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Global Warming 1990 (1 of 2) [5]

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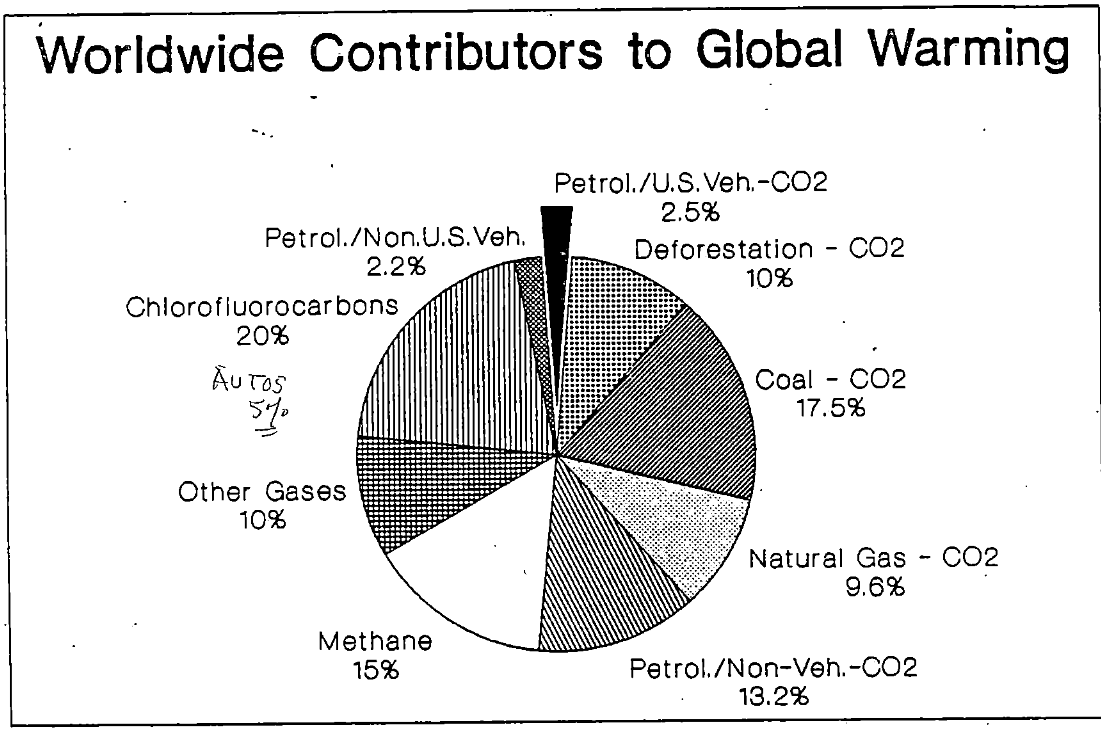
ELIMINATING CFCs OFFERS GREATEST POTENTIAL FOR GLOBAL WARMING



Air Cycle Air Cond

- Extensive research on global climate change is necessary to ensure definitive information on which to base U.S. policy directions.

  - U.S. policy should be developed in concert with international efforts because the best efforts of any one nation can be negated if all nations are not working to the same objective.

- Ford supports an aggressive plan to eliminate the use of chlorofluorocarbons (CFCs) in all applications -- foam, solvents, propellants and refrigerants.

- Elimination of CFCs is the most significant contribution the auto industry can make in addressing global warming concerns.

  - Ford is working to redesign its air conditioning systems to accommodate CFC substitutes as its suppliers make them available. System redesign includes a significantly larger condenser, a larger evaporator, and other vehicle design modifications to accommodate these larger components.

  - Suppliers must develop non-toxic substitutes, develop new lubricants, and build new facilities to provide adequate capacity.

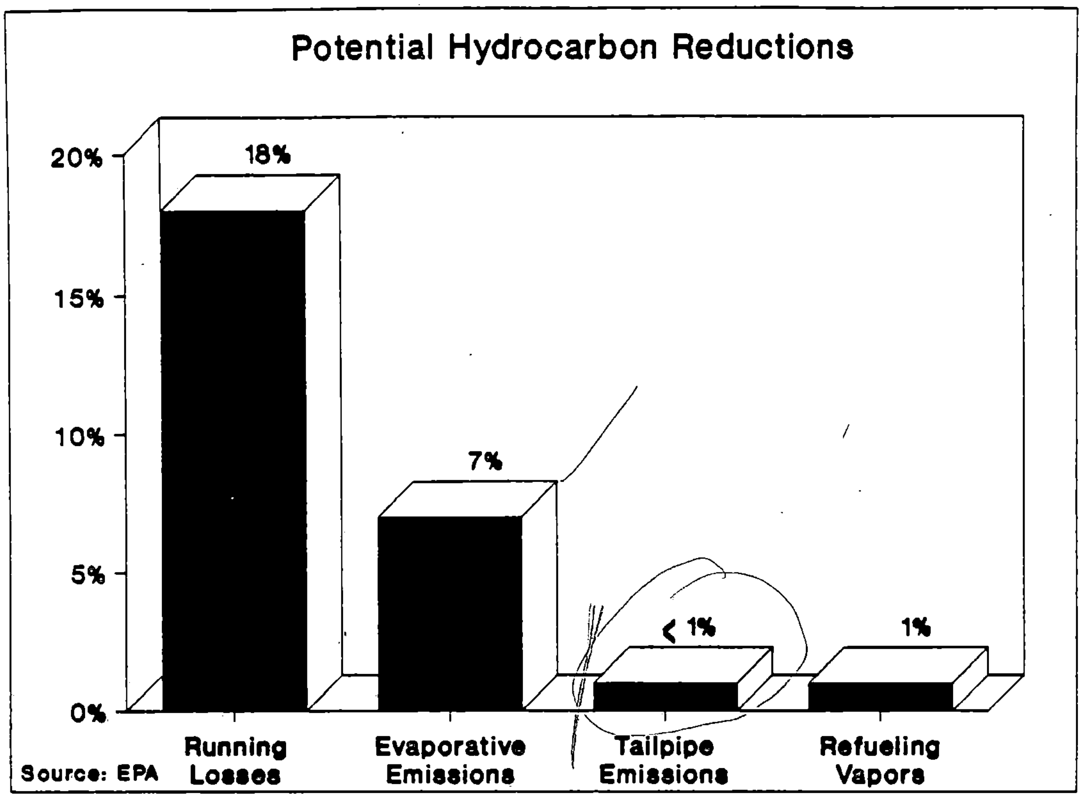
  - Ford also is investigating recycling methods to prevent CFCs from escaping into the atmosphere during servicing procedures at assembly plants and dealer service centers.

- Carbon dioxide emissions cannot be eliminated without abandoning fossil fuels as an energy source.

  - Because U.S. motor vehicle CO2 emissions account for only 2.5 percent of global warming gases worldwide, raising automotive CAFE standards would have a minimal impact.

  - Even doubling the fuel economy of the entire U.S. car fleet would reduce global warming gases by only 0.5 percent.

CONTROL OF EVAPORATIVE EMISSIONS AND RUNNING LOSSES OFFER THE GREATEST POTENTIAL FOR REDUCING AMBIENT HYDROCARBONS



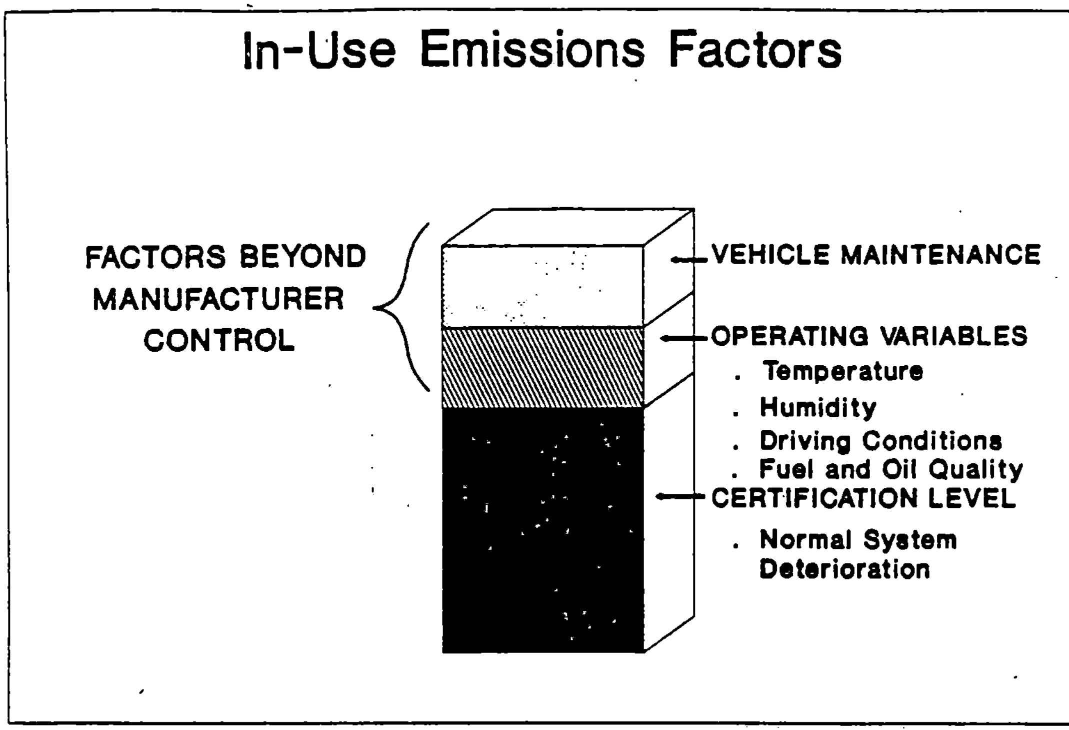
- Air quality will continue to improve as new vehicles replace older ones on the road today. By the year 2000, emissions of the average vehicle in service will be 50 percent less than today's for hydrocarbons and carbon monoxide and 33 percent lower for nitrogen oxides -- just by turning over the fleet.

- Reducing evaporative emissions and running losses are the most significant contributions the auto industry can make to improving air quality.

  - Ford supports efforts to redesign fuel and emissions control systems to reduce running losses and evaporative emissions. These actions will reduce fuel evaporation that occurs on hot days when ozone formation is likely to be highest. If the oil companies take simultaneous action to reduce gasoline volatility, total ambient hydrocarbons could be reduced by 25 percent.

  - By contrast, a tightening of exhaust hydrocarbon standards to 0.25 gpm or controlling refueling vapors either by onboard systems or Stage II service stations controls would only reduce fleet hydrocarbons by about 1 percent. This is because of the progress that has been made to date in reducing hydrocarbons from vehicle exhaust and the minimal contribution of refueling vapors.

FACTORS OUTSIDE A MANUFACTURER'S CONTROL REQUIRE SEPARATE IN-USE STANDARDS



- There are many factors outside a manufacturer's control that affect emissions performance, such as:

  - Operating Variables -- differences between controlled "certification" test and "in use" conditions (e.g., temperature, humidity, fuel and oil quality, and driving conditions).

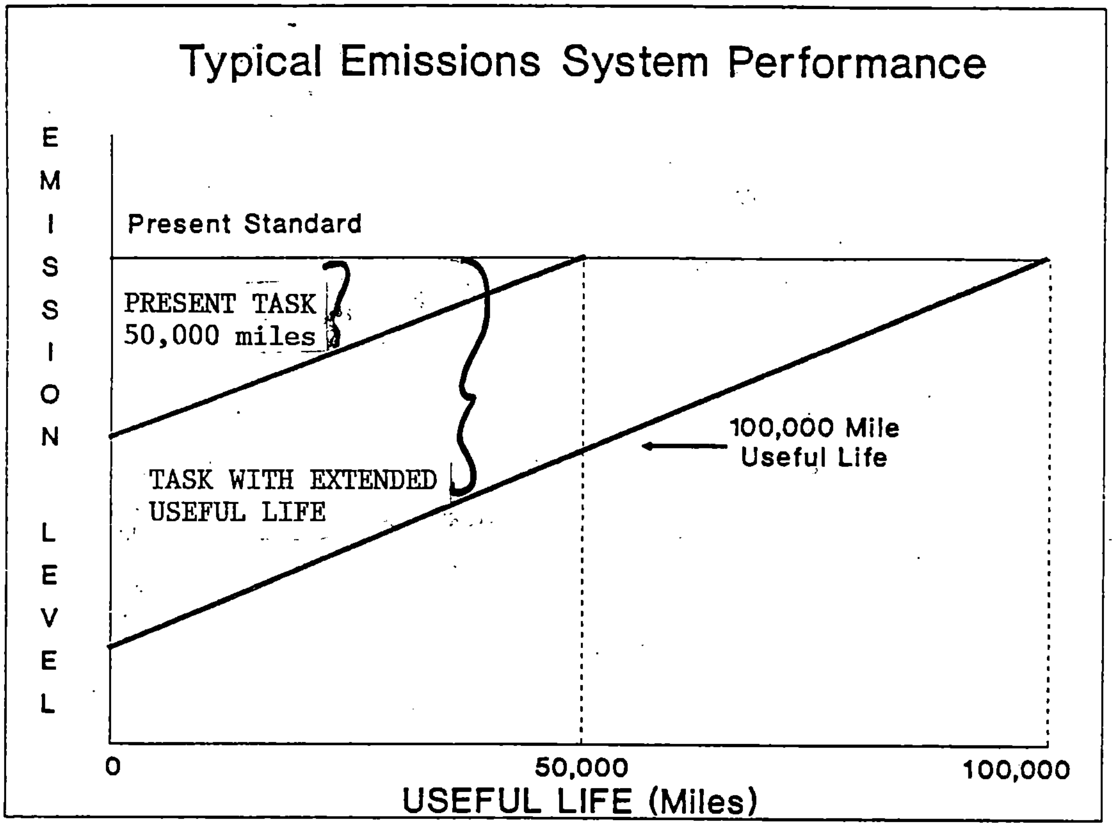
  - Vehicle Maintenance -- differences in how a vehicle is maintained and serviced throughout its lifetime. These factors have major impacts even if obvious tampering with emissions systems is excluded.

- Today the same emissions standard applies to a vehicle during the certification process and when the vehicle is in the customer's hands (in-use).

- If tighter tailpipe standards are adopted, there must be different standards for certification and in-use.

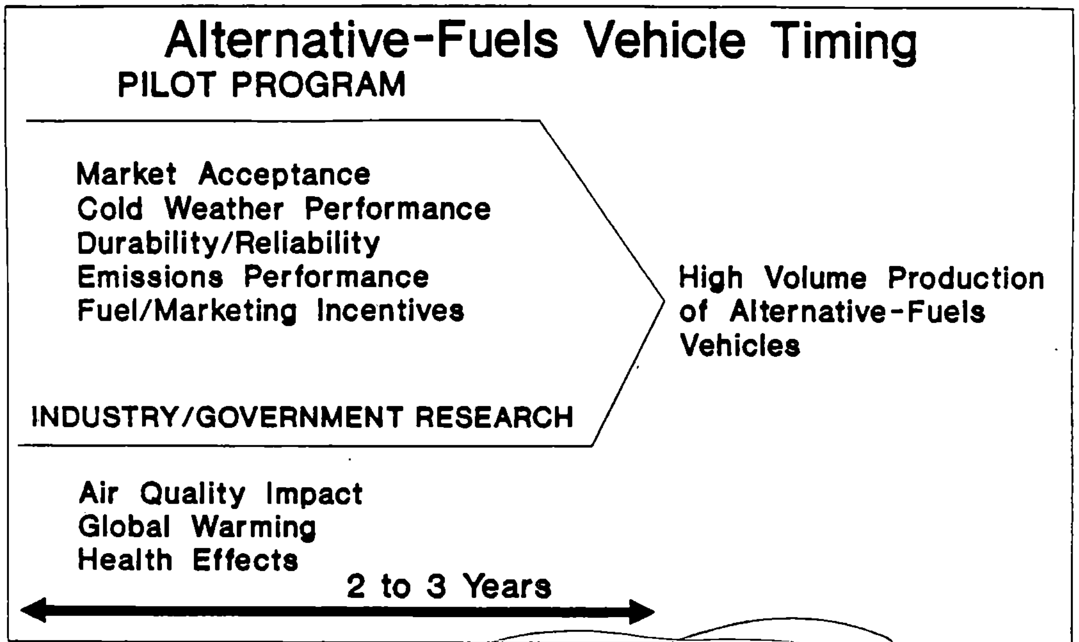
  - Aggregate automotive tailpipe emissions have already been reduced by 90 percent, making much more stringent in-use standards unproductive.

"USEFUL LIFE" REQUIREMENTS SHOULD NOT BE CHANGED



- To assure compliance for five years or 50,000 miles (today's requirement), manufacturers must design their vehicles to meet levels far more stringent than the standard.
  - This safety margin is necessary to take into account the normal deterioration of catalysts and other emission control components as well as operating and maintenance variables.
- Without any change in emissions standards, the task of meeting tailpipe standards would effectively double if "useful life" requirements were increased to 10 years or 100,000 miles -- which may be unachievable.
  - Meeting extended useful life requirements in combination with tighter tailpipe standards is unlikely to prove feasible, given the low emissions design levels manufacturers already meet today.

ALTERNATIVE FUELS COULD HELP CLEAN AIR, PETROLEUM CONSERVATION AND GLOBAL WARMING



- Alternative fuels -- such as ethanol and methanol -- promise benefits in the areas of clean air, petroleum conservation and global warming.

- Ford strongly supports continued development of alternative fuels and alternative-fuels vehicles.

  - For years, the problem in such development has been the "chicken and egg" dilemma of whether the vehicles or the fuel systems should come first.

  - But there is a technology that can resolve this dilemma. Flexible-fuel vehicles can operate on gasoline, ethanol, methanol or any combination of these fuels. A sensor automatically recognizes what fuel is being used without any action on the part of a driver.

  - Such vehicles permit drivers to use alternative fuels and make an environmental contribution in problem air quality areas such as Southern California -- but the driver can use gasoline when such fuels are not available on a cross-country trip.

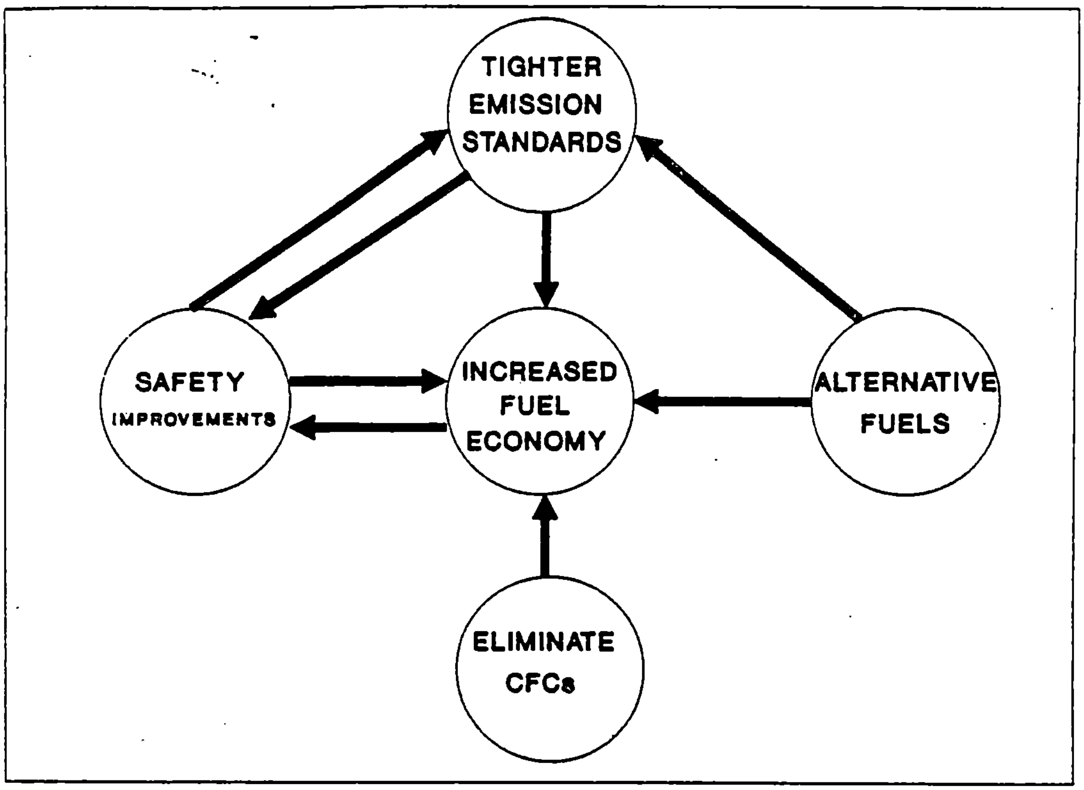
- An alternative-fuels pilot program should be established to evaluate whether the nation should commit the resources to convert the motor vehicle fleet to alternative fuels.

  - Such a program should include broad-based participation by governments, vehicle manufacturers, energy companies and users; and be focused on two or three key non-attainment areas.

  - It should provide a minimum of two to three years' experience that can yield answers to technical, market acceptance, fuel availability and economic issues.

  - It should be accompanied by joint industry/government research to resolve open scientific issues with regard to anticipated air quality benefits, the impact of such fuels on global warming and health effects.

AUTOMOTIVE POLICY ISSUES ARE INTERRELATED



- Ford supports programs that are both technologically feasible and resource-efficient and that will have a positive impact on clean air, global warming, fuel conservation and safety.

- Most of these goals are interrelated and some, in fact, conflict with one another. For example, Ford projects that its average fuel economy will decrease by more than 1 mpg as a result of tighter hydrocarbon and nitrogen oxide emissions requirements, elimination of chlorofluorocarbons (CFCs) and planned safety improvements.

- Recent and anticipated safety and emissions control actions could raise the cost of cars to consumers by $1,500 to $2,000. An increase of this magnitude would make all cars less affordable and is likely to encourage consumers to keep older vehicles that are less-fuel efficient and do not meet today's stringent auto emission standards.

- The nation's resources should be focused on the approaches with the most potential -- avoiding unrealistic timetables and conflicting requirements that could harm consumers and hinder manufacturers' ability to compete internationally.

CONTINUOUS FUEL ECONOMY IMPROVEMENT IS THE RIGHT APPROACH TO CONSERVATION

Fuel Economy Improvement

- **EQUAL PERCENTAGE TASK FOR ALL MANUFACTURERS**
- **INDEPENDENT OF SALES MIX**
- **LONG-TERM GOAL**
- **AUTO~~MATICALLY ADJUSTED~~ FOR OTHER REGULATIONS**

- Higher Corporate Average Fuel Economy (CAFE) standards have been proposed as an approach to clean air, global warming and petroleum conservation.

  - CAFE standards will not help in improving air quality -- small and large cars both meet the same emissions standards for each mile driven. They would provide minimal help on global warming because automobile CO_2 emissions are such a small contributor to global warming. 27.5

- Continuous improvement of the fuel economy of individual car and truck models is a better approach to petroleum conservation than CAFE.

- CAFE standards place U.S. full-line manufacturers at a competitive disadvantage with foreign manufacturers.

  - Domestic full-line producers must sell enough small cars -- the segment of the market where foreign manufacturers are most competitive -- to offset sales of family and full-size cars where domestic manufacturers are most competitive.

  - Domestic manufacturers must meet CAFE standards separately on vehicles with 75 percent North American content and those with less than 75 percent. This can lead to inefficient balancing actions that have no energy benefit. Foreign producers have only one fleet because they fall short of the 75 percent threshold even in transplant operations.

  - Foreign manufacturers that sell a large mix of small cars -- Japanese and Korean producers -- are able to sell increasing numbers of larger cars and high-performance engines and still meet CAFE. If foreign models substitute for domestic ones, there are no energy savings to the nation. These imports are no more fuel efficient than comparable domestic models -- for example, the all-new Nissan Infiniti has been reported as falling under the gas guzzler tax.

- CAFE standards do not necessarily reflect a manufacturer's technical effort because market changes can offset the application of new technology.

- To remedy these flaws in the CAFE concept, any new fuel economy targets should reflect equal percentage tasks that are independent of consumer purchasing patterns. It is also important that they be long-term goals, rather than arbitrary year-over-year requirements that can introduce inefficiencies, and that they be automatically adjusted for changes in other requirements that adversely affect fuel economy.

Recent Developments in the Science of Global Change

CEES

Working Group on Global Change

SCIENCE TRENDS - SUMMER 1990

Committee
On
Earth and Environmental Sciences

Federal Coordinating Council for Science, Engineering, and Technology

August 1990

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Talking Points
for
Science Briefing on Global Change

**KEY SCIENCE ISSUES
(AUGUST 1990)**

- o **GREENHOUSE GASES--SOURCES AND SINKS**
    - **CARBON SEQUESTRATION (CARBON DIOXIDE)**
        - * **OCEAN VS TERRESTRIAL BIOSPHERE**
    - **METHANE**
- o **ROLE OF OCEAN IN CLIMATE**
    - **EL NINO/SOUTHERN OSCILLATION--INTERANNUAL TIMESCALES**
    - **AIR-SEA COUPLING--DECADAL TIMESCALES**
        - * **NORTH ATLANTIC--THERMOHALINE FORCING**
        - * **ANTARCTIC HEAT UPTAKE**

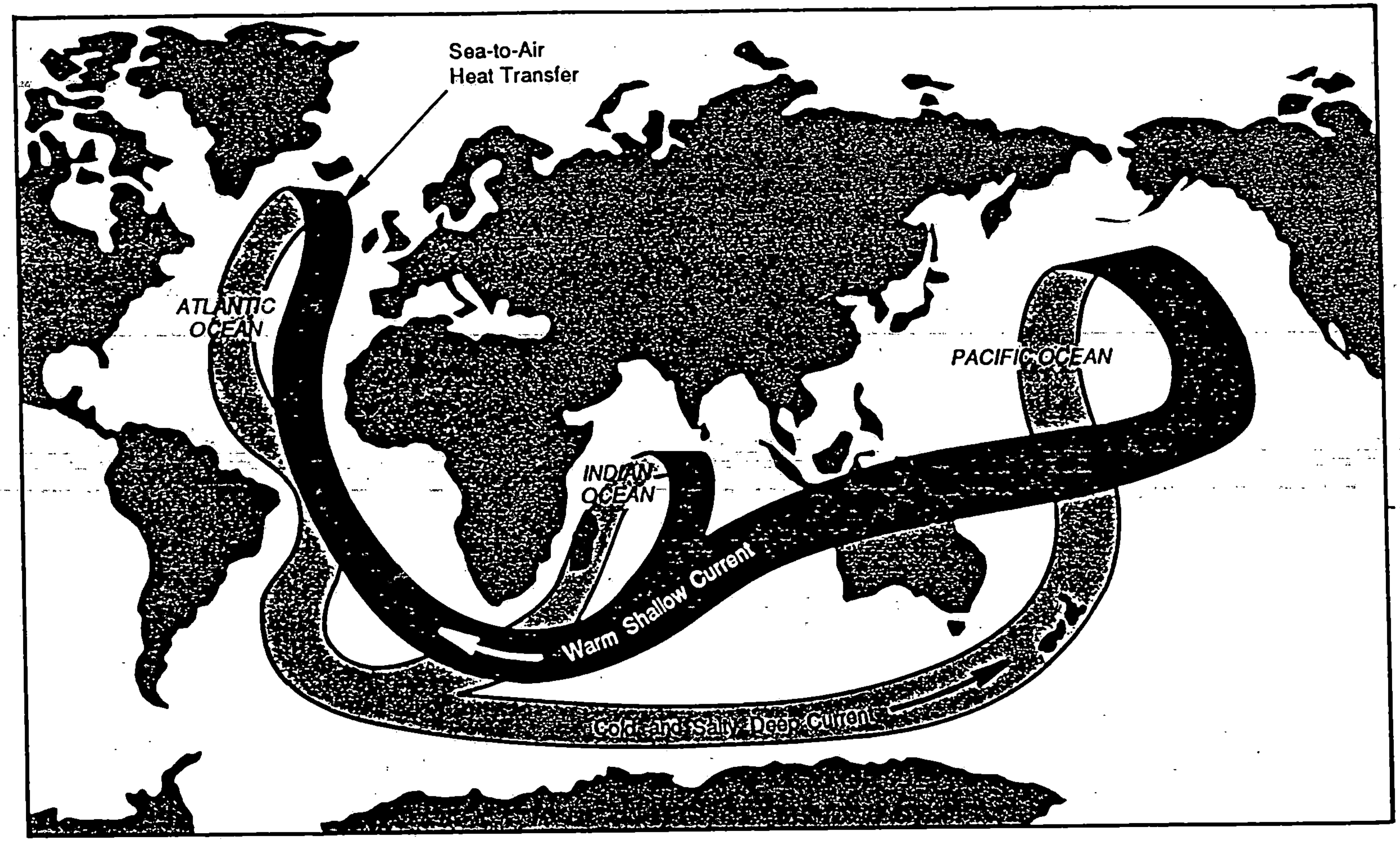
- EARTH SYSTEM MODELLING

  - CLOUDS/RADIATION BALANCE/LIQUID WATER
  - VALIDATION OF QUANTITATIVE OCEAN MODELS
  - REGIONAL MODELS/PREDICTION--ENSO READY TO GO
  - ROLE OF LAND SURFACE IN TRACE GASES, ENERGY AND WATER EXCHANGE
  - SNOWMASS INSTITUTE--FULLY-COUPLED SIMPLE MODELS AVAILABLE IN 4-5 YEARS

