HALOCARBONS

Chlorine and bromine containing halocarbons have been shown to deplete O₃ in the stratosphere. In addition, it has been recognized that they are important greenhouse gases. Their sources, sinks, atmospheric distributions, and role in perturbing stratospheric O₃ and the Earth's radiative balance have been reviewed in detail (WMO 1985, 1989a, 1989b). Many governments, recognizing the harmful effects of halocarbons on the environment, signed the "Montreal Protocol on Substances that Deplete the Ozone Layer" (UNEP, 1987) in 1987 to limit the production and consumption of a number of fully halogenated CFCs and halons. The control measures of the Montreal Protocol freeze the production and consumption of CFCs 11, 12, 113, 114, and 115 in developed countries at their 1986 levels from the year 1990, a reduction to 80% of their 1986 levels from the year 1993, with a further reduction to 50% of their 1986 levels from the year 1998. Developing countries, with a per capita use of CFCs of less than 0.3 kg per capita, are allowed to increase their per capita use up to this limit and can delay compliance with the control measures by 10 years. All major producing and consuming developed countries, and many developing countries, have signed and ratified the Montreal Protocol.

1.4.1 Atmospheric Distribution of Halocarbons

The mean atmospheric concentrations of the most abundant radiatively active halocarbons are shown in Table 3. The atmospheric concentrations of the halocarbons are currently increasing more rapidly on a global scale (on a percentage basis) than the other greenhouse gases (Figure 1.12). The concentrations of the fully halogenated chlorofluorocarbons (CFCs), slightly greater in the northern hemisphere than in the southern hemisphere, are consistent with the geographical distribution of releases (>90% from the industrialized nations), a 45 °N - 45 °S mixing time of about 1 year, and their very long atmospheric lifetimes.

1.4.2 Sinks for Halocarbons

There is no significant tropospheric removal mechanism for the fully halogenated halocarbons such as CCl₃F (CFC-11), CCl₂F₂ (CFC-12), C₂Cl₃F₃ (CFC-113), C₂Cl₂F₄ (CFC-114), C₂ClF₅ (CFC-115), carbon tetrachloride (CCl₄), and halon 1301 (CBrF₃). They have long atmospheric lifetimes, decades to centuries, and are primarily removed by photodissociation in the mid - upper stratosphere. There is currently a significant imbalance between the sources and sinks giving rise to a rapid growth in atmospheric concentrations. To stabilize the atmospheric concentrations of CFCs 11, 12 and 113 at current levels would require reductions in emissions of approximately 70-75%, 75-85%, and 85-95%, respectively (see Annex).

Non-fully halogenated halocarbons containing a hydrogen atom such as methyl chloride (CH₃Cl), methylchloroform (CH₃CCl₃), CHClF₂ (HCFC-22), and a number of other HCFCs and HFCs being considered as substitutes for the current CFCs (c.f. Section 1.4.4) are primarily removed in the troposphere by reaction with OH. These hydrogen containing species

1 have atmospheric lifetimes ranging from about one to forty years, much shorter on average
 2 than the fully halogenated CFCs. To stabilize the atmospheric concentrations of HCFC-22 at
 3 current levels would require reductions in emissions of approximately 40-50%.
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 5

TABLE 3					
Halocarbon Concentrations and Trends (1990) †					
Halocarbon		Mixing Ratio pptv	Annual Rate of Increase pptv %		Lifetime Years
CCl ₃ F	(CFC-11)	280	9.5	4	65
CCl ₂ F ₂	(CFC-12)	484	16.5	4	130
CClF ₃	(CFC-13)	5			400
C ₂ Cl ₃ F ₃	(CFC-113)	60	4-5	10	90
C ₂ Cl ₂ F ₄	(CFC-114)	15			200
C ₂ ClF ₅	(CFC-115)	5			400
CCl ₄		146	2.0	1.5	50
CHClF ₂	(HCFC-22)	122	7	7	15
CH ₃ Cl		600			1.5
CH ₃ CCl ₃		158	6.0	4	7
CBrClF ₂	(halon 1211)	1.7	0.2	12	25
CBrF ₃	(halon 1301)	2.0	0.3	15	110
CH ₃ Br		10-15			1.5

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 7 † There are a few minor differences between the lifetimes reported in this table and the equivalent table in
 8 WMO 1989b. These differences are well within the uncertainty limits. The 1990 mixing ratios have
 9 been estimated based upon an extrapolation of measurements reported in 1987 or 1988, assuming that
 10 the recent trends remained approximately constant.
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13 1.4.3 Sources of Halocarbons

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 15 Most halocarbons, with the notable exception of CH₃Cl, are exclusively of industrial origin.
 16 Halocarbons are used as aerosol propellants (CFCs 11, 12, and 114), refrigerants (CFCs 12
 17 and 114, and HCFC-22), foam blowing agents (CFCs 11 and 12), solvents (CFC-113,
 18 CH₃CCl₃, and CCl₄), and fire retardants (halons 1211 and 1301). The atmospheric
 19 concentration of methyl chloride is about 0.6 ppbv, and is primarily released from the oceans
 20 and during biomass burning. There is no evidence that the atmospheric concentration of
 21 CH₃Cl is increasing. Methyl bromide (CH₃Br) is produced by oceanic algae, and there is
 22 evidence that its atmospheric concentration has been increasing in recent times due to a
 23 significant anthropogenic source (Penkett et al., 1985; Wofsy et al., 1975).
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26 1.4.4 Future Atmospheric Concentration of Halocarbons

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 28 Future emissions of CFCs 11, 12, 113, 114, and 115 will be governed by the Montreal
 29 Protocol on "Substances that Deplete the Ozone Layer" as discussed in Section 1.4. In
 30 addition, international negotiations are currently in progress that will likely (i) result in a
 31 complete global phase-out of production of these chemicals by the year 2000, and (ii) enact
 32 limitations on the emissions (via production and consumption controls) of CCl₄, and
 33 CH₃CCl₃. However, even with a complete cessation of production of CFCs 11, 12 and 113 in
 34 the year 2000 their atmospheric concentrations will still be significant for at least the next

century because of their long atmospheric lifetimes. It should be noted that emissions of these gases into the atmosphere will continue for a period of time after production has ceased because of their uses as refrigerants, foam blowing agents, fire retardants, etc.

A number of hydrofluorocarbons (HFCs) and hydrochlorofluorocarbons (HCFCs) are being considered as potential replacements for the long-lived CFCs (11, 12, 113, 114, and 115) that are regulated under the terms of the Montreal Protocol. The HFCs and HCFCs primarily being considered include: HCFC-22, HCFC-123 (CHCl_2CF_3), HCFC-124 (CHClF_2), HFC-125 (CHF_2CF_3), HFC-134a (CH_2FCF_3), HCFC-141b ($\text{CH}_3\text{CCl}_2\text{F}$), HCFC-142b (CH_3CClF_2), HFC-143a (CH_3CF_3), and HFC-152a (CH_3CHF_2). The calculated atmospheric lifetimes of these chemicals are controlled primarily by reaction with tropospheric OH and range between about 1 and 40 years. It has been estimated (UNEP, 1989) that a mix of HFCs and HCFCs will replace the CFCs currently in use at a rate of about 0.4 kg of substitute for every kg of CFCs currently produced, with an annual growth rate of about 3%. Because of their shorter lifetimes, and expected rates of substitution and emissions growth rates, the atmospheric concentrations of HFCs and HCFCs will be much lower for the next several decades than if CFCs had continued to be used, even at current rates. However, continued use, accompanied by growth in the emission rates, of HFCs and HCFCs for more than several decades would result in atmospheric concentrations that would be radiatively important.

1.4.5 Conclusions

The atmospheric concentrations of the industrially produced halocarbons, primarily CCl_3F , CCl_2F_2 , $\text{C}_2\text{Cl}_3\text{F}_3$, and CCl_4 are about 280 pptv, 484 pptv, 60 pptv, and 146 pptv, respectively. Over the past few decades their concentrations (except CCl_4) have increased more rapidly (on a percentage basis) than the other greenhouse gases, currently at rates of at least 4% per year. The fully halogenated CFCs and CCl_4 are primarily removed by photolysis in the stratosphere, and have atmospheric lifetimes in excess of 50 years.

Most halocarbons, with the notable exception of methyl chloride, are exclusively anthropogenic and their sources (solvents, refrigerants, foam blowing agents, and aerosol propellants) are well understood.

To stabilize, and then reduce, the current atmospheric concentrations of the fully halogenated CFCs (e.g., 11, 12 and 113) would require approximate reductions in emissions of 70-75%, 75-85%, and 85-95%, respectively. Future emissions of CFCs and CCl_4 will, most likely, be eliminated or be significantly lower than today's because the stringency, scope, and timing of international regulations on chlorine and bromine containing chemicals, i.e., the Montreal Protocol on Substances that Deplete the Ozone Layer, are currently being renegotiated. However, the atmospheric concentrations of CFCs 11, 12 and 113 will still be significant (30 - 40% of current) for at least the next century because of their long atmospheric lifetimes.

1.5 NITROUS OXIDE

Nitrous oxide is a chemically and radiatively active trace gas that is produced from a wide variety of biological sources in soils and water and is primarily removed in the stratosphere by photolysis and reaction with electronically excited oxygen atoms.

1.5.1 Atmospheric Distribution of Nitrous Oxide

The mean atmospheric concentration of N_2O in 1990 is about 310 ppbv, corresponding to a reservoir of about 1500 Tg N, and increasing at a rate of 0.2 - 0.3% per year (Figure 1.13; Weiss, 1981; Prinn et al. 1990; Robinson et al. 1988; Elkins and Rossen, 1989; Rasmussen and Khalil, 1986). This observed rate of increase represents an atmospheric growth rate of about 3 to 4.5 Tg N per year. The atmospheric concentration of N_2O is higher in the northern

hemisphere than in the southern hemisphere by about 1 ppbv. Ice core measurements show that the pre-industrial value of N_2O was relatively stable at about 285 ppbv for most of the past 2000 years, and started to increase around the year 1700 (Figure 1.14; Pearman et al., 1986; Khalil and Rasmussen, 1988a; Etheridge et al. 1988; Zardini et al. 1989). Figure 1.14 shows that the atmospheric concentrations of N_2O may have decreased by a few ppbv during the period of the mini ice-age.

1.5.2 Sinks for Nitrous Oxide

The major atmospheric loss process for N_2O is photochemical decomposition in the stratosphere, and is calculated to be $10 \pm 3 \text{ Tg N per year}$ (Table 4). Nitrous oxide has an atmospheric lifetime of about 150 years. The observed rate of growth represents a 30% imbalance between the sources and sinks (Hao et al. 1987). Tropospheric sinks such as surface loss in aquatic and soil systems are considered to be small (Elkins et al. 1978, Blackmer and Bremner, 1976).

1.5.3 Sources of Nitrous Oxide

1.5.3.1 Oceans

The oceans are a significant, but not dominant source of N_2O (McElroy and Wofsy, 1986). Based on measurements of the concentration gradients between the atmosphere and surface waters (Butler et al. 1989, and NOAA GMCC unpublished data), and on estimates of the gas exchange coefficient, the current estimate of the magnitude of the ocean source ranges from 1.4 - 2.6 Tg N per year, significantly lower than earlier estimates (Elkins et al. 1978; Cohen and Gordon, 1979; Cline et al. 1987). An accurate determination of the global annual ocean flux is difficult because of uncertainties associated with quantifying the gas exchange coefficient and because the partial pressure of N_2O in the surface waters is highly variable, both spatially and temporally. The partial pressure of N_2O in surface waters varies considerably, ranging from being supersaturated by up to 40% in upwelling regions to being undersaturated by a few percent in areas around Antarctica and within gyres. Data suggest that during El Niño events, where upwelling in the Pacific ocean is suppressed, the ocean fluxes of N_2O are significantly lower (Cline et al. 1987; Butler et al. 1989). It is still unclear whether N_2O is primarily produced from nitrification in near surface waters, or denitrification in oxygen deficient deep waters. Based on vertical profile measurements of oceanic N_2O (NOAA GMCC, unpublished) the oceanic reservoir has been estimated to be between 900 and 1100 Tg N, comparable to the atmosphere. Consequently, changes in the exchange fluxes of N_2O between the ocean and the atmosphere could significantly impact its atmospheric concentration.

1.5.3.2 Soils

Denitrification in aerobic soils is thought to be a dominant source of atmospheric N_2O (Keller et al. 1986; Matson and Vitousek, 1987; Matson and Vitousek, 1989; Slemr et al. 1984). Nitrification under anaerobic conditions could, however, produce higher yields of N_2O per unit of transformed nitrogen. Quantification of global N_2O emissions from soils is difficult because of the heterogeneity of terrestrial ecosystems and the variability in environmental conditions that control the fluxes of N_2O .

Estimates of global fluxes of N_2O from tropical forests range from 2.2 - 3.7 Tg N per year. The impact of deforestation on the emissions of N_2O from tropical soils is unclear, with some studies suggesting that the emissions of N_2O from deforested land are enhanced by as much as a factor of three (Luizao et al. 1989), whereas other studies concluded that N_2O fluxes decreased if vegetation did not return (Robertson and Tiedje, 1988).

Quantifying the roles of temperate forest soils and grasslands in the N_2O budget is difficult because of the paucity of data, and conflicting results. Estimates of N_2O fluxes from temperate forest soils range from 0.7 - 1.5 Tg N per year in one study (Schmidt et al. 1988), to almost none in another study (Bowden et al. 1989). One study also reported that deforestation in temperate forests would lead to enhanced emissions of N_2O (Bowden and Bormann, 1986).

Reliable global N₂O fluxes from grasslands are impossible to derive from the fragmented data available. One study (Ryden, 1981) concluded that English grassland soils, with no fertilization, are a sink for N₂O, whereas limited studies of tropical grasslands and pastures suggest that they may be a moderate to significant source of N₂O (Luizao et al. 1989; Robertson and Tiedje, 1988).

1.5.3.3 Combustion

Until recently, the combustion of fossil fuels was thought to be an important source of atmospheric N₂O (Pierotti and Rasmussen, 1976; Weiss and Craig, 1976; Hao et al. 1987). However, a recent study has shown that the earlier results are incorrect because N₂O was being artificially produced in the flasks being used to collect N₂O from combustion sources (Muzio and Kramlich, 1988). The latest estimate of the global flux of N₂O from combustion sources is between 0.1 and 0.3 Tg N per year, compared to earlier values which were as high as 3.2 Tg N per year.

1.5.3.4 Biomass Burning

Biomass burning is now thought to be a minor source of atmospheric N₂O with a global flux of less than 0.2 Tg N per year (Muzio and Kramlich, 1988; Crutzen 1989; Elkins et al. 1990; Winstead et al. 1990; Griffith et al. 1990). This value is 1-2 orders of magnitude less than previous estimates (Crutzen et al. 1979, 1985) which were influenced by artifacts involving N₂O analysis (Crutzen et al., 1985) and N₂O production in sampling flasks (Muzio and Kramlich, 1988).

1.5.3.5 Fertilizer / Ground-Water

Nitrous oxide production from the use of nitrate and ammonium fertilizers is difficult to quantify because the N₂O fluxes are dependent upon numerous factors including type of fertilizer, soil type, soil temperature, weather, and farming practices (e.g., ploughing, sowing, irrigating). Conversion of fertilizer N to N₂O ranges from 0.01 - 2.0% (Conrad et al. 1983; Bremner et al. 1981). This range, coupled with a global fertilizer production of 55 Tg N per year in 1980, results in a total N₂O emission of between 0.01 - 1.1 Tg N per year (Conrad et al. 1983). Leaching of nitrogen fertilizers from soils into groundwater may result in additional fluxes of N₂O up to 1.1 Tg N per year (Conrad et al. 1983; Ronen et al. 1988). Consequently, a range of 0.01 - 2.2 Tg N per year can be derived for the flux of N₂O from fertilizer use.

1.5.4 Conclusions

Nitrous oxide is a greenhouse gas whose atmospheric concentration, at 310 ppbv, is now about 8% greater than in the pre-industrial era, and is increasing at a rate of about 0.2 - 0.3% per year, corresponding to about 3 - 4.5 Tg N per year. This represents an excess of 30% of current global emissions over current sinks. The major sink for N₂O is photolysis in the stratosphere, resulting in a relatively long atmospheric lifetime of about 150 years. The magnitude of the sink for N₂O is relatively well known ($\pm 30\%$). In order to stabilize concentrations at present day levels, an immediate reduction of 70 - 80% of the additional flux of N₂O that has occurred since the pre-industrial era would be necessary.

Quantification of the various natural and anthropogenic sources is uncertain. Since the latest studies indicate that the total combined flux of N₂O from combustion and biomass burning is between 0.1 to 0.5 Tg N per year, in contrast to earlier estimates of about 5 Tg N per year, and production of N₂O from fertilizer (including groundwater) is believed to be less than or equal to 2.2 Tg N per year, it is difficult to account for the annual increase based on known sources. Stimulation of biological production due to agricultural development may account for the missing anthropogenic emissions. Estimates of the removal rate of N₂O by photodissociation in the stratosphere range from 7 - 13 Tg N per year. Therefore, the total source needed to account for the observed annual atmospheric growth is 10 - 17.5 Tg N per year against a flux of N₂O from known sources of 4.4 - 10.5 Tg N per year. These data suggest that there are missing sources of N₂O, or the strengths of some of the identified sources have been underestimated. Despite these uncertainties, it is believed that the observed increase in N₂O concentrations is caused by human activities.

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TABLE 4	
Estimated Sources and Sinks of Nitrous Oxide	
	Range (TgN per year)
Source	
Oceans	1.4 - 2.6
Soils (tropical forests)	2.2 - 3.7
(temperate forests)	0.7 - 1.5
Combustion	0.1 - 0.3
Biomass burning	0.02 - 0.2
Fertilizer (including ground-water)	0.01 - 2.2
TOTAL:	4.4 - 10.5
Sink	
Removal by soils	?
Photolysis in the stratosphere	7 - 13
Atmospheric Increase	3 - 4.5

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1.6 STRATOSPHERIC OZONE

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Stratospheric O₃ is an important constituent of the Earth's atmosphere. It protects the Earth's surface from harmful solar ultraviolet radiation and it plays an important role in controlling the temperature structure of the stratosphere by absorbing both incoming solar ultraviolet radiation and outgoing terrestrial (longwave) radiation. Part of the absorbed outgoing longwave radiation is then re-radiated back to the surface-troposphere system. Reductions in stratospheric O₃ can modify the surface temperature via two competing processes: more solar radiation is transmitted to the surface-troposphere system, thereby contributing to a surface warming; on the other hand, the cooler stratosphere (due to decreased solar and long-wave absorption) emits less to the troposphere which would tend to cool the surface. The solar warming (a function of total column amount of O₃) and longwave cooling (a function of the vertical distribution of O₃) are similar in magnitude. Therefore, the magnitude as well as the sign of the change in surface temperature depends critically on the magnitude of the O₃ change, which in turn is depends strongly on altitude, latitude and season.

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The concentration and distribution of stratospheric O₃ is controlled by dynamical, radiative and photochemical processes. Stratospheric O₃ is photochemically controlled by chemically active species in the (i) oxygen, (ii) hydrogen, (iii) nitrogen, (iv) chlorine, and (v) bromine families. The precursors for the photochemically active species are (i) O₂, (ii) H₂O and CH₄; (iii) N₂O; (iv) CFCs, CCl₄, CH₃CCl₃, CH₃Cl, and (v) halons and CH₃Br, respectively.

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1.6.1 Stratospheric Ozone Trends

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1.6.1.1 Total column ozone trends

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The Antarctic ozone hole, which formed during the mid to late 1970s recurs every springtime. To determine O₃ trends more widely, data from the ground-based Dobson network have been re-evaluated, station by station, and used to determine changes in total column O₃ over the past two decades. Unfortunately, the network and data are adequate for only a limited geographical

region, i.e., 30 - 64°N. They are inadequate to determine total column O₃ changes in the Arctic, tropics, subtropics, or southern hemisphere apart from Antarctica. Satellite data can provide the desired global coverage, but the current record is too short (about one solar cycle, 1978 to present), to differentiate the effects of natural and human-influenced processes on O₃. The re-evaluated data was analysed for the effects of known natural geophysical processes [seasonal variation, the approximately 26-month quasi-biennial-oscillation, and the 11-year solar cycle] and possible human perturbations. After allowing for natural variability, the analyses, using a variety of statistical models and assumptions, showed measurable zonal mean O₃ decreases in the range 3.4% to 5.1% between 30 and 64°N latitude for the winter months (December - March) between 1969 and 1988, with the larger decreases at the higher latitudes (WMO, 1989a,b). No statistically significant zonal trends were found for the summer period (May - August). Lastly, within longitudinal sectors, regional differences in the O₃ trends were indicated, with the largest values over North America and Europe and the smallest over Japan.

1.6.1.2 Changes in the vertical distribution of ozone

Substantial uncertainties remain in defining changes in the vertical distribution of O₃. Analysis of SAGE I and II satellite data, averaged over 20 to 50°N and S latitudes, indicates that near 40 km O₃ decreased by $3 \pm 2\%$ between February 1979 - November 1981 and October 1984 - December 1988 (WMO, 1989b). Because the SAGE record is so short (i.e., less than one solar cycle), no attempt has been made to distinguish between solar-induced and human-influenced contributions to these changes. A thorough analysis of data from 10 ground-based Umkehr stations in the Northern Hemisphere for the period 1977 to 1987 indicates a statistically significant decrease in O₃ between 30 and 43 km. The decrease near 40 km of $4.8 \pm 3.1\%$, after allowing for seasonal and solar-cycle effects and correcting the data for aerosol interferences, is broadly consistent with theoretical predictions. Based on satellite, ground-based, and ozonesonde data, there are indications of a continuing stratospheric O₃ decrease since the late 1970s of a few percent at 25 km and below. Photochemical models, which do not take into account heterogeneous processes) do not predict these changes, but the measurements are qualitatively consistent with those required for compatibility with the total column measurements.

1.6.2 Future Changes

Future changes in stratospheric O₃ are critically dependent upon future emissions of CFCs, other halocarbons, CH₄, N₂O, and CO₂. Assuming that the current regulatory measures agreed under the Montreal Protocol are not strengthened, then the chlorine loading of the atmosphere is predicted to reach about 9 ppbv by the year 2060, about three times today's level, and a bromine loading of about 30 pptv, about twice today's level. Models predict column O₃ reductions of 0 to 4% in the tropics, and from 4 to 12% at high latitudes in late winter. These predictions do not include the effects of heterogeneous processes, which play a critical role in the formation of the Antarctic ozone hole. Consequently, models that include the effects of heterogeneous processes would predict larger O₃ depletions, at least in polar regions. Ozone is predicted to decrease by 25 - 50% at 40 km and result in stratospheric temperature decreases of 10 to 20 K. If, as expected, the Montreal Protocol is modified to eliminate the emissions of CFCs 11, 12, 113, 114, 115, halons 1211 and 1301, and restrict the emissions of CCl₄ and CH₃CCl₃, by the year 2000, then the chlorine loading of the atmosphere by the year 2060 will probably lie between 2.5 and 4 ppbv (depending upon the emissions of CCl₄ and CH₃CCl₃, and HCFCs). Models that do not include heterogeneous processes predict that global O₃ levels would be similar to today. However, if the atmospheric chlorine loading approaches 4 ppbv the implications for polar O₃, and its subsequent impacts on O₃ at mid-latitudes, are unknown.

1.7 TROPOSPHERIC OZONE AND RELATED TRACE GASES (CARBON MONOXIDE, NON-METHANE HYDROCARBONS, AND REACTIVE NITROGEN OXIDES).

1.7.1 Tropospheric Ozone

Tropospheric O₃ is a greenhouse gas, of particular importance in the upper troposphere in the tropics and sub-tropics. Its distribution is controlled by a complex interplay between chemical, radiative, and dynamical processes. Ozone is: (i) transported down into the troposphere from the stratosphere; (ii) destroyed by vegetative surfaces; (iii) produced by the photo-oxidation of CO, CH₄, and NMHC in the presence of reactive nitrogen oxides (NO_x), and (iv) destroyed by uv-photolysis and by reaction with hydrogen oxide radicals (HO₂) (Danielsen, 1968; Mahlman and Moxim, 1978; Galbally and Roy, 1980; Crutzen, 1974; Isaksen et al., 1978). Chemical processes in clouds could have a strong influence on O₃ production and destruction rates (Lelieveld and Crutzen, 1990).

Consequently, while CO, NMHC, and NO_x are not important greenhouse gases in themselves, they are important precursors of tropospheric O₃ and they are therefore treated in some detail in this subsection.

1.7.1.1 Atmospheric distribution

Ozone in the troposphere has a lifetime of at most several weeks, hence its concentration varies with latitude, longitude, altitude and season (Chatfield and Harrison, 1977; Logan, 1985). Near the surface monthly mean concentrations (30 - 50 ppbv) are highest in spring and summer at northern mid-latitudes (Figure 1.15). In the middle troposphere at northern mid-latitudes values are highest also in spring and summer, 60 - 65 ppbv. The summer maximum results from photo-oxidation of O₃ precursors from fossil fuel combustion and industrial activity (Isaksen et al., 1978; Fishman et al., 1985; Logan 1985). Ozone values are highest in winter and spring at other latitudes, in part because the stratospheric source is largest then (Levy et al., 1985). There is 35% more O₃ at 40°N than at 40°S, in the middle troposphere (Logan, 1985).

Concentrations of O₃ tend to be smaller in the tropics than in mid-latitudes, except in the dry season, when emissions of O₃ precursors from biomass burning provide a photochemical source (Delany et al., 1985; Crutzen et al., 1985; Logan and Kirchoff, 1986; Fishman et al., 1990). Ozone values during the southern spring over South America can reach almost as high values as found over the industrialized mid-latitudes in summer. Large regions of the tropical troposphere appear to be influenced by sources of O₃ from biomass burning (Fishman et al., 1990). Remote marine air and continental air during the wet season may provide a photochemical sink for O₃ in the tropics; mean surface concentrations as low as 4 - 12 ppbv have been measured (Liu et al., 1980; Oltsmans and Komhyr, 1986; Kirchoff, 1990).

1.7.1.2 Trends

Most long-term measurements of O₃ have been made at northern mid-latitudes, from surface sites and from balloons. Only sporadic data are available before the 1970s. A comparison of data obtained in Paris from 1876-1910 (Volz and Kley, 1988) with rural data from the present day from Europe and North America (Logan, 1985, 1989) suggests that surface O₃ has increased by a factor of 2-3 on average; the increase is largest in summer, the factor then being 4-6 (Figure 1.15). Ozone values in Europe in the 1970s appear to be about twice those found between 1930 and 1950 (Crutzen, 1988). Data from Europe suggest an increase of 1-2% per year from the mid 1950s to the early 1980s, with increases in winter and summer (Feister and Warmbt, 1987; Bojkov, 1988). Since the mid-1970s O₃ has increased by 0.8% per year at remote sites in Alaska and Hawaii; shown no annual trend at Samoa, but has decreased by 0.5% per year at the South Pole (Oltmans et al., 1988). Decreases of 1.8% per year are found at both Samoa and South Pole in summer. Trend data are lacking for tropical continental sites.

Ozonesonde data for northern mid-latitudes between 1965 and 1986 suggest that O₃ has increased by about 1% per year below 8 km, primarily over N. Europe and Japan (Angell and Korshover, 1983; Logan 1985; Tiao et al., 1986; WMO, 1989a,b), but there are no clear trends in the upper troposphere. By contrast O₃ has decreased in the lower stratosphere (below 25 km), the crossover in the trend being near the tropopause. There is no trend in O₃ at the single sonde station at southern mid-latitudes, and long term sonde data are lacking in the tropics.

1.7.1.3 Relationships between ozone and its precursors

The concentration of tropospheric O_3 is dependent in a very non-linear manner on the atmospheric concentrations of its precursor gases, i.e., CO , CH_4 , $NMHC$, and, in particular NO_x ($NO_x = NO + NO_2$). Nitrogen oxide concentrations and trends control changes in the concentration of O_3 (Dignon and Hameed, 1985). At low NO_x concentrations (where NO_x is less than 5 - 30 pptv; this threshold depends on the concentrations of O_3 and hydrocarbons) increases in CO , CH_4 , and $NMHC$ lead to a decrease in O_3 , whereas at high NO_x concentrations increases in CO , CH_4 , and $NMHC$ lead to significant enhancements in O_3 . Therefore, no simple relationship exists between increases in the precursor gases and changes in tropospheric O_3 . Several model calculations have been performed to investigate the sensitivity of O_3 changes to changes in the precursor gases, both individually and collectively. All models that have attempted to simulate changes in O_3 during the past century have calculated increases in Northern Hemisphere O_3 by up to a factor of two, broadly consistent with observations, depending upon the assumptions made regarding the initial concentration, distribution, and changes in precursor gas concentrations, particularly NO_x .