- TRENDS

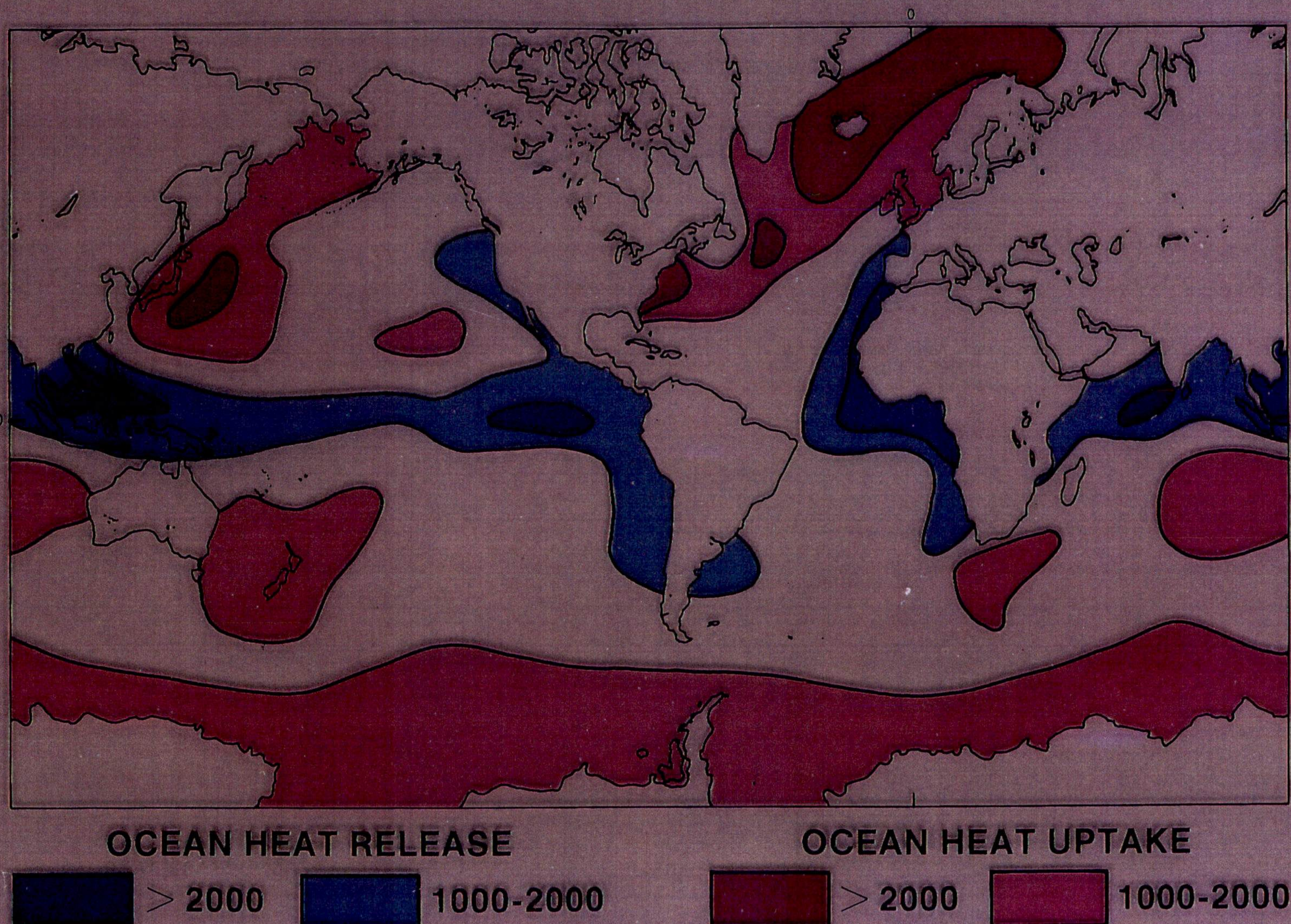
  - LONG-TERM VARIABILITY IN TEMPERATURE WELL-DOCUMENTED
  - OBSERVED INCREASE IN TEMPERATURE OVER LAST ONE HUNDRED YEARS; CAUSE AND EFFECT NOT ESTABLISHED
  - DETECTION OF PREDICTED ANTHROPOGENIC WARMING WILL TAKE 10 TO 50 YEARS
  - SEA LEVEL RISE--ROLE OF POLAR ICE SHEETS UNCERTAIN
  - RESPECTIVE CONTRIBUTIONS OF DEFORESTATION--TROPICAL AND MID-LATITUDE UNCERTAIN

- **RESEARCH PRIORITIES**
  - **CARBON CYCLE**
  - **GLOBAL WATER AND ENERGY CYCLES**
  - **GLOBAL MODELS WITH REGIONAL RESOLUTION**
  - **ECOSYSTEMS AND POPULATION DYNAMICS**

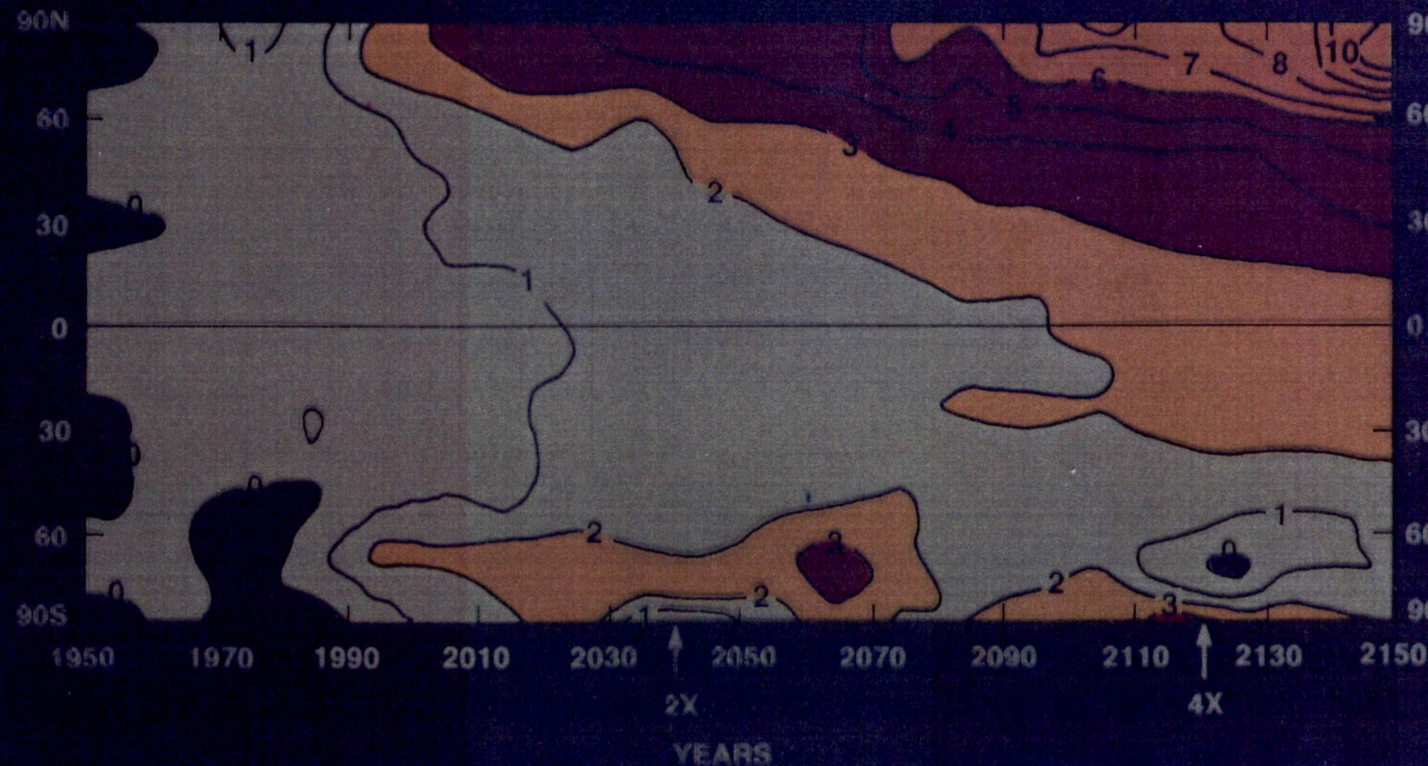


Sea-to-Air
Heat Transfer
ATLANTIC
OCEAN
PACIFIC OCEAN
INDIAN
OCEAN
Warm Shallow Current
Cold and Salty Deep Current

(JOULES/M²/YEAR)



Southern Hemisphere Resistance to Temperature Change (°C)



(CO_2 Increasing 1% per Year)

STUDIES IN GEOPHYSICS

Sea-Level Change

Geophysics Study Committee
Commission on Physical Sciences, Mathematics,
and Resources
National Research Council

NATIONAL ACADEMY PRESS
Washington, D.C. 1990

Overview and Recommendations

EXECUTIVE SUMMARY

This study addresses current scientific understanding of sea-level change—particularly the processes of sea-level change, their rates, and the record of past change. An important part of such an assessment is an evaluation of the adequacy of the geophysical knowledge base and the opportunities to improve upon it. Discussion of engineering and societal responses to sea-level change is not included in this study as these issues are fully discussed in a report, *Responding to Changes in Sea Level: Engineering Implications* (NRC, 1987).

Average sea level over the oceans has never been constant throughout earth history, and it is changing slightly today. The entirety of civilization has occurred within a single high stand of the sea, and yet global sea level was more than 100 m lower than it is at present only 18,000 yr ago. And during the geologic past there have been repeated variations of more than 100 m from present sea level, both during times of intense glaciation and during times of an ice-free Earth.

Relative sea level (RSL; i.e., sea level relative to a fixed point on land) at any particular place in the ocean varies over a wide range of time and space scales. The direct causes of these variations are vertical motions of the land to which the tide gauge or other measuring instrument is attached, and changes in the volume of sea water in which the tide gauge is immersed. But changes in climate, plate tectonics, ice and snow, and ocean circulation are all indirect causes of changing sea level. The relative importance of the forcing functions varies with the time scales of interest.

On the basis of estimates of global warming of the atmosphere and ocean resulting from increasing concentrations of carbon dioxide and other greenhouse gases, it is possible to make an approximate forecast of global rise in sea level during the next 100 yr. Two processes will be principally involved: thermal expansion of ocean waters as they become warmer and changes in the mass of land ice in both continental ice sheets and mountain glaciers. One hundred years from now it is likely that sea level will be 0.5 to 1 m higher than it is at present.

There are two principal uncertainties at present about global sea level. (1) What if any is the value of the long-term trend over the next few centuries? (2) If a secular trend exists, what proportion of this trend results from changes in the specific volume of sea water (called steric changes) and what from changes in the total mass of water in the oceans?

The apparent trend of sea level at any particular place as measured by a tide gauge is the sum of the trend in motion of the gauge itself as the land on which it is mounted moves vertically, the trend of change in steric sea level, and the trend of change in water mass under the tide gauge. To understand what is happening, one needs to be able to make measurements that will separate these three components of the observed sea level. In principle, a combination of inverted echo sounders (which in effect measure thermal expansion or contraction of the water column) or systematic observations of ocean temperature as a function of depth with conductivity-temperature-depth recorders to determine the steric component, plus one or more of three methods (very-long-baseline interferometry, global positioning system, and absolute gravimetry) for measuring the vertical motions of the tide gauge, plus tide-gauge measurements at the sea surface should allow us to separate the three major components of changes in RSL. Within the next few years, it should be possible to measure accurately the combined effects on global sea level of steric changes and changes in the mass of sea water from laser or radar altimeters mounted on Earth-orbiting satellites.

Recommendation 1. Long-term sea-level measurements of sufficient accuracy over the world's oceans could provide one of the most significant data sets for understanding global change, particularly climatic change resulting from greenhouse warming. It is for this reason that the planning committees for the World Climate Research Program and the Intergovernmental Oceanographic Commission of UNESCO have given a very high priority to extending the global sea-level network in the Indian, South Atlantic, and South Pacific oceans. **We strongly recommend that national oceanographic and meteorological communities lend moral and intellectual support to this sea-level program.**

Recommendation 2. Possible changes in the mass balance of the Antarctic and Greenland ice sheets are fundamental gaps in our understanding and are crucial to the quantification and refinement of sea-level forecasts (the probable contribution from ice wastage makes up more than half of various forecasts). **A polar-orbiting satellite altimeter would be invaluable in monitoring the mass balance of these ice sheets.**

Recommendation 3. To refine estimates of sea-level change related to greenhouse warming, it is necessary **to develop and improve coupled atmosphere–ocean–cyrosphere global circulation models in which greenhouse gas concentrations in the atmosphere are gradually increased.**

Recommendation 4. The Cretaceous period offers special opportunities to understand global processes and their variations, in particular, large, long-term changes in sea level. One of the major projects of the Global Sedimentary Geology Program, under the auspices of the International Union of Geological Sciences, is entitled "Cretaceous Resources, Events, and Rhythms." **We urge national and international support of this and similar programs that will improve our understanding of past sea-level changes and the processes that produced them.**

Recommendation 5. To separate epeirogeny from eustasy and steric components, it is important to measure repeatedly the absolute heights of tide gauges. **We recommend that global measurements of absolute heights of these gauges be undertaken using absolute gravimetry and space-based techniques.**

INTRODUCTION

It is easy to think of sea level as a stable baseline against which changes on land can be measured, and it is so used by politicians, engineers, land planners, and households. But as we have become more aware of the Earth and its metabolism, we have recognized that sea level is highly unstable both in time and in space. We can regain our former confidence in its usefulness only by learning how and why it varies and compensating for these variations.

Average sea level over the globe has never been constant throughout earth history; and it is changing slightly today. The entirety of civilization has occurred within a single high stand of the sea, and yet global sea level was more than 100 m lower than it is at present during the maximum of the last glacial period only 18,000 yr ago (yrBP).

Relative sea level (RSL; i.e., sea level relative to a fixed point on land) at any particular place in the ocean varies over a wide range of time and space scales. Among the causes of these variations are vertical motions of the land to which the tide gauge or other measuring instrument is attached and changes in the volume of sea water in which the tide gauge is immersed. Wind-driven waves produce the shortest-period variations in the height of the sea surface. Variations with periods of 12 hours or more are caused by lunar and solar tides. Variations in atmospheric pressure cause inverse variations in sea level—an atmospheric pressure differential of 1 mbar is equivalent to a sea-level differential of 10 mm. A series of depressions in atmospheric pressure can cause a rise in sea level in a shallow ocean basin of 0.3 m or more (Hekstra, 1988). Variations in the runoff of large rivers can result in local sea-level changes of as much as 1 m. In relatively shallow water, large variations in sea level are also caused by offshore and onshore winds that pile up water against the shore or drive it away from the shore. (This process is called wind setup.) In exceptional circumstances, in the North Sea, along the Chinese coast, and in the Bay of Bengal, sea level may rise by 5 m or more in a "storm surge" under the action of strong winds (Hekstra, 1988). Both irregular and seasonal variations in temperature or salinity of the upper ocean layers cause expansion or contraction of the water volume in different regions. These relatively short-term steric changes in sea level may persist for a few days, several months, or even several years, and the magnitude may be as much as 50 to 150 mm.

Changes in sea level have many practical consequences, often disruptive but sometimes beneficial. The disruptive consequences can sometimes be avoided and the benefits enhanced if the ways and means by which sea level varies are understood. Although disruptive consequences, especially from sea-level rise, are far more common, occasional examples of benefits are being obtained as a result of changes in policy. For example, in the Wadden Sea on the northwest coast of the Netherlands, where the land has subsided as much as 260 mm during the past 20 yr, the Dutch government has decided not to follow age-old tradition by attempting to reclaim more land for agriculture. The area has been designated as a "Declared UNESCO Biosphere Reserve" in which only those economic activities are allowed that do not conflict with natural conditions or processes. In view of agricultural surpluses in the Netherlands, and indeed throughout the European community, the Wadden Sea can be used much more effectively as a nursery for young fish and shrimp and other valuable invertebrates (Hekstra, 1988) than as reclaimed land for agriculture.

Changes of sea level on the order of 300 mm can have significant implications for coastal communities and coastal engineering practices. Engineering responses to sea-level change are largely a function of the rate of change. Many of these issues and engineering responses are described in the report *Responding to Changes in Sea Level: Engineering Implications* prepared by the Marine Board of the National Research Council (NRC, 1987). A 1-m change in average sea level can translate into major shifts in shoreline positions, positions that have both economic and legal significance.

Sea-level change, seemingly so simple and straightforward, is in fact the product of many interrelated processes. Insight into these processes can be gained by intensive study of sea-level change in the context of related environmental phenomena, remembering that changes in sea level are an integrated measure of environmental change, in terms of both causes and consequences. Changes occur on all space and time scales, from local to global, and from a few seconds to geologic ages.

This volume is primarily concerned with future sea-level changes over the next few centuries and past changes over a wide range of times from which greater understanding can be gained and used as an aid in the prediction of future changes and in the search for fossil fuels and other natural resources. Also covered are the mechanisms and processes involved in past changes, in order to gain greater knowledge of the Earth as a dynamic system, i.e., how the Earth works.

Climate, plate tectonics, the cryosphere, and ocean circulation all contribute to changing sea level. The relative importance of the forcing functions varies with the time scales of interest. The effects of changing sea level are also broad—on the one hand with direct feedbacks to the causative forcing functions, e.g., albedo change, and on the other hand with effects on other processes such as sedimentation or coastal ecology.

PROCESSES AND FEEDBACKS

Many processes can cause a change in RSL at any particular location. They include the following:

1. local or regional uplift or subsidence of the land;
2. changes in atmospheric pressure, winds, or ocean currents;
3. changes in the mass of ocean water brought about by wastage or accumulation of ice sheets and mountain glaciers (glacio-eustatic) or by increased or decreased retention of liquid water within or upon the continents, and possibly also by a slow release through geologic time of juvenile water from the Earth's interior;
4. steric changes in the volume of ocean water without changes in water mass (Patullo *et al.*, 1955) in response to temperature or salinity changes (also called thermohaline changes in Table 1); and
5. changes in the volume of the ocean basins owing to changes in the rate of plate divergence (seafloor spreading), plate convergence (subduction, overthrusting), epeirogenic changes in the elevation of the seafloor (largely from mid-ocean volcanism), marine sedimentation, or isostatic adjustment of the Earth's crust under the sea resulting from glaciation or deglaciation on land.

The latter three processes can affect global mean sea level or eustatic sea level. But all processes need to be considered, even though, depending on the time scale of interest or the magnitude of the sea-level change, some of them may be insignificant. In a following section on forecasting sea-level change due to greenhouse-induced climate warming, a projected global sea-level rise of 0.5 ± 1 m by the year A.D. 2100 is ascribed to a combination of thermal expansion of ocean water and melting of glaciers and ice sheets.

Time Scales of Sea-Level Change

Sea-level change encompasses a broad range of time scales, with different mechanisms associated with change over different times. The oceanographer concerned with storm tides will not have much interest in the factors explaining Cretaceous sea levels; likewise, the geologist's glacio-eustatic theories have little application to seasonal events. The problem of sorting out time scales and processes afflicts studies of climate change in general. Complexities multiply in attempts to link different processes together.

Table 1 summarizes mechanisms of sea-level change by time scale and magnitude.

TABLE 1 Some Mechanisms of Sea-Level Change

	Time Scale (years)	Order of Magnitude of Change (mm)
Ocean Steric (thermohaline) Volume Changes		
Shallow (0 to 500 m)	10^{-1} to 10^{2}	10^{0} to 10^{3}
Deep (500 to 4000 m)	10^{1} to 10^{4}	10^{0} to 10^{4}
Glacial Accretion and Wastage		
Mountain Glaciers	10^{1} to 10^{2}	10^{1} to 10^{3}
Greenland Ice Sheet	10^{2} to 10^{5}	10^{1} to 10^{4}
East Antarctic Ice Sheet	10^{3} to 10^{5}	10^{4} to 10^{5}
West Antarctic Ice Sheet	10^{2} to 10^{4}	10^{3} to 10^{4}
Liquid Water on Land		
Groundwater Aquifers	10^{2} to 10^{5}	10^{2} to 10^{4}
Lakes and Reservoirs	10^{2} to 10^{5}	10^{0} to 10^{2}
Crustal Deformation		
Lithosphere Formation and Subduction	10^{5} to 10^{8}	10^{3} to 10^{5}
Glacial Isostatic Rebound	10^{2} to 10^{4}	10^{2} to 10^{4}
Continental Collision	10^{5} to 10^{8}	10^{4} to 10^{5}
Sea Floor and Continental Epeirogeny	10^{5} to 10^{8}	10^{4} to 10^{5}
Sedimentation	10^{4} to 10^{8}	10^{3} to 10^{5}

Heat exchange is relatively rapid within the uppermost few hundred meters of the oceans. A one-dimensional treatment implies that thermal expansion of these waters can occur on time scales of months to decades. Sea level will rise about 100 mm for every degree of temperature increase throughout the uppermost 500 m. Heat exchange with ocean deep waters is slower (Chapter 13). If the deep ocean were to warm everywhere by 10°C, as was perhaps the case during the early Tertiary and Cretaceous, sea level could rise by about 10 m.

The time scales and magnitudes of melting ice can be estimated from both historical data and mass balance considerations. The present Greenland and Antarctic ice caps are remnants of the late Pleistocene ice sheets that increased sea levels about 100 m by disintegrating over a period of several thousand years encompassing the end of the Pleistocene (Chapters 4 and 5). Mass balance estimates suggest modern ice residence times on the order of 10^2 to 10^5 yr. The Antarctic Ice Sheet contains enough water to raise sea level by about 60 m, and the Greenland Ice Sheet contains water equivalent to a 6-m sea-level rise.

Sea Level and the Geoid

The sea surface departs significantly from the geoid, owing to waves, the tidal attraction of the Sun and Moon, ocean circulation that tilts the ocean surface, atmospheric disturbances, and steric regional variations in water temperature and salinity. If these effects can be taken into account, the resulting mean sea level accurately follows the geoid, which however is far from being an ideal spheroidal shape. Deep mantle phenomena and gravity anomalies associated with subduction can cause deviations from the spheroid of many tens of meters. Local crustal geological features, such as seamounts, fracture zones, and other abrupt topographic forms that are not in isostatic equilibrium, can cause deviations of several meters. In fact, it is possible to derive a partial picture of ocean floor topography from a knowledge of the mean shape of the sea surface.

If the effects of vertical tectonic movements, in addition to the local or regional variations of sea-surface topography, are removed from tide-gauge records, presumably any remaining trend is global (eustatic sea-level change). Eustatic changes can result from processes that change the mass or volume of water in the ocean basin or those that change the volume of the ocean basin itself. These processes are discussed in later sections.

Land Elevation Changes

Tide gauges are anchored to the land, which itself can be moving vertically at rates comparable to sea-level change. These vertical tectonic movements of the land will result in an RSL change as measured by a tide gauge. If the land where the tide-gauge station is located is subsiding, RSL will show a rise; likewise, uplift will result in an RSL fall. The Viking city now called Old Uppsala was a major port for ships sailing Lake Malaren; however, the port and the city of Uppsala were forced to move downstream around the eleventh century because of a 10-mm/yr uplift in the Fennoscandian region while the lake remained connected with the worldwide mean sea level.

Vertical tectonic motions can be very localized, such as subsidence along a coastline owing to sediment load or from the withdrawal of groundwater or hydrocarbons. Subsidence is occurring along the Louisiana gulf coast because of the deposition and dewatering of sediment from the Mississippi River system. On the other hand, vertical motions can be systematically related to one another through isostasy.

Land areas that were ice covered during the last glacial episode (~18,000 yrBP) were depressed; this depression created a forebulge in adjacent areas (see Peltier, Chapter 4). The North American or Laurentide ice sheets began disintegrating about 15,000 yrBP and by about 7000 yrBP had all but vanished. Along the east coast of North America, those areas that were ice covered are still being uplifted due to isostatic rebound at rates of up to about 10 mm/yr. In the peripheral area where the forebulge is collapsing, the land is subsiding at more than 1 mm/yr. The east coast of the United States is perhaps the best location illustrating this "drowning" effect, which is responsible for many of the unique features of its nearshore environment, including the extensive occurrence of salt marshes. In these areas and with current models of isostasy and the Earth, the relative vertical motions are predictable; thus, their relative contribution to sea-level changes as measured by tide gauges can be taken into account in arriving at eustatic sea-level changes.

Effects of Atmospheric Pressure, Winds, and Ocean Currents

Local RSL variability can result from several forcing functions including air pressure, wind stress, ocean circulation, and thermohaline changes. These processes contribute to the high-frequency noise in tide-gauge data and need to be compensated for in extracting eustatic changes.

The difference between the average air pressure over the world ocean and the local barometric pressure can result in a change of sea level at annual and shorter periods. One millibar of pressure differential is equivalent to 10 mm of sea-surface change.

Wind stress can have an important effect on the sea level. The coastal sea-level response to a steady longshore wind stress can be similar in magnitude to the air-pressure effect. The time to achieve a steady state is of the order of a few days. Sea-surface changes caused by wind stress are localized and are not a major contributor to low-frequency sea-level change.

Ocean circulation can also result in sea-level variability from long-period waves, as Sturges (Chapter 3) shows. Sea-level signals of about 50 to 150 mm with periods of 5 to

10 yr and longer are coherent between the U.S. Pacific coast and Hawaii, and on both sides of the Atlantic. These signals are out of phase, with delays of several years, and are consistent in part with baroclinic wave propagation across the ocean.

Also coherent are sea-surface variabilities resulting from El Niño and related effects (Figure 1). From these, sea level can vary on the time scale of a few years by 50 to 150 mm resulting from warming of the ocean subsurface waters by a few degrees Centigrade (°C).

Changes in the Mass of Ocean Water

For all intents and purposes, the mass of water at or near the Earth's surface is constant over time scales of less than 10^4 yr. It is how the water is partitioned between the major hydrologic reservoirs that is of importance to sea-level change. The four major reservoirs, in order of abundance, are the oceans (1370×10^6 km^3), ice (30×10^6 km^3), ground and surface waters (8×10^6 to 19×10^6 km^3), and atmospheric moisture (0.01×10^6 km^3). The principal exchange of water over the past several million years involved ice. Sea level was over 100 m lower during the peak of the most recent glacial 18,000 yrBP. The melting of the northern continental ice sheets between 15,000 and 7000 yrBP probably accounted for most of the rise of the sea to present levels. Indeed, sea-level change within the next few thousand years probably will also be dominated by water within the global ice budget and by ice sheet dynamics. If the polar ice sheets were to disappear, sea level would be some 60 to 70 m higher than at present.

Mountain glaciers make up about 1 percent of the volume of land ice. Their potential contribution to sea level is about 1 m if they were to melt totally and all the meltwater reached the sea. Data cited by Meier (Chapter 10) suggests that wastage of the world's

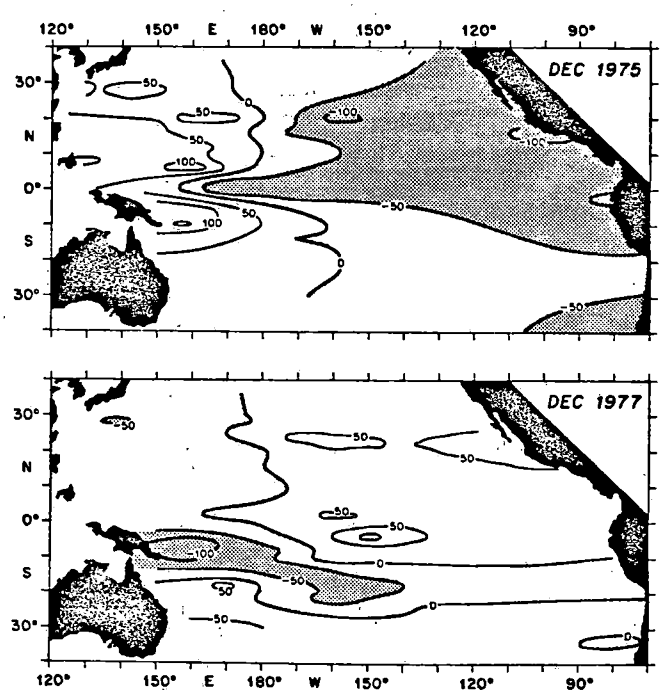


FIGURE 1 Maps of sea-level anomaly for December 1975 and December 1977. Contours show sea-level anomalies in millimeters after removal of seasonal cycle. The two cases were selected for their large contrast. From Wyrtki and Nakahoro (1984).

mountain glaciers and small ice caps contributed about 0.46 ± 0.26 mm/yr to higher sea level between 1900 and 1961 corresponding to a total sea-level rise of 28 ± 16 mm, which is about a third of the estimated sea-level rise during that period.