Understanding the feedbacks among O_3 and its precursor gases is essential to understand tropospheric OH , which controls the atmospheric lifetimes of CH_4 and the $NMHC$ s. The global concentration of OH , which determines the oxidizing capacity of the troposphere, can be either enhanced because of elevated levels of tropospheric O_3 , NO_x or water vapour (associated with a global warming) or suppressed because of increases in CH_4 , CO , and $NMHC$ (Crutzen, 1987; Thompson et al., 1989). Prediction of regional and global trends in OH concentrations requires an understanding of regional emissions of CH_4 , CO , $NMHC$ and NO_x , as well as transport of O_3 between its source regions and the remote troposphere. One key point is that a continued increase in levels of CO would reduce the global concentration of OH because NO_x is too short-lived to counteract that effect over much of the globe. This would increase the atmospheric lifetime of CH_4 .

1.7.2 Carbon Monoxide

1.7.2.1 Atmospheric distribution of carbon monoxide

The atmospheric concentration of CO exhibits significant spatial and temporal variability because of its short atmospheric lifetime (2 - 3 months). The short atmospheric lifetime, coupled with an inadequate monitoring network, means that the global spatial variability and long-term trends in CO are not well documented. The limited observational data base (Heidt et al. 1980; Dianov-Klovov and Yurganov, 1981; Seiler and Fishman, 1981; Seiler et al. 1984; Khalil and Rasmussen, 1984, 1988a; Fraser et al. 1986 a, c; Newell et al. 1989; Zander et al., 1989; Kirchoff and Marimho, 1989; Kirchoff et al., 1989) has demonstrated that the concentration of CO , (i) is about a factor of two greater in the Northern than in the Southern Hemisphere where the annual average is about 50 - 60 ppbv, (ii) increases with latitude in the northern hemisphere, (iii) exhibits strong seasonal variations in both hemispheres at mid to high latitudes, and (iv) decreases with altitude. CO appears to be increasing at about 1% per year in the northern hemisphere, but the evidence for increases in the southern hemisphere is ambiguous.

1.7.2.2 Sources and sinks for carbon monoxide

The total annual source of CO is about 2400 Tg CO , being about equally divided between direct anthropogenic (incomplete combustion of fossil fuels and biomass) and atmospheric (oxidation of natural and anthropogenic CH_4 and $NMHC$) sources (Logan et al. 1981; Cicerone, 1988). Atmospheric concentrations of CO may have increased in the northern

hemisphere because of the fossil fuel source, and because of changes in the rate of oxidation of CH_4 , whose atmospheric concentration has increased since pre-industrial times. Fossil fuel sources of CO are presently decreasing in North America (EPA, 1989) and possibly in Europe, but may be increasing elsewhere.

The major removal process for atmospheric CO is reaction with OH (Logan et al. 1981). The observed seasonal variability in the southern hemisphere, distant from seasonally varying sources, can be explained by the seasonal variability in the concentration of tropospheric OH. Soils may provide a minor sink for CO (Conrad and Seiler, 1985).

1.7.3 Reactive Nitrogen Oxides

The key constituents of tropospheric NO_y , defined as the sum of all nitrogen oxide species except for N_2O , are NO_x , nitric acid (HNO_3), peroxyacetylnitrate (PAN: $\text{CH}_3\text{CO}_3\text{NO}_2$), and organic nitrates. Most primary sources of nitrogen oxides release NO_x (mainly NO); the other species are produced by photochemical reactions in the atmosphere. While the atmospheric lifetime of NO_x is short (about 1 day), the atmospheric lifetime of NO_y can range up to several weeks. Thus NO_y can transport nitrogen compounds away from source regions to more remote locations, where photolysis of HNO_3 and PAN, and thermal decomposition of PAN, can regenerate NO_x .

1.7.3.1 Atmospheric distribution of nitrogen oxides

The atmospheric concentrations of NO_x exhibit significant spatial and temporal variability, reflecting the complex distribution of sources and the short atmospheric lifetime. The near surface and free tropospheric concentrations of NO_x each vary by several orders of magnitude, highly influenced by the proximity of source regions. Near surface concentrations of NO_x range from as low as 0.001 ppbv in remote maritime air to as high as 10 ppbv in Europe and Eastern N. America (excluding urban areas), while free tropospheric concentrations range from 0.02 ppbv in remote regions to more than 5 ppbv over populated areas (Fehsenfeld et al., 1988).

The spatial inhomogeneity, coupled with a sparsity of measurements, means that the spatial and temporal distribution and long-term trends in NO_x and NO_y are not adequately documented, although reconstructed emissions inventories of NO_x suggest large increases throughout this century (Dignon and Hameed, 1989). Data from a Greenland ice core have shown that the concentration of nitrate ions (dissolved nitrate from HNO_3) remained constant from 10,000 years ago to about 1950, then doubled by the late 1970's, consistent with the increase in industrial emissions (Neftel et al. 1985b). Data from glacier ice in Switzerland indicates that nitrate ions increased by a factor of 4-5 between 1900 and the 1970's in Western Europe (Wagenbach et al. 1988).

1.7.3.2 Sources and sinks of nitrogen oxides

The sources of atmospheric NO_x are about equally divided between anthropogenic (combustion of fossil fuels: 21 Tg N per year, and biomass burning: 2-5 Tg N per year), and natural (microbial processes in soils: 20 Tg N per year; lightning: 2-8 Tg N per year, and transport from the stratosphere: 1 Tg N per year) (Galbally, 1989). Emissions of NO_x (6.3 Tg N per year) from the combustion of fossil fuels have not increased in N. America since 1970 (EPA, 1989). Soil emissions of NO are stimulated by agricultural activity (e.g., addition of fertilizer, manure, etc.), hence, agricultural soil emissions may provide significant sources of NO_x in many areas.

The dominant removal processes for NO_x are (i) conversion to HNO_3 , PAN, and organic nitrates by photochemical mechanisms, (ii) reactions involving NO_3 radicals, and possibly (iii) deposition of NO_2 on vegetation. The resulting NO_y species are then removed from the atmosphere by wet and dry deposition, or by conversion back to NO_x .

1.7.4 Non-Methane Hydrocarbons

1.7.4.1 Atmospheric distribution of non-methane hydrocarbons

The NMHC can be classified by atmospheric lifetime: (i) relatively long-lived (lifetimes > week) where the highest concentrations (up to 3 ppbv for ethane) are observed at middle to high northern latitudes; (ii) more reactive (lifetimes between half a day and one week) such as C₂ - C₅ alkenes whose concentrations exhibit significant temporal and latitudinal variability from <0.1 ppbv in remote areas to a few ppbv close to source regions, and (iii) extremely short lived (lifetimes of hours) such as terpenes or isoprene whose local concentrations may reach about 10 ppbv very close to their sources. Trends in the atmospheric concentrations of NMHC have not been established due to a lack of measurements.

1.7.4.2 Sources and sinks for non-methane hydrocarbons

The oceans are a major source of NMHC, mainly alkenes. Estimates of the source strength of ethene and propene range from 26 Tg C per year (Bonsang et al., 1988) to as high as 100 Tg C per year (Penkett, 1982). Emissions of NMHC from terrestrial vegetation are dependent upon environmental factors as well as the type of vegetation. Isoprene is primarily emitted from deciduous plants, whereas conifer trees are primarily a source of terpenes. Isoprene and terpene emission rates are very large, about 500 Tg per year for each (Rasmussen and Khalil, 1988). The source strength of NMHC from anthropogenic activities such as biomass burning, solvents and fossil fuel combustion has been estimated to be about 100 Tg per year.

The dominant loss mechanism for most NMHC is rapid (much faster than CH₄) reaction with OH. The products of these reactions are capable of forming O₃ in the presence of NO_x.

1.7.5 Feedbacks Between Climate and the Methane / Non-Methane Hydrocarbon / Carbon Monoxide / Oxides of Nitrogen / Tropospheric Ozone System

There are numerous potentially important feedbacks between climate change and tropospheric O₃ and OH. Changes in cloud cover, precipitation, and circulation patterns, as well as changes in the biospheric source strengths of CH₄, CO, NMHC and NO_x, will induce changes in homogeneous and heterogeneous reactions controlling O₃ and OH. In addition, changes in stratospheric O₃ may induce changes in tropospheric processes, through changes in ultraviolet radiation. Stratospheric O₃ depletion is likely to increase tropospheric O₃ when the levels of CO, NO_x, and NMHC are high, but reduce it in regions of very low NO_x. The importance of these feedback processes remains to be determined.

1.7.6 Conclusions

Tropospheric O₃ is a greenhouse gas that is produced photochemically through a series of complex reactions involving CO, CH₄, NMHC and NO_x. Hence, the distribution and trends of tropospheric O₃ depend upon the distribution and trends of these gases whose atmospheric concentrations are changing.

The short atmospheric lifetimes of O₃ (several weeks), and many of its precursor gases, coupled with inadequate observational networks, leave their distributions and trends inadequately documented. Most data support positive trends of about 1% per year for O₃ below 8 km altitude in the Northern Hemisphere (consistent with positive trends in several of the precursor gases, especially NO_x, CH₄, and CO), and a similar trend for CO in the Northern Hemisphere, but not in the Southern Hemisphere. While there is no systematic series of data that allow quantitative estimates of trends in NMHC and NO_x to be made, their atmospheric concentrations are likely to have increased during the past few decades because of increased anthropogenic sources. The ice core records of nitrate levels provide indirect evidence for a Northern Hemisphere increase in atmospheric NO_x.

1.8 AEROSOL PARTICLES

1.8.1 Concentrations and Trends of Aerosol Particles in the Troposphere

Aerosol particles play an important role in the climate system because of their direct interaction (absorption and scattering) with solar and terrestrial radiation, as well as through their influence on cloud processes and thereby, indirectly, on radiative fluxes. These processes are discussed in more detail in Sections 2.3.2 and 2.3.3. Two separate issues should be identified. The first is the effect of increasing or decreasing anthropogenic emissions of aerosol particles and their precursors in regions impacted by these emissions. The second is the role of feedback processes linking climate change and natural (biological) production of particles in unpolluted regions, especially over the oceans (cf. Section 10.8.3).

Total suspended particulate matter in air varies from less than $1 \mu\text{g m}^{-3}$ over polar ice caps or in the free mid ocean troposphere to 1 mg m^{-3} in desert dust outbreaks or in dense plumes from e.g., forest fires. In a typical sample of continental air mineral dust, sulphuric acid, ammonium sulphate as well as organic material and elemental carbon (soot) may be found both as pure or mixed particles. Most of the soluble particles become solution droplets at relative humidities above 80%; thus the radiative properties of aerosol particles even vary with relative humidity at constant dry aerosol mass.

A large part of the aerosol mass in submicron size particles is derived from gas-to-particle conversion through photochemical processes involving gaseous sulphur and hydrocarbon compounds. Such conversion may take place through photochemical processes involving the oxidation of sulphur dioxide (SO_2) and other sulphur gases to sulphuric acid (H_2SO_4) by reaction with OH. The H_2SO_4 so formed, having a low equilibrium vapour pressure, immediately condenses onto existing aerosol particles or forms new ones. Transformation to sulphuric acid and sulphate also takes place in cloud droplets, the majority of which eventually evaporate leaving the sulphate in the aerosol phase. Trends in the emission of these gaseous precursors, especially the sulphur gases, are therefore of great importance for the regional aerosol burden and thereby potentially for climate.

Large quantities of aerosol particles are also emitted from the burning of savannas and forests in tropical regions. The directly emitted particles consist largely of carbonaceous materials including black carbon (soot) (Andreae et al., 1988). In addition, particles are formed from precursor gases like SO_2 and hydrocarbons emitted by fires.

The average tropospheric life time of aerosol particles, and of their precursor gases, is of the order of only days or weeks. This is much shorter than the lifetime of most greenhouse gases. It implies that the atmospheric loading at any one time reflects the emissions that have taken place during the past few weeks only. No long-term accumulation in the troposphere is thus possible and any reduction in anthropogenic emissions will immediately result in a corresponding reduction in tropospheric concentrations. The short lifetime also implies large spatial and temporal variability in the concentrations of aerosol particles.

It has been established from analyses of Greenland ice cores that the amounts of sulphate, nitrate and trace metals, derived mainly from atmospheric aerosols, have been increasing since industrialisation began (Neftel et al., 1985 b; Mayewsky et al., 1986). However, there are almost no long-term, continuous direct observations of aerosol parameters in the atmosphere outside urban and industrial areas (Charlson, 1988). Indirect evidence from visibility observations indicates that the concentration of submicron aerosols over much of the eastern part of the U.S. has increased during the period 1948-1978 (Husar et al., 1981).

Another example of a trend analysis of atmospheric aerosols is due to Winkler and Kaminski (1988), who concluded that submicrometer aerosol mass outside Hamburg has increased by a factor of nearly two between 1976 and 1988 due to long range transport from industrialized centres in the region.

The hypothesis by Charlson et al. (1987) of a connection between climate and phytoplankton activity in ocean surface waters is based on the role played by soluble aerosol particles in determining the microphysical properties of clouds. The proposed climate-phytoplankton feedback rests on the facts that cloud condensation nucleus (CCN) concentrations in air are low over oceans far from land, that the CCN available in clean maritime air are composed almost totally of sulphate particles, and that this sulphur originates almost entirely from emissions of reduced sulphur gases (principally dimethylsulphide (DMS)) from the ocean surface. There is a significant non-linearity in the effect on cloud microphysics of given changes in CCN concentration, depending on the starting CCN concentration characteristics of clean oceanic air.

There is abundant evidence in the literature to confirm the role played by CCN concentration in determining cloud droplet size distribution. However, at this stage neither the sign nor magnitude of the proposed climate feedback can be quantitatively estimated, though preliminary calculations based on plausible scenarios indicate that this hypothesis merits careful consideration. Preliminary attempts to test this hypothesis using existing historical data of various types have been inadequate and have yielded only equivocal conclusions.

1.8.2 The Atmospheric Sulphur Budget

Current estimates of the global sulphur cycle show that anthropogenic emissions of SO₂ are likely to be at least as large as natural emissions of volatile sulphur species, cf. Table 5 (based essentially on Andreae, 1989). Within the industrialized regions of Europe and North America, anthropogenic emissions dominate over natural emissions by about a factor of ten or even more (Galloway et al., 1984; Rodhe, 1976). The anthropogenic SO₂ emissions have increased from less than 3 TgS per year globally in 1860, 15 in 1900, 40 in 1940 and about 80 in 1980 (Ryaboshapko, 1983). It is evident from these numbers that the sulphur fluxes through the atmosphere have increased very substantially during the last century, especially in the Northern Hemisphere. During the past decade the anthropogenic sulphur emissions in North America and parts of Europe have started to decline.

TABLE 5	
Estimates of global emission to the atmosphere of gaseous sulphur compounds †	
Source	Annual Flux (TgS)
Anthropogenic (mainly SO ₂ from fossil fuel combustion)	80
Biomass burning (SO ₂)	7
Oceans (DMS)	40
Soils and plants (H ₂ S, DMS)	10
Volcanoes (H ₂ S, SO ₂)	10
Total	147

† The uncertainty ranges are estimated to be about 30% for the anthropogenic flux and a factor of two for the natural fluxes.

Small amounts of carbonyl sulphide (COS) are also emitted into the atmosphere. They do not significantly affect the sulphur balance of the troposphere but they are important in maintaining an aerosol layer in the stratosphere.

1 Because of the limited atmospheric lifetime of most sulphur compounds, the augmentation of
2 the sulphur concentrations brought about by industrialization is not evenly distributed around
3 the globe. This is illustrated by Figure 1.16, which shows an estimate of how much more
4 aerosol sulphate there is at present in the lower atmosphere (900 hPa level) than in the pre-
5 industrial situation (Langner and Rodhe, 1990). Over the most polluted regions of Europe and
6 North America the sulphate levels have gone up by more than a factor of 10. Smaller increases
7 have occurred over large parts of the oceans.
8
9

10 1.8.3 Aerosol Particles in the Stratosphere

11
12 The vertical profile of aerosol particle concentration normally exhibits a marked decline up
13 through the troposphere followed by a secondary maximum in the lower stratosphere at around
14 20 km. The stratospheric aerosol layer is maintained by an upward flux of gaseous precursors,
15 mainly carbonyl sulphide (COS). Concentrations may be greatly enhanced over large areas for
16 a few years following large volcanic eruptions, such as El Chichon in 1982. No significant
17 trends have been detected in the global background aerosol layer in the stratosphere during
18 periods of low volcanic activity (WMO, 1989a). The potential impact on climate of
19 stratospheric aerosols is discussed in Section 2.3.2.
20
21

22 1.8.4 Conclusions

23
24 Aerosol particles have a lifetime of at most a few weeks in the troposphere and occur in highly
25 variable concentrations. A large proportion of the particles which influence cloud processes
26 and for the radiative balance, are derived from gaseous sulphur emissions. These emissions
27 have more than doubled globally, causing a large increase in the concentration of aerosol
28 sulphate especially over and around the industrialized regions in Europe and North America. If
29 anthropogenic sulphur emissions are indeed a major contributor to cloud condensation nuclei
30 concentrations on a global scale, then any climate prediction must take account of future trends
31 in regional and global anthropogenic sulphur emission, which may be quite different from
32 those of the greenhouse gases.
33

34 Aerosol particles derived from natural (biological) emissions may contribute in important ways
35 to climate feedback processes. During a few years following major volcanic eruptions the
36 concentration of aerosol particles can be greatly enhanced.
37

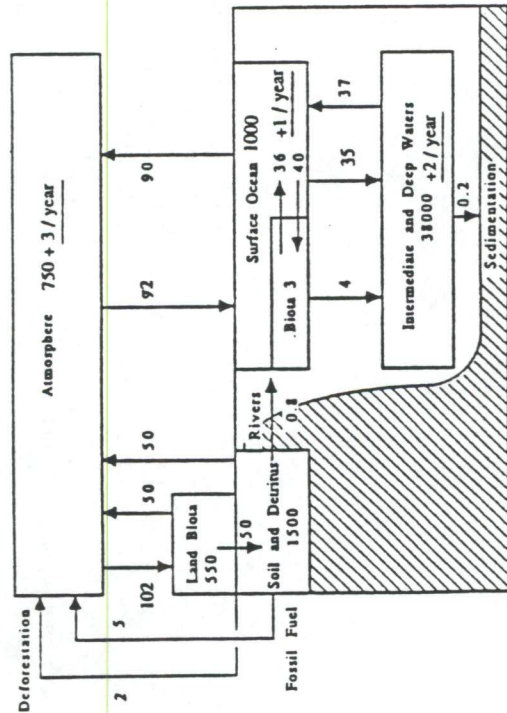


Figure 1.1: Global carbon reservoirs and fluxes. The numbers apply for the present-day situation and represent typical literature values. Fluxes, e.g. between atmosphere and surface ocean, are gross annual exchanges. Numbers underlined indicate net annual CO₂ accumulation due to human action. Units are gigatons of carbon (GtC; 1Gt = 10⁹ metric tons = 10¹²kg) for reservoir sizes and GtC yr⁻¹ for fluxes. More details and discussions are found in several reviews (Sundquist, 1985; Trabalka, 1985; Bolin, 1986; Siegenthaler, 1986).

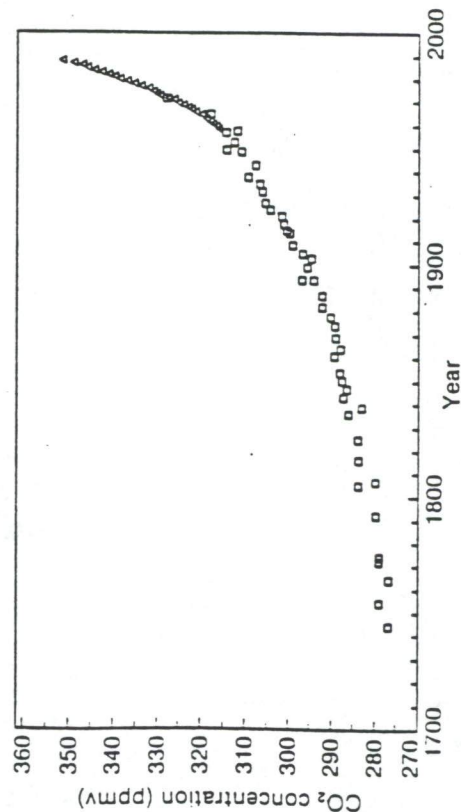


Figure 1.3: Atmospheric CO₂ increase in the past 250 years, as indicated by measurements on air trapped in ice from Siple Station, Antarctica (squares; Neftel et al., 1985a; Friedli et al., 1986) and by direct atmospheric measurements at Mauna Loa, Hawaii (crosses; Keeling et al., 1989a).

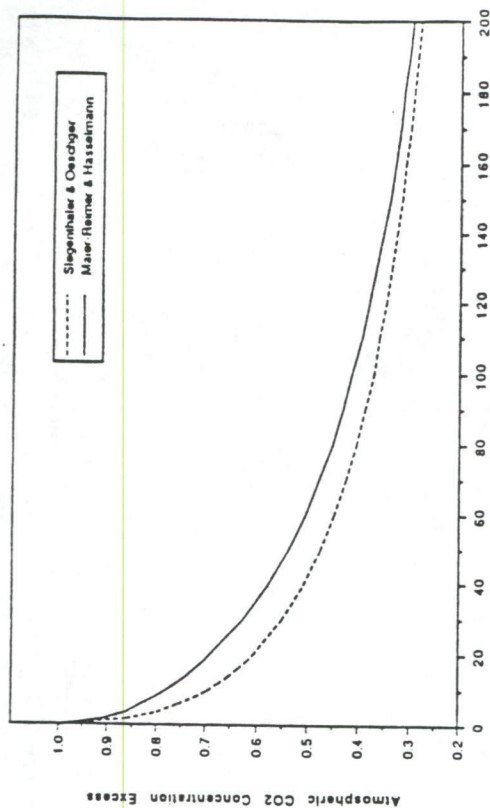


Figure 1.2: Atmospheric CO₂ concentration excess after a pulse input at time 0 (initially doubling the atmospheric CO₂ concentration), as calculated with two ocean-atmosphere models. Solid line: 3-dimensional ocean-circulation model of Maier-Reimer and Hasselmann (1987), dashed line: 1-dimensional box-diffusion model of Siegenthaler and Oeschger, 1987). The adjustment towards a new equilibrium does not follow an exponential curve; it is very fast during the first decade, then slows down more and more. The concentration excess does not go to zero; after a long time, a new equilibrium partitioning between atmosphere and ocean will be reached, with about 15 percent of the input residing in the atmosphere.

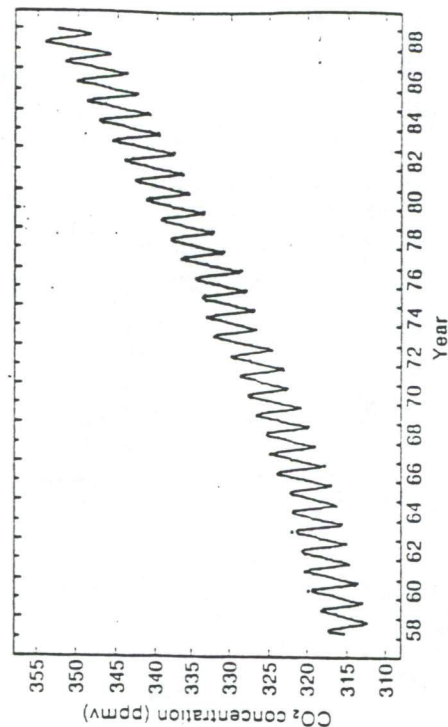


Figure 1.4: Monthly average CO₂ concentration in parts per million of dry air, observed continuously at Mauna Loa, Hawaii (Keeling et al., 1989a). The seasonal variations are due primarily to the withdrawal and production of CO₂ by the terrestrial biota.

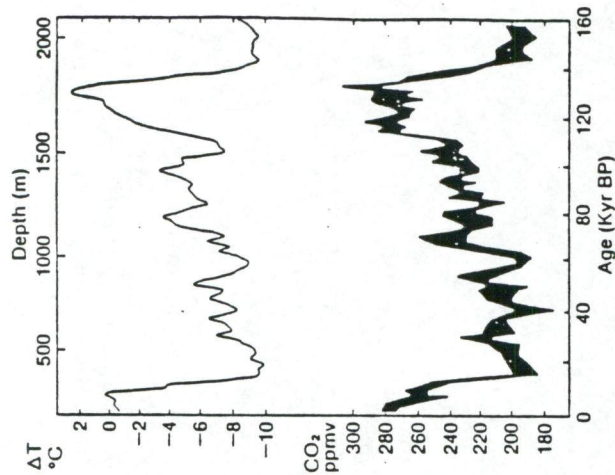


Figure 1.5: Global annual emissions of CO₂ from fossil fuel combustion and cement manufacturing, expressed in GtC yr⁻¹ (Rotty and Marland, 1986; Marland, 1989). The average rate of increase in emissions between 1860 and 1910 and between 1950 and 1970 is about 4% per year.

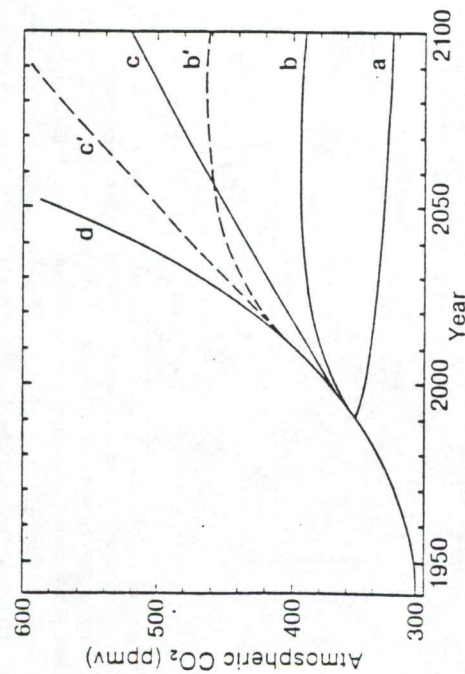


Figure 1.7: Future atmospheric CO₂ concentrations as simulated by means of a box-diffusion carbon cycle model (Enting and Pearman, 1982, 1987) for the following scenarios: a) - d): anthropogenic CO₂ production rate p prescribed after 1990 as follows: a) $p = 0$; b) p decreasing by 2% per year; c) $p = \text{constant}$; d) p increasing at 2% per year. Scenarios b') and c'): p grows by 2% per year from 1990-2010, then decreases by 2% per year (b') or is constant (c'). Before 1990, the concentrations are those observed (cf. Figure 3), and the production rate was calculated by to fit the observed concentrations.

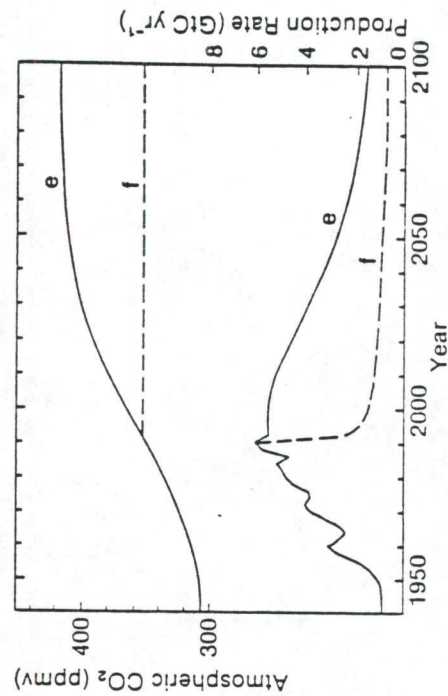


Figure 1.8: Future CO₂ production rates calculated by means of a box-diffusion carbon cycle model (Eniting and Pearman, 1982, 1987) so as to yield the prescribed atmospheric CO₂ concentrations after 1990. e) concentration increasing steadily (logistic function of time) to 420 ppmv. f) concentration constant after 1990.