The Antarctic and Greenland ice sheets gain material mainly through the accumulation of snow and lose material through several processes. These processes include surface melt and runoff of meltwater, calving (discharge) of icebergs, and melting of the underside of floating ice shelves. Surface melt/runoff is a minor process for the Antarctic Ice Sheet, but is important to the balance of the Greenland Ice Sheet. Iceberg calving is an important loss process for both ice sheets, and is predominant in Antarctica. Melting of the underside of ice shelves has no effect on sea level, but does control the speed of discharge of land-based ice from Antarctica and is important for predicting the behavior of that ice sheet in the next century.

The increase in air temperature caused by a rise in concentration of CO_2 and other greenhouse gases may cause increased snow and ice melting, and some of this meltwater may run off to the oceans, causing a rise in sea level. Increased air temperature and/or meltwater production may also cause some outlet glaciers and ice streams to flow faster, transferring land-based ice to the ocean and causing a further rise in sea level. On the other hand, a rise in CO_2 concentration may, in some regions, lead to increased snow precipitation on glaciers and ice sheets, which will have the opposite effect on sea level. Predicting the effects of climate change on ice growth and wastage is a complex problem because several different, interacting processes must be considered.

Five important factors in future changes of glaciers and ice sheets are (1) the variation of energy and mass balance components with altitude, (2) the warming of cold firn to allow meltwater runoff, (3) the dynamic response of ice masses to changes in thickness, (4) increased flow and iceberg calving of tidal glaciers due to increased meltwater, and (5) the stability of ice-sheet/ice-stream/ice-shelf systems. The time frame is restricted to the next 100 yr, approximately the time of doubling of the present level of CO_2. The first and fifth points are briefly discussed below; see Chapter 10 for additional details on all five processes.

For glaciers, many of the mass and energy fluxes related to melting are known or observable functions of altitude. The two most sensitive of these are absorbed solar radiation and precipitation of snow. The first is critically dependent on the albedo (reflectivity) of the surface, which in turn depends on how long the surface is covered with high-albedo snow during the course of the melt season. The persistence of the snow cover depends, of course, on the amount of snowfall and the intensity of melt processes, both of which also depend on altitude. The altitudinal dependence of the mass and energy fluxes leads to a potential instability, which is one of the reasons for the sensitivity of glaciers to slight climate changes. An increase in melting causes a lowering of the ice surface, which in turn may cause a further increase in melting or decrease in snow accumulation, leading to further changes accentuating the melting.

Some have suggested that a climatic change due to increased CO_2 in the atmosphere could lead to disintegration of the West Antarctic Ice Sheet, most of which is grounded below sea level, causing a 6-m rise in global sea level. The discharge of ice from this ice sheet is mainly through rapidly moving ice streams, which flow into floating ice shelves. An ice sheet that rests on a flat bed situated below sea level can be inherently unstable. Floating ice shelves at its seaward edge, which are "pinned" in position by shallow bottom areas, act as buttresses that prevent the ice sheet from quickly flowing out into the ocean. The rise of 100 m in sea level during the past 15,000 yr caused a substantial collapse of a large part of the West Antarctic Ice Sheet until it approximately stabilized at its present size. A relatively small further sea-level rise could act to "unpin" the buttressing ice shelves that allow the remnant ice sheet to exist (Thomas and Bentley, 1978).

If the climate in the future becomes warmer with the result that warmer ocean water intrudes under the ice shelves causing increased melting under the shelves, then the back pressure exerted on the ice streams by the shelves will be reduced and the ice streams will accelerate, draining the ice sheet itself.

The critical questions then are: (1) How rapidly will the temperature of Antarctic subsurface waters rise in response to increased atmospheric concentrations of greenhouse gases? (2) How much sub-ice melting will be caused by the circulation of this warmer water under the ice shelves? (3) How will the changed conditions affect calving rates and thus the dimensions of ice shelves? (4) How rapidly will the ice streams react to changes in the back pressure? For the next one or two centuries we need to consider only the first question because warming of the ocean south of the Antarctic convergence is likely to be markedly delayed. Bryan *et al.* (1988) used a combined oceanic and atmospheric general circulation model to show that convective mixing from the surface down to 4000 m will slow the rate of ocean warming because the entire water column must be warmed by the same amount. If we consider the added heat energy of around 4 watts/m^2 transferred from the atmosphere to the surface ocean layers, a time of several hundred years would be required to accomplish this warming. Bentley (1985) has summarized a number of other reasons why disintegration of the West Antarctic Ice Sheet should not occur within the next one or two centuries. Accepting Bentley's arguments, sea level will not rise catastrophically in the near future resulting from the demise of this ice sheet.

Scientific concern about possible disappearance of the West Antarctic Ice Sheet has largely been based on Mercer's (1978) hypothesis that this body of ice disappeared during the last interglacial, 125,000 yrBP, with the result that a terrace about 5 m above present sea level was created around many shorelines around the world. An alternative hypothesis is that the surface of the East Antarctic Ice Sheet was lower by some 300 to 350 m than today (Robin, 1987). This idea is supported by the investigation by Lorius *et al.* (1985) of $\delta^{18}O$ from the ice core collected by Soviet engineers at Vostok in the East Antarctic Ice Sheet. At 125,000 yrBP, the oxygen isotope values were about 2 ‰ higher than present. Robin points out that this apparently higher temperature could be caused by a reduction in surface elevation of about 300 m. Surface lowering of this amount for the East Antarctic Ice Sheet would correspond to a volume of ice of about 2 million km^3 and a corresponding rise of sea level of 5 to 7 m during the last interglacial.

Eustatic Effects of Changes in Liquid Water on Land

In the absence of large-scale glaciation and deglaciation, a possible mechanism for relatively rapid eustatic sea-level change could be changes in the mass of liquid water sequestered on the continents, both above and below the ground surface. Such a mechanism is needed to explain the apparently eustatic sea-level changes in a virtually ice-free Earth during Mesozoic and early Cenozoic time described in papers by Haq *et al.* (1987) and by Christie-Blick *et al.* (Chapter 7). The topic presented first is the possible variations in groundwater.

A global climate change toward less precipitation will lower the water table in groundwater aquifers, transfer water from the land to the sea, and raise sea level. Less precipitation on land should result from atmospheric cooling, lower wind velocity, or changes in atmospheric circulation which would alter the balance between precipitation over the ocean and over the land. Increased precipitation resulting from atmospheric warming or other causes will raise the water table and lower sea level.

Removal of fresh-water-bearing porous sediments—chiefly sands and calcareous deposits—by erosion should have the same effect as a decline of precipitation, while accretion of such sediments will create a greater potential aquifer volume and hence be roughly equivalent to a rise in the water table.

Land subsidence, which may now be occurring in many coastal areas, will reduce the volume of sedimentary aquifers above sea level, and thus have much the same effect as erosion of sediments or a decline in the level of the water table due to decreased precipitation. Emergence of previously submerged aquifers should have an opposite effect, especially in carbonate terrains where karst formation can occur.

It is also possible that infiltration or discharge rates could change with time, leading to larger or smaller accumulations of groundwater as the balance between infiltration and discharge approaches a new equilibrium determined by changes in hydrostatic pressure in the aquifer. For example, canyon cutting during times of low sea level should increase discharge rates while deposition of unconsolidated coarse sediments should increase infiltration rates.

According to Hay and Leslie (Chapter 9) the late Cenozoic fluctuations in sea level caused by glaciation and deglaciation resulted in an offloading of sediments from coastal plain regions and continental shelves into the continental slopes and abyssal plains of the oceans. Thus the potential groundwater reservoir that exists today may well represent a minimum for much of geologic history. In past times, the potential change in sea level resulting from fluctuations in groundwater storage could have been double that which exists today.

These statements may be roughly quantified by assuming an equivalent rise or fall of the water table by 100 m, 13.3 percent of the average height of continental surfaces above sea level. Hay and Leslie (Chapter 9) estimate that porosities are close to 40 percent in the upper layers of the sands and calcareous sediments resting on the continents in geosynclines (intracratonic basins), coastal plains, and cratonic platforms. They calculate that these two types of sediments make up 38 to 47 percent of the total deposits in these three sedimentary environments, and that they are the only sediments that take part in significant water exchange with the environment outside the aquifers.

Table 2 shows the volumes and pore space of the top 100 m of sands and calcareous sediments in cratonic platforms, geosynclines, and coastal plains, computed from these estimates by Hay and Leslie and supplemented by estimates made by Southam and Hay (1981) and by Ronov (1982).

The total volume of pore space in the top 100 m of the continental sediments—2.5×10^6 km^3—is equivalent to a rise or fall of sea level by 7 m. Hay and Leslie assume that the rates of filling or discharge in these coarse sediments would be less than 13.5×10^3 km^3/yr. Thus more than 185 yr would be required to fill or empty an aquifer 100 m thick, corresponding to a rate of sea-level change of less than 4 mm/yr. Where only a slight imbalance exists between infiltration and discharge the times required for filling or emptying an aquifer 100 m thick could be tens to hundreds of thousands of years. For example, Meier (1984)

TABLE 2 Aquifers and Pore Space in Top 100 Meters of Sediments on Land

Location	Area covered by sediments (10^6 km^2)	Area covered by sandy or calcareous sediments (percent)	Volume of sandy and calcareous sediments in top 100 m of aquifers (10^6 km^3)	Volume of pore space in top 100 m of aquifers[a] (10^6 km^3)	Sea-level equivalent (meters)
Cratonic platforms	55	47	2.6	1.0	2.9
Geosynclines	59	38	2.3	0.9	2.6
Coastal plains and shelves	31	47	1.4	0.6	1.7
Total	145		6.3	2.5	7.2

[a]Assuming 40 percent porosity

estimates that global depletion of groundwater during this century has been between 1600 and 2400 km^3/yr or 20 to 30 km^3/yr. At this rate, filling or emptying a 100-m-thick global aquifer would take 85,000 to 130,000 yr and the corresponding rate of rise or fall of sea level would be less than 0.1 mm/yr.

Baumgartner and Reichel (1975) and Woods (1984) estimate the global volume of groundwater at 8×10^6 km^3, about 22 percent of the Earth's fresh water, equivalent to 22 m of sea level. This may be compared to 70 m of sea-level equivalent for the Greenland and Antarctic ice sheets.

From data given by Hay and Leslie, assuming an average porosity of 20 percent for sandy and calcareous sediments, a pore volume in sediments has been computed at 64.8×10^6 km^3. But 55.5×10^6 km^3 of this pore space is below sea level and hence presumably saturated with water. The pore space above sea level, which could be drained or filled by the various mechanisms discussed herein, is 9.4×10^6 km^3, equivalent to 27 m of sea level, very close to the estimate of groundwater volume given by Baumgartner and Reichel. However, as Hay and Leslie suggest, the porosity of the 750 m of sediments above sea level may be 30 to 40 percent, corresponding to a pore volume of 15×10^6 to 19×10^6 km^3.

Geological evidence shows that large quantities of shallow water carbonates and non-marine sands were laid down in mid-Paleozoic, late Paleozoic to early Mesozoic, and mid-Cretaceous times. During these periods, their abundance was probably more than twice that of the present time and the potential for groundwater storage was higher. Insofar as these porous sediments occupied a greater percentage of the land area than their present counterparts, infiltration rates must also have been higher than today, compared with rates of discharge, which can occur only at the edges of sedimentary columns.

The mass of liquid water above ground in lakes and rivers is only a small fraction of the mass of groundwater. Residence times in rivers vary from less than a week to about a year depending on size and length and on the slope of the river bed.

The volume of water stored in lakes is about 0.22×10^6 km^3 (Robin, 1987), probably about 50 times the volume in rivers but only about 1 percent of the volume of groundwater above sea level. Changes in lake volume result from climate variation and change and from human activities, primarily diversion of inflows for irrigation or other purposes. The Aral Sea in the Uzbek Republic of the Soviet Union is a striking example (Micklin, 1971). This lake without an outlet, fed by the Amur Darya and Syr Darya rivers, was formerly the world's fourth largest lake, behind the Caspian Sea, Lake Superior, and Lake Victoria. In 1960 its area was 68,000 km^2, its average depth was 16 m, and its volume was 1090 km^3. Beginning in 1960, there was a large increase in diversions of the river flows for irrigation caused by expansion and intensification of the irrigated areas. These increased diversions were not compensated for by conservation measures as previously, and the lake began to shrink rapidly. By the beginning of 1970, the area had decreased by 40 percent, the volume had decreased by 66 percent, and the water depth had dropped to 9 m. This change in lake volume must have been accompanied by a eustatic rise in sea level of slightly less than 2 mm. Destruction of the lake is still occurring; without drastic changes in irrigation practices, it will have largely disappeared by the early part of the twenty-first century, and there will be a further eustatic rise in sea level of about 1 mm.

On a worldwide basis, Robin (1987) estimates lake volumes are diminishing by 72 km^3/yr. He bases his estimates on the observed annual decline of the Caspian Sea by 10.9 km^3, assuming that this decline in the Caspian, the world's largest lake, is 15 percent of the global diminution of lake volumes. The corresponding eustatic rise in sea level should be 0.2 mm/yr or 20 mm/century. To this should probably be added the sea-level rise of about 1 mm due to the future decline of the Aral Sea.

Robin (1987) also points out that a long-term warming of the atmosphere by 3°C at low latitudes and 6°C at higher latitudes should result in an increase of the atmospheric content of water vapor, and a corresponding fall in sea level of about 7 mm.

Changes in the Volume of Water Without Changes in Mass

Steric height change (i.e., temperature and/or salinity variation) contributes to global sea-level fluctuations from year to year and decade to decade. It has been shown in regional studies that steric height changes account very well for sea-level variations (on the order of 100 mm) observed on seasonal and interannual time scales. On the other hand, sea-level fluctuations over periods of tens of thousands of years are so large that steric expansion can be ruled out as a major contributor. A warming of the entire ocean from 0°C to the currently observed temperatures would involve a thermal expansion of only about 10 m.

The thermal expansion coefficient increases significantly both with increasing temperature and with increasing pressure. The steep increase with increasing temperature has sometimes been used as an argument that deep cold water may be ignored in steric height calculations relative to warm surface water. As Roemmich (Chapter 13) shows, this is not so, because a parcel of water at 4°C at about 2000 m depth has a thermal expansion coefficient 60 percent as large as that of a parcel at the ocean surface at 20°C.

We need to discover how much variability in steric height is contained in time scales of decades or longer and to determine what vertical and horizontal scales of variability correspond to these long-term changes. These essential questions must be answered first before we can ask how long the ocean must be sampled at a single location, how many locations must be sampled, and to what depths the sampling should extend to determine the relationship of steric height to the long-term trend in sea level. Trends in steric height over three decades in the central north Pacific appear to be about 1 mm/yr, but these trends are so small in comparison with the large interannual variations of 10 to 100 mm that a much longer time series is needed to confirm the trend (Thomson and Tabata, 1987).

On time scales of tens to hundreds of years, the effects of salinity changes on global sea level are slight. The total salt content of the oceans remains approximately constant, and sea-level changes due to vertical redistribution of salt in the water column are likely to be small. Heat, unlike salt, is freely exchanged with the atmosphere, and even the combined heat content of the ocean–atmosphere system may change significantly over a few decades or centuries. Thus global expansion or contraction of ocean waters must be dominated by temperature change rather than by salinity change except under some circumstances in high latitudes and semi-isolated basins. In general, there is a fairly tight temperature/salinity correlation; a temperature increase of 1°C or above in the thermocline can be accompanied by a salinity increase of approximately 0.1 parts per thousand. This temperature change produces an increase in specific volume whose magnitude is more than three times the corresponding decrease in specific volume caused by the salinity change. Thus the effects are of opposite sign, and although salinity is not negligible, the temperature effects dominate.

Steric changes in the water column on time scales of a decade and longer are not confined to or concentrated in the upper ocean. In the subtropical North Atlantic such changes extend to depths of at least 3000 m (Chapter 13). For observational programs the large vertical extent of the signal means that the entire depth of the ocean should be sampled. In the upper water layers, both vertical diffusion and advection are important, as are water mass transport along quasi-horizontal isopycnal (constant density) surfaces from subsurface regions in higher latitudes toward the mid-latitude ocean interior. Critical processes are air-sea interactions in the regions where the isopycnals intercept the sea surface, and such cross-pycnal mixing processes as double diffusion (diffusion of heat and salinity at different molecular diffusion rates) and intermittent internal wave breaking. In the deeper layers (below the permanent thermocline in the North Atlantic) we are concerned with the formation of North Atlantic deep water by convective mixing and sinking in the Greenland and Norwegian seas and by horizontal advection and mixing as this water mass moves southward. The critical problems are the mechanisms and rates by which waters of lower density can gradually replace higher density water in the deep ocean basins.

Roemmich (Chapter 13) shows that at mid-latitudes in the North Atlantic the thermocline fluctuations in steric height exhibit greater variance in the time domain than the deeper changes and also more variability around the subtropical gyre. For the 27 yr of Panuliris hydrographic data taken by the Bermuda Biological Station (discussed by Roemmich, Chapter 13), the thermocline fluctuations make a greater contribution to steric height change than the deep changes, but on a longer time scale, it is not clear whether mid-depth or deep variations would dominate. The thermocline comes to the surface along the northern rim of the subtropical gyre, and these waters are therefore subject to atmospheric forcing relatively near the Panuliris area. The deep water is farther removed from contact with the atmosphere, which can account for the more gradual changes and larger spatial scale of the deep signal.

Volume of Ocean Basins

Changes in the volume of the ocean basins can have very significant long-term effects on sea level—effects that can be as great as or greater than that caused by ice. The dominant cause of these large-scale changes is the variable volume (Figure 2) of ocean ridge material from seafloor spreading processes. If spreading increases, the volume of the ridge crest increases, displacing water and flooding the continental areas. The critical quantity is the area of seafloor produced per unit time, which can be changed either by an increase in spreading rate over a ridge crest of constant length, or by increasing the length of the ridge crest, or by a combination of the two. Harrison (Chapter 8) shows that during the past 80 million yr (m.y.) a diminution in the volume of ocean-ridge crests by 9.55×10^{16} m^3 could have produced an increase in freeboard (essentially equal to a fall in sea level) of about 180 m, with a very large uncertainty of +140 and −180 m.

The continents themselves can undergo vertical motion termed epeirogeny. Evidence shows that epeirogenic movements of the continents can result in several hundreds of meters of elevation over very long time periods. Similar elevation changes occurring in ocean basins can significantly affect the volume of the ocean basin and, thus, sea level. For example, there is evidence for a thermally induced swelling or uplift in the equatorial Pacific during the Cretaceous, which was accompanied by large amounts of extrusive

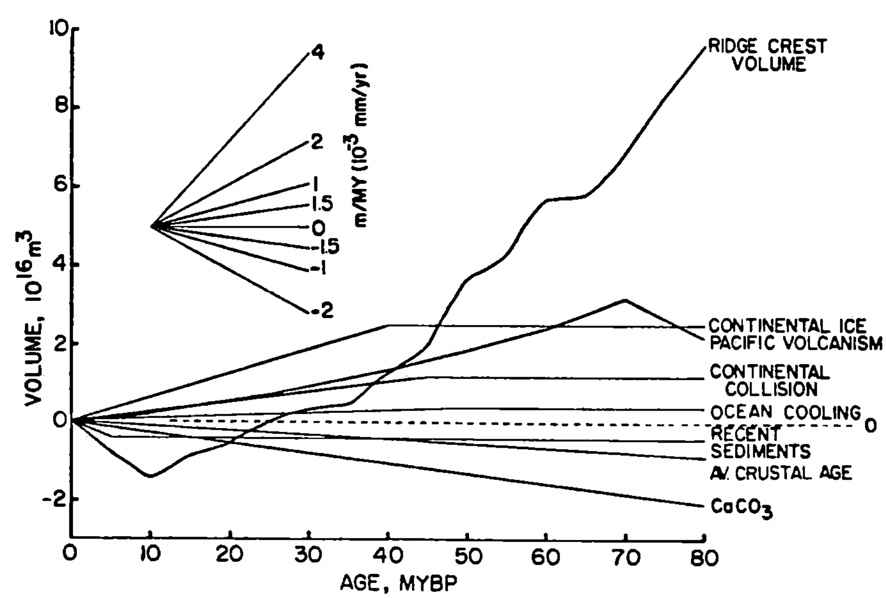


FIGURE 2 Volume estimates for six tectonic processes affecting the volume of the ocean basins. An increase in volume as time progresses produces a sea-level fall (or freeboard increase). See Harrison (Chapter 8) for additional detail.

volcanic activity. A decrease of the volume of the ocean basins of about 2×10^{16} m^3 was produced by this "ocean-floor epeirogeny" between 110 m.y. ago (Ma) and 70 Ma.

Many other possibilities exist for changing sea level by tens to hundreds of meters. For instance, a change of the pattern of subduction can cause the ocean basins to become, on average, younger or older and so produce a sea-level change that is in principle similar to that caused by varying the amount of seafloor produced per unit time interval. If the area of the ocean basin changes due to continental growth or continental destruction (e.g., India colliding with Asia and decreasing the continental surface), a change in ocean volume will also occur.

Harrison (Chapter 8) suggests three processes by which the quantity or distribution of sediments in the deep sea could be changed: (1) the evolution of pelagic foraminifera in the Late Cretaceous, (2) the greater average age of the ocean basins today than 80 Ma, and (3) higher sedimentation rates during the past 5 m.y. than earlier times. The last two processes would tend to counteract about one-eighth of the effect of greater ridge areal volume in earlier times compared with today. The evolution of pelagic foraminifera may not have changed the rates of deposition of calcareous deposits, which depend almost entirely on the rate of weathering of calc-silicates and volcanic emission of CO_2, but it should have resulted in a redistribution of these sediments from mainly shallow-water reef deposits to a more or less uniform blanket over the deep-sea floor. This would have had a significant isostatic effect, replacing marked subsidence of shallow water areas and probably a forebulge under both adjacent land areas and the offshore deep ocean, with a relatively uniform sinking of the deep-sea floor and a rise under the continents.

Harrison (Chapter 8) calculates that the total increase in the average freeboard of the continents from all causes over the past 80 m.y., not counting the possible effects of redistribution of calcareous sedimentation, should have been about 260 m. But estimates based on paleogeographic sediment distribution for the past 80 m.y. indicate a much smaller value—150 m. He suggests that the discrepancies could be resolved if the continental hypsometric curves were steeper in the past than they are today. However, the uncertainties of the estimates of the various volume effects are sufficient to account for this freeboard discrepancy.

RECORD OF CURRENT AND PAST CHANGE

Sea level has changed dramatically over the course of earth history, with repeated variations of more than a 100 m from present sea level. These large sea-level changes occurred both during times of intense glaciation and during times of an ice-free Earth. Even during times of glacio-eustatic change, there were other broad-scale variations resulting from long-term climatic and tectonic changes. The details of sea-level variations are well documented for much of the record of the past few hundred thousand years but gradually deteriorates with older records, mainly because the time resolutions of dating and correlation methods become less precise. In addition, the numbers of assumptions needed to extract sea-level change from proxy records increase with age.

Historic and Tide-Gauge Records—The Past 100 Years

Large sections of the world's coastlines are experiencing changes in RSL that have been analyzed as being spatially and temporally coherent, but there is no unique way to average the existing sea-level data from different regions to obtain a global measure of RSL change. Variations on the order of 50 percent in the estimate of relative change can be induced simply by use of different averaging methods.

Barnett (Chapter 1), using both a key tide-gauge station approach and a grouping of records from different regions, has estimated that RSL has increased by 1 to 2 mm/yr over

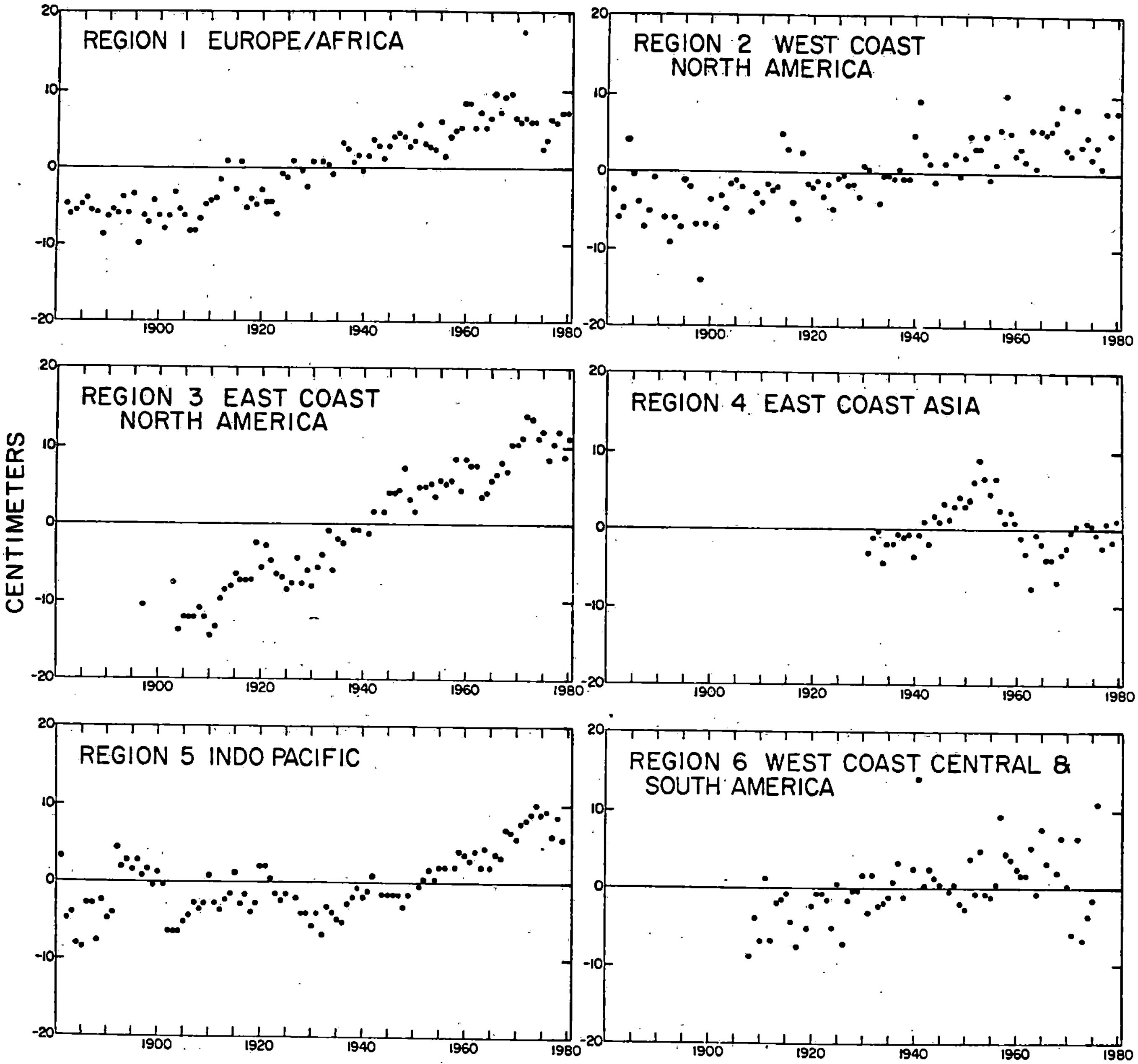


FIGURE 3 Regional averages of annual sea-level anomaly (see Chapter 1, for further discussion). From Barnett (1984) with permission of the American Geophysical Union.

the past 80 yr or so (Figure 3). There is no strongly coherent change on a global scale (coherence is primarily regional), nor is there an obviously accelerating rate of RSL increase in recent years.

Interpretation of sea-level records from tide-gauge measurements faces many problems. The principal problems are summarized below:

1. Station locations have been subject to position changes as, for example, during harbor development, which can result in discontinuity of the time-series record.
2. Tectonic uplift/subsidence of the land mass on which the tide gauge sits will induce

an RSL change. These vertical tectonic movements are further discussed in a previous section and in the discussion by Peltier (Chapter 4).

3. A variety of physical processes that include changes in water temperature and salinity, river runoff, local and nonlocal currents, winds, waves, and atmospheric pressure all can affect RSL, sometimes in conflicting ways. Thompson (Chapter 2) found that the standard error of an estimate of longer term trend in North Atlantic sea-level records can be halved (from 0.4 to 0.2 mm/yr) on removal of short-term meteorological and oceanographic effects. Munk *et al.* (Chapter 14) discuss subsidiary measurements that can be used to take account of some of the oceanographic effects.

Sturges (Chapter 3) finds that the dominant sea-level signals at periods of 5 to 10 yr and longer are coherent between the U.S. Pacific coast and Hawaii, and on both sides of the Atlantic. The amplitudes, at these periods, are ~50 to 150 mm. The signals are out of phase, with delays of several years, but are partly consistent with baroclinic wave propagation across the ocean. The longest records appear to be "contaminated" by this ocean-wave propagation. A wave of 10-yr period and 50-mm amplitude has a rate of rise of the order of 20 mm/yr, which is about an order of magnitude larger than Barnett's estimate of the mean rate of sea-level rise during the past 80 yr. Consequently, records of short duration, even if they have high precision, can hide the existence of long-term variability or change.

The Past 18,000 Years

Eighteen thousand years before present, huge continental ice sheets covered most of Canada, parts of the northern United States, Greenland, northwestern Europe and Antarctica, and sea level was more than 100 m lower than it is today; this is known as the Weichsel-Wisconsin glacial maximum (see discussion by Matthews, Chapter 5). The ice sheets stored the water that corresponded to nearly all of this lowering of global sea level. The mass of the ice sheets also deformed the shape of the Earth, and affected the volume of the ocean basins.

The ice sheets in North America and Europe began to dissipate by 15,000 yrBP. The meltwater was returned to the ocean basins with a concomitant rise in sea level. The causes of ice-sheet disintegration are not clearly understood, but cyclical Earth-orbital variations in the inclination of the Earth's axis and in the seasonal Earth-Sun distance (COHMAP, 1988) together with ice-sheet dynamics and changes in ocean circulation probably formed a complex set of interrelated forcing functions that initiated breakup and continued wastage of the ice sheets between 15,000 and 7000 yrBP.

The evidence for past sea levels is partly geological—the existence of coral reef terraces containing the littoral zone coral, *A. palmata*; submerged or buried strand-line complexes; peat deposits just above subaerial exposure surfaces; and articulated valves of shallow water or tidal zone pelecypods. Evidence of this kind includes a series of paired pelecypod valves off the Texas gulf coast at –57 m with a carbon-14 date of 12,900 ± 400 yrBP and a brackish water peat from New England at almost the same depth and age. Similar data are found at younger ages, but the only high-quality data between about 12,500 and 18,000 yrBP are the deep-sea oxygen-isotope records in planktonic and benthic foraminifera of pelagic sediments. The oxygen-isotope ratios reflect both the quantity of light oxygen sequestered in the ice sheets and the temperature of the waters in which the foraminifera lived. The relative importance of these two effects is uncertain. Matthews shows that this uncertainty gives a range of 120 to 108 m for the maximum sea-level lowering in the tropical North Atlantic near Barbados. Chappell and Shackleton (1986) estimate a value of –130 m for the Huon Peninsula in New Guinea, and they cite a maximum sea-level lowering estimate of 150 m on the broad shelf off northern Australia. They state that such variations around the globe "are to be expected as a result of adjustment to glacial/interglacial change of ice and ocean volumes."

During most of the time between 15,000 and 7000 yrBP, the rate of dissipation of the ice caps was very rapid, and the average rate of rise in sea level was 12.5 mm/yr. But there were two brief periods of renewed cold in the European, though not the American, record at about 11,000 to 12,000 yrBP. These are called the Older and Younger Dryas. Pollen records from European bogs show that the forests that had begun to repopulate northern Europe after the ice departed were destroyed and replaced by Arctic shrubs. These colder intervals lasted for several hundred years. They may have resulted from a reduction in the temperature of the near-surface water lying to the west of Europe (Broecker, 1987).

By 7000 to 8000 yrBP the North American and European ice sheets had disappeared. In high northern latitudes, RSL was 10 to 30 m higher than it is today because of the isostatic depression of the land under the previous weight of ice. In England and southern New England and in lower latitudes, the isostatic "forebulge" had produced an RSL 10 to 20 m lower than today's sea level.

Bloom and Yonekura (Chapter 6) examined 13 well-documented Holocene sea-level records from published literature that covers the past 8000 radiocarbon years and a range of latitudes from the Arctic to Panama. In the north, RSL declined exponentially at initial rates of ≥10 mm/yr and a final rate during the past 1000 yr of the order of 1 or 2 mm/yr. South of the "hinge line" in central New England, RSL rose at an exponentially declining rate. Interestingly enough, the amount of emergence or submergence for each 1000-yr interval at any particular place is proportional to the total amount of emergence or submergence at that place over the past 8000 yr. This would imply that there has been no eustatic change in sea level during the later Holocene, only isostatic adjustment—upward RSL rise in the south and downward RSL fall in the north, which are strictly proportional to each other. If so, in the absence of climatic change, future sea-level change could be controlled by residual isostatic response to the loading and unloading completed 7000 to 8000 yrBP.

This suggestion is consistent with Peltier's model of isostatic adjustment in the mantle (see Chapter 4) and with a recent discussion by Stewart (1989). It is given further credibility by Barnett's (1984) observation that sea level is falling on the east coast of Asia, where presumably different isostatic adjustments are taking place.

The Past 250,000 Years

The record of sea-level change for this time period can be interpreted from two principal proxy sources: (1) oxygen isotope records in sediment cores from the deep sea and (2) the study of ancient shorelines (coral reefs) that have been tectonically uplifted. Although each type of record presents several interpretive problems and requires assumptions in deriving sea level, a recent correlation of an oxygen isotope record and a flight of raised coral terraces in the southwest Pacific has provided improved estimates of sea level (Table 3 and Figure 4) (Chappell and Shackleton, 1986).

Coral terraces from the Huon Peninsula (Papua New Guinea) are described by Bloom and Yonekura in Chapter 6. These uplifted terraces are like a continuous tape recorder; each coral reef developed when the rising sea level overtook the rising land, and the reef crests represent approximately the peaks of each sea-level rise. RSL can be estimated from the current heights of the terraces. It has been assumed that sea level at ~125,000 yrBP was +6 m above the present sea level—a number derived from several different terrace sequences around the world. On the basis of this assumption, the uplift rates for different portions of the Huon Peninsula can be derived. Further from the assumption that these uplift rates have remained essentially constant for the past 125,000 yr, sea level can be calculated (Figure 4) for the other Huon Peninsula terraces as described in Chapter 6.

Oxygen isotope ratios, $\delta^{18}O$ [$\delta = (^{18}O/^{16}O \div {}^{18}O/^{16}O_{SMOW}) - 1$, where SMOW is standard

TABLE 3 Sea Level and Calculated Rates of Average Change for the Past 135,000 Years

Stage	Age[a] (1000 yrBP)	Sea Level[a] (m)	$\delta^{18}O$[a]	Change	Rate (mm/yr)
Modern	0	0	3.42		
				—	—
I	6	0[b]	3.47		
				+118	+10.7
I/II	17	−130	5.09		
				−84	−7.0
II	29	−46	4.48		
				+20	+3.3
II/IIIb	35	−65	4.67		
				−24	−4.8
IIIb	40	−41	4.47		
				+11	+5.5
IIIb/a	42	−52	4.66		
				−7	−3.5
IIIa	44	−45	4.51		
				(−15)	(−2.5)
IVb	53	−30	4.30		
				+14	+4.7
IVb/a	56	−44	4.58		
				−16	−5.3
IVa	59	−28	4.20		
				+33	+6.6
IVa/Vb	64	−61	4.66		
				−25	−3.1
Vb	72	−36	4.21		
				+8	+4.0
Vb/a	74	−44	4.33		
				−25	−3.6
Va	81	−19	3.99		
				+25	+3.6
Va/VIb	88	−44	4.37		
				−18	−2.2
VIb	96	−26	4.09		
				(+7)	(+0.7)
VIa	106	−19	4.03		
				+43	+7.2
VIa/VIIb	112	−62	4.38		
				−62	−10.3
VIIb	118	0	3.49		
				+8	+2.0
VIIb/a	122	−8	—		
				−14	−7.0
VIIa	124	+6	3.37		
				+136	+12.4
VII/VIII	135	−130	5.20		

[a]Data from Chappell and Shackleton (1986).

[b]Sea level at 6000 yrBP has been within a meter or so with fluctuations arising from isostatic and tectonic adjustments (Bloom and Yonekura, Chapter 6).

mean ocean water], of foraminifera vary with the temperature, salinity, and $\delta^{8}O$ of the water from which their carbonate tests are deposited. Ocean water $\delta^{18}O$ varies with the quantity of isotopically light ice stored on the continents so that the record from foraminifera is a blend of global ice volume and local water temperature and salinity components. For the late Pleistocene, temperature and salinity components of the signal are demonstrably small compared to the ice volume signal. A $\delta^{18}O$ record from the east equatorial Pacific (core V19-30; Shackleton *et al.*, 1983) is shown in Figure 4. Peaks represent the influx of isotopically light water derived from continental ice and are correlated with high stands of sea level indicated by the Huon Peninsula terraces.

If these sea-level change magnitudes and timing are representative of global sea level, then rates of eustatic sea-level change during the period from 250,000 yrBP to about 6000 yrBP varied between −10 mm/yr and +12 mm/yr, with typical rates being on the order of ±3 to ±7 mm/yr. These rates exceed those that have prevailed over the past few thousand years but may be approached by those projected for the next millennium or so (see section on Forecasting Changes in Sea Level Related to Greenhouse Gases).

At the peak of glaciation 18,000 and 135,000 yrBP, sea level near New Guinea was lower than it is today by about 130 m (Table 3). But between these extremes, during glacial episodes, sea level oscillated between 19 and 65 m below the interglacial levels over time intervals of 2000 to 12,000 yr. These oscillations resulted from variations in the area and thickness of both the continental ice sheets in northern Europe and northern North America and the Antarctic Ice Sheet. As shown by Labeyrie *et al.* (1986), at least some intervals of rising sea level may have been initiated by surges of ice streams behind ice shelves in Antarctica, which greatly increased the rates of iceberg calving. The resulting rise of a few meters in global sea level may have then destabilized the marine portions of the Northern Hemisphere ice sheets, leading to the much larger rise in sea level shown in the oxygen isotope and Huon Peninsula terrace records.

The Past 250 Million Years

The sea-level record becomes much more cloudy as we look back in time. Much of the past 250 m.y. was dominated by an ice-free Earth or at least a world lacking extensive continental ice caps (see Frakes and Francis, 1988).

Oxygen isotope ratios of tropical planktonic foraminifera show that continental ice volumes during only the past 40 to 45 m.y. were consistently as large as or larger than that of today (Prentice and Matthews, 1988). Between 50 and 65 Ma the world was intermit-

FIGURE 4 (A) $\delta^{18}O$ record for the past 240,000 yr from east equatorial Pacific core V19-30; and (B) sea-level curve for the Huon Peninsula recalculated to correlate with the $\delta^{18}O$ record from core V19-30. Modified after Chappell and Shackleton (1986).

tently ice free; and continental ice was apparently absent throughout the Mesozoic (65 to 250 Ma). Consequently, sea level was considerably higher than in later times, and vast continental areas on the order of 50 million km^2 (one third of present land area) were flooded, e.g., the Cretaceous Interior Seaway in North America.

Sea-level changes were dominated by three major forcing factors: climatic change, tectonic processes, and sedimentation. Climate was much warmer than it is at present, with ocean-bottom temperatures at some 10 to 15°C. Thermal contraction of seawater (steric effect) could account for a sea-level fall of more than 10 m between the climatically warm and equable Cretaceous (144 to 66 Ma) and the onset of an ice-dominated system in the Cenozoic. Tectonic processes such as seafloor spreading, subduction of oceanic crust, broad-scale ocean-floor epeirogeny, continental collisions, and other activity affected the volume of the ocean basins. Epeirogenic uplift or subsidence of the continents also affected RSL. The volume of the ocean basins was also changed by changes in sedimentation. The evolution of the calcareous algae called *Coccolithophoridae* at the end of the Jurassic and of pelagic foraminifera in the Middle Cretaceous resulted in deposition, perhaps for the first time, of a thick blanket of calcareous sediments over the deep seafloor and in a possible rise in sea level.

Direct evidence for sea level at different times during the past 250 m.y. comes from paleogeographic maps, from which the area of the continents covered by marine sediments can be estimated. Sea level was highest, about 180 ± 40 m above the present level, at 100 Ma, having risen from 80 m about 180 Ma. There was also a pronounced regression of sea level (commonly associated with a fall) during the Paleocene at 60 Ma.

For the past 80 m.y., there is evidence concerning changes in the volume of the ocean basins, particularly in the volume of ancient mid-ocean ridges, based on estimates of spreading rates between times of magnetic reversals and estimates of ridge lengths. From the data analyzed by Kominz (1984), ridge-crest volume changes since 80 Ma would have resulted in a fall of sea level of 180 ± ~150 m. If ice volume changes, the effects of Pacific volcanism, and changes in sedimentation are added, the calculated sea level fall since 80 Ma is about 230 m, some 40 percent greater than that computed from modern paleogeographic maps. Although the errors in both sets of estimates are more than sufficient to account for the discrepancy, it is possible, as Harrison has suggested (Chapter 8), that the continental hypsographic curves were steeper in the past than they are today, so that a greater change of freeboard was necessary to produce a given increase in the areas of flooding.

Little or no oceanic crust older than about 80 m.y. still exists in the present oceans; hence there is little direct evidence concerning the volume of the ocean basins. We must rely on paleogeographic reconstructions of the continents and analysis of their sediment covers to infer ancient sea levels.

Within the past dozen or so years, seismic stratigraphic methods have been developed and applied to the analysis of scores of events of coastal onlap/offlap within stratigraphic successions. P. R. Vail and his colleagues, e.g., Haq *et al.* (1987), have ascribed most of the boundaries between offlapping and onlapping stratal units to sea-level change (Figure 5) and have implied that the vast majority reflect eustatic or global sea-level variation. Although this approach has proven valuable in exploration for hydrocarbon resources, many investigators question whether these boundaries are necessarily of eustatic origin. Some boundaries may be caused by variations in the local rate of subsidence or by changes in the rate of sediment supply. Vail and his co-workers have estimated magnitudes of sea-level change that are on the order of 10 to 100 m and are superimposed on the long-term trends from climatic and tectonic processes discussed above, but questions also remain about whether such large variations are indicated by the data in hand, even if the signal is a eustatic one. It is particularly difficult to explain sea-level changes in excess of 100 m in intervals of 1 m.y. or less during the Cretaceous, a time when the Earth was largely ice free. These and related issues are discussed by Christie-Blick *et al.* in Chapter 7.

Unconformity-bounded sequences, which form largely in response to changes in de-

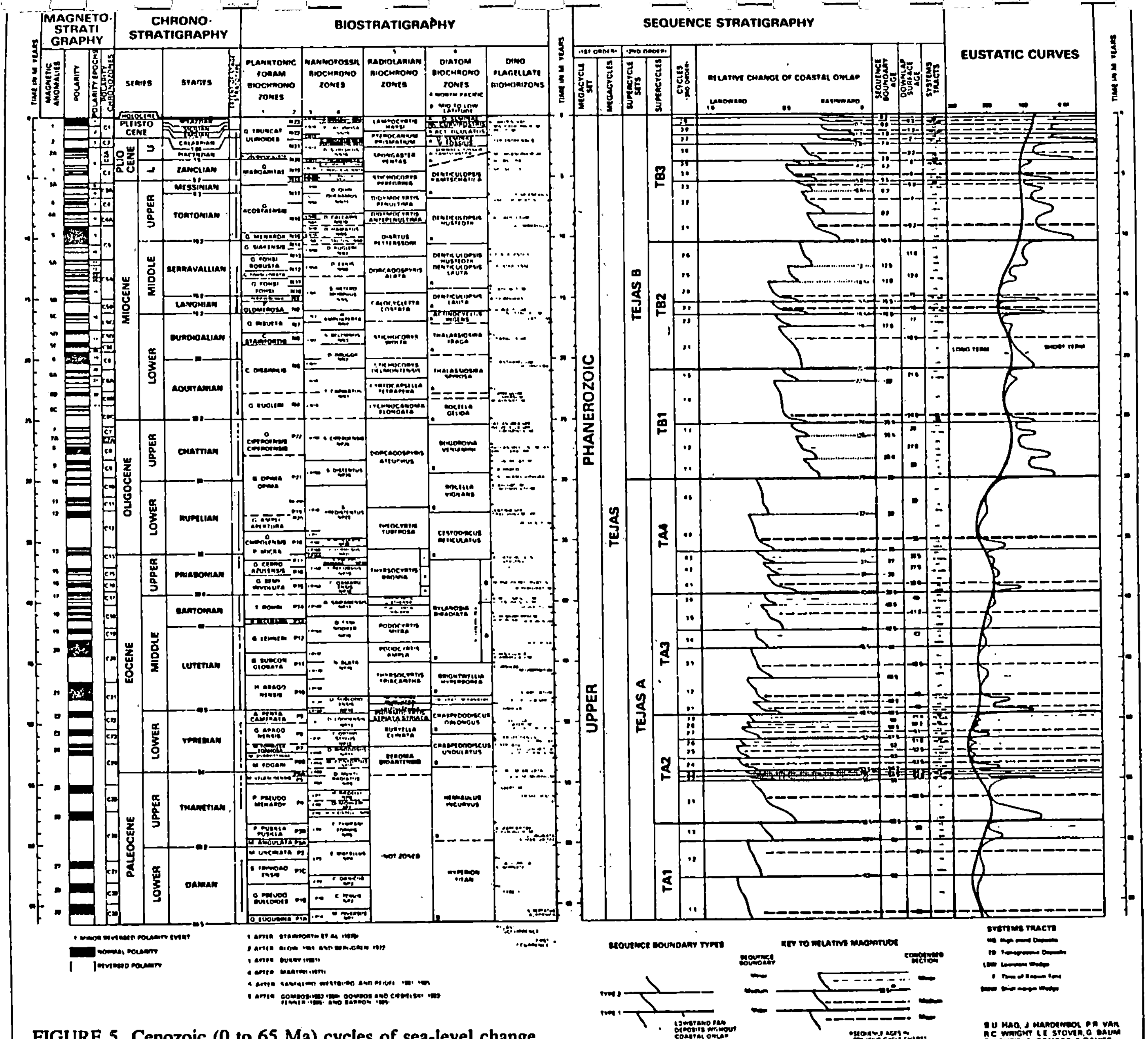


FIGURE 5 Cenozoic (0 to 65 Ma) cycles of sea-level change. After Haq *et al.* (1987). (See Chapter 7 for additional discussion.)

positional base level, can be identified in outcrops and boreholes as well as in seismic profiles. Unlike transgressions and regressions of the shoreline or changes in paleobathymetry, which are used in classical stratigraphy to gauge sea-level fluctuations, the formation of regional unconformities is relatively insensitive to the rate of sediment supply, and this constitutes the main advantage of the sequence stratigraphic approach. According to Christie-Blick *et al.* (Chapter 7) and Christie-Blick (1989) most sequence boundaries record times at which the rate of sea-level fall increased (or reached a maximum) or the rate of tectonic subsidence decreased. Those of eustatic origin should be formed in all basins connected to the open ocean and should be nearly correlative, whatever the local tectonic history. Unconformities of tectonic origin may also be present, but these are not expected to extend beyond a region more than a few hundred or a few thousand kilometers across. The identification of a global sea-level signal therefore depends on demonstrating that particular unconformities are present in widely separated basins, and are synchronous (ideally to within 0.5 m.y.).

More difficult than establishing the timing of sea-level change is the problem of estimating amplitudes and rates of change. Amplitudes of sea-level oscillations may be estimated through a combination of sequence stratigraphy and geophysical modeling of subsidence history, but practical difficulties and sensitivity of results to model assumptions may limit estimates to no better than a factor of 2 or 3 larger or smaller than true amplitudes (Christie-Blick *et al.*, Chapter 7). The principal uncertainties are in age control, paleobathymetry, compaction history, the effects of sediment loading on the lithosphere, and the tectonic subsidence that must be subtracted from corrected stratigraphic data to obtain the sea-level signal.

Harrison (Chapter 8) has suggested that 10 transgressive-regressive cycles of sedimentation with average periods of 7 to 10 m.y. during the Cretaceous resulted from alternating increases and decreases in the rates of seafloor spreading. These would have caused eustatic rises and falls of sea level of about 15 m. The eustatic sea-level changes were greatly amplified in the Western Interior Seaway by subsidence of the basin resulting from subduction of the Pacific plate under the western part of the North American continent. In Chapter 9, Hay and Leslie discuss possible changes in the volume of liquid water stored as groundwater as a cause of some of these Cretaceous sea-level changes. These and many other problems reviewed in Chapters 7 and 8 need to be solved before these curves can be read as eustatic changes.

If stratigraphic sequence boundaries do turn out to be related to eustatic sea-level change, the record from the Triassic (about 250 Ma) through to the present becomes rich with events that need to be considered in the light of possible causal processes.

From 250 to 2500 Ma

Although there is some direct stratigraphic record of sea-level change that extends back into the Paleozoic and Precambrian (Christie-Blick *et al.*, 1988; Bond *et al.*, 1988), most of the data are derived from continental freeboard consideration. Geologic evidence on the area of the continents covered with marine sediments indicates that continental freeboard, the average elevation of the continents above sea level, has been approximately constant since the Archean, 2500 Ma. Continental freeboard today, taking the mean of the continents, is about 750 m, and changes of more than a few hundred meters are apparently ruled out by the geologic evidence.

Provided that the mass of seawater (plus water on land) has not varied significantly, an approximately constant continental freeboard since the Archean, combined with a 50 percent decrease of heat flow from the mantle, has been shown by Schubert and Reymer (1985) to require a net growth of the continents of about 25 percent (1 km^3/yr) over the past 2500 m.y. They believe that post-Archean changes in the mean thickness of the continental crust are highly unlikely, hence the continents must have grown in area. If freeboard has been within ±200 m of its present value since the end of the Archean, the continents have grown in area by anywhere from 10 to 40 percent. On the other hand, if the mass of seawater has increased significantly since the Archean, isostasy requires that the thickness of the continental crust should have likewise increased. For example, if the flux of water to the ocean from the Earth's interior has been constant during the Earth's lifetime, the entire growth of continental volume estimated by Schubert and Reymer can be accounted for by an increase in continental thickness rather than continental area. Continental growth is required as long as freeboard at the end of the Archean was not more than 400 m smaller than it is today.

Continental ice ages, which must have caused a fall in sea level, occurred at intervals during the Proterozoic and the Paleozoic. According to Crowell (1982), evidence from what is now southern Ontario shows that there were large-scale glaciations between 2500 and 2100 Ma. One ice sheet may have reached southern Wyoming. A similar glaciation probably occurred in South Africa sometime between 2700 and 2200 Ma, and possible tillites are reported from Western Australia. During the span of 600 m.y. between 2700 and 2100 Ma, a period as long as the entire time in which multicelled animals and plants have existed, several widespread glaciations, and presumably concomitant falls in sea level, apparently occurred, separated by ice-free eras.

There is no record of glaciation anywhere on Earth during the ensuing 1100 m.y. until the Late Precambrian, when continental glaciation flourished intermittently on all the continents, except possibly Antarctica, from about 950 to 560 Ma. Apparently there were three glaciation peaks, one about 940 Ma, another around 770 Ma, and a third about 615 Ma, in regions around the present North Atlantic, in central and southwestern Africa, and in Brazil, western North America, and Australia. Each of these extreme glacial times must have been a period of markedly lower sea level in comparison with earlier or later periods.

These Late Precambrian glaciations apparently occurred at low latitudes, distant from the Earth's poles, for which no satisfactory explanation has come forth. Perhaps, there was a reduced greenhouse effect or the patterns of oceanic and atmospheric circulation were markedly different during Precambrian times from those of today, or of the past half billion years.

The next widespread glacial epoch began in the supercontinent Gondwana near the end of the Ordovician lasting into the Silurian, from about 450 to 400 Ma. The record extends in scattered localities from northern Europe to South Africa and from the Sahara region to Bolivia and Peru.

For 90 m.y., from 330 Ma to 240 Ma, a Late Paleozoic ice age existed in large areas of the Gondwana supercontinent, beginning in what is now South America, reaching a climax in South America, Africa, and Antarctica, and ending in Australia. There were apparently several moderately sized ice caps instead of one huge one, because it does not appear that sea level was drastically lowered. The center of the glaciated regions was near the South Pole, as the supercontinent of Gondwana drifted across it.

There was no continental glaciation between Late Permian and Paleogene time, even though Antarctica lay near the South Pole. This may have been the result of a greenhouse effect caused by a high atmospheric CO_2 content, related to high rates of seafloor spreading and undersea volcanism (Berner *et al.*, 1983).

DO CHANGES IN SEA LEVEL CAUSE CHANGES IN CLIMATE?

Climate variability and change cause local or regional variations in sea level over months, years, or decades. Quantitatively the seasonal oscillation is the most important of these, followed by interannual variations of 5 to 10 yr, related to the Southern Oscillation in atmospheric pressure between the two sides of the Pacific Ocean, and manifested in El Niño in the eastern Pacific and in many other phenomena elsewhere.

Longer-term global climatic changes can be reflected in eustatic sea-level change through three processes: the waxing and waning of continental ice caps and alpine glaciers; variations in the quantity of liquid water stored on land in lakes, rivers, and underground aquifers; and steric changes in the volume of seawater resulting from warming or cooling.

It is also often suggested that a change in sea level can produce a change in climate. As Barron and Thompson (Chapter 11) point out, this suggestion is consistent with the apparent correlation of climate and sea level over geologic time. Several lines of evidence show that during the past 70 m.y. globally averaged surface temperatures have declined by 6° to 12°C, global sea level has fallen by perhaps 200 m, and the total land area above sea level has increased by about one-third.

There is a variety of physical mechanisms by which sea-level changes could directly affect global climate: changes in albedo, regional changes in atmosphere-surface coupling, changes in ocean circulation, changes at ice-sheet ocean margins, and changes in ocean/atmosphere chemical composition.

Virtually every aspect of the Earth's climate is affected by the exchange of heat, moisture, and momentum between the atmosphere and the underlying surface. Changes in the surface energy balance are determined by changes in surface albedo and surface wetness. A rise of sea level will increase the area covered by ocean water, which has a much lower albedo and a much higher wetness than the land. A change in global albedo by 0.01 will produce about a 1°C change in surface temperature. Such an albedo change would require a major change in the land/sea ratio, caused by a rise or fall of 100 m or more in sea level.

The change in wetness resulting from this assumed change in relative areas of sea and land would lead to marked changes in evaporation, and therefore in precipitation, although not necessarily in the same region, and also to a significant change in average summer surface temperatures because of the much larger thermal inertia of the ocean compared to the land. However, model calculations indicate that when both surface albedo and moisture availability are altered simultaneously, they produce nearly complete compensating effects. For example, modeled deforestation of the Amazon Basin, which would produce the same effects on albedo and wetness as a fall in sea level, produced only a small net surface temperature change in the deforested area, and no detectable global scale climate effects (Henderson-Sellars and Gornitz, 1985).

In general, surface roughness is an order of magnitude higher over the land than over the ocean. Consequently, a change in ocean/land proportions caused by a rise or fall of sea level may result in a marked redistribution of the surface areas of momentum exchange between the land and ocean and the atmosphere. This could have a marked effect on atmospheric circulation systems.

In some regions, for example, the Blake Plateau in the Western North Atlantic, it can be shown from studies of seafloor erosion that the position of the Gulf Stream has varied repeatedly by hundreds of kilometers with changes of sea level during the past 20 m.y. But the extent of possible climatic change caused by such movements in current position is unknown. The same lack of knowledge about climatic effects afflicts examples of seafloor subsidence such as that of the Greenland-Iceland-Faroe Ridge between the Arctic and the Atlantic oceans, and the Walvis Ridge off South Africa. Sea-level change can also isolate or reconnect small basins along the ocean margins. Enhanced evaporation in partially isolated basins can produce water masses of markedly different density than those of the main ocean, and thereby affect deep water and mid-water formation.

A marked increase in atmospheric CO_2 accompanied the deglaciation of the Northern Hemisphere ice sheets about 10,000 yrBP; it probably had a significant warming effect on the lower atmosphere. One hypothesis to explain at least part of this increase in atmospheric CO_2 involves the submergence of the continental shelves by the rise in sea level. In tropical waters this resulted in a large increase in the growth of coral reefs and precipitation of other calcareous sediments with a corresponding release of CO_2 to the subsurface ocean layers and the atmosphere.

If bottom water temperatures were unchanged, the fall of the sea level by 20 to 130 m during the last glaciation should also have resulted in the release of methane (a potent greenhouse gas) from methane ices (clathrates) in the upper layers of continental slope sediments. These sediments are believed to contain 2.2 mg/cm^3 of methane, or a total of 3400 gigatons of methane (Revelle, 1983a). A fall of 100 m in sea level without a change in ocean-bottom temperature should release 425 gigatons of methane. If the sea-level change took 5000 yr, 0.085 gigatons would have been released each year. Methane in the atmosphere has a half life of about 8 yr, implying that the release of methane from continental slope sediments should have increased the preindustrial methane level of less than 2 gigatons by about 50 percent. In fact, evidence from ice cores indicates that the methane content of the atmosphere decreased by nearly 50 percent during the latter half of the last glacial epoch and was at no time higher than the postglacial, preindustrial level of 650 parts per billion (Stauffer, 1988). This information suggests that during the glacial period the temperature of the ocean waters bathing the continental slope sediments decreased by more than 1°C than it is at present. This would prevent destabilization of the sedimentary clathrates by the release of pressure that resulted from the drop in sea level.

As far as the apparent correlation between the 70-m.y. fall in eustatic sea level and the drop in temperature over the same period are concerned, it now seems most reasonable to ascribe both phenomena to the same set of processes within the Earth and not to attempt to forge a causal relation between them. The high sea levels of late Cretaceous time were most probably caused by intense oceanic lithosphere formation and seafloor spreading from the mid-ocean ridges (see Harrison, Chapter 8). This same process resulted in perhaps a tenfold higher level of atmospheric CO_2 and a correspondingly warmer climate, perhaps 10°C warmer (Berner *et al.*, 1983).