Figure 1.6: CO₂ concentrations (bottom) and estimated temperature changes (top) during the past 160,000 years, as determined on the ice core from Vostok, Antarctica (Barnola et al., 1987). Temperature changes were estimated based on the measured deuterium concentrations.

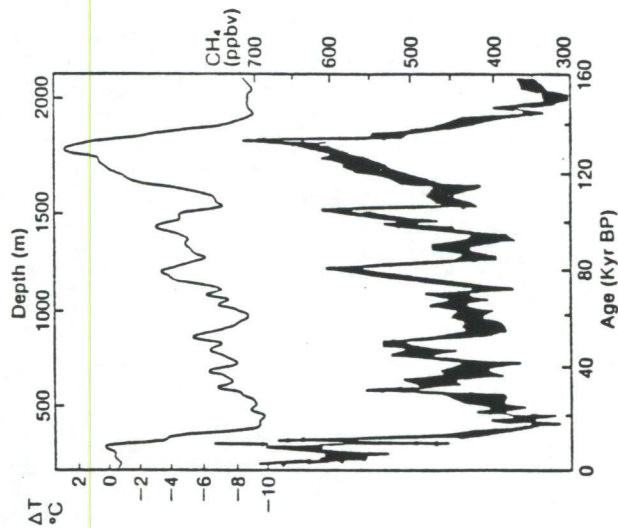


Figure 1.9: Methane concentrations (bottom) and estimated temperature changes (top) during the past 160,000 years as determined on the ice core from Vostok, Antarctica (Chappelaz et al., 1990). Temperature changes were estimated based on the measured deuterium concentrations.

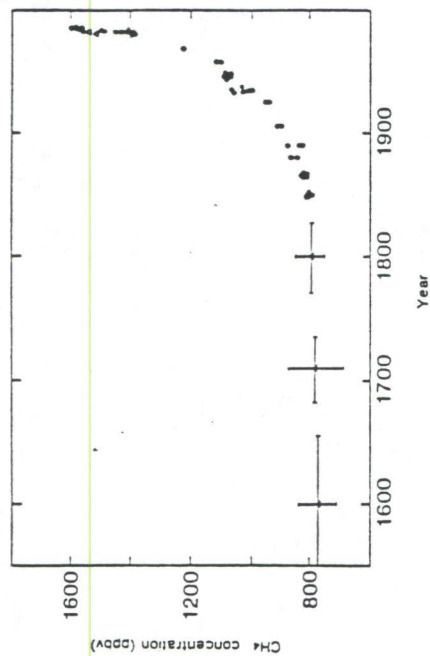


Figure 1.10: Atmospheric methane variations in the past few centuries measured from air in dated ice cores (Etheridge et al., 1988; Pearman and Fraser, 1989).

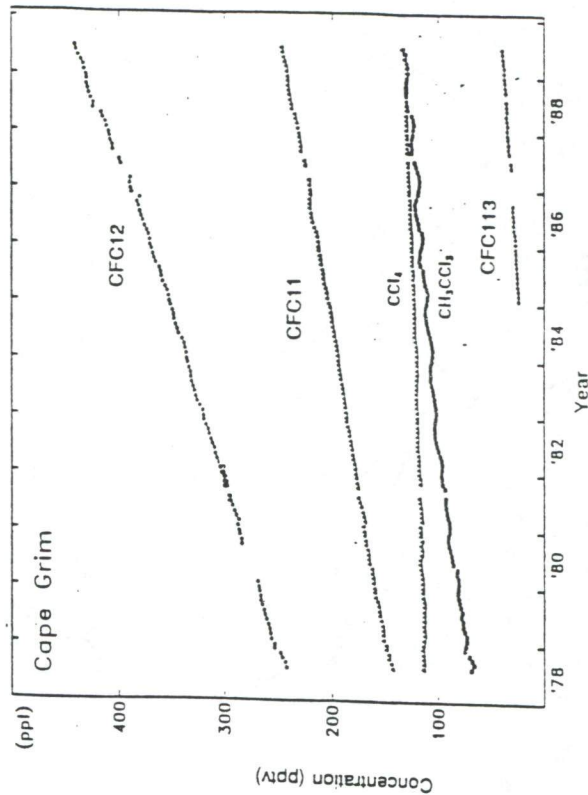


Figure 1.12: Halocarbon concentrations measured at Cape Grim, Tasmania during the period 1978-1989 (Fraser and Derek, 1989, and unpublished data).

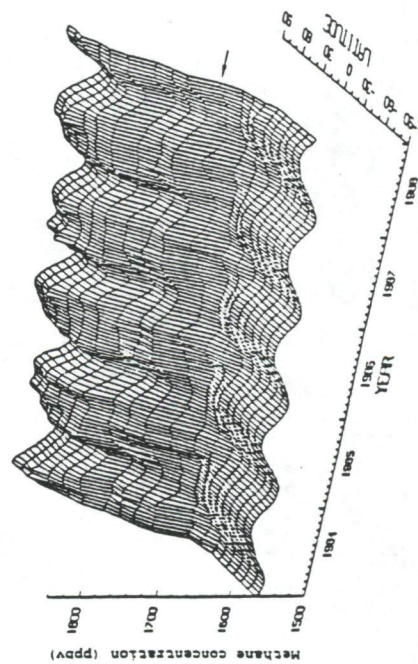


Figure 1.11: The global distribution, seasonality, and trend of methane from the GMCC network (Steele et al., 1987, and unpublished data)

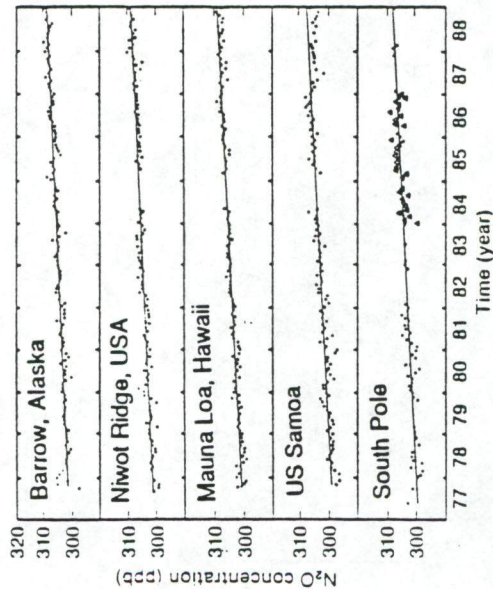


Figure 1.13: Atmospheric measurements of nitrous oxide from the NOAA/GMCC network (Elkins and Rossen, 1989).

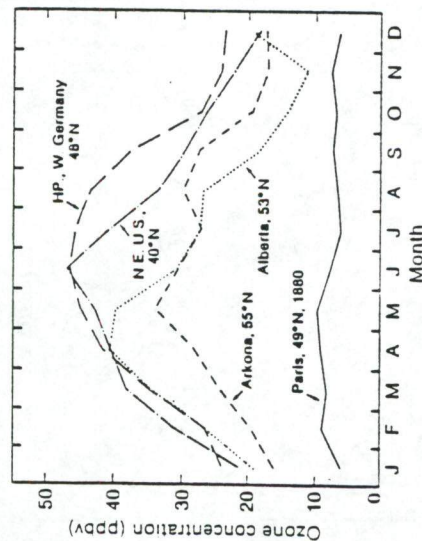


Figure 1.15: The seasonal variation of surface ozone. The solid line shows data from Montsouris, France, for 1876-86 (Volz and Kley, 1988). All other data are from the 1970s and 1980s: dashed line, Arkona, GDR (Feister and Warmb, 1987); dotted line, Ellerslie, Alberta, Canada (Angle and Sandhu, 1986); dot-dash line, average of eight rural sites in the northeastern U.S., the SUTE sites (Logan, 1988); long dashed line, Hohenpeissenberg, FRG (Logan, 1985). All the recent data are shown as monthly means of daily average values.

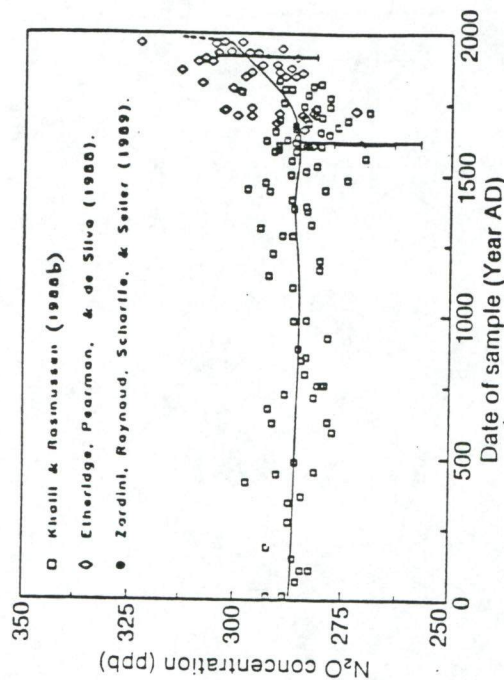


Figure 1.14: Nitrous oxide measurements from ice-core samples.

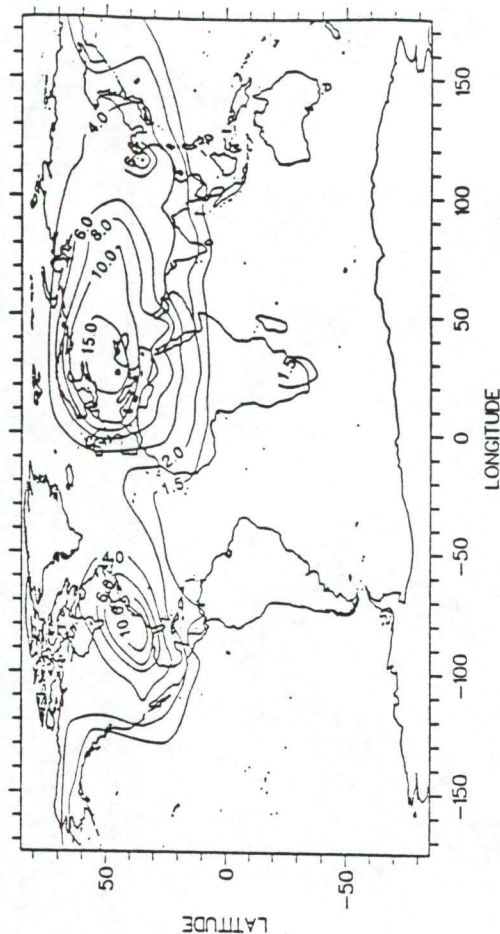


Figure 1.16: Simulated concentration of sulphate at 900 hPa: Ratio of concentrations based on total emissions (natural plus anthropogenic) divided by concentrations based on natural emissions, in July (Langner and Rodhe, 1990).

Section 1

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- 1
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- 3
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Environmental and Scientific Affairs

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 General
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Ram

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OVERVIEW AND CONCLUSIONS

Climate Change as One of Many Global Issues

A possible change of climate due to man's increasing emissions of greenhouse gases into the atmosphere is but one of several major threats that will confront mankind as the 21st century begins. The world population is increasing by more than 1.6% per year; the exploitation of natural resources is taking place at an enhanced rate; industrialization and urbanization is rapidly increasing the number of cities in the world with more than 10 million, in recent years even 20 million inhabitants, particularly in developing countries. At the same time the poor and developing countries are striving for improving living conditions for their populations. The World Commission on Environment and Development addressed many of these problems in its 1987 report but as yet rather few steps have been taken to resolve these serious problems.

The present report addresses one of these issues, that of a future change of climate. An attempt has been made to assess the present scientific view about the likelihood and characteristics of possible future man-induced climatic changes as well as the impacts of such changes. These studies also include analyses of their possible socio-economic consequences. Alternative options for actions that might be needed in order to slow down a climatic change or to stabilize climate are given. It is of course important to recognize that such interventions also might have far reaching socio-economic effects. Also, possible means for mitigating the effects of a climate change are presented.

The IPCC objective has been to provide a scientific-technical basis for political action. It is clear from the present overview, supported by the extensive working group reports and their policymakers' summaries, that some very specific general conclusions already can be drawn from this first phase of a broad assessment of the issue, although many uncertainties remain. On this basis the IPCC calls on the leaders of the nations of the world to consider carefully the conclusions reached so far and to take some first steps towards slowing down the present accelerating rate of change. The report also outlines long term targets that should serve as general guidelines in developing a first short term action plan with such an aim.

Still, further work is required to establish a more comprehensive basis for future policy decisions. The Panel considers it essential that its activity as a scientific-technical assessment group continues in order to keep policymakers as well informed as possible about the further advance of our knowledge and alternative actions for prevention and mitigation and also to respond to the possible needs of addressing specific and urgent issues as they may arise in forthcoming negotiations.

The following overview of the IPCC assessment presents a synthesis of the most important conclusions that have been reached by the Working Groups and that are judged to be of particular importance for the political analyses and the negotiations that will begin after the Fourth IPCC Planetary Meeting. For more specific questions, that also arise in an analysis of this major environmental issue, reference is made to the Working Group Reports and their Policymakers' Summaries. These have been completed after extensive discussions at a large number of Working Group and Subgroup meetings, which have been open to all countries that have expressed an interest to take part in the IPCC work.

The scientific issue

Increasing Greenhouse Gas concentrations

It is first of all important to emphasize that although many uncertainties about the likelihood and the character of a possible man-induced climate change, we do know some things with considerable certainty. The Working Group I has structured its Policymakers' Summary in distinguishing clearly between

- what we know with quite some certainty about future man-induced climatic change due to greenhouse gas emissions,
- what we calculate with confidence and
- what our best judgement is on matters that are more uncertain.

We are certain that

"there is a natural greenhouse effect which already keeps the Earth warmer than it otherwise would be."

"emissions resulting from human activities are substantially increasing the atmospheric concentrations of the greenhouse gases: carbon dioxide, methane, the chlorofluorocarbons (CFCs) and nitrous oxide. These increases will enhance the greenhouse effect, resulting on average in an additional warming of the Earth's surface."

There is thus no controversy about the ongoing changes of the composition of the atmosphere, i.e. an increase of the abundance of the major greenhouse gases. The

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carbon dioxide concentration has increased by 26-28% during the last two hundred years, half of which has occurred since the early 1950's. Methane has more than doubled during this same time. The concentrations of CFC-gases are well established. They were absent before the 1920's. Dinitrogen oxide has increased by about 10%. The integrated increase of all man-induced greenhouse gas emissions is at present equivalent to an increase of the preindustrial carbon dioxide concentrations by about 45%.