FORECASTING CHANGES IN SEA LEVEL RELATED TO GREENHOUSE GASES

On a time scale of 10^2 to 10^5 yr, global changes in sea level (eustatic changes) can result mainly from the buildup or decay of alpine or continental glaciers, and from long-term ocean volume or steric changes caused by temperature or salinity changes in waters below the thermocline. The concern here is with forecasting eustatic changes related to the rise of CO_2 and other greenhouse gas concentrations in the atmosphere.

The past few thousand years have been a time of a high and relatively stable stand of sea level after 100 millennia of rapidly varying levels during the last ice age. Regional steric and other irregular variations of 10 to 20 cm from year to year, or from decade to decade, have been common, but there has been at most only a small unequivocally detected long-term trend in eustatic sea level. This situation can be expected to change with the advent of greenhouse-gas-induced climate change.

Sundquist (Chapter 12) examined the probable future course of atmospheric CO_2 concentrations over the next 1000 yr. It might be expected that oceanic biogeochemical processes related to the dissolution of calcium carbonate sediments on the deep-sea floor would, within a few centuries, reduce the content of free CO_2 in seawater, and hence the atmospheric content. This turns out not to be so, provided sufficient CO_2 has been generated by fossil fuel combustion.

The total remaining reserves of coal, oil, natural gas, oil shale, and oil sands that are ultimately recoverable for human use are believed to correspond to about 7500 billion tons of carbon. Of this amount approximately 3500 billion tons have already been identified. Sundquist assumes that 2500 billion tons will ultimately be consumed in human activities, with a peak rate of production in the middle of the next century of 16.8 billion tons/yr compared to the annual carbon emissions in 1985 of 6 billion tons. He assumes that the production rate will decline to 6 billion tons by about A.D. 2150 and go nearly to zero by A.D. 2350. With this assumed history of hydrocarbon combustion, Sundquist finds that the

CO_2 content of the air rises from 350 parts per million by volume (ppmv)(700 billion tons) in 1985 to 800 ppmv (1600 billion tons) around A.D. 2160, and slowly declines thereafter to 550 ppmv (1100 billion tons) by A.D. 2700. The peak concentration is approximately 2.9 times the base concentration of 280 ppmv in 1880, from which increases of atmospheric CO_2 content are usually calculated. Also to be taken into account are the increasing concentrations of methane, nitrous oxide, tropospheric ozone, and other minor greenhouse gases that can be expected 150 yr from now.

General circulation models of the atmosphere constructed by the National Center for Atmospheric Research, the NASA Goddard Institute of Space Sciences, the Geophysical Fluid Dynamics Laboratory of NOAA, and others indicate that the estimated increased concentrations of greenhouse gases should result in an average global temperature rise of 3° to 6°C in the atmosphere near the Earth's surface in the next 100 yr. [Recent modeling studies show that more sophisticated parameterization of clouds, taking into account both ice and liquid droplets, greatly reduces the sensitivity of the climate models to increased CO_2 (Cess *et al.*, 1989; Mitchell *et al.*, 1989). These newer models indicate a smaller temperature increase than the 3° to 6°C range given above.]

The next question to ask is: How will this atmospheric temperature change affect the world's oceans. This question has recently been studied by Frei *et al.* (1988). They use two kinds of models: a "pure diffusion" (PD) model in which heat is carried downward by eddy diffusion, assuming vertical diffusion coefficients of 1.3 and 2 cm^2/s and a modified diffusion model in which cold, polar water sinks to the bottom of the ocean and mass is conserved by assuming a slow, global upwelling. Frei *et al.* (1988) call this model an upwelling-diffusion (UD) model; they assume that the coefficient of vertical eddy diffusion is about 0.65 cm^2/s and that the global average upwelling rate to balance deep and bottom water formation is about 4 m/yr. A considerably smaller rate of upwelling and, correspondingly, a higher rate of downward vertical diffusion would correspond to estimates by Whitehead (1989) that cold deep and bottom water is formed at a rate of only 5 million to 10 million m^3/s instead of the rate of 40 million m^3/s assumed by Frei *et al.* (1988). Measurements of tritium distribution in the North Atlantic made by the "GEOSECS" Expedition in 1972 (Ostlund *et al.*, 1974) and 10 yr later by the "Transient Tracers in the Ocean" Expedition (PCODF, 1981; Ostlund, 1983) indicate that the tritium "front" in the Atlantic deep water between depths of 2500 and 5000 m moved about 800 km south during this 10-yr period, indicating that the mass of deep water sinking in the Norwegian Sea and cascading downward through the Denmark Strait, is 10 million to 20 million m^3/s. Hence, the rate of upwelling is probably smaller and the rate of downward diffusion greater than assumed by Frei *et al.* (1988).

Basically, a PD model transports heat relatively rapidly into the oceans, which slows the atmospheric temperature response to the rising CO_2 concentration but increases the rate of sea-level rise. The UD model reduces heat penetration into the ocean, allowing the climate to warm relatively rapidly but reducing the sea-level response.

Frei *et al.* estimate that the rise in sea level during the past 100 yr caused by thermal (steric) expansion was between 3 and 8 cm, with the lower value corresponding to the UD model. They project a rise of 10 to 50 cm during the next century resulting from thermal expansion, the range arising from uncertainties in CO_2 and trace gas concentrations and in estimates of climate sensitivity to greenhouse warming.

To this estimate of steric sea-level rise in the next century must be added the NRC (1985) Committee on Glaciology's estimate of the contribution to sea-level rise by ice wastage in a CO_2-enhanced environment. This would come from three sources: glaciers and small ice caps, the Greenland Ice Sheet, and the Antarctic Ice Sheet. This NRC committee estimates a sea-level rise by the year 2100 (the assumed time for a doubling of atmospheric CO_2) from ice wastage of 0.1 to 1.6 m with "most likely" values of 0.2 to 0.9 m; the "most likely" scenario can be expressed as 0.55 ± 0.21 m if one assumes that this range expresses a standard deviation from the mean and if the "errors" are considered to be independent.

Revelle (1983b), using a two-dimensional vertical diffusion model for ocean thermal expansion, estimated a total sea-level rise of about 0.7 m. Also, in 1983, Hoffman *et al.* (1983) forecast a larger global rise, between 1.44 m and 2.17 m by A.D. 2100. This was estimated to result from thermal expansion of ocean waters and from ablation and partial melting of alpine glaciers and the ice caps of Antarctica and Greenland. Of the total rise, an average 0.72 m was estimated to result from ocean thermal expansion, and 0.72 to 1.45 m were added from ice discharge to maintain the relative contributions thought to have contributed to sea-level rise over the past 100 yr. Because the range of rise related solely to ocean thermal expansion was calculated to be 0.28 to 1.15 m, the ratio approach led to an extreme upper limit of more than 3 m (of this amount, mountain glacier melting could contribute, at most, 0.3 to 0.5 m). Robin (1987) forecast a rise of 0.80 m by A.D. 2100 with a range of 0.20 to 1.65 m. Gornitz *et al.* (1982) calculated the component of sea-level rise resulting from ocean thermal expansion between 1980 and 2050 as 0.20 to 0.30 m. MacCracken *et al.* (1989), incorporating the Frei *et al.* (1988) model results and, with some adjustments, the findings of Revelle (1983b), estimate a total sea-level change by the year 2100 of less than 0.5 to 1 m.

In all these estimates, the possible change, largely a result of human activities, in the quantity of water stored in lakes, man-made reservoirs, and underground aquifers has been neglected. Robin (1987) estimated the net effect of these three sources together at the present time as causing an annual rise in sea level of 0.08 mm, or 8 mm/century.

Most calculations indicate that sea level will continue to rise for at least several hundred years at average rates of centimeters per year or less. Of course, a considerably more rapid rate would ensue if the West Antarctic ice cap should disintegrate during the next 1000 yr. As we have seen, such a dissolution is probably impossible during the next several hundred years because, as Bryan *et al.* (1988) have shown, deep ocean convection on the southern side of the great circumpolar Antarctic Current may markedly delay much Antarctic temperature change in the upper ocean layers.

The assumed rate of carbon combustion is highly uncertain. In order to reach an atmospheric level of 800 ppmv by A.D. 2180, an average rate of CO_2 production corresponding to 9 billion tons of carbon per year during the next 200 yr is assumed. This is at least 50 percent higher than the present rate of 6 billion tons/yr, which is estimated to include tropical deforestation of greater than about 1 billion tons/yr (Machta, 1983). On the one hand, economic and social development of the now developing countries, which make up the vast majority of mankind, may well require a considerable increase above present levels of carbon combustion, and consequently increase the worldwide rate, perhaps by a factor of 3 or more. On the other hand, as Goldemberg *et al.* (1987), Mintzer (1987), and Bach (1988) have persuasively argued, it may be possible, through increases in energy use efficiency and substitution of other energy sources for fossil fuels, to reduce considerably the influx of CO_2 to the atmosphere and ultimately to the oceans. With these alternative energy sources in use, the atmospheric burden of CO_2 might remain at all times much below our estimated figure of 800 ppmv, and the consequent rise in air and sea temperature and steric sea level would be considerably reduced.

In these calculations it has been tacitly assumed that the circulation of the deep water of the world ocean will continue relatively unchanged, except that the deep water will be somewhat warmer because of vertical and lateral mixing. In other words, the future ocean circulation will be "surprise free." However, the warming of the atmosphere in high latitudes, and the corresponding warming of the subsurface ocean waters, may greatly reduce the volume of water sinking to the depths. The projected rise in sea level from steric expansion would then be intermediate between the range estimated from the UD model (10 to 50 cm), and the range computed from the PD model (20 to 110 cm).

HOW CAN WE IMPROVE THE MEASUREMENT OF SEA-LEVEL CHANGE?

The apparent trend of sea level at a particular place as measured by a tide gauge is the sum of trends in motion of the gauge itself as the land on which it is mounted moves vertically, the trend of change in steric sea level, and the trend of change in water mass under the tide gauge. To understand what is happening one needs to be able to make measurements that will separate these three components of the observed sea level. This problem is addressed by Munk *et al.* (Chapter 14). Each component of the observed sea level is considered separately.

The vertical motion of the tide gauge can be measured with fair accuracy in four different ways: by measuring changes in the acceleration of gravity at the sea surface under the gauge, by very-long-baseline interferometry (VLBI), by satellite laser ranging, and by use of the satellite signals of the global positioning system (GPS). Without repeated measurements over several years, none of these methods is sufficiently accurate to determine the vertical position of the tide gauge.

The acceleration of the Earth's gravity, g, can be measured with an uncertainty of about 1 part in 10^8 by the methods described in Chapter 14. This accuracy corresponds to a sensitivity to height changes of the tide gauge of about 30 mm. VLBI observing stations yield estimates of intercontinental baselines with an rms scattering of 20 to 30 mm. Up to the present, the scatter of the vertical components has been 3 or 4 times larger than this. One of the major sources of error is atmospheric refraction caused by water vapor. Improved water vapor radiometers are being developed and placed in operation. The rms scatter of all components of the baseline should then be reduced to the 10- to 20-mm level. However, reducing the scatter below the 10-mm level may be very difficult (Carter *et al.*, 1986).

Global Positioning System surveys of benchmarks separated by 8 to 50 km agree in the height components by ±10 to 30 mm. Carter *et al.* (1986) believe that these results are about as good as can be expected reliably for the foreseeable future. However, there are plans for increasing the accuracy of space-based geodetic techniques. It is possible that satellite laser ranging (SLR) can improve position accuracy from the present 1-cm value to 1 mm in the next decade. If vertical positions can achieve this accuracy, we should be in a position to improve our knowledge of eustatic change considerably. The Geodynamic Laser Ranging System (GLRS) of the Earth Observing System (EOS) will allow a large number of tide gauges to be included in the network.

At present, the single measurement errors in all four methods of measuring the changes in the elevation of the tide gauge are comparable at ±10 to 30 mm, and it seems likely that they will remain for some time to come. In contrast, it is desirable that the motion of the tide gauge be determined to a fraction of 1 mm/yr. Hence observational strategies will have to rely on repetitive measurements spanning intervals of several years and even then the desired accuracy can only be achieved by GLRS.

The change in the steric height of sea level can perhaps best be monitored with bottom-mounted upward-looking fathometers plus tide gauges. (The alternative of measuring bottom pressure with sufficient accuracy presents many difficulties, because of seemingly inevitable unpredictable drift of the pressure gauges at pressures of a few tens of atmospheres.) Because of the marked variation in the velocity of sound in water with changes in temperature (some 23 times larger than the changes of specific volume or density with temperature), it should be quite practical to estimate the integrated changes in temperature over the water column above the fathometer. The main problems in measuring steric changes, as Roemmich (Chapter 13) points out, are that they tend to extend over great depths and are not confined to or concentrated in the upper ocean. In the subtropical North Atlantic, steric changes extend to at least 3000 m with maxima in the thermocline at depths of 300 to 700 m and below the thermocline at about 1800 m. Hence the inverted fathometer is likely to be most useful in the depths of the open ocean far from land. The logistical problems of maintaining operating fathometers in such depths and locations and combining their measurements with surface mounted tide gauges may be difficult to solve.

In principle, however, the combination of inverted echo sounders plus one or more of the four methods for measuring the vertical motions of the tide gauge, plus tide-gauge measurements at the sea surface should allow us to separate the 3 major components of changes in RSL (volume of water, mass of water, and changes in the elevation of the tide gauge).

A further difficulty arises, as Roemmich shows in Chapter 13: steric height variations occur over such a range of space and time scales that possible trends cannot be identified, even in a 30-yr time series at a single station, no matter how dense the sampling. However, the major steric changes seem to be relatively coherent over distances of several thousands of kilometers. Munk *et al.* (Chapter 14) therefore suggest that 10 spatially independent stations, each forming 5 independent samples over a 25-yr period, could be combined to give a ±7-mm standard deviation in a long-term trend of 10 to 25 mm in steric level. It should be possible within the next few years to measure change in sea level with the precision of 1 cm or less from satellite altimeter measurements such as those planned for the TOPEX/POSEIDON experiment (B. Tapley, University of Texas, personal communication, 1989). Born *et al.* (1986) listed the magnitude of different sources of error in such measurements. The accuracy of the measurements will be limited by the calibration of the altimeter and of the estimated heights of fixed points on land. The effects of ionospheric and tropospheric refraction can be eliminated through use of two-frequency altimeters and a water vapor radiometer in the satellite. The drift of the altimeter, and not the absolute calibration, is important for measurement in changes of sea level during the life of the instrument.

RECOMMENDATIONS

1. Long-term sea-level measurements of sufficient accuracy over the world's oceans could provide one of the most significant data sets for understanding global change, particularly climatic change resulting from the greenhouse effect. It is for this reason that the planning committees for the World Climate Research Program and the Intergovernmental Oceanographic Commission of UNESCO have given a very high priority to extending the global sea-level network in the Indian, South Atlantic, and South Pacific oceans. This effort is being supported, insofar as available funds will allow, by the U.S. National Oceanic and Atmospheric Administration (NOAA). **We strongly recommend that the national oceanographic and meteorological communities lend moral and intellectual support to this sea-level program and to develop satellite altimeter methods for changes in sea level.**

2. **A polar orbiting satellite equipped with a radar, preferably a laser, altimeter should be operated on a continuing basis to measure changes in volume of the Antarctic and Greenland ice sheets.** These ice sheets may be the principal sources of variations in sea level during the next century.

As reviewed by the Topographic Science Working Group (1988), detailed and repeated height measurements by near polar-orbiting satellites are required to study the mass balance and dynamics of ice sheets. Repeat surveys of the ice sheets at 1- to 5-yr intervals with a vertical resolution of 10 cm are required for determination of elevation changes indicative of changes in ice volume, thus providing a measurement of net mass balance. Only refined radar altimeters or laser altimeter systems are capable of global coverage with the requisite accuracy.

The output of ice in the mass balance equation occurs through iceberg discharge, surface melting near the margin, melting at the bottom of ice shelves, evaporation, and ablation. A 1-m difference in surface elevation of the ice shelves reflects a nearly 10-m difference in ice shelf thickness (Topographic Science Working Group, 1988). The position of the grounding line can also be observed in elevation data because of a marked change in slope. The output of the ice sheets is not known better than 30 to 100 percent of the total snow accumulation, thus its measurement is critical in an assessment of the mass balance of the ice sheets.

3. A geological record of sea-level change is well preserved in numerous basins for at least the past 200 m.y., a span that includes the nearly ice-free Cretaceous Period and the present ice age. Comparison of the record between glacial and nonglacial times will provide an improved understanding of how depositional systems respond to sea-level change, as well as insights about nonglacial mechanisms of sea-level change.

Support of programs, both national and international, that address the questions of the sea-level record during the past 250 m.y. should be vigorously pursued. One such program, the Global Sedimentary Geology Program (of the International Union of Geological Sciences) on Cretaceous Resources, Events and Rhythms, addresses a variety of questions that we have raised, viz., (a) Is there a global correlation of sequences? (b) Are sequences caused by eustatic fluctuations and/or global tectonic variations, or are sequences developed as a result of regional and local tectonic adjustments? (c) What are the relationships between subsidence, sea level, sediment supply, erosion, and other factors in mid-Cretaceous sedimentary basins?

4. To improve estimates of future steric changes in ocean volume caused by greenhouse warming of the ocean water, **coupled ocean-atmospheric general circulation models should be improved and used to trace probable changes in ocean and atmospheric temperature as the greenhouse gas concentrations in the atmosphere gradually increase.**

5. **To measure absolute elevation of tide gauges, measurements of position using satellite laser ranging, the global positioning system, and very-long-baseline interferometry techniques and absolute gravity should be started.**

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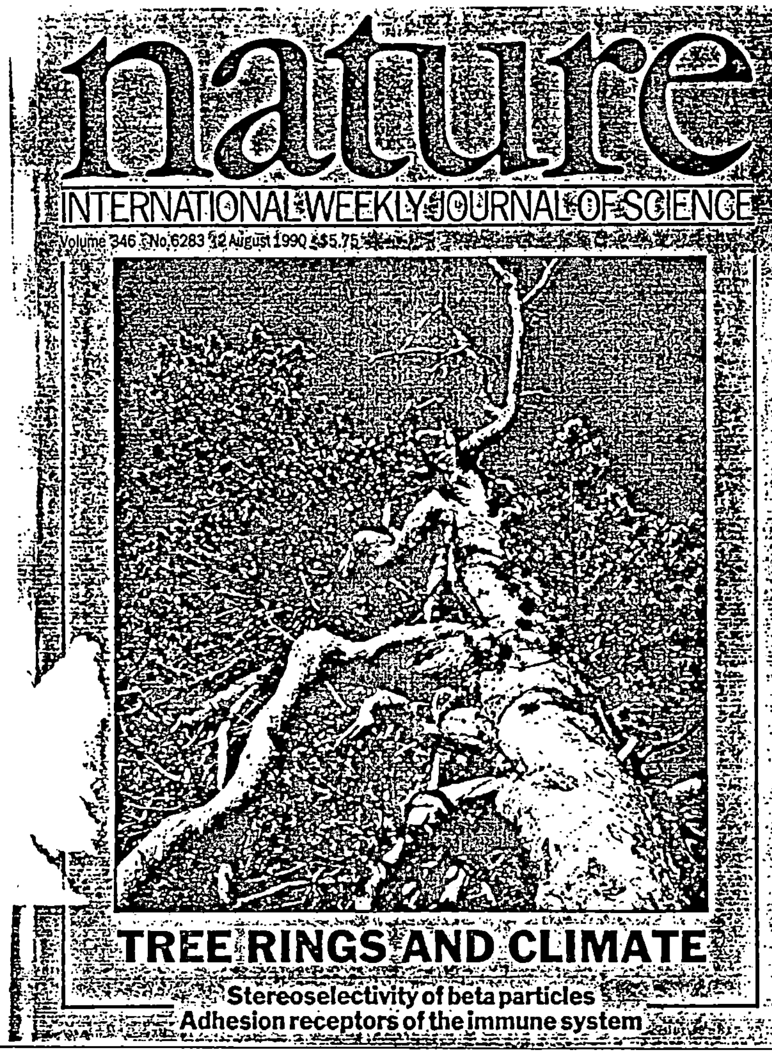

ARTICLES

A 1,400-year tree-ring record of summer temperatures in Fennoscandia

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Tree-ring data have been used to reconstruct the mean summer (April–August) temperature of northern Fennoscandia for each year from AD 500 to the present. Summer temperatures have fluctuated markedly on annual, decadal and century timescales. There is little evidence for the existence of a Medieval Warm Epoch, and the Little Ice Age seems to be confined to the relatively short period between 1570 and 1650. This challenges the popular idea that these events were the major climate excursions of the first millennium, occurring synchronously throughout Europe in all seasons. An analysis of past warming trends suggests that any summer warming induced by greenhouse gases may not be detectable in this region until after 2030.

THE collection of living and remnant Scots pine (*Pinus sylvestris* L.) from the Torneträsk region in northern Sweden has been reported previously[1]. This material has yielded absolutely dated series of raw mean ring widths[2] and maximum latewood densities[3] extending unbroken from the present back to AD 436 and AD 443, respectively. At high latitudes, both of these variables are known to correlate well with climate data (specifically, summer temperatures[4-7]), but only qualitative inferences based on these particular long ring-width and density chronologies have been made to date[3,8]. Here we describe the reprocessing of the raw ring-width and density data to produce new chronologies and how they have been used to derive yearly estimates of mean Fennoscandian temperatures for a statistically defined 'summer' (April–August) season, reaching back to AD 500. These are the longest annually resolved climate reconstructions from tree rings yet published.

Torneträsk tree-ring chronologies

Standardization in dendroclimatology is a method for removing variability in tree-ring series that is not related to climate. It involves removing part of the low-frequency variability from the series of raw measurements[9]. The rationale behind this is a need to account for differences in the general growth rate or vitality of individual trees and to remove trends in individual tree-ring series that arise as a consequence of non-climate-

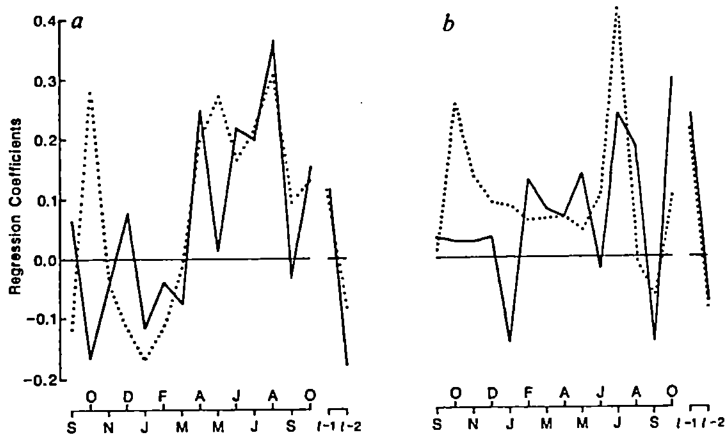


FIG. 1 Patterns of growth response (*a*, maximum latewood density; *b*, total ring width) in Torneträsk pine to intra-annual temperature forcing for the periods 1876–1925 (dotted line) and 1926–1975 (solid line), derived using multiple regression. The predictors include 14 monthly temperature series for the successive months from September of the year preceding ring formation, through to October immediately following the growing season. Two other variables, the relevant ring width or maximum latewood density for the two preceding years ($t-1$ and $t-2$) are also included to quantify the strengths of any biological preconditioning of growth in one summer by conditions in previous seasons.

related growth processes such as tree ageing or forest-stand development. Various deterministic, digital filtering and time-series approaches may be used[10]. Just what fraction of low-frequency variability is lost depends on the method applied. As a general rule, if efficient methods are used, the longer the constituent series, the longer is the timescale of potential climate variability that can be reconstructed from the mean standardized chronology.

The 65 ring-width and 65 maximum-density series that make up the chronologies produced here range in length from 92 to 612 years with a mean of 290 years (27 series are more than 300 years long). We have standardized the constituent data series using smoothing splines with a frequency-response cutoff set at two-thirds of the length of each series[10]. For example, in a 300-yr series, most of the variance above 200 years would be lost. The mean length of the splines fitted to all of the series is 194 years, but a number are long enough to preserve significant variance on timescales up to ~300 years. Although this standardization procedure effectively precludes any possibility of reconstructing climate variability on timescales much over 300 years, it is capable of preserving much of the variability on timescales approaching this.

The statistical reliability of a chronology is a function of the common growth signal (measured as the mean inter-series correlation coefficient) and the number of series (or replication depth)[11]. The best-replicated sections of the ring-width and density chronologies (the data from the nineteenth and twentieth centuries), explain over 90% of the variance of a hypothetical infinitely replicated chronology. This figure remains above 77% back to about 550 and is only marginally worse before this. The chronologies may therefore be considered statistically reliable and homogeneous time series over virtually their whole lengths.

Temperature effects

To define an optimum season for reconstruction and to help suggest the form of linear regression equation with which to model it, we carried out temperature-response function analysis for both the ring-width and densitometric chronologies[9]. This involves regressing principal components of the monthly temperature and previous growth data on the annual tree growth indices, to derive a set of regression coefficients that indicate the direction and relative strength of the temperature forcing on tree growth. The temperature data used were area averages formed by merging gridded data from 65° and 70° N at 10°, 20° and 30° E, producing mean monthly anomalies with respect to a base period of 1951–70[12]. The resulting series is considered homogeneous back to 1876.

Response functions were calculated and verified using data from 1876–1975, split into equal halves (see Fig. 1 and Table 1). Similar patterns of response are apparent over the two subperiods, although the similarity and strength of the response is somewhat greater in the case of the densitometric data. The largest regression coefficients show that maximum density is apparently enhanced by warm conditions during all of the months from April to August. The same is generally true for ring widths, except that the regression coefficients for April, May and June are noticeably smaller than those for density and the biggest width response is in July and August. The July temperature coefficient is the largest. The importance of July temperature in the growth of high-latitude pines is well documented[6,13,14].

These results suggest that April–August mean 'summer' temperatures may be usefully reconstructed using these tree-ring data. The relatively large coefficient on the previous year's ring width in the response function, which reflects the significant one-year-lag autocorrelation (r_1) in the ring-width series (ring width $r_1=0.38$, density $r_1=0.15$), means that a lagged ring-width variable must be included in the regression model for climate reconstruction. We therefore used a regression equation of the form

$$T_t = b_1\,\mathrm{TRW}_t + b_2\,\mathrm{TRW}_{t+1} + b_3\,\mathrm{MXD}_t + b_4\,\mathrm{MXD}_{t+1} \quad (1)$$

where T_t is the mean April–August temperature of northern Fennoscandia in year t; TRW_t and TRW_{t+1} are the ring-width data in the years t and $t+1$ respectively and MXD_t and MXD_{t+1} are the equivalent maximum latewood densities.

Statistical reconstruction equation

Having established the potential for reconstructing the 5-month summer season (April–August) and chosen the form of regression model (equation (1)) we undertook a cross-calibration/verification exercise to test the general form of regression equation to be used for the reconstruction back to 500. The mean summer temperature series was divided into two 50-year periods 1876–1925 and 1926–1975. Equation (1) was then fitted over one period using a principal-component regression technique[15] and the derived coefficients applied to the tree-growth data over the other period to give estimates of temperature which could be compared with the observed climate data (see Table 2). This process was then repeated with the periods reversed. The early- and late-calibrated regression equations are very similar with a large positive weight on current-year maximum density. The results of the verification tests show that 44–49% of the independent temperature variance is recovered in the estimated data. All of the verification tests give results that are comparable to the best tree-ring-based climate reconstructions made to date.

Having satisfied ourselves of the validity of our general regression model we re-calibrated the reconstruction equation using all 100 years of climate data so as to maximize the timescale of variability against which the final regression equation could be fitted. The recalibrated regression coefficients (Table 2) are

TABLE 1 Response function analyses on Torneträsk chronologies

	Ring width			Maximum latewood density		
Calibration period	1876–1925	1926–1975	1876–1975	1876–1925	1926–1975	1876–1975
Verification period	1926–1975	1876–1925	—	1926–1975	1876–1925	—
Calibration						
R^2	0.49	0.32	0.51	0.66	0.58	0.56
R^2_{adj}	0.42	0.24	0.48	0.59	0.50	0.51
Verification						
r^2	0.31	0.29	—	0.37	0.44	—
RE	0.54	0.61	—	0.40	0.53	—
CE	0.09	0.13	—	0.25	0.44	—
PMt	5.3	3.6	—	3.3	4.1	—
First difference sign						
correct	32	27	—	35	31	—
incorrect	15	20	—	11	16	—

Significant regression coefficients ($P=0.05$) and their signs

Oct. ($t-1$)	+	Feb. ($t-1$)	+	Oct. ($t-1$)	+	Oct. ($t-1$)	−	Oct. ($t-1$)	−	Jan. (t)	−
Nov. ($t-1$)	+	Apr. (t)	+	Feb. (t)	+	Jan. (t)	−	Jan. (t)	−	Apr.	+
Jan. (t)	+	July	+	Mar.	−	Apr.	+	Apr.	+	May	+
Mar.	−	Oct.	+	May	+	May	+	June	+	June	+
Apr.	+			June	+	July	+	July	+	July	+
June	+			July	+	Aug.	+	Aug.	+	Aug.	+
July	+			Aug.	+	Oct.	+	Oct.	+	Density ($t-2$)	+
				Oct.	+						
Ring width ($t-1$)	+			Ring width ($t-1$)	+						

R^2, square of the correlation coefficient calculated between actual and the estimated data; R^2_{adj} is the R^2 value adjusted to take account of the effective number of predictors; r^2 is the square of the actual/estimated correlation over the verification period; RE is the reduction of error[9]; CE of the coefficient of efficiency[7]; PMt is the t value derived using the product mean test and first difference sign is a test for comparing the signs of the first differences for the estimated and actual data[9].

essentially the same as those for the two subperiods. The three possible reconstructions (based on the early, late and overall periods) have ≥95% variance in common over the period 500–1980. Supporting evidence for the reliability of our temperature estimates comes from comparisons with fragmentary instrumental temperature records[16]. These exist for a number of sites in northern Fennoscandia. At least 20 years of data in the first half of the nineteenth century exist for the three stations, Övertorneå (66.4° N 23.8° E), Hailuoto Carlö (65.0° N 24.7° E) and Vöyri (63.2° N 22.0° E). The correlations between these early data and our reconstructed series are 0.64 (1802–1832), 0.48 (1817–1836) and 0.73 (1800–1824), respectively. These values would be expected to occur by chance one time in twenty in the case of Hailuoto Carlö and one time in a thousand for Övertorneå and Vöyri. There is also a very short early record (1830–1838) for Karesuando, which is near to the Torneträsk tree sites. The correlation between these temperature data and our reconstructions, 0.85, has a chance probability of below one in a hundred.

Summer temperature reconstructions

The complete reconstruction from 500 is plotted as temperature anomalies from the mean of 1951–70 in Fig. 2*a*. This long record shows that alternating periods of generally cool and generally warm conditions are typical of northern Fennoscandian summers throughout the last millennium. It is also clear that the length and amplitude of these temperature 'cycles' has varied dramatically during this time (Fig. 2*b*–*d*). There are only two periods (on a timescale of decades and above) when temperatures remained near the long-term mean: one lasting through the sixth and seventh centuries and a much shorter period in the latter half of the thirteenth century.

The smoothed line superimposed on the recent section of the reconstructed curve in Fig. 2*a* shows 10-year filtered values of the instrumental record for north Fennoscandia. This smooth line can be used as a modern yardstick against which to compare the magnitude and duration of summer temperature changes over the past 15 centuries. The relative warmth of the 1930s and 1940s is clear. Measured summer temperatures over these 20 years were on average 0.56 °C higher than the 1951–70 mean whereas the mean anomaly for the preceding 50 years (1880–1929) was −0.31 °C. The reconstruction shows that this recent cycle is part of a quasi 80–90 year oscillation which has continued since ~1700 (see Fig. 2*d*).

The mean of the reconstructed values for 1930–49 is 0.31 °C, and that for 1880–1929 is −0.20 °C. There are eight periods (based on 20-year running means) with warmth at least

TABLE 2 Calibration of reconstruction equation

		1876–1925	1926–1975	1876–1975
Calibration Period		1876–1925	1926–1975	1876–1975
Verification Period		1926–1975	1876–1925	—
Calibration				
Explained Variance				
Overall		0.49	0.44	0.51
<10 years		0.49	0.53	0.51
>10 years		0.49	0.19	0.54
Verification				
Explained variance				
Overall		0.42	0.46	—
<10 years		0.48	0.49	—
>10 years		0.23	0.49	—
Reduction of error		0.56	0.59	—
Coefficient of efficiency		0.33	0.43	—
First-difference Sign test				
Correct signs		38	33	72
Incorrect signs		11	16	27
Regression weights (equation (1))	b_1	−0.026	0.122	0.081
	b_2	0.073	0.080	0.109
	b_3	0.675	0.536	0.600
	b_4	−0.014	0.113	0.052

Summary of the regression results for separate 50-year calibrations and the final reconstruction equation fitted over the full 100 years of overlap between the temperature and tree-ring data.

equivalent to 1930–49 persisting over two or more decades in the extended record, 750–780, 920–940, 960–1000, 1160–1190, 1400–1440, 1540–1570, 1750–1780 and 1840–1860. The warmest 20-year means occurred in the twelfth and the eighteenth centuries (Table 3). The second half of the twelfth century was also the warmest 50-year period (an anomaly of 0.35 °C over 1152–1201). Periods of cool conditions similar to those that prevailed in the later decades of the nineteenth century include 780–830, 850–870, 1110–1160, 1330–1360, 1570–1620 and 1800–1820. Interestingly, both the warmest 20-year period and the coolest 20-year period occur in the twelfth century. Both the coolest and warmest 50-year periods occur in this century (Table 3).

The cycle of temperature change that occurred in the twelfth century (cooling to ~1140, rapid warming to ~1160 and subsequent cooling) is the most notable feature of the reconstruction (Fig. 2). The reconstructed anomaly of 1.59 °C in 1092 was the warmest up to that date. The lowest value in the whole record, −1.65 °C, occurs in 1139. Between these dates temperatures decrease rapidly and almost monotonically. The largest 50-year cooling trend in the record, −1.78 °C, occurred between 1090 and 1139. After this an even more dramatic warming took place. Within 30 years temperatures rose by >2 °C, with anomalies >1.6 °C occurring in 1161 and 1164. The period with the largest positive 50-year trend in the reconstruction is 1127–1176 with a value of 2.43 °C. From ~1160 temperatures again decreased linearly, a 50-year trend of −1.76 °C being registered from 1161–1210 (the third largest 50-year cooling of the record). Although the largest oscillation in the record, the temperature cycle that occurred through the twelfth century is not unique. A similar cycle occurred in the second half of the eighth and the early part of the ninth centuries (Fig. 2*b*–*d*).

Medieval Warm Epoch and Little Ice Age

It is widely believed that two periods with strongly contrasting annual temperatures occurred during the late Holocene. The first, thought to have been a period of prolonged warmth, has been described as the Medieval Warm Epoch or the Little Climatic Optimum. It is believed to have lasted, in Europe at

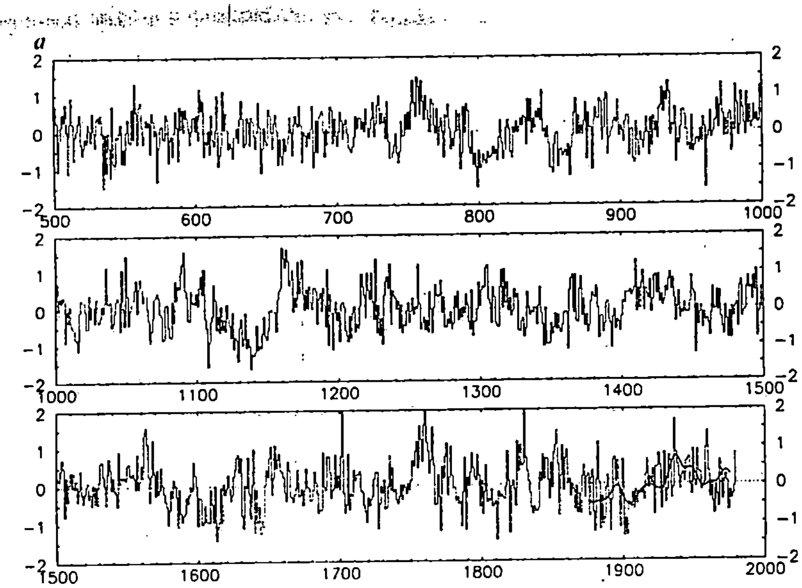
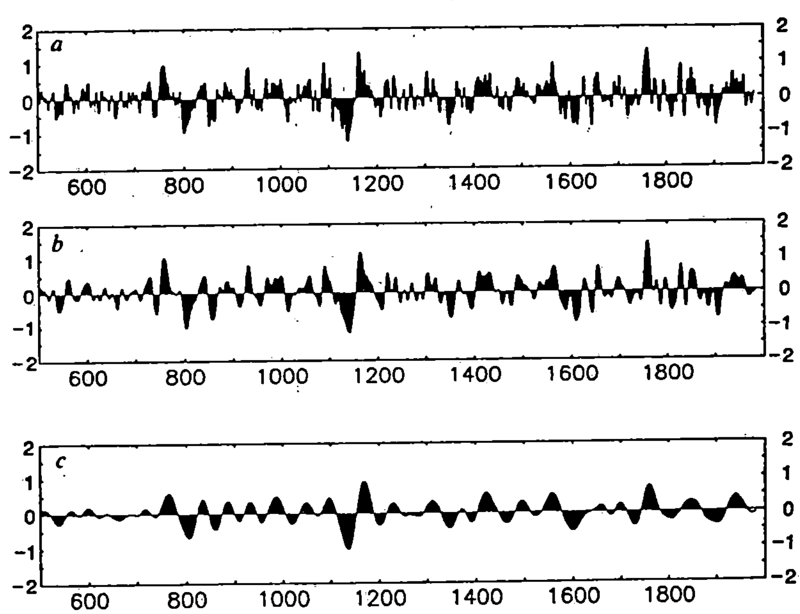


FIG. 2 *a*, Reconstructed northern Fennoscandian temperature anomalies (°C) (April–August mean) relative to 1951–70. Measured temperatures, smoothed with a 10-year low-pass filter, are also shown after 1876. *b*–*d*, The reconstructed summer temperatures after filtering with different gaussian and band-pass filters. *b*, 10-year low-pass values; *c*, variations between 25 and 40 years; *d*, variations between 50 and 150 years.



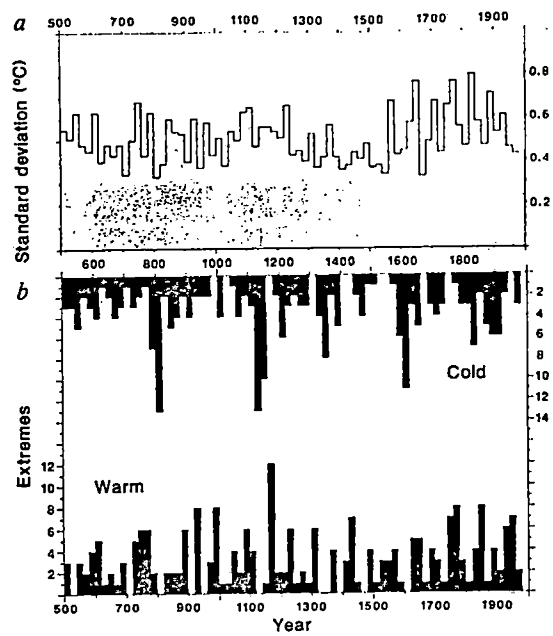


FIG. 3 *a*, Non-overlapping 20-year totals of negative and positive extremes. Extremes are defined as absolute anomalies ≥0.57 °C (one standard deviation of the full reconstructed series). *b*, Standard deviations calculated over the same periods.

least, from ~1000 to 1300[17]. Our reconstruction dispels any notion that summers in Fennoscandia were consistently warm throughout that period. Although the second half of the twelfth century was very warm the first half was very cold. For most of the eleventh and thirteenth centuries, summers were near normal (relative to the mean for 1951–70).

The other extensively cited period has been termed the Little Ice Age. During this time temperatures in many areas of both hemispheres are thought to have fallen to their lowest levels since the end of the last glaciation[17]. Perhaps not surprisingly, given the natural spatial variability of climate and the poor resolution of much of the data, there is confusion associated with the use of this term and there is considerable uncertainty as to the severity or synchroneity of the various cool events which have been ascribed to it[18]. Even in Europe the Little Ice Age has been said to have started at dates ranging between the twelfth to the sixteenth centuries and to have ended in the late seventeenth, the middle nineteenth and the early twentieth centuries[19]. There is, however, probably some concensus for a main European phase between 1550 and 1700 or 1800[17]. Our reconstruction offers strong evidence that summer temperatures in northern Fennoscandia did fall abruptly, to a level ~0.5 °C below the long-term mean, at ~1570. But they remained at this level only until 1650. Between 1600 and 1650, there were 13 summers with mean temperature anomalies below −1 °C. This is the largest number of such cold summers in any 50-year period in the reconstruction, approached only by the 12 that occurred between 1100 and 1150. The 1550s and the 1650s, however, were warm. Conditions were near normal from 1660 to 1750 and the 20 years between 1750 and 1770 were extremely warm, the equal warmest 20 years of the whole reconstruction (Table 3).

Much of the evidence for a Medieval Warm Epoch and the Little Ice Age comes from documentary historical data[17]. A more direct comparison with such data can be made by using counts of extremely warm or cold summers. When this is done (Fig. 3*a*), the results lead to essentially the same conclusions. Our reconstructed summer temperatures show little evidence for a Medieval Warm Epoch in Fennoscandia and the only evidence for the Little Ice Age comes from the period ~1570–1650. As regards the latter period, similar periods of cold summers occurred during the late eighth/early ninth century, the first half of the twelfth century and the second half of the fourteenth century. The first half of the twelfth century was almost certainly much more severe than the 1570–1650 period. Absence of a prolonged Medieval Warm Epoch or Little Ice Age in northern Fennoscandia is the most direct interpretation of our results. This implies either that northern Fennoscandia has experienced climate fluctuations that differed markedly from

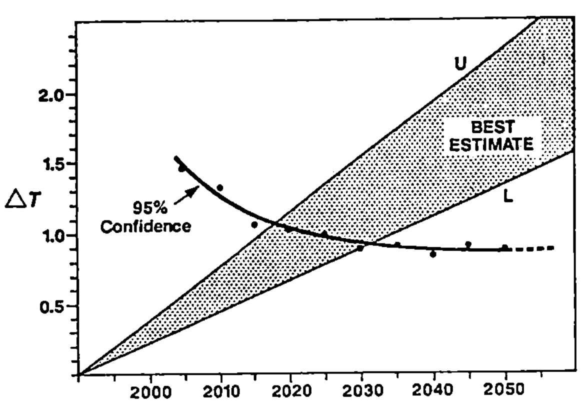


FIG. 4 The stippled area shows the 'best estimate' range of northern Fennoscandian summer temperature increase expected to result from increases in greenhouse gases (from ref. 21; some more recent estimates, unpublished, give lower values). The curved line can be considered a minimum threshold over which future summer temperature trends must pass before a greenhouse signal can be said to be distinguishable from natural climate variability at the regional scale. This detection threshold is drawn by hand through a series of empirically derived points each of which is the upper-95-percentile point of the distribution of trends of length up to that point on the horizontal axis. For example, the point plotted at 2030 (40 years after 1990) comes from the distribution of 40-year trends in the reconstruction. The illustrated threshold is a minimum because not all of the observed climate variance is reconstructed. Therefore, this graph probably provides lower-bound estimates of the detection time.

TABLE 3 Largest temperature anomalies and trends

Positive Temperature anomaly (°C)	Year	Negative Temperature anomaly (°C)	Year
Individual summers			
2.07	1761	−1.65	1139
2.05	1831	−1.58	1109
1.99	1703	−1.57	961
1.71	1937	−1.53	800
1.67	1161	−1.50	536
20-year means			
Positive Temperature anomaly (°C)	Period	Negative Temperature anomaly (°C)	Period
0.78	1158-1177	−0.93	1127-1146
0.78	1748-1767	−0.73	795-814
0.67	749-768	−0.65	1601-1620
0.52	1087-1106	−0.55	848-867
0.49	1551-1570	−0.53	1344-1363
50-year means			
Positive Temperature anomaly (°C)	Period	Negative Temperature anomaly (°C)	Period
0.35	1152-1201	−0.58	1108-1157
0.34	1403-1452	−0.42	1576-1625
0.33	1750-1799	−0.34	780-829
0.29	719-768	−0.31	1867-1916
0.27	962-1011	−0.23	1345-1394
50-year linear trends			
Positive Temperature change (°C)	Period	Negative Temperature change (°C)	Period
2.43	1127-1176	−1.78	1090-1139
1.77	1718-1767	−1.77	757-806
1.63	797-846	−1.76	1161-1210
1.17	850-899	−1.38	1560-1609
1.13	1385-1434	−1.27	1754-1803

the rest of Europe, or that the significance of these climate periods in Europe has been overstated, or both. In support of the latter possibility, we note that the proxy evidence for spatially coherent century-timescale climate fluctuations around the North Atlantic basin is relatively weak[18] and the historical evidence is equally sketchy owing to problems of data reliability[20].

Changes in interannual variability

It is commonly believed that the Little Ice Age was marked by high interannual temperature variablity. The same is often thought true of cool periods in general. We have tested this idea by calculating the standard deviations of non-overlapping 20-year periods (Fig. 3*b*) and comparing them with the means for the same 20-year periods. The expected relationship would yield a negative correlation. Actually, the correlation is positive (0.23) and marginally significant at the 5% level. The three reconsructed 20-year periods with largest standard deviations (in the range 0.77-0.80 °C) all have means that are either positive or near zero. The coolest reconstructed period during the Little Ice Age (−0.65 °C in 1601-20) has a standard deviation of only 0.45 °C. Interestingly, there does seem to have been a sharp increase in the magnitude and variability of the 20-year standard deviations for the period after the second half of the sixteenth century (Fig. 3*b*). This is, however, strongly accentuated by the distinctly low standard deviations for the 20-year periods in the preceding 300 years.

Greenhouse-warming detectability

Our extended reconstruction allows us to evaluate possible future changes in the climate of our study region in the context of natural variability, and so to address the issue of detection of greenhouse-gas-induced climate change at the regional level. We do this by comparing expected trends in summer temperature with those that have occurred over the past 1,481 years. A recent study[21] of possible greenhouse-gas-forced temperature change predicts that between 1990 and 2030 the global mean temperature will increase by 0.4-2.3 °C, with a 'best estimate' in the range 1.0-1.7 °C. Another study, integrating the regional patterns of climate change produced in a number of doubled CO_2 experiments using equilibrium general circulation models, suggests that the summer temperature change in northern Fennoscandia is likely to be somewhat less than this, between 0.75 and 1.0 times the annnual mean global change[22]. Therefore, a rough estimate of the expected summer temperature trend in northern Fennoscandia resulting from the greenhouse effect is somewhere between ~0.9 and 1.5 °C over the next 40 years.

We have used our reconstruction to generate the observed frequency distributions of trends over periods of different length up to 60 years, using only those trends that originate from the currently prevailing level, that is, within a range of temperatures 0.1 °C on either side of the long-term mean. This avoids distortions that may arise from warming and cooling trends that represent 'recovery' from very anomalous conditions. Plotting the upper-95-percentile point from each of the observed distributions of 15-, 20-, 25-, and so on, year trends, along with the range of predicted local greenhouse warming allows us to estimate when the predicted changes will rise above the 'noise' of natural variablity (Fig. 4).

Even if the climate sensitivity is at the top of the generally accepted range (U in Fig. 4), it will still not be possible to detect a regional greenhouse signal (with 95% confidence) until around 2020. If climate sensitivity is at the lower end of the range (L), as is indicated by the empirical evidence of recorded global temperatures over recent decades[23], then we will be unable to identify the signal in Fennoscandian temperatures with high confidence until some time after the year 2030. □

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RESEARCH PRIORITIES FOR FY92 BUDGET

I. Economic Growth and Structure Relating to Global Change (Regional and Worldwide)

A. Economic activities as one source of global change

B. Sensitivity of socio-economic consequences of global change to scientific, technological, demographic, and market conditions

C. Impacts of global change on environmental non-market assets

D. Market and individual adaptation to global change

E. Sources and sectoral direction of economic growth in developing and centrally planned economies

II. Technological Change Relating to Global Change

A. Technological change as a determinant of global change

B. Impacts of global change on technological change

C. Impacts of government action on technological change

D. Economic assessment of potential investment opportunities in research and development of technologies to prevent, offset, and adapt to global change

III. Government Policy Options and Global Change

A. Impacts of existing government policies and institutional structures on activities relating to global change

B. Government actions to adapt human activities to global change

C. Government action to prevent or abate physical system responses induced by economic activities

D. Government action to improve market allocation mechanisms

E. Interaction among governments addressing a common threat of global change: strategy, gaming, negotiation, compliance and enforcement

IV. Links Between Research in Economics and Other Disciplines

A. Scientific variables associated with global change that have significant economic implications

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[Being re-ordered somewhat]

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INTERGOVERNMENTAL PANEL ON
CLIMATE CHANGE

POLICYMAKERS SUMMARY

OF THE SCIENTIFIC ASSESSMENT OF CLIMATE CHANGE

WMO

UNEP

Report Prepared for IPCC by Working Group 1

June 1990

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EXECUTIVE SUMMARY

We are certain of the following:

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

- emissions resulting from human activities are substantially increasing the atmospheric concentrations of the greenhouse gases: carbon dioxide, methane, chlorofluorocarbons (CFCs) and nitrous oxide. These increases will enhance the greenhouse effect, resulting on average in an additional warming of the Earth's surface. The main greenhouse gas, water vapour, will increase in response to global warming and further enhance it.

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 enhanced 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. Continued emissions of these gases at present rates would commit us to increased concentrations for centuries ahead. The longer emissions continue to increase at present day rates, the greater reductions would have to be for concentrations to stabilise at a given level.

- the long-lived gases would require immediate reductions in emissions from human activities of over 60% to stabilise their concentrations at today's levels; methane would require a 15-20% reduction.

Based on current model results, we predict:

- under the IPCC Business-as-Usual (Scenario A) emissions of greenhouse gases, a rate of increase of global mean temperature during the next century of about 0.3°C per decade (with an uncertainty range of 0.2°C to 0.5°C per decade); this is greater than that seen over the past 10,000 years. This will result in a likely increase in global mean temperature of about 1°C above the present value by 2025 and 3°C before the end of the next century. The rise will not be steady because of the influence of other factors.

- under the other IPCC emission scenarios which assume progressively increasing levels of controls, rates of increase in global mean temperature of about 0.2°C per decade (Scenario B), just above 0.1°C per decade (Scenario C) and about 0.1°C per decade (Scenario D).

- that land surfaces warm more rapidly than the ocean, and high northern latitudes warm more than the global mean in winter.

- regional climate changes different from the global mean, although our confidence in the prediction of the detail of regional changes is low. For example, 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. There are less consistent predictions for the tropics and the southern hemisphere.

- under the IPCC Business as Usual emissions scenario, an average rate of global mean sea level rise of about 6cm per decade over the next century (with an uncertainty range of 3 - 10cm per decade), mainly due to thermal expansion of the oceans and the melting of some land ice. The predicted rise is about 20cm in global mean sea level by 2030, and 65cm by the end of the next century. There will be significant regional variations.

There are many uncertainties in our predictions particularly with regard to the timing, magnitude and regional patterns of climate change, due to our 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
- polar ice sheets which affect predictions of sea level rise

These processes are already partially understood, and we are confident that the uncertainties can be reduced by further research. However, the complexity of the system means that we cannot rule out surprises.

Our judgement is that:

- Global - mean surface air temperature has increased by 0.3°C to 0.6°C over the last 100 years, with the five global-average warmest years being in the 1980s. Over the same period global sea level has increased by 10-20cm. These increases have not been smooth with time, nor uniform over the globe.
- The size of this 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.
- There is no firm 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 affect climate, and will be affected by a changing climate and by increasing carbon dioxide concentrations. Rapid changes in climate will change the composition of ecosystems; some species will benefit while others will be unable to migrate or adapt fast enough and may become extinct. Enhanced levels of carbon dioxide may increase productivity and efficiency of water use of vegetation. The effect of warming on biological processes, although poorly understood, may increase the atmospheric concentrations of natural greenhouse gases.

To improve our predictive capability, we need:

- to **understand** better the various climate-related processes, particularly those associated with clouds, oceans and the carbon cycle
- to **improve** the systematic observation of climate-related variables on a global basis, and further investigate changes which took place in the past
- to **develop** improved models of the Earth's climate system.
- to **increase** support for national and international climate research activities, especially in developing countries
- to **facilitate** international exchange of climate data

Introduction: what is the issue ?

There is concern that human activities may be inadvertently changing the climate of the globe through the enhanced greenhouse effect, by past and continuing emissions of carbon dioxide and other gases which will cause the temperature of the Earth's surface to increase - popularly termed the "global warming". If this occurs, consequent changes may have a significant impact on society.

The purpose of the Working Group I report, as determined by the first meeting of IPCC, is to provide a scientific assessment of:

1) the factors which may affect climate change during the next century especially those which are due to human activity.

2) the responses of the atmosphere - ocean - land - ice system.

3) current capabilities of modelling global and regional climate changes and their predictability.

4) the past climate record and presently observed climate anomalies.

On the basis of this assessment, the report presents current knowledge regarding predictions of climate change (including sea level rise and the effects on ecosystems) over the next century, the timing of changes together with an assessment of the uncertainties associated with these predictions.

This Policymakers Summary aims to bring out those elements of the main report which have the greatest relevance to policy formulation, in answering the following questions:

- What factors determine global climate?
- What are the greenhouse gases, and how and why are they increasing?
- Which gases are the most important?
- How much do we expect the climate to change?
- How much confidence do we have in our predictions?
- Will the climate of the future be very different ?
- Have human activities already begun to change global climate?
- How much will sea level rise?
- What will be the effects on ecosystems?
- What should be done to reduce uncertainties, and how long will this take?

This report is intended to respond to the practical needs of the policymaker. It is neither an academic review, nor a plan for a new research programme. Uncertainties attach to almost every aspect of the issue, yet policymakers are looking for clear guidance from scientists; **hence authors have been asked to provide their best-estimates wherever possible**, together with an assessment of the uncertainties.

This report is a summary of our understanding in 1990. Although continuing research will deepen this understanding and require the report to be updated at frequent intervals, basic conclusions concerning the reality of the enhanced greenhouse effect and its potential to alter global climate are unlikely to change significantly. Nevertheless, the complexity of the system may give rise to surprises.

What factors determine global climate ?

There are many factors, both natural and of human origin, that determine the climate of the earth. We look first at those which are natural, and then see how human activities might contribute.

What natural factors are important?

The driving energy for weather and climate comes from the sun. The Earth intercepts solar radiation (including that in the short-wave, visible, part of the spectrum); about a third of it is reflected, the rest is absorbed by the different components (atmosphere, ocean, ice, land and biota) of the climate system. The energy absorbed from solar radiation is balanced (in the long term) by outgoing radiation from the Earth and atmosphere; this terrestrial radiation takes the form of long-wave invisible infra-red energy, and its magnitude is determined by the temperature of the Earth-atmosphere system.

There are several natural factors which can change the balance between the energy absorbed by the Earth and that emitted by it in the form of longwave infra-red radiation; these factors cause the **radiative forcing** on climate. The most obvious of these is a change in the output of energy from the Sun. There is direct evidence of such variability over the 11-year solar cycle, and longer period changes may also occur. Slow variations in the Earth's orbit affect the seasonal and latitudinal distribution of solar radiation; these were probably responsible for initiating the ice ages.

One of the most important factors is the greenhouse effect; a simplified explanation of which is as follows. Shortwave solar radiation can pass through the clear atmosphere relatively unimpeded. But long-wave terrestrial radiation emitted by the warm surface of the Earth is partially absorbed and then re-emitted by a number of trace gases in the cooler atmosphere above. Since, on average, the outgoing long wave radiation balances the incoming solar radiation, both the atmosphere and the surface will be warmer than they would be without the greenhouse gases.

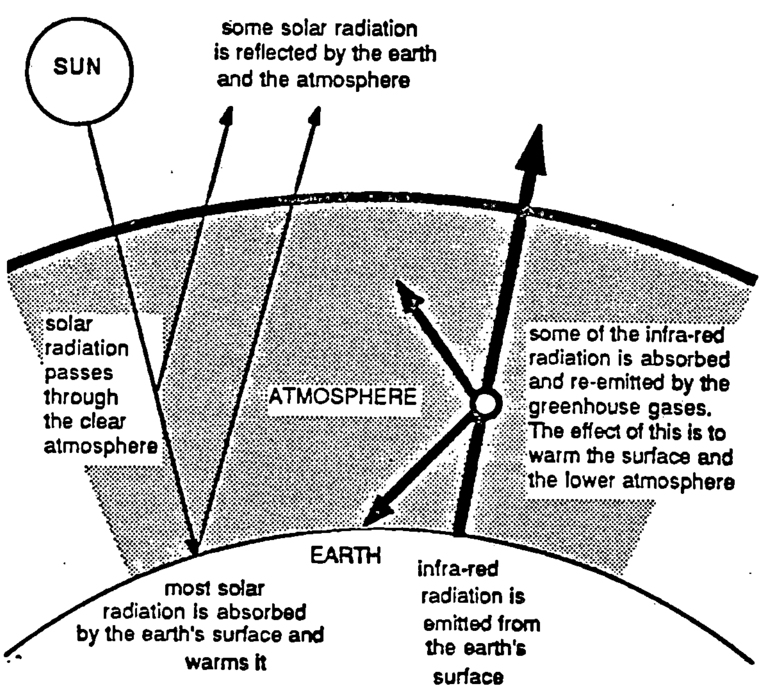
The main natural greenhouse gases are not the major constituents, nitrogen and oxygen, but water vapour (the biggest contributor), carbon dioxide, methane, nitrous oxide, and ozone in the troposphere (the lowest 10-15km of the atmosphere) and stratosphere.