Man-Induced Climate Change

There is, however, considerable uncertainty about the magnitude of a climatic change caused by an increase of greenhouse gases in the atmosphere or, in other words, we know only approximately how rapidly a climatic change would be realized. If we compare the estimated change of the global mean surface temperature to prevailing conditions towards the end of last century i.e. before a significant impact of man's emissions could have occurred, and also include the uncertainty ranges assigned to the assessments as given by Working Group I the IPCC summarizes that:

The rate of change of the global mean surface temperature during the next 100 years is estimated to be 0.2 - 0.5 °C per decade, if assuming future emissions of greenhouse gases as given in a scenario Business-as-Usual, Scenario A. The global mean surface temperature is expected to have increased by 1 - 3°C from late last century to 2025, and by 2.5 - 6 °C towards the end of next century.

The conclusion above is based on computer experiments using climate models, which show that a doubling of the amount of carbon dioxide in the atmosphere (or an equivalent increase of a mix of greenhouse gases) would lead to an increase of the mean global surface temperature by 1.5 - 4.5 °C.

A climate prediction is the average result of computer experiments using a number of climate models, that differ with regard to the simplifying assumptions that have been made in order to be able to study the behaviour of the climate system quantitatively by using computers. The given uncertainty range for such a prediction of the future climate does not express a statistical confidence limit of the traditional kind, for example given as an outcome of a set of laboratory experiments. The range of the increase of the mean global surface temperature if doubling the amount of carbon dioxide simply includes the results of all model experiments reported in the scientific literature so far (except for a few that yield values even somewhat above the given upper limit). This does not exclude, however, the possibility that future experiments might give values outside this range, since we know for certain that all possible physical mechanisms have not been included adequately even in the most advanced climate models. *Nobody can either tell if further experiments preferentially might yield values for the change of the mean global surface temperature that would be less than the lower limit of the range or larger than the upper limit.*

This question of uncertainty were raised by several scientists in the peer review process of the Working Group I Report. Ideas were brought forward about possible

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mechanisms that might yield results outside the given range of uncertainty. They were, however, not supported by model experiments that permitted a real check on the implications of the additional processes that were suggested to be important. Some of these ideas can be tested without much delay, but the more general criticism, that the models have too coarse a resolution to permit the inclusion of sufficient details of the climatic system, requires considerable more efforts to resolve. Particularly when concerned with an improved treatment of cloud processes in the atmosphere and including a more detailed treatment of oceanic processes, it is necessary to design new models and to employ more powerful computers. Although interesting, these peer review suggestions do not at present justify a change of the range of uncertainty as agreed upon by the Working Group.

The IPCC concludes that

the uncertainty range given for the predicted change of the global mean surface temperature due to the emission of greenhouse gases into the atmosphere includes the results of experiments using essentially all climate models that are presently available for quantitative analysis and that have been reported in the scientific literature.

One What basis?

The scenario *Scenario A, Business-as-Usual* (of the reports by Working Groups I and III) is a scenario of world development with continued economic growth in the absence of policy interventions to reduce greenhouse gas emissions. The implicit assumption has been made that there will be no major long-lasting disturbances of the global economy and further that a rather rapid growth of GNP will take place, being somewhat quicker in developing countries than in the industrialized world. The world population is assumed to be above 8 000 million people by 2030 and gradually level off at about 11 000 million people towards the end of the 21st century.

Although oil and natural gas, according to this scenario, will play an important role as energy sources for several decades into next century, the major energy supply will be from coal, while on the other hand different forms of renewable energy as well as nuclear energy are judged not to play a more important role globally until well into the 21st century. Account has been taken of the Montreal Protocol for reducing the emissions of the CFC-gases but not the possibility of phasing them out. The increase of greenhouse concentrations will be equivalent of a doubling of the carbon dioxide concentration by about 2030 and a tripling or quadrupling during the latter part of next century. The long term projections are of course only indications of a possible development.

It is interesting to note that national plans for future expansion of the energy supply systems as revealed in an inquiry to nations (*Reference Scenario*; Working Group III) show an even somewhat more rapid increase of carbon dioxide emissions until 2025 than implied by the Scenario A.

The annual increase of carbon dioxide emissions by about 2% as implicit in both these scenarios is about the same as the observed increase since 1984. Although any of these scenarios can only provide a general idea about the character of the future, if no actions would be taken to prevent a build up of greenhouse gases in the atmosphere, their similarity is of considerable interest for the following analysis.

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The values for the changes of the mean global surface temperature, as given in the conclusion above, have been derived with due regard taken to the delay in the realization of the induced changes because of the inertia of the climatic system. A stabilization of the concentrations of greenhouse gases at the level reached in 2025 for the two given scenarios, i.e. at a level equivalent to a doubling of the carbon dioxide concentration, would ultimately lead to a change of the global mean temperature between 1.5 and 4.5 °C, i.e. a significantly larger change than the transient changes expected for the year of 2025. It should be pointed out, that there is no disagreement between scientists about the reality of such a lag due to the inertia of the climate system, although there are some different views about its magnitude.

Testing Model Computations against Observations

It is of course most essential to test model calculations against reality in order to validate the results. We know from observations that the global mean temperature has increased between 0.3 and 0.8 °C, since the middle of last century and that most of this change has occurred primarily during two periods, i.e. the first four decades of this century and since 1975 (Cf. Working Group I Report). Working Group I has carefully considered the criticism that local warming at observation sites within or in the vicinity of densely populated regions may explain some part of the observed change and accounted for this possibility when assigning the range of uncertainty for the observed change.

When trying to validate the model computations with the aid of these observations a further complication arises in that the global mean temperature has not been increasing steadily. Actually even a slight decline was observed during the period 1940 - 1975. Model computations show that such irregular (possibly unpredictable) variations of the global mean temperature by several tenths of a degree are to be expected. These, of course, further obscure the warming trend that may result from increasing amounts of greenhouse gases in the atmosphere. We can therefore not yet conclude firmly that a change of the global mean surface temperature has occurred because of increasing concentrations of greenhouse gases in the atmosphere. However, the IPCC confirms the view as expressed by Working Group I, i.e.

"the size of the observed warming is broadly consistent with predictions of climate models, but it is also of the same magnitude as natural climate variability. Thus the observed increase could be largely due to this natural variability; alternatively this variability and other human factors could have offset a still larger human-induced greenhouse warming. The unequivocal detection of the enhanced greenhouse effect from observations is not likely for a decade or more. The commitment to future climate change will then be considerably larger than today."

*Needs
clarification!*

Inertia of the Climate System and Delay of Warming

There is general agreement among scientists that an increase of the atmospheric carbon dioxide concentration will not return to pre-industrial concentrations in less than one or several centuries and that the residence times of the most abundant CFC-gases as well as nitrous oxide are also of the order of a century or more. The IPCC therefore emphasizes that

the changes of climate that might be induced by increasing greenhouse gas concentrations most likely will remain for more than a hundred years, i.e. for human practical purposes these changes of climate may be considered as irreversible.

We shall later return to this characteristic time scale for the response of the climatic system to changes of atmospheric composition in relation to the inertia of human society, particularly the time required for the development of new energy supply systems.

Regional Differences of a Climate Change

A change of climate does not only mean a change of the global average temperature. It also brings with it changes of cloudiness and precipitation as well as changes of other climatic parameters. Evaporation from the ocean surface would be enhanced as would precipitation over both sea and land. The global hydrological cycle would intensify. It is, however, difficult to conclude more firmly what average changes of such climatic features of climate would occur.

It is important to emphasize that regional climate changes will be different from the global mean. The geographical distribution of changing climatic conditions is of prime interest for nations and for people in different trades of life. Although model computations do yield maps of the likely regional distributions of change, the results from different models differ to the extent that as yet only some rather general conclusions can be drawn. We cannot therefore give reliable regional predictions at the smaller scales which are required for more precise impact assessments. The IPCC emphasizes the Working Group I view, that

our confidence in more precise predictions of regional changes is low, though in general these changes will be different from the global mean. For example, the temperature increases in Southern Europe and central North America are predicted to be higher than the global mean, accompanied on average by reduced summer precipitation and soil moisture.

Also, the North polar regions probably will experience more of a temperature increase than the globe in general, while predictions of changes in the Antarctic region are more uncertain because of as yet insufficient consideration of air sea interaction. Soil moisture in the interior of the continents at middle latitudes might generally also decrease.

So what?
What is so
magic re
pre industrial
values?

Do we have a
general
statement!

A different and potentially useful way of predicting patterns of future climate is to search for periods in the past when the global mean temperatures were similar to those we expect in the future and then use the spatial patterns as analogs when predicting the regional changes. Although some such analogs can be found (cf. Working Group II), their usefulness is not generally accepted, since validation against independent data is not as yet sufficiently convincing (cf. Working Group I).

Improvements of our capability to predict likely regional changes of climate will come only slowly. The models must be developed to permit a more detailed treatment of land surface features as well as atmospheric and oceanic processes. This in turn will require the use of more powerful computers. Detailed predictions may become available in the course of the 1990's, but their reliability must be validated in some way. This will require comparison with observed changes in the past. Data from the last tens of thousands of years may be useful for general testing of the models. Such paleoclimatic data are, however, still too few and of insufficient accuracy to serve this purpose adequately.

The IPCC concludes:

It is probable that validation of regional and in particular local predictions of climate change will only be gradually achieved by comparison of model results with observed changes in the future. Advanced climate models will permit early validation as spatial patterns of change become detectable in spite of the natural climatic variability. It will be particularly important to establish in this way the internal consistency of regional differences of predicted changes.

It must finally be emphasized that our imperfect understanding of climate processes makes us vulnerable to surprises. It is of interest to recall in this context that the man-made seasonal depletion of ozone over the Antarctica was entirely unpredicted. In particular, the arctic circulation, changes of which are thought to have led to periods of comparatively rapid climate change at the end of the last ice age, is not well observed, understood or modelled. This inadequate knowledge and understanding also calls for precaution in formulating environmental policy for the future.

Environmental impact of climate change and socio-economic consequences

Impact and sensitivity

It is clear from the previous section that we are not yet able to foresee future likely changes of the regional distribution of climate with much accuracy. We therefore cannot either predict more specifically the probable impacts of regional changes, but are limited to assessing them in rather general terms. Quantitative assessments of

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the socio-economic implications of a climate change for specific regions and countries are not either yet possible. We can, however, analyze the sensitivity of the environment to a given change of climate and as a matter of fact much of what is brought out in the report of Working Group II should be interpreted in that way. It is also important to recognize that the many possible impacts of climate change as described in the Working Group II report will not necessarily all occur for a given global scenario of climate change.

Some improvements of this situation will probably occur gradually during the 1990's, but as emphasized before even an increased research effort will hardly resolve this difficulty quickly. On the basis of the Working Group II Report the IPCC concludes that:

The uncertainty of the regional predictions of climate change at present limits our possibility to assess quantitatively the impacts of such a change. For the time being we rely on sensitivity analyses for the assessment of the likely impacts, which should, however, yield rather clear ideas about the vulnerability of nations to such changes. This uncertainty about the spatial characteristics of regional and local changes of climate must not overshadow the earlier conclusion that a significant change of the global mean temperature brings about major changes on the regional and local scale.

It is likely that there will be both winners and losers in case of a global change of climate because the probable differences of the change between one part of the globe and another. It seems most likely from the assessment that has been made by Working Group II, that the latter will dominate. In any case, because of the very major capital investments that have been made in the development of the present national infrastructures, the losses will be larger the more quickly a change will occur.

A rising sea level and its impacts

A warmer climate will induce a rise of the sea level, the magnitude of which would not be obscured because of regional uncertainties with regard to the change of climate. It might, however, be counterbalanced or aggravated by changes that are dependent on changes of the ocean circulation as well as slow vertical movements of the Earth's crust. The former cannot yet be predicted, while the latter are reasonably well known in parts of the world, however less so in developing countries. This rise of sea level is partly due to the expansion of sea water when being warmed, partly due to the melting of glaciers. The response of the ice sheets over Greenland and in the Antarctica is more uncertain, since both accumulation (due to change of snowfall) and melting is influenced by a change of climate. Working Group I concludes that:

"For the IPCC Scenario A, Business-as-Usual, sea level is expected to be between 10 and 30 cm higher than today by the year 2030, with a best-estimate of 20 cm, compared with a rise by 10 to 15 cm during the last 100 years, i.e. the rate of rise is expected to be 2 to 6 times more rapid than in the

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recent past. By the end of next century the rise is estimated to be between 30 and 100 cm, with a best-estimate of 65 cm."

These values are less than those given in earlier assessments, particularly those giving the upper bounds. The lower values do not give rise to an immediate concern except in coastal areas of some countries with extensive coastlines of low lying land and for some island states. The Working Group II recommends a more precise assessment of the risks to coastal areas and islands and to living resources caused by a 0.3 - 0.5 m rise of sea level which might be expected towards the latter part of next century, if no preventive measures were taken. It is further important to note that:

A rise of sea level would probably not cease even if a climate change were stopped. A prevailing warmer climate would warm the sea water down to greater depths and decrease glaciers and possibly also the Greenland ice sheet, the more rapidly the larger the change of climate would be at the time of climate stabilization.

Further:

The slowly rising mean sea level is not an immediate threat. Increased damage will rather be experienced during storms, when short term variations of sea level induced by winds and by high waves, would cause more destructive flooding. Delta regions of major rivers are obviously sensitive and often also heavily exploited for agricultural production.

The rise of sea level might threaten some low island and coastal zones by the middle of next century. A one meter rise, which might occur in 100 to 150 years time, would impact 300 000 kilometers of coastline (Working Group III), would render some island countries uninhabitable, displace tens of millions of people, seriously threaten low-lying urban areas, flood productive land and contaminate fresh water supplies.

The West Antarctic ice sheet is of special concern because a large part of it is grounded far below sea level. Although the Working Group concludes that "within the next century, it is not likely that there will be a major outflow of ice from the West Antarctica due directly to global warming", the likelihood of this happening probably will increase during later centuries in the case of a prevailing climate considerably warmer than that of today.