Aerosols (small particles) in the atmosphere can also affect climate because they can reflect and absorb radiation. The most important natural perturbations result from explosive volcanic eruptions which affect concentrations in the lower stratosphere. Lastly, the climate has its own **natural variability** on all timescales and changes occur without any external influence.

How do we know that the natural greenhouse effect is real?

The greenhouse effect is real; it is a well understood effect, based on established scientific principles. We know that the greenhouse effect works in practice, for several reasons.

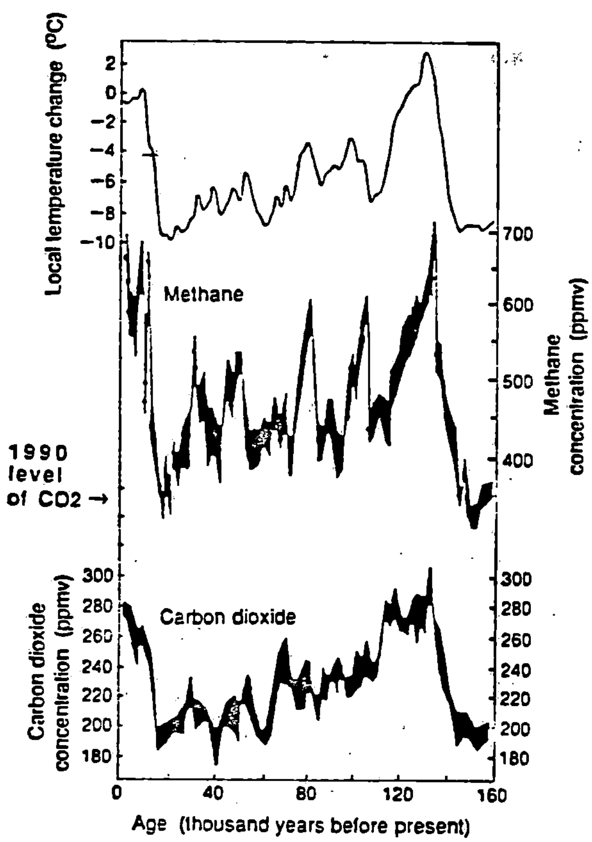
Firstly, the mean temperature of the Earth's surface is already warmer by about 33°C (assuming the same reflectivity of the earth) than it would be if the natural greenhouse gases were not present. Satellite observations of the radiation emitted from the earth's surface and through the atmosphere demonstrate the effect of the greenhouse gases.



A simplified diagram illustrating the greenhouse effect.

Secondly, we know the composition of the atmospheres of Venus, Earth and Mars are very different, and their surface temperatures are in general agreement with greenhouse theory.

Thirdly, measurements from ice cores going back 160,000 years show that the earth's temperature closely paralleled the amount of carbon dioxide and methane in the atmosphere. Although we do not know the details of cause and effect, calculations indicate that changes in these greenhouse gases were part, but not all, of the reason for the large (5-7°C) global temperature swings between ice ages and interglacial periods.



Analysis of air trapped in Antarctic ice cores shows that methane and carbon dioxide concentrations were closely correlated with the local temperature over the last 160,000 years. Present day concentrations of carbon dioxide are indicated

How might human activities change global climate ?

Naturally occurring greenhouse gases keep the Earth warm enough to be habitable. By increasing their concentrations, and by adding new greenhouse gases like chlorofluorocarbons (CFCs), humankind is capable of raising the global-average annual-mean surface-air temperature (which, for simplicity, is referred to as the "global temperature"), although we are uncertain about the rate at which this will occur. Strictly, this is an enhanced greenhouse effect - above that occurring due to natural greenhouse gas concentrations; the word "enhanced" is usually omitted, but it should not be forgotten. Other changes in climate are expected to result, for example changes in precipitation, and a global warming will cause sea levels to rise; these are discussed in more detail later.

There are other human activities which have the potential to affect climate. A change in the albedo (reflectivity) of the land, brought about by desertification or deforestation affects the amount of solar energy absorbed at the Earth's surface. Human-made aerosols, from sulphur emitted largely in fossil fuel combustion, can modify clouds and this may act to lower temperatures. Lastly, changes in ozone in the stratosphere due to CFCs may also influence climate.

What are the greenhouse gases and why are they increasing?

We are certain that the concentrations of greenhouse gases in the atmosphere have changed naturally on ice-age time-scales, and have been increasing since pre-industrial times due to human activities. The table below summarizes the present and pre-industrial abundances, current rates of change and present atmospheric lifetimes of greenhouse gases influenced by human activities. Carbon dioxide, methane, and nitrous oxide all have significant natural and human sources, while the chlorofluorocarbons are only produced industrially.

SUMMARY OF KEY GREENHOUSE GASES AFFECTED BY HUMAN ACTIVITIES	Carbon Dioxide	Methane	CFC-11	CFC-12	Nitrous Oxide
Atmospheric concentration	ppmv	ppmv	pptv	pptv	ppbv
Pre-industrial (1750-1800)	280	0.8	0	0	288
Present day (1990)	353	1.72	280	484	310
Current rate of change per year	1.8 (0.5%)	0.015 (0.9%)	9.5 (4%)	17 (4%)	0.8 (0.25%)
Atmospheric lifetime (years)	(50-200)†	10	65	130	150

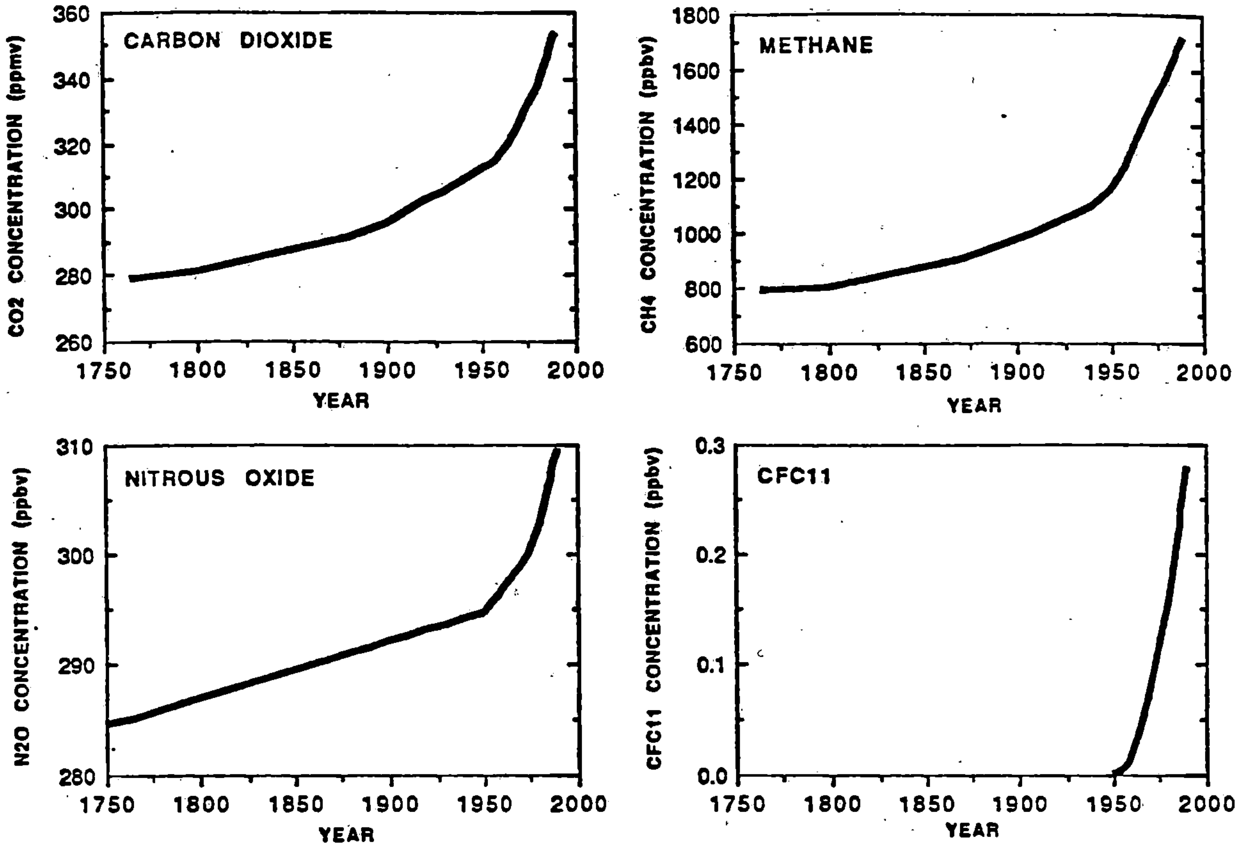
ppmv = parts per million by volume;
ppbv = parts per billion (thousand million) by volume;
pptv = parts per trillion (million million) by volume.
† The way in which CO_2 is absorbed by the oceans and biosphere is not simple and a single value cannot be given; refer to the main report for further discussion.

Two important greenhouse gases, water vapour and ozone, are not included in the table above. Water vapour has the largest greenhouse effect, but its concentration in the troposphere is determined internally within the climate system, and, on a global scale, is not affected by human sources and sinks. Water vapour will increase in response to global warming and further enhance it; this process is included in climate models. The concentration of ozone is changing both in the stratosphere and the troposphere due to human activities, but it is difficult to quantify the changes from present observations.

For a thousand years prior to the industrial revolution, abundances of the greenhouse gases were relatively constant. However, as the world's population increased, as the world became more industrialized and as agriculture developed, the abundances of the greenhouse gases increased markedly. The figures below illustrate this for carbon dioxide, methane, nitrous oxide and CFC-11.

Since the industrial revolution the combustion of fossil fuels and deforestation have led to an increase of 26% in carbon dioxide concentration in the atmosphere. We know the magnitude of the present day fossil-fuel source, but the input from deforestation cannot be estimated accurately. In addition, although about half of the emitted carbon dioxide stays in the atmosphere, we do not know well how much of the remainder is absorbed by the oceans and how much by terrestrial biota. Emissions of chlorofluorocarbons, used as aerosol propellants, solvents, refrigerants and foam blowing agents, are also well known; they were not present in the atmosphere before their invention in the 1930s.

The sources of methane and nitrous oxide are less well known. Methane concentrations have more than doubled because of rice production, cattle rearing, biomass burning, coal mining and ventilation of natural gas; also, fossil fuel combustion may have also contributed through chemical reactions in the atmosphere which reduce the rate of removal of methane. Nitrous oxide has increased by about 8% since pre-industrial times, presumably due to human activities; we are unable to specify the sources, but it is likely that agriculture plays a part.



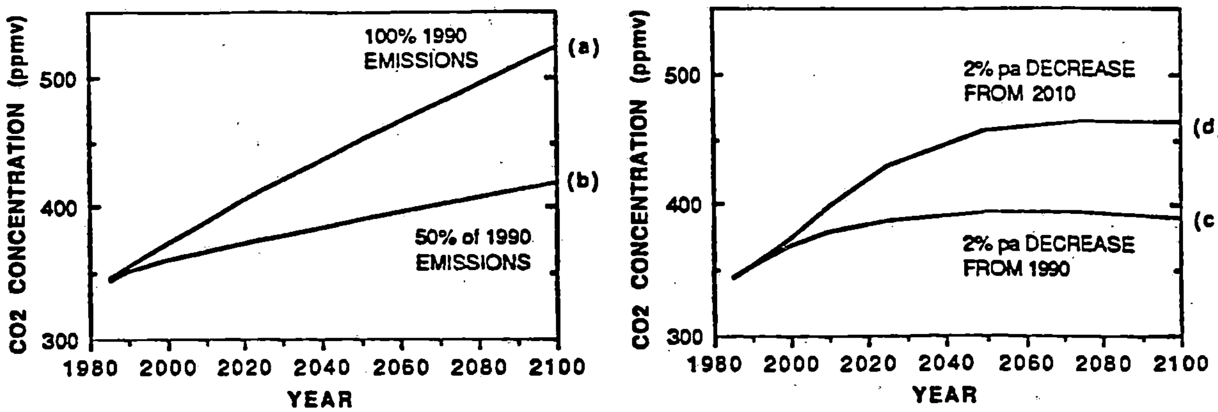
Concentrations of carbon dioxide and methane after remaining relatively constant up to the 18th century, have risen sharply since then due to man's activities. Concentrations of nitrous oxide have increased since the mid-18th century, especially in the last few decades. CFCs were not present in the atmosphere before the 1930s

The effect of ozone on climate is strongest in the upper troposphere and lower stratosphere. Model calculations indicate that ozone in the upper troposphere should have increased due to human-made emissions of nitrogen oxides, hydrocarbons and carbon monoxide. While at ground level ozone has increased in the northern hemisphere in response to these emissions, observations are insufficient to confirm the expected increase in the upper troposphere. The lack of adequate observations prevents us from accurately quantifying the climatic effect of changes in tropospheric ozone.

In the lower stratosphere at high southern latitudes ozone has decreased considerably due to the effects of CFCs, and there are indications of a global-scale decrease which, while not understood, may also be due to CFCs. These observed decreases should act to cool the earth's surface, thus providing a small offset to the predicted warming produced by the other greenhouse gases. Further reductions in lower stratospheric ozone are possible during the next few decades as the atmospheric abundances of CFCs continue to increase.

Concentrations, lifetimes and stabilisation of the gases

In order to calculate the atmospheric concentrations of carbon dioxide which will result from human-made emissions we use computer models which incorporate details of the emissions and which include representations of the transfer of carbon dioxide between the atmosphere, oceans and terrestrial biosphere. For the other greenhouse gases, models which incorporate the effects of chemical reactions in the atmosphere are employed.



The relationship between hypothetical fossil fuel emissions of carbon dioxide and its concentration in the atmosphere is shown in the case where (a) emissions continue at 1990 levels, (b) emissions are reduced by 50% in 1990 and continue at that level, (c) emissions are reduced by 2% pa from 1990, and (d) emissions, after increasing by 2% pa until 2010, are then reduced by 2% pa thereafter.

The atmospheric lifetimes of the gases are determined by their sources and sinks in the oceans, atmosphere and biosphere. Carbon dioxide, chlorofluorocarbons and nitrous oxide are removed only slowly from the atmosphere and hence, following a change in emissions, their atmospheric concentrations take decades to centuries to adjust fully. Even if all human-made emissions of carbon dioxide were halted in the year 1990, about half of the increase in carbon dioxide concentration caused by human activities would still be evident by the year 2100.

In contrast, some of the CFC substitutes and methane have relatively short atmospheric lifetimes so that their atmospheric concentrations respond fully to emission changes within a few decades.

To illustrate the emission-concentration relationship clearly, the effect of hypothetical changes in carbon dioxide fossil fuel emissions is shown below: (a) continuing global emissions at 1990 levels; (b) halving of emissions in 1990; (c) reductions in emissions of 2% per year (pa) from 1990 and (d) a 2% pa increase from 1990-2010 followed by a 2% pa decrease from 2010.

Continuation of present day emissions are committing us to increased future concentrations, and the longer emissions continue to increase, the greater would reductions have to be to stabilise at a given level. If there are critical concentration levels that should not be exceeded, then the earlier emission reductions are made the more effective they are.

STABILISATION OF ATMOSPHERIC CONCENTRATIONS

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

Carbon Dioxide	>60%
Methane	15 - 20%
Nitrous Oxide	70 - 80%
CFC-11	70 - 75%
CFC-12	75 - 85%
HCFC-22	40 - 50%

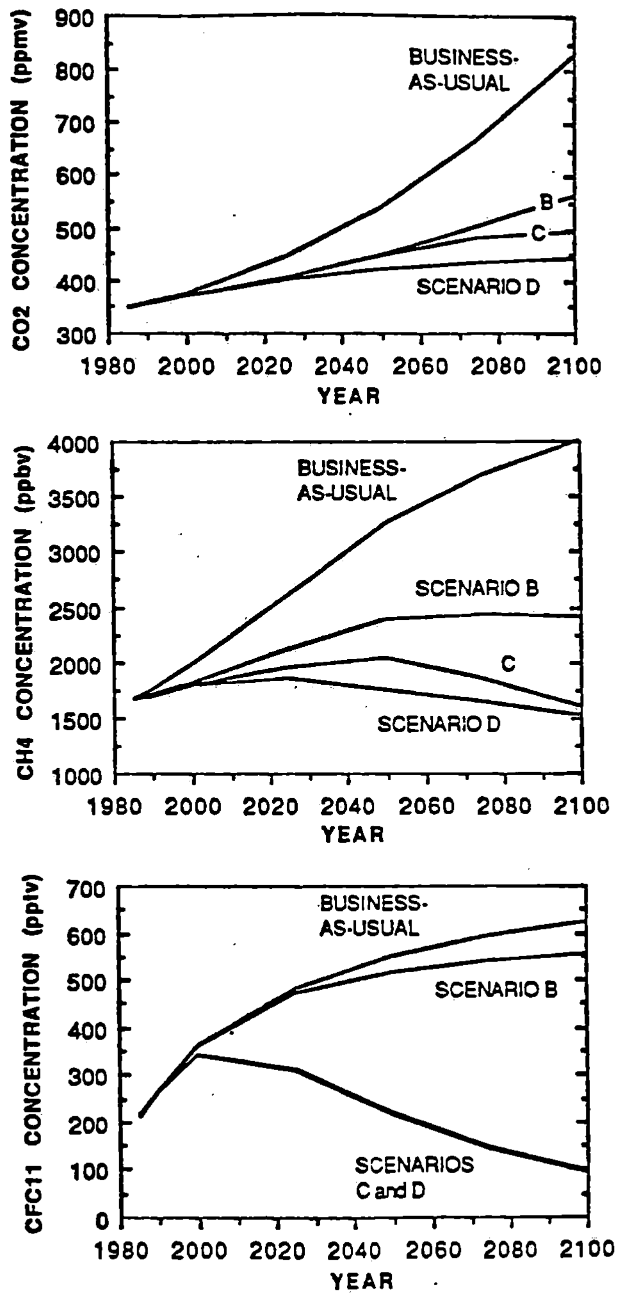
Note that the stabilisation of each of these gases would have different effects on climate, as explained in the next section.

The term "**atmospheric stabilisation**" is often used to describe the limiting of the concentration of the greenhouse gases at a certain level. The amount by which human-made emissions of a greenhouse gas must be reduced in order to stabilise at present day concentrations, for example, is shown in the box opposite. For most gases the reductions would have to be substantial.

How will greenhouse gas abundances change in the future?

We need to know future greenhouse gas concentrations in order to estimate future climate change. As already mentioned, these concentrations depend upon the magnitude of human-made emissions and on how changes in climate and other environmental conditions may influence the biospheric processes that control the exchange of natural greenhouse gases, including carbon dioxide and methane, between the atmosphere, oceans and terrestrial biosphere - the greenhouse gas "feedbacks".

Four scenarios of future human-made emissions were developed by Working Group III. The first of these assumes that few or no steps are taken to limit greenhouse gas emissions, and this is therefore termed Business-as-Usual (BaU). (It should be noted that an aggregation of national forecasts of emissions of carbon dioxide and methane to the year 2025 undertaken by Working Group III resulted in global emissions 10-20% higher than in the BaU scenario.) The other three scenarios assume that progressively increasing levels of controls reduce the growth of emissions; these are referred to as scenarios B, C, and D. They are briefly described in the Annex. Future concentrations of some of the greenhouse gases which would arise from these emissions are shown opposite.



Atmospheric concentrations of carbon dioxide, methane and CFC-11 resulting from the four IPCC emissions scenarios

Greenhouse gas feedbacks

Some of the possible feedbacks which could significantly modify future greenhouse gas concentrations in a warmer world are discussed in the following paragraphs.

The net emissions of carbon dioxide from terrestrial ecosystems will be elevated if higher temperatures increase respiration at a faster rate than photosynthesis, or if plant populations, particularly large forests, cannot adjust rapidly enough to changes in climate.

A net flux of carbon dioxide to the atmosphere may be particularly evident in warmer conditions in tundra and boreal regions where there are large stores of carbon. The opposite is true if higher abundances of carbon dioxide in the atmosphere enhance the productivity of natural ecosystems, or if there is an increase in soil moisture which can be expected to stimulate plant growth in dry ecosystems and to increase the storage of carbon in tundra peat. The extent to which ecosystems can sequester increasing atmospheric carbon dioxide remains to be quantified.

If the oceans become warmer, their net uptake of carbon dioxide may decrease because of changes in (i) the chemistry of carbon dioxide in seawater (ii) biological activity in surface waters and (iii) the rate of exchange of carbon dioxide between the surface layers and the deep ocean. This last depends upon the rate of formation of deep water in the ocean which, in the North Atlantic for example, might decrease if the salinity decreases as a result of a change in climate.

Methane emissions from natural wetlands and rice paddies are particularly sensitive to temperature and soil moisture. Emissions are significantly larger at higher temperatures and with increased soil moisture; conversely, a decrease in soil moisture would result in smaller emissions. Higher temperatures could increase the emissions of methane at high northern latitudes from decomposable organic matter trapped in permafrost and methane hydrates.

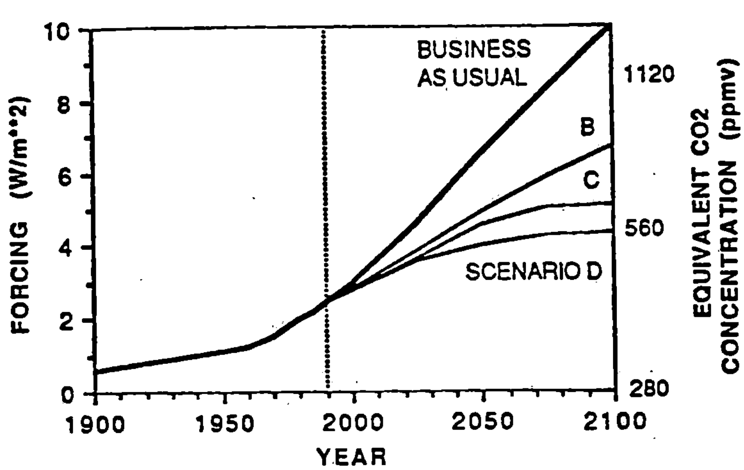
As illustrated earlier, ice core records show that methane and carbon dioxide concentrations changed in a similar sense to temperature between ice ages and interglacials.

Although many of these feedback processes are poorly understood, it seems likely that, overall, they will act to increase, rather than decrease, greenhouse gas concentrations in a warmer world.

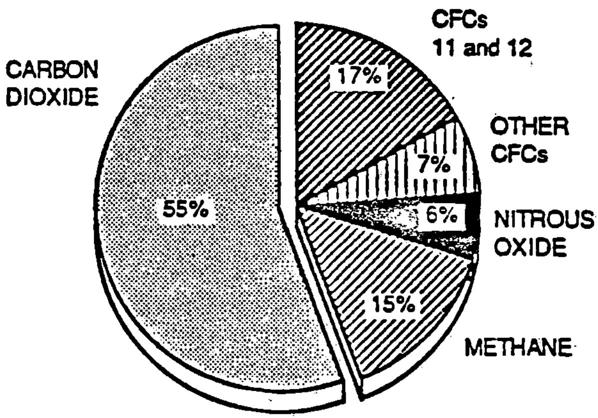
Which gases are the most important?

We are certain that increased greenhouse gas concentrations increase radiative forcing. We can calculate the forcing with much more confidence than the climate change that results because the former avoids the need to evaluate a number of poorly understood atmospheric responses. We then have a base from which to calculate the relative effect on climate of an increase in concentration of each gas in the present-day atmosphere, both in absolute terms and relative to carbon dioxide. These relative effects span a wide range; methane is about 21 times more effective, molecule-for-molecule, than carbon dioxide, and CFC-11 about 12,000 times more effective. On a kilogram-per-kilogram basis, the equivalent values are 58 for methane and about 4,000 for CFC-11, both relative to carbon dioxide. Values for other greenhouse gases are to be found in the full report.

The total radiative forcing at any time is the sum of those from the individual greenhouse gases. We show in the figure below how this quantity has changed in the past (based on observations of greenhouse gases) and how it might change in the future (based on the four IPCC emissions scenarios). For simplicity, we can express total forcing in terms of the amount of carbon dioxide which would give that forcing; this is termed the **equivalent carbon dioxide concentration**. Greenhouse gases have increased since pre-industrial times (the mid-18th century) by an



Increase in radiative forcing since the mid-18th century, and predicted to result from the four IPCC emissions scenarios, also expressed as equivalent carbon dioxide concentrations



The contribution from each of the human-made greenhouse gases to the change in radiative forcing from 1980 to 1990. The contribution from ozone may also be significant, but cannot be quantified at present.

amount that is radiatively equivalent to about a 50% increase in carbon dioxide, although carbon dioxide itself has risen by only 26%; other gases have made up the rest.

The contributions of the various gases to the total increase in climate forcing during the 1980s is shown above as a pie diagram; carbon dioxide is responsible for about half the decadal increase. (Ozone, the effects of which may be significant, is not included)

How can we evaluate the effect of different greenhouse gases?

To evaluate possible policy options, it is useful to know the relative radiative effect (and, hence, potential climate effect) of equal emissions of each of the greenhouse gases. The concept of relative Global Warming Potentials (GWP) has been developed to take into account the differing times that gases remain in the atmosphere.

This index defines the time-integrated warming effect due to an instantaneous release of unit mass (1 kg) of a given greenhouse gas in today's atmosphere, relative to that of carbon dioxide. The relative importances will change in the future as atmospheric composition changes because, although radiative forcing increases in direct proportion to the concentration of CFCs, changes in the other greenhouse gases (particularly carbon dioxide) have an effect on forcing which is much less than proportional.

The GWPs in the following table are shown for three time horizons, reflecting the need to consider the cumulative effects on climate over various time scales. The longer time horizon is appropriate for the cumulative effect; the shorter timescale will indicate the response to emission changes in the short term. There are a number of practical difficulties in devising and calculating the values of the GWPs, and the values given here should be considered as preliminary. In addition to these direct effects, there are indirect effects of human-made emissions arising from chemical reactions between the various

GLOBAL WARMING POTENTIALS

The warming effect of an emission of 1kg of each gas relative to that of CO_2

These figures are best estimates calculated on the basis of the present day atmospheric composition

	Time Horizon		
	20 yr	100 yr	500 yr
Carbon dioxide	1	1	1
Methane (including indirect)	63	21	9
Nitrous oxide	270	290	190
CFC-11	4500	3500	1500
CFC-12	7100	7300	4500
HCFC-22	4100	1500	510

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

THE RELATIVE CUMULATIVE CLIMATE EFFECT OF 1990 MAN-MADE 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%
CFCs	Various	0.9	11%
HCFC-22	1500	0.1	0.5%
Others*	Various		8.5%

*These values include the indirect effect of these emissions on other greenhouse gases via chemical reactions in the atmosphere. Such estimates are highly model dependent and should be considered preliminary and subject to change. The estimated effect of ozone is included under "others". The gases included under "others" are given in the full report.

† 26 000 Tg (teragrams) of carbon dioxide = 7 000 Tg (=7 Gt) of carbon

constituents. The indirect effects on stratospheric water vapour, carbon dioxide and tropospheric ozone have been included in these estimates.

The table indicates, for example, that the effectiveness of methane in influencing climate will be greater in the first few decades after release, whereas emission of the longer-lived nitrous oxide will affect climate for a much longer time. The lifetimes of the proposed CFC replacements range from 1 to 40 years; the longer lived replacements are still potentially effective as agents of climate change. One example of this, HCFC-22 (with a 15 year lifetime), has a similar effect (when released in the same amount) as CFC-11 on a 20 year timescale; but less over a 500 year timescale.

The table shows carbon dioxide to be the least effective greenhouse gas per kilogramme emitted, but its contribution to global warming, which depends on the product of the GWP and the amount emitted, is largest. In the example in the box below, the effect over 100 years of emissions of greenhouse gases in 1990 are shown relative to carbon dioxide. This is illustrative; to compare the effect of different emission projections we have to sum the effect of emissions made in future years

GAS	MAJOR CONTRIBUTOR?	LONG LIFETIME?	SOURCES KNOWN?
Carbon dioxide	yes	yes	yes
Methane	yes	no	semi-quantitatively
Nitrous oxide	not at present	yes	qualitatively
CFCs	yes	yes	yes
HCFCs, etc	not at present	mainly no	yes
Ozone	possibly	no	qualitatively

There are other technical criteria which may help policymakers to decide, in the event of emissions reductions being deemed necessary, which gases should be considered. Does the gas contribute in a major way to current, and future, climate forcing? Does it have a long lifetime, so earlier reductions in emissions would be more effective than those made later? And are its sources and sinks well enough known to decide which could be controlled in practice? The table opposite illustrates these factors.

How much do we expect climate to change?

It is relatively easy to determine the direct effect of the increased radiative forcing due to increases in greenhouse gases. However, as climate begins to warm, various processes act to amplify (through positive feedbacks) or reduce (through negative feedbacks) the warming. The main feedbacks which have been identified are due to changes in water vapour, sea-ice, clouds and the oceans.