Impacts on terrestrial systems

Turning next to the more difficult question of impacts of climate change on terrestrial ecosystems, it is first important to recognize that a warmer climate is not necessarily harmful, if precipitation is adequate. The ecosystem composition might of course change, which in turn would imply the loss of plant species. It is likely, however, that the distribution of precipitation would change and that soil moisture predominantly might decrease in continental regions. A change of the

distribution of climatic zones would probably also occur and be associated with large regional and local changes. Some ecosystems might well have difficulties to adjust in this process. Working Group II Policymakers' Report gives a more detailed analysis of the expected impacts in the case of Scenario A. Some key features of these impacts as described by Working Group II have been extracted for the present overview:

Changes of climate would have an important effect on agriculture, including livestock. Studies have not yet conclusively determined whether, on average, global agricultural potential would increase or decrease. Negative impacts will be felt at the regional level and there may be severe effects in some regions, particularly decline in production in regions of high present-day vulnerability, that are least able to adjust technologically to such effects. Patterns of agricultural trade could be altered by decreased cereal production in some of the currently high-production areas. Although not analyzed more specifically, the supply of food would become more vulnerable to disturbances caused by changes of climate in a world with a rapidly growing population.

The rotation period of forests is long and current forests will mature and decline in a climate, to which they are increasingly more poorly adapted. Major forest-type zones and species ranges in the Northern Hemisphere could shift significantly as a result of climate change. The most sensitive areas will be where species are close to their biological limits in terms of temperature and precipitation, as for example in semi-arid areas. Social stresses can be expected to increase and consequent damage to forests may occur. These increased and non-sustainable uses will place more pressure on forest investments, forest conservation and sound forest management.

Natural terrestrial ecosystems would be markedly influenced because the rates of projected climatic change are likely to be faster than the ability of some species to respond. Flora and fauna would lag behind these climatic shifts and could find themselves in a different climatic regime. Major vegetation zones could face severe disturbances and disruptions, influencing plant and associated animal populations. Species could disappear and lead to substantial reductions in biological diversity. The socioeconomic consequences of these impacts will be significant, especially for those regions of the globe where societies and related economies are dependent on natural terrestrial ecosystems for their welfare.

Enhanced carbon dioxide concentrations in the atmosphere would increase the water efficiency in the process of photosynthesis which might enhance growth for certain plants in some regions of the globe, which might be beneficial in agriculture and forestry as well as for natural ecosystems. Man-induced changes of the chemical composition of the

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atmosphere and ecosystem feedbacks could reduce this positive effect over time.

Relatively small climate changes can cause large water resource problems in many areas, especially in arid and semi-arid regions and those humid areas where demand or pollution has led to water scarcity. While many areas will have increased precipitation, soil moisture will decrease in other areas, particularly in those that are already in marginal situations such as the Sahelian zone in Africa. Increased drought risk represents potentially the most serious impact of climate change on agriculture at both the regional and the global level. Future design in water resources engineering will need to take possible impacts into account when considering structures with life span of 80 to 100 years, but necessary analyses are uncertain at present because of the lack of adequate regional predictions.

Global warming can be expected to affect the availability of water and biomass, both major energy sources in a large number of developing countries.

The most vulnerable human settlements are in developing countries, in the lower income groups, residents of coastal lowlands and islands, in semi-arid grasslands and the urban poor in squatter settlements, slums and shanty towns, especially in mega-cities.

It is obviously difficult to judge how serious a climatic change might be to an individual nation, or to humankind as a whole, on the basis of the still largely qualitative findings as given in the Working Group II report. Means to integrate more quantitatively the implications of these many possible impacts, both positive and negative, are urgently needed, but can hardly be used before we can predict expected regional changes of climate with some better skill.

Impacts on people

Direct impacts of a climate change on people are modest. A warmer climate may imply, however, that precious hot places become even hotter. Even though this may cause problems for elderly and sensitive people, it usually cannot be considered as a serious problem.

Changing climate could, on the other hand, change the geographical distribution of diseases in the world and in some cases stimulate a spread of epidemic diseases. The IPCC recognizes the fears as expressed by working group II, that

Changes in precipitation and temperature could radically alter the patterns of vector-borne and viral diseases by shifting them to higher latitudes and thereby put large populations at risk. This could initiate migrations of people in turn leading to

But quantitatively
how many are
required to bring
South Am diseases
to North Am or
African ones to
Europe?

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disruptions of settlement pattern and social instability in some areas.

It is obvious that decisions on measures to delay or prevent climatic change must be based on a judgement of the seriousness of its impact on the environment, nations, people and its consequences for a sustainable development in the world as a whole, on one hand, and the resources required and their costs, on the other. As has already been emphasized there will undoubtedly be winners and losers, but the latter most likely will dominate. It is important to recognize that we should not only consider the direct impacts but also the indirect ones due to the increasing interdependence of nations and people in the world. Assessments of this latter kind have hardly yet begun. A basic question will undoubtedly be:

How will people in the countries of the world respond to negative and positive local impacts brought about by changes of climate? How will the solidarity between people and nations be influenced by predominantly negative global effects?

In principle, but also because we shall have to rely on an overall, largely qualitative, evaluation of the prospects of a possible climate change induced by man for a good part of the 1990's, we must view this issue in the broader perspective of human activities on Earth in general. It cannot be dealt with separately from a series of other fundamental issues such as those of the continuing rapid growth of world population, mismanagement of land and water resources, overexploitation of natural resources, rapid urbanization, etc. The climatic issue is linked to the sustainable development of human life on Earth as examined by the World Commission on Environment and Development in its report "Our Common Future". In conclusion, it is essential that the spirit, as expressed in the following paragraph from the introduction of their assessment, also can serve as a basis for the development of necessary analysis and action programs that will be required to battle against climate change.

"The members of the World Commission on Environment and Development came from 21 very different nations. In our discussions, we often disagreed on details and priorities. But despite our widely differing backgrounds and varying national and international responsibilities, we were able to agree to the lines along which change must be drawn. We were unanimous in our conviction that the security, well-being and very survival of the planet depend on such changes, now."

What changes.

Slowing down climate change and/or mitigating its consequences and/or adapting to them.

As pointed out above mans emissions of greenhouse gases have enhanced their concentrations in the atmosphere so that their present integrated effect is equivalent to a concentration of carbon dioxide in the atmosphere of about 405 parts

How do we know that they are predominantly negative? If only 2°C this is clearly not the case!

Scale factor.

WMU/UNEP/IPCC

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*what magic**Need to add data on natural fluxes to put this in perspective
What steps?*

per million, i.e. an increase by about 45% of the preindustrial carbon dioxide concentration. Half of this increase has occurred in less than 20 years. The analysis of different scenarios of the future increase of greenhouse gases in the atmosphere shows that the equivalent carbon dioxide concentration in the atmosphere probably would double in about 40 years (or even less) if no measures to limit future emissions were taken. It is an international political issue to reach agreements, based on the analysis of the expected impacts, on how far reaching measures are justified.

It is obvious that our present knowledge, particularly with regard to the socio-economic impacts of preventive actions as well as of a policy of no action, is not as yet sufficient in order to reach a firm agreement on a long-term policy for dealing with the issue of climate change. IPCC still concludes,

that the characteristics of man-induced climate change and its impacts require serious attention now. Although we do not know very well the magnitude of such a change, nor the seriousness of its impacts, the practical irreversibility of the expected change as well as the fact that there will be a considerable lag in its realization (an ongoing change will be hidden by natural variability of the climate for still some time) call for some early steps of action to be taken.

Before proceeding to such a discussion it is most important to recognize fully the very major efforts that are required in order to slow down and in particular to halt an on-going climate change.

The scope of the problem

It is important to clarify first the relative importance of the different greenhouse gases in bringing about a climate change. We need to understand the difficulties that will be encountered in efforts to reduce their emissions and thereby to slow down their rates of increase or ultimately to stabilize their atmospheric concentrations. Working Group I and II have presented their past relative contributions to the additional forcing (heating) of the atmosphere during the period 1875 - 1990 and during last decade as well as their expected contributions for the period 1990 - 2025 in the Scenario A, Business-as-Usual, (in percent)

	1875-1990	1990-90	Scenario A 1990-2025
Carbon dioxide	55	49	60
Methane	15	9	14
CFC-gases	11	20	11
Dinitrogen oxide	4	4	5
Tropospheric ozone	10	14	5
Stratospheric water (due to decay of methane)	5	5	5

The recent (1980 - 1990) reduction of the forcing due to carbon dioxide is due to a slower increase of the carbon dioxide emissions since the early 1970's, but particularly to the very rapid increase of the role of the CFC-gases as forcing agents. The role of the CFC-gases is, however, again expected to decrease because

Clearly need to add a category for natural fluxes

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of the partial application of the Montreal Ozone Protocol, it is important to recognize, however, that this decline will be rather slow, since their life times in the atmosphere are long. In addition, the CFC-11 and CFC-12 will probably be replaced by CFC-22, which is essentially harmless with regard to the ozone layer, but will cause significant climate forcing (heating). Still, the relative role of carbon dioxide as a heating agent will increase throughout the 21st century according to this scenario.

A stabilization of the concentrations of greenhouse gases in the atmosphere in the case of the Scenario A (Business-as-Usual) can not be achieved until the 22nd century and only at a level above doubling of the equivalent carbon dioxide concentrations.

The IPCC has developed three other scenarios;

Ways not nuclear too

Scenario B, Low Emissions, equivalent of doubling carbon dioxide concentrations by the middle of the 21st century, and an ultimate stabilization also above doubling; the energy supply mix of fossil fuels shifts towards natural gas, large efficiency increases are achieved, deforestation is reversed and the Montreal Protocol on CFC-gases is fully implemented;

Scenario C, Control Policies, stabilization of greenhouse gas concentrations at a level equivalent of doubling the carbon dioxide concentration towards the end of next century; shift towards renewable energies and nuclear energy takes place in latter part of the next century, CFC-gases are phased out and agricultural emissions (methane and dinitrogen oxide) are limited;

Scenario D, Accelerated Policies; stabilization of the greenhouse concentrations before the end of next century at a level equivalent of 170 % of the preindustrial carbon dioxide concentration; a rapid shift to renewable energies and nuclear energy takes place early in the next century, stringent emission controls in industrial countries and moderate growth of emissions in developing countries stabilize atmospheric concentrations.

All these three scenarios require more or less stringent measures in order to slow down the rate of increase of greenhouse gas concentrations in the atmosphere. We next analyse the possibilities to restrict the emissions of these gases.

Reducing emissions of CFC-gases

The Working Group I assessment quite clearly shows that the CFC-gases are very much more effective in heating the atmosphere than any of the other greenhouse gases. IPCC notes that

even the simplest cost-benefit analysis shows that a reduction of the use of CFC-gases would, by orders of magnitude, be the most cost-effective measure to slow down global warming.

Because of their additional role in decreasing the ozone layer in the stratosphere, much more stringent measures for their use may soon be agreed upon. It should be recognized that

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this would also be an effective measure for reducing the climate forcing only if no CFC-gases, particularly not CFC-22, were permitted for more extensive future use.

Reducing methane and dinitrogen oxide emissions

The Working Group I analysis further shows that reductions of the emissions of methane and particularly dinitrogen oxide are difficult to achieve, since they are primarily due to wide spread activities in society related to agriculture, forestry and waste management. Our present knowledge of what to do is inadequate and prevailing techniques can only be changed slowly. Some part of the methane emissions are released in association with mining, transportation and use of fossil fuels. The possibility to eliminate such leakage should be carefully studied, but efficient measures are probably difficult to implement at least with regard to leakage from coal mines. The projected increase of water vapour in the stratosphere is also caused by the increasing concentrations of methane in the atmosphere. Of course further work to try to reduce these emissions should be given high priority.

The increase of ozone in the troposphere is due to enhanced levels of air pollution and a further increase can be stopped by more stringent measures for reduction of air pollution, in particular the emission of hydrocarbons and nitrogen oxides released by the traffic.

Reducing fossil fuel combustion and deforestation

It is clear from the analysis above that the major source of greenhouse emissions is due to the emissions of carbon dioxide.

The IPCC notes that

assuming that effective steps to limit the future emissions of CFC-gases into the atmosphere are taken, the Scenario A (Business-as-Usual) implies that the climate forcing by emissions of carbon dioxide soon will be above 80% of the projected total forcing due to greenhouse gas emissions and may become 70 - 75% in the latter part of the 21st century; prevention of a change of climate necessarily requires a stabilization of atmospheric carbon dioxide concentrations.

Carbon dioxide emissions are due to fossil fuel burning, at present 5.5 - 6.0 billion tons of carbon per year, and a net decrease of the terrestrial biosphere due to deforestation and changing land use, 1.0 - 2.5 billion tons of carbon per year in 1980 (at present probably somewhat larger). The total emissions are thus at least 6.5 - 8.5 billion tons of carbon at present. The world population is about 5.2 billion and the emissions per capita thus about 1.1 tons of carbon due to fossil fuel emissions and 0.35 billion tons due to deforestation and changing land use.

The emissions due to fossil fuel burning are, however, very unevenly distributed between regions of the world as shown in the following table:

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*Need to include
natural fluxes here
and label these as anthropogenic*

Year: 1985	Carbon emissions billion tons of carbon per year	Carbon emissions tons of carbon per capita and year
Global Totals	5.15	1.06
Developed	3.83	3.12
North America	1.34	5.08
Western Europe	0.85	2.14
OECD Pacific	0.51	2.14
Centr. Planned Europe	1.93	3.19
Developing	1.33	0.36
Africa	0.17	0.29
Centr. Planned Asia	0.54	0.47
Latin America	0.22	0.55
Middle East	0.13	1.20
South and East Asia	0.27	0.18

It is not possible to deduce reasonably accurate values for the distribution of emissions due to deforestation and changing land use. It seems clear, however, from the reports by Working Groups II and III that the net emission from the tropics about equals the total global emissions which amounted to about 1.7 billion tons of carbon per year (range 1.0 - 2.5) in 1980, while the net emissions from the rest of the world were about zero. In some regions there is an accumulation of carbon in the forests, while the carbon content of other forests decline somewhat. The average emissions per capita in the tropical countries is about 1 ton of carbon per year while some with more intense exploitation of their forests account for a per capita emission that approaches the amounts emitted by fossil fuel combustion in industrial countries.

IPCC summarizes, that

- the globally averaged per capita emissions of carbon dioxide due to fossil fuel carbon is at present about 1.1 ton of carbon/year and that due to deforestation and changing land use about 0.35 tons of carbon/year,

and recalls the conclusion reached by Working Group II, that

- carbon dioxide emissions would have to be reduced by at least 60% in order to stabilize atmospheric concentrations of carbon dioxide, i.e. to totally not more than 3 billion tons of carbon per year.

On this basis the IPCC emphasizes that if

*Here need to
note the
possibility of
anthropogenic
modification
of the natural
fluxes to
the same
end.*

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- considering the projected increase of world population to about 8 billion people by 2025, this implies that the per capita emissions at that time must not exceed about 0.35 ton per year in order to stabilize atmospheric carbon dioxide concentrations, a value that should be compared with present rates of emissions as given in the table above.

- IPCC concludes that stabilization of the atmospheric carbon dioxide concentrations would require very major efforts by the global society.

Naturally all countries, and developing countries in particular, are striving for higher standards of living and energy is an important prerequisite for achieving such goals. It is, however, obvious that the creation of a supply scheme for energy use in developing countries, based primarily on the use of fossil fuels, as is now in existence in industrial countries, would beyond the year 2025 imply about a 4 - 5 fold increase of present global carbon dioxide emissions, if also accounting for the increasing world population. Even if the emissions due to deforestation and changing land use could be stopped, this magnitude of the energy supply, still primarily by using fossil fuels, would mean at least 10 times larger emissions than acceptable in order to stabilize the atmospheric carbon dioxide concentrations.

It is important in this context also to compare the possible importance of reducing the rate of deforestation and expand reforestation. We note first that the total emissions of carbon dioxide due to man's use of the terrestrial ecosystems are about 25% of the total man-induced emissions of carbon dioxide into the atmosphere. A major decrease of the rate of deforestation certainly would contribute significantly to slowing down the rate of carbon dioxide increase in the atmosphere, but not stop it.

It has been proposed that a major program for reforestation should be initiated. The Noordwijk Ministerial Declaration from November 1989 requested the IPCC to explore the effects of an annual reforestation of 12 million hectares. Working Group I has calculated roughly the transfer of carbon from the atmosphere into the terrestrial biosphere that thereby would be achieved. If such a policy would be maintained for a period of 40 years an area about 50% of the total area of the United States would have been reforested. By 2030, i.e. 40 years from now, about 25 billion tons of carbon would have been sequestered into the terrestrial biosphere, possibly somewhat more if also the soil carbon would increase. After 100 years the total uptake might be 50 - 75 billion tons of carbon. This is 5 - 10 % of the emissions of carbon dioxide into the atmosphere that is expected during these periods of time due to fossil fuel burning according to the Scenario A and the Reference Scenario. An analysis of the practical difficulties of implementing such a policy has not been analyzed by the Working Group III.

The IPCC considers it important, that

early steps are taken to reduce the rate of deforestation in tropical countries in order to initiate attempts to limit the increase of carbon dioxide in the atmosphere, but also and not the least for maintaining biological diversity,

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major programs of reforestation might play some role in slowing down the rate of increase of carbon dioxide in the atmosphere, but would at most reduce it by 5 - 10 %.

Strategies for the future

The major dilemma

The outcome of the work of Working Group I is clearly that the scientific community, although not unanimously, essentially agrees about the likely climate changes that would occur if the present increases of greenhouse gas emissions would continue well into the next century. Also, Working Group II envisages far reaching environmental consequences of such changes.

The discussions in Working Group III on the other hand shows that the uncertainties of the socio-economic impacts of a climate change, as well as the consequences of the measures required in order to slow down and ultimately stop such a change, present us with serious difficulties when trying to develop a strategy to deal with the greenhouse gas issue. A main reason is obviously the fact that we cannot as yet predict well enough how a change of climate would influence different countries directly. On the other hand, the implications, and particularly the costs, of the specific actions that are being considered to slow down or prevent a change of climate can be envisaged more easily by economists and politicians. It is not either certain that more stringent and therefore costly measures as yet would find approval amongst a broad majority of people even in countries with a population being reasonably well informed about environmental problems.

It is further clear that the rather immediate effects on the national economies of specific preventive actions, being considered, vary much from one country to another. At present fossil fuels account for 80 - 85% of the primary energy supply in the world, but different nations are very differently dependent on their use; for several countries the income from exports of fossil fuels represents an important component of their national economy; many developing countries have extensive plans for providing the increased energy needed for development by the use of fossil fuels; the infrastructure of most countries to a considerable extent has been based on transport systems using cars and trucks; deforestation is to a considerable part an expression for the need to expand agriculture or the dependence on forest products for export; measures to limit methane production in wet-land agriculture may imply major changes of traditional management practices with unpredictable consequences for the farmers involved.

As a matter of fact it is not surprising that controversies arise about as complex an issue as the present one, in particular when a more detailed understanding of the likely problems as yet is very incomplete. The major dilemma is of course, as has been said repeatedly, that actions may be required well before many of these specific issues have been resolved.

It is appropriate to recall in this context the lesson we learn from the well-known essay by Professor Hardin: "The tragedy of the commons". In the present context it can be formulated as follows: If individual nations can optimize their benefits by

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letting the compellers share their costs (i.e. pay for pollution) without they themselves taking the necessary precautions (reducing pollution), the economy of the total system becomes unstable and a catastrophic development may occur. We have barely learned how to deal with this problem at the national level when concerned with environmental threats, but we certainly have not as yet established adequate means for managing such problems internationally. We must also recognize the vast differences in economic strength between the players on the international scene, i.e. the nations and the large multinational companies in the world. How can a sustainable and improving future for the developing countries, i.e. for 3/4 of the world population, as well as for the industrial countries, be achieved under these circumstances?

In the light of the importance of understanding better the economics of the greenhouse gas issue the IPCC recommends that

enhanced resources are made available for intensified research on problems of international economy with particular regard to the emerging global environmental issues also recognizing especially the differences between developing and developed countries. It is further important to develop mechanisms, by which the international market can find optimal means to recognize the implicit environmental costs of human activities.

Working Group III also emphasizes in this context the importance of public information and education and IPCC supports the view that

because climate change would affect, either directly or indirectly, almost every sector of society, broad global understanding of the issue will facilitate the adoption and the implementation of such response options as deemed necessary and appropriate. Further efforts to achieve this are urgently needed

Steps towards action

Scenario A (Business as Usual) shows very clearly that a significant change of climate will be experienced a few decades into next century. If no preventive measures are taken. Working Group I points out that, the later such measures are initiated, the more stringent will they have to be in order to prevent a development that later may be found unacceptable. It can also be seen from the other scenarios (B, C and D), that global actions only slowly changes an ongoing development. This is also demonstrated by the fact that the establishment of new energy supply systems in the past (e.g. transition from coal to oil as the major primary source of energy) has taken three to five decades. In this perspective it is already late for actions that prevent a significant change of climate to occur. The IPCC concludes that

some change of the global climate cannot be avoided. In order to prevent a more far reaching change, steps towards adopting an action plan are urgently needed. These should be considered as precautionary measures, the costs of which should be viewed as insurance premiums and the magnitude of which will have to

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be decided on the basis of judgements of the evidence that is presently available.

Energy policy has so far almost exclusively been viewed as the policy required to secure the "necessary" or sufficient energy supply for the future. Although increased efficiency over the years has lead to the use of less energy per unit of production, the total use of energy has increased steadily in almost all countries. Energy has been comparatively cheap and there have been few incentives to save energy. It is clear from the Working Group III report that there are a number of actions that also would be beneficial for other environmental reasons and that would cost comparatively little, as a matter of fact in some cases rather imply savings. These now deserve attention. An action program to take advantage of such possibilities has been called a "no regret policy".

We should thus analyse more closely the "energy end-use efficiency" side of the problem. This is particularly important with regard to developing countries which now expand their energy supply systems and usually follow the pattern of development that the industrialized countries have already past. Much more efficient techniques are available that could be used rather than older and more wasteful technology.

The need that industrialized countries reduce their fossil fuel use substantially is partly hampered by the existing complex and energy demanding infrastructure of society which has developed, which people are very dependent upon for their basic needs and which is difficult and costly to change quickly. Still, there are very significant potentials for more efficient use of energy also in developed countries.

The IPCC concludes that

Increased end-use-efficiency should be aimed for in order to slow down the present rather rapid increase of fossil fuel use both in the developed and developing world.

In order to avoid a continued increase of the carbon dioxide emissions into the atmosphere we must necessarily invest in new non-fossil fuel supply systems, i.e. nuclear energy or/and renewable energy. We know from history that the characteristic introduction time for new supply systems is of the order of several decades to half a century and there are at present no economic incentives for the build up of such alternative energy systems. Plenty of fossil fuel is available at comparatively low costs.

It is, however, equally clear that there are a number of political issues that arise, particularly those related to the developing countries in the world. Major efforts to increase economic assistance and technology transfer to the developing countries will be required and it is in the interest of the industrialized countries that agreements of this kind can be reached.

Negotiations about a climatic convention and possible protocols must proceed in parallel with these research and development efforts. As a matter of fact some of the discussions within working group III can actually be viewed as explorations of the border line between a scientific-technical assessment and the phase of international negotiations. However, only by confronting the political issues directly

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can we can we penetrate the major difficulties that will be encountered at the international level. The IPCC considers it essential

that negotiations to develop a climate convention and possible associated protocols should begin soon with the purpose to define more clearly the key political issues that should be the subject for international agreement and to initiate further scientific and technical assessments as judged desirable in order to penetrate these more specifically,

that a precautionary principle is adopted to permit agreements on early actions that simultaneously would be beneficial for other national or international reasons and that would be of moderate cost, and

that technical development for the benefit of the climate issue and transfer of technology from developed to developing countries be given high priority

Global climate change and developing countries

A long term strategy

It should be recognized that the early steps for action that have been outlined above will not prevent a climate change from occurring, but slow down the rate of change. There is obviously a need for some long term strategy, but too many issues are still unresolved to aim for an early international agreement on a specific agreement of this kind. Some general aims could be included in a convention, but some time will be required to formulate and agree on a set of specific long term targets. The key question will of course be: How to develop a policy that in a balanced way considers both the long term environmental and economic consequences of a global change of climate, and on the other hand also the implications of major, internationally agreed actions to slow down and ultimately stop such a change. There obviously is a need to address the needs of individual nations, which is, however, not possible at present because of lack of knowledge both of the regional characteristics of a climate change and the future economical development in the world. Some provisional long term targets for the international cooperation would, however, be of great value.

The IPCC agrees that

it would be desirable, that in a second phase the forthcoming international negotiations could be devoted to the issue of the scientific and technical information that would be required in order to conduct meaningful negotiations on long term, internationally agreed targets for dealing with the climate change issue

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Global Climate Change and developing countries

(To be completed)

Future IPCC activities

(To be completed)

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The conduct of the IPCC assessment

The IPCC at its first meeting in November 1988 decided that a prime task for the Panel would be to prepare a comprehensive report on our present knowledge about possible future changes of climate due to mans activities on earth, particularly the emissions of greenhouse gases into the atmosphere. It was recognized that a major effort and considerable time would be required in order to do so and that a large number of scientists and other specialists would have to become engaged. On the other hand the discussion of the World Commission on Environment and Development Report "Our Common Future" in the UN General Assembly in 1987 had clearly shown that many nations would like to have an analysis of this issue as soon as possible. The IPCC therefore set the target to complete a first report already in 1990, but it was recognized that this would have to be an interim report. It was thought, however, that it still could help in providing a common basis for the discussions at the international level and thereby assist in attempts to arrive at some early agreements on first preventive actions if so would be judged as desirable. It is, however, essential that the work of the IPCC be continued in order to supply new information of importance in an orderly form for the development of long term policy with regard to the greenhouse issue.

Three Working Groups were formed:

Working Group 1 on the Scientific Assessment of Climate Change.

Working Group 2 on the Potential Environmental and Socio-economic Impacts of Climate Change.

Working Group 3 on Response Strategies.

At the second plenary session of the IPCC the Panel set up a

Special Committee on Participation of Developing Countries.

On the same occasion it was also decided that the first IPCC Report should be available for discussions in the UN General Assembly in the fall of 1990 and also

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serve as the basis for the Second World Climate Conference at about the same time, particularly for its second part conceived as a ministerial conference.

Due to the different characteristics of the tasks given to the three Working Groups and the Special Committee their work has progressed differently:

Working Group I selected a group of 33 lead authors for the eleven chapters planned for the report. They were chosen among the leading scientists in the fields to be covered and were given the task to prepare authoritative summaries of our present state of knowledge. In order to involve additional active researchers in the field altogether 11 specialist meetings were arranged for discussions of the subject matter for the different subsections of the report. The first drafts were subject to internal reviews and finally scrutinized at a lead authors meeting. The second draft of the report that emerged was circulated to a wide group of scientists, many of whom were chosen by participating nations in order to ascertain, that possible critical views from scientists outside the group of selected lead authors were received. More than 300 scientists became engaged in this way. A third, revised draft of the report was prepared and further discussed and amended before the Working Group meeting 23 - 25 May 1990.

Working Group II created seven subcommittees with two co chairmen for each such committee. First drafts were discussed in these committees and were revised accordingly. After further review the chairman and vice chairmen of the Working Group as well as the co-chairmen of the sub-committees met for a final drafting session, the outcome of which served as the input to the plenary meeting of the working group. Extensive comments were received and a revised draft was available for discussion at a plenary meeting of the Working Group 28 - 31 May 1990. (To be modified on the basis of information from Working Group II)

Working Group 3 formed four sub-groups, two each for adaptive and preventive measures respectively and also special committees, one for the formulation of alternative scenarios for future emissions and another for legal matters. Information about national plans for the future, particularly concerning the supply of energy, were collected. National comments on first drafts of the different chapters of the report and a series of sub-group meetings lead to the final draft of the working group report that served as a basis of its final plenary session. A large number of comments were submitted and a considerably modified version served as a basis for discussions at the final plenary meeting 5 - 8 June 1990. (To be modified on the basis of information from Working Group III)

The Special Committee started its work later than the three working groups. The key issues were discussed at two meetings and the final report of the committee was considered at a final meeting of the committee 2 - 3 June 1990.

All four group meetings have been open for participation by any country that has asked for association with the IPCC. The present overview and set of conclusions have been worked out by a drafting committee consisting of the chairman, vice chairman and rapporteur for the IPCC and the four chairmen and a few additional representative of the working groups and the special committee.