The best tools we have which take the above feedbacks into account (but do not include greenhouse gas feedbacks) are three-dimensional mathematical models of the climate system (atmosphere-ocean-ice-land), known as General Circulation Models (GCMs). They synthesise our knowledge of the physical and dynamical processes in the overall system and allow for the complex interactions between the various components. However, in their current state of development, the descriptions of many of the processes involved are comparatively crude. Because of this, considerable uncertainty is attached to these predictions of climate change, which is reflected in the range of values given; further details are given in a later section.

The estimates of climate change presented here are based on

i) the "best estimate" of equilibrium climate sensitivity (i.e the equilibrium temperature change due to a doubling of carbon dioxide in the atmosphere) obtained from model simulations, feedback analyses and observational considerations (see later box: "What tools do we use?")

ii) a "box diffusion upwelling" ocean-atmosphere climate model which translates the greenhouse forcing into the evolution of the temperature response for the prescribed climate sensitivity. (This simple model has been calibrated against more complex atmosphere-ocean coupled GCMs for situations where the more complex models have been run).

How quickly will global climate change?

a. If emissions follow a Business-as-Usual pattern

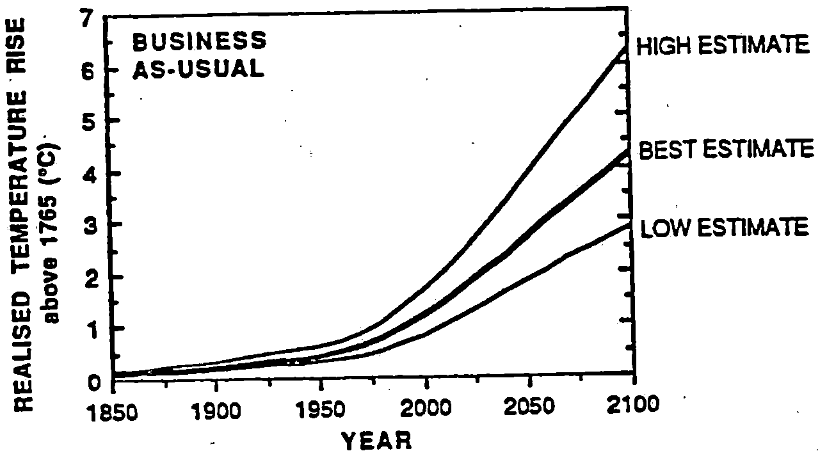
Under the IPCC Business-as-Usual (Scenario A) emissions of greenhouse gases, the average rate of increase of global mean temperature during the next century is estimated to be about 0.3°C per decade (with an uncertainty range of 0.2°C to 0.5°C). This will result in a likely increase in global mean temperature of about 1°C above the present value (about 2°C above that in the pre-industrial period) by 2025 and 3°C above today's (about 4°C above pre-industrial) before the end of the next century.

The projected temperature rise out to the year 2100, with high, low and best-estimate climate responses, is shown in the diagram below. Because of other factors which influence climate, we would not expect the rise to be a steady one.

The temperature rises shown above are realised temperatures; at any time we would also be committed to a further temperature rise toward the equilibrium temperature (see box: "Equilibrium and Realised Climate Change"). For the BaU "best estimate" case in the year 2030, for example, a further 0.9°C rise would be expected, about 0.2°C of which would be realised by 2050 (in addition to changes due to further greenhouse gas increases); the rest would become apparent in decades or centuries.

Even if we were able to stabilise emissions of each of the greenhouse gases at present day levels from now on, the temperature is predicted to rise by about 0.2°C per decade for the first few decades.

The global warming will also lead to increased global average precipitation and evaporation of a few percent by 2030. Areas of sea-ice and snow are expected to diminish.

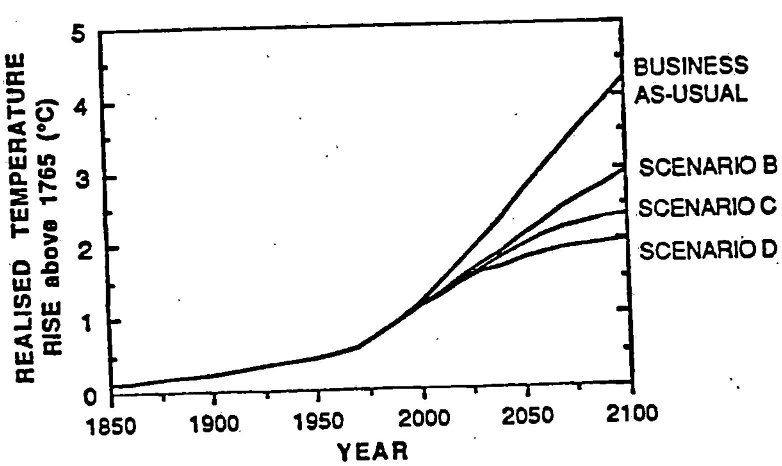


Simulation of the increase in global mean temperature from 1850-1990 due to observed increases in greenhouse gases, and predictions of the rise between 1990 and 2100 resulting from the Business-as-Usual emissions.

b. If emissions are subject to controls

Under the other IPCC emission scenarios which assume progressively increasing levels of controls, average rates of increase in global mean temperature over the next century are estimated to be about 0.2°C per decade (Scenario B), just above 0.1°C per decade (Scenario C) and about 0.1°C per decade (Scenario D). The results are illustrated opposite with the Business-as-Usual case for comparison. Only the best-estimate of the temperature rise is shown in each case.

The indicated range of uncertainty in global temperature rise given above reflects a subjective assessment of uncertainties in the calculation of climate response, but does not include those due to the transformation of emissions to concentrations, nor the effects of greenhouse gas feedbacks.



Simulations of the increase in global mean temperature from 1850-1990 due to observed increases in greenhouse gases, and predictions of the rise between 1990 and 2100 resulting from the IPCC Scenario B,C and D emissions, with the Business-as-Usual case for comparison.

What tools do we use to predict future climate, and how do we use them?

The most highly developed tool which we have to predict future climate is known as a **general circulation model or GCM**. These models are based on the laws of physics and use descriptions in simplified physical terms (called parameterisations) of the smaller-scale processes such as those due to clouds and deep mixing in the ocean. In a climate model an atmospheric component, essentially the same as a weather prediction model, is coupled to a model of the ocean, which can be equally complex.

Climate forecasts are derived in a different way from weather forecasts. A weather prediction model gives a description of the atmosphere's state up to 10 days or so ahead, starting from a detailed description of an initial state of the atmosphere at a given time. Such forecasts describe the movement and development of large weather systems, though they cannot represent very small scale phenomena; for example, individual shower clouds.

To make a climate forecast, the climate model is first run for a few (simulated) decades. The statistics of the model's output is a description of the model's simulated climate which, if the model is a good one, will bear a close resemblance to the climate of the real atmosphere and ocean. The above exercise is then repeated with increasing concentrations of the greenhouse gases in the model. The differences between the statistics of the two simulations (for example in mean temperature and interannual variability) provide an estimate of the accompanying climate change.

The long term change in **surface air temperature** following a doubling of carbon dioxide (referred to as the **climate sensitivity**) is generally used as a benchmark to compare models. The range of results from model studies is 1.9 to 5.2°C. Most results are close to 4.0°C but recent studies using a more detailed but not necessarily more accurate representation of cloud processes give results in the lower half of this range. Hence the models results do not justify altering the previously accepted range of 1.5 to 4.5°C.

Although scientists are reluctant to give a single best estimate in this range, it is necessary for the presentation of climate predictions for a choice of best estimate to be made. Taking into account the model results, together with observational evidence over the last century which is suggestive of the climate sensitivity being in the lower half of the range, (see section: "Has man already begun to change global climate?") a value of climate sensitivity of 2.5°C has been chosen as the best estimate. Further details are given in Section 5 of the report.

In this Assessment, we have also used much simpler models, which simulate the behaviour of GCMs, to make predictions of the evolution with time of global temperature from a number of emission scenarios. These so-called box-diffusion models contain highly simplified physics but give similar results to GCMs when globally averaged.

A completely 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 future, and then use the past spatial patterns as **analogues** of those which will arise in the future. For a good analogue, it is also necessary for the forcing factors (for example, greenhouse gases, orbital variations) and other conditions (for example, ice cover, topography, etc.) to be similar; direct comparisons with climate situations for which these conditions do not apply cannot be easily interpreted. Analogues of future greenhouse-gas-changed climates have not been found.

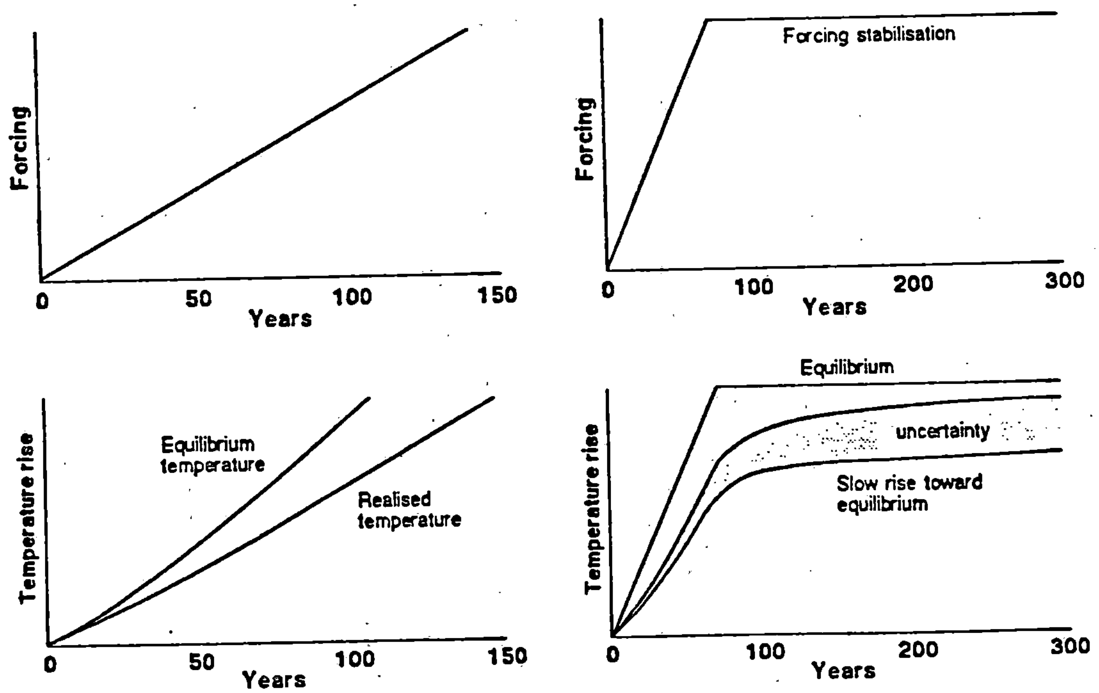
We cannot therefore advocate the use of palaeo-climates as predictions of regional climate change due to future increases in greenhouse gases. However, palaeo-climatological information can provide useful insights into climate processes, and can assist in the validation of climate models.

Equilibrium and realised climate change

When the radiative forcing on the earth-atmosphere system is changed, for example by increasing greenhouse gas concentrations, the atmosphere will try to respond (by warming) immediately. But the atmosphere is closely coupled to the oceans, so in order for the air to be warmed by the greenhouse effect, the oceans also have to be warmed; because of their thermal capacity this takes decades or centuries. This exchange of heat between atmosphere and ocean will act to slow down the temperature rise forced by the greenhouse effect.

In a hypothetical example where the concentration of greenhouse gases in the atmosphere, following a period of constancy, rises suddenly to a new level and remains there, the radiative forcing would also rise rapidly to a new level. This increased radiative forcing would cause the atmosphere and oceans to warm, and eventually come to a new, stable, temperature. A commitment to this equilibrium temperature rise is incurred as soon as the greenhouse gas concentration changes. But at any time before equilibrium is reached, the actual temperature will have risen by only part of the equilibrium temperature change, known as the **realised** temperature change.

Models predict that, for the present day case of an increase in radiative forcing which is approximately steady, the realised temperature rise at any time is about 50% of the committed temperature rise if the climate sensitivity (the response to a doubling of carbon dioxide) is 4.5°C and about 80% if the climate sensitivity is1.5°C. If the forcing were then held constant, temperatures would continue to rise slowly, but it is not certain whether it would take decades or centuries for most of the remaining rise to equilibrium to occur



What will be the patterns of climate change by 2030?

Knowledge of the global mean warming and change in precipitation is of limited use in determining the impacts of climate change, for instance on agriculture. For this we need to know changes regionally and seasonally.

Models predict that surface air will warm faster over land than over oceans, and a minimum of warming will occur around Antarctica and in the northern North Atlantic region.

There are some continental-scale changes which are consistently predicted by the highest resolution models and for which we understand the physical reasons. The warming is predicted to be 50-100% greater than the global mean in high northern latitudes in winter, and substantially smaller than the global mean in regions of sea ice in summer. Precipitation is predicted to increase on average in middle and high latitude continents in winter (by some 5 - 10% over 35-55°N).

Five regions, each a few million square kilometres in area and representative of different climatological regimes, were selected by IPCC for particular study (see map below). In the box below are given the changes in temperature, precipitation and soil moisture, which are predicted to occur by 2030 on the Business-as-Usual scenario, as an average over each of the five regions. There may be considerable variations within the regions. In general, confidence in these regional estimates is low, especially for the changes in precipitation and soil moisture, but they are examples of our best estimates. We cannot yet give reliable regional predictions at the smaller scales demanded for impacts assessments.

How will climate extremes and extreme events change?

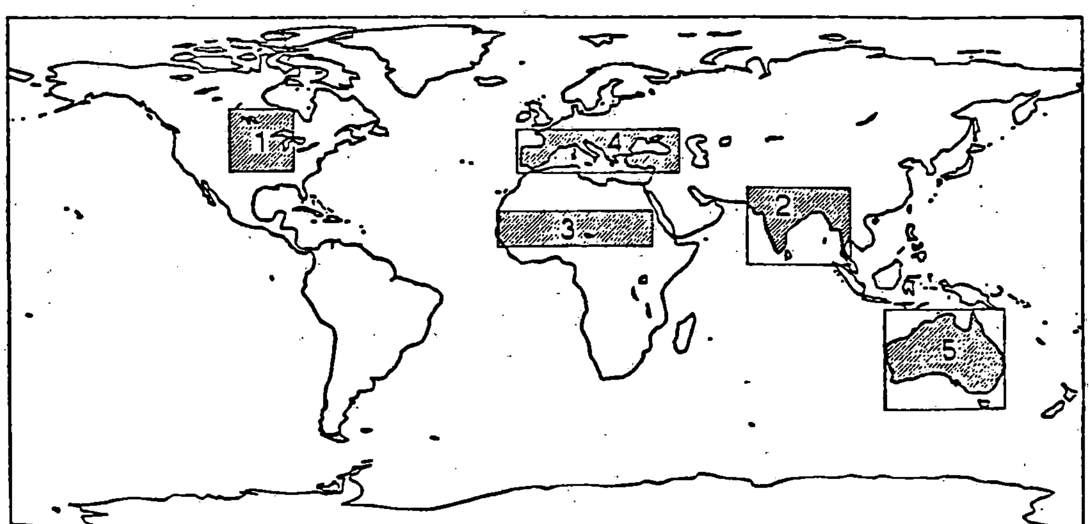
Changes in the variability of weather and the frequency of extremes will generally have more impact than changes in the mean climate at a particular location. With the possible exception of an increase in the number of intense showers, there is no clear evidence that weather variability will change in the future. In the case of temperatures, assuming no change in variability, but with a modest increase in the mean, the number of days with temperatures above a given value at the high end of the distribution will increase substantially. On the same assumptions, there will be a decrease in days with temperatures at the low end of the distribution. So the number of very hot days or frosty nights can be substantially changed without any change in the variability of the weather. The number of days with a minimum threshold amount of soil moisture (for viability of a certain crop, for example) would be even more sensitive to changes in average precipitation and evaporation.

If the large scale weather regimes, for instance depression tracks or anticyclones, shift their position, this would effect the variability and extremes of weather at a particular location, and could have a major effect. However, we do not know if, or in what way, this will happen.

Will storms increase in a warmer world?

Storms can have a major impact on society. Will their frequency, intensity or location increase in a warmer world?

Tropical storms, such as typhoons and hurricanes, only develop at present over seas that are warmer than about 26°C. Although the area of sea having temperatures over this critical value



Map showing the locations and extents of the five areas selected by IPCC

ESTIMATES FOR CHANGES BY 2030

(IPCC Business-as-Usual scenario; changes from **pre-industrial)**

The numbers given below are based on high resolution models, scaled to be consistent with our best estimate of global mean warming of 1.8°C by 2030. For values consistent with other estimates of global temperature rise, the numbers below should be reduced by 30% for the low estimate or increased by 50% for the high estimate. Precipitation estimates are also scaled in a similar way.

Confidence in these regional estimates is low

Central North America (35°-50°N 85°-105°W)
The warming varies from 2 to 4°C in winter and 2 to 3°C in summer. Precipitation increases range from 0 to 15% in winter whereas there are decreases of 5 to 10% in summer. Soil moisture decreases in summer by 15 to 20%.

Southern Asia (5°-30°N 70°-105°E)
The warming varies from 1 to 2°C throughout the year. Precipitation changes little in winter and generally increases throughout the region by 5 to 15% in summer. Summer soil moisture increases by 5 to 10%.

Sahel (10°-20°N 20°W-40°E)
The warming ranges from 1 to 3°C. Area mean precipitation increases and area mean soil moisture decreases marginally in summer. However, throughout the region, there are areas of both increase and decrease in both parameters throughout the region.

Southern Europe (35°-50°N 10°W- 45°E)
The warming is about 2°C in winter and varies from 2 to 3°C in summer. There is some indication of increased precipitation in winter, but summer precipitation decreases by 5 to 15%, and summer soil moisture by 15 to 25%.

Australia (12°-45°S 110°-115°E)
The warming ranges from 1 to 2°C in summer and is about 2°C in winter. Summer precipitation increases by around 10%, but the models do not produce consistent estimates of the changes in soil moisture. The area averages hide large variations at the sub-continental level.

will increase as the globe warms, the critical temperature itself may increase in a warmer world. Although the theoretical maximum intensity is expected to increase with temperature, climate models give no consistent indication whether tropical storms will increase or decrease in frequency or intensity as climate changes; neither is there any evidence that this has occurred over the past few decades.

Mid-latitude storms, such as those which track across the North Atlantic and North Pacific, are driven by the equator-to-pole temperature contrast. As this contrast will probably be weakened in a warmer world (at least in the northern hemisphere), it might be argued that mid-latitude storms will also weaken or change their tracks, and there is some indication of a general reduction in day-to-day variability in the mid-latitude storm tracks in winter in model simulations, though the pattern of changes vary from model to model. Present models do not resolve smaller-scale disturbances, so it will not be possible to assess changes in storminess until results from higher resolution models become available in the next few years.

Climate change in the longer term

The foregoing calculations have focussed on the period up to the year 2100; it is clearly more difficult to make calculations for years beyond 2100. However, while the timing of a predicted increase in global temperatures has substantial uncertainties, the prediction that an increase will eventually occur is more certain. Furthermore, some model calculations that have been extended beyond 100 years suggest that, with continued increases in greenhouse climate forcing, there could be significant changes in the ocean circulation, including a decrease in North Atlantic deep water formation.

Other factors which could influence future climate

Variations in the output of **solar energy** may also affect climate. On a decadal time-scale solar variability and changes in greenhouse gas concentration could give changes of similar magnitudes. However the variation in solar intensity changes sign so that over longer timescales the increases in greenhouse gases are likely to be more important. **Aerosols** as a result of volcanic eruptions can lead to a cooling at the surface which may oppose the greenhouse warming for a few years following an eruption. Again, over longer periods the greenhouse warming is likely to dominate.

Human activity is leading to an increase in aerosols in the lower atmosphere, mainly from sulphur emissions. These have two effects, both of which are difficult to quantify but which may be significant particularly at the regional level. The first is the direct effect of the aerosols on the radiation scattered and absorbed by the atmosphere. The second is an indirect effect whereby the aerosols affect the microphysics of clouds leading to an increased cloud reflectivity. Both these effects might lead to a significant regional cooling; a decrease in emissions of sulphur might be expected to increase global temperatures.

Because of long-period couplings between different components of the climate system, for example between ocean and atmosphere, the earth's climate would still vary without being perturbed by any external influences. This **natural variability** could act to add to, or subtract from, any human-made warming; on a century timescale this would be less than changes expected from greenhouse gas increases.

How much confidence do we have in our predictions?

Uncertainties in the above climate predictions arise from our imperfect knowledge of:

- future rates of human-made emissions
- how these will change the atmospheric concentrations of greenhouse gases
- the response of climate to these changed concentrations

Firstly, it is obvious that the extent to which climate will change depends on the rate at which greenhouse gases (and other gases which affect their concentrations) are emitted. This in turn will be determined by various complex economic and sociological factors. Scenarios of future emissions were generated within IPCC WGIII and are described in the annex.

Secondly, because we do not fully understand the sources and sinks of the greenhouse gases, there are uncertainties in our calculations of future concentrations arising from a given emissions scenario. We have used a number of models to calculate concentrations and chosen a best estimate for each gas. In the case of carbon dioxide, for example, the concentration increase between 1990 and 2070 due to the Business-as-Usual emissions scenario spanned almost a factor of two between the highest and lowest model result (corresponding to a range in radiative forcing change of about 50%)

Furthermore, because natural sources and sinks of greenhouse gases are sensitive to a change in climate, they may substantially modify future concentrations (see earlier section: "Greenhouse gas feedbacks"). It appears that, as climate warms, these feedbacks will lead to an overall increase, rather than decrease, in natural greenhouse gas abundances. For this reason, climate change is likely to be greater than the estimates we have given.

Thirdly, climate models are only as good as our understanding of the processes which they describe, and this is far from perfect. The ranges in the climate predictions given above reflect the uncertainties due to model imperfections; the largest of these is cloud feedback (those factors affecting the cloud amount and distribution and the interaction of clouds with solar and terrestrial radiation), which leads to a factor of two uncertainty in the size of the warming. Others arise from the transfer of energy between the atmosphere and ocean, the atmosphere and land surfaces, and between the upper and deep layers of the ocean. The treatment of sea-ice and convection in the models is also crude. Nevertheless, for reasons given in the box below, we have substantial confidence that models can predict at least the broad-scale features of climate change.

Furthermore, we must recognise that our imperfect understanding of climate processes (and corresponding ability to model them) could make us vulnerable to surprises; just as the human-made ozone hole over Antarctica was entirely unpredicted. In particular, the ocean circulation, changes in 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.

Will the climate of the future be very different?

When considering future climate change, it is clearly essential to look at the record of climate variation in the past. From it we can learn about the range of natural climate variability, to see how it compares with what we expect in the future, and also look for evidence of recent climate change due to man's activities.

Climate varies naturally on all time scales from hundreds of millions of years down to the year to year. Prominent in the Earth's history have been the 100,000 year glacial-interglacial cycles when climate was mostly cooler than at present. Global surface temperatures have typically varied by 5-7°C through these cycles, with large changes in ice volume and sea level, and temperature changes as great as 10-15°C in some

Confidence in predictions from climate models

What confidence can we have that climate change due to increasing greenhouse gases will look anything like the model predictions? Weather forecasts can be compared with the actual weather the next day and their skill assessed; we cannot do that with climate predictions. However, there are several indicators that give us some confidence in the predictions from climate models.

When the latest atmospheric models are run with the present atmospheric concentrations of greenhouse gases and observed boundary conditions their simulation of present climate is generally realistic on large scales, capturing the major features such as the wet tropical convergence zones and mid-latitude depression belts, as well as the contrasts between summer and winter circulations. The models also simulate the observed variability; for example, the large day-to-day pressure variations in the middle latitude depression belts and the maxima in interannual variability responsible for the very different character of one winter from another both being represented. However, on regional scales (2,000km or less), there are significant errors in all models.

Overall confidence is increased by atmospheric models' generally satisfactory portrayal of aspects of variability of the atmosphere, for instance those associated with variations in sea surface temperature. There has been some success in simulating the general circulation of the ocean, including the patterns (though not always the intensities) of the principal currents, and the distributions of tracers added to the ocean.

Atmospheric models have been coupled with simple models of the ocean to predict the equilibrium response to greenhouse gases, under the assumption that the model errors are the same in a changed climate. The ability of such models to simulate important aspects of the climate of the last ice age generates confidence in their usefulness. Atmospheric models have also been coupled with multilayer ocean models (to give coupled ocean-atmosphere GCMs) which predict the gradual response to increasing greenhouse gases. Although the models so far are of relatively coarse resolution, the large scale structures of the ocean and the atmosphere can be simulated with some skill. However, the coupling of ocean and atmosphere models reveals a strong sensitivity to small scale errors which leads to a drift away from the observed climate. As yet, these errors must be removed by adjustments to the exchange of heat between ocean and atmosphere. There are similarities between results from the coupled models using simple representations of the ocean and those using more sophisticated descriptions, and our understanding of such differences as do occur gives us some confidence in the results.

middle and high latitude regions of the northern hemisphere. Since the end of the last ice age, about 10,000 years ago, global surface temperatures have probably fluctuated by little more than 1°C. Some fluctuations have lasted several centuries, including the Little Ice Age which ended in the nineteenth century and which appears to have been global in extent.

The changes predicted to occur by about the middle of the next century due to increases in greenhouse gas concentrations from the Business-as-Usual emissions will make global mean temperatures higher than they have been in the last 150,000 years.

The rate of change of global temperatures predicted for Business-as-Usual emissions will be greater than those which have occured naturally on earth over the last 10,000 years, and the rise in sea level will be about three to six times faster than that seen over the last 100 years or so.

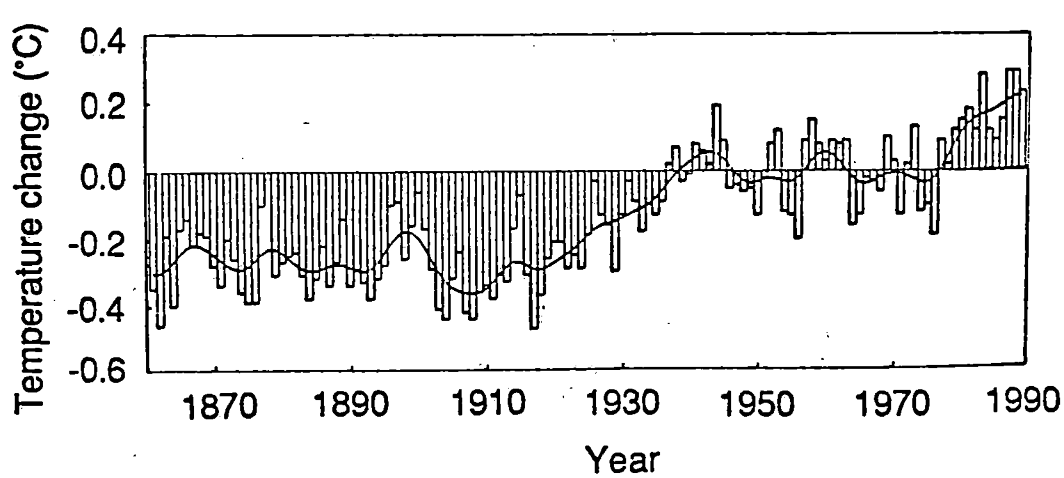
Has man already begun to change the global climate?

The instrumental record of **surface temperature** is fragmentary until the mid-nineteenth century, after which it slowly improves. Because of different methods of measurement, historical records have to be harmonised with modern observations, introducing some uncertainty. Despite these problems we believe that a real warming of the globe of 0.3°C - 0.6°C has taken place over the last century; any bias due to urbanisation is likely to be less than 0.05°C.

Moreover since 1900 similar temperature increases are seen in three independent data sets: one collected over land and two over the oceans. The figure below shows current estimates of smoothed global mean surface temperature over land and ocean since 1860. Confidence in the record has been increased by their similarity to recent satellite measurements of mid-tropospheric temperatures.

Although the overall temperature rise has been broadly similar in both hemispheres, it has not been steady, and differences in their rates of warming have sometimes persisted for decades. Much of the warming since 1900 has been concentrated in two periods, the first between about 1910 and 1940 and the other since 1975; the five warmest years on record have all been in the 1980s. The northern hemisphere cooled between the 1940s and the early 1970s when southern hemisphere temperatures stayed nearly constant. The pattern of global warming since 1975 has been uneven with some regions, mainly in the northern hemisphere, continuing to cool until recently. This regional diversity indicates that future regional temperature changes are likely to differ considerably from a global average.

The conclusion that global temperature has been rising is strongly supported by the retreat of most **mountain glaciers** of the world since the end of the nineteenth century and the fact that global **sea level** has risen over the same period by an average of 1 to 2mm per year. Estimates of thermal expansion of the oceans, and of



Global mean combined land-air and sea-surface temperatures, 1861 - 1989, relative to the average for 1951-80.

increased melting of mountain glaciers and the ice margin in West Greenland over the last century, show that the major part of the sea level rise appears to be related to the observed global warming. This apparent connection between observed sea level rise and global warming provides grounds for believing that future warming will lead to an acceleration in sea level rise.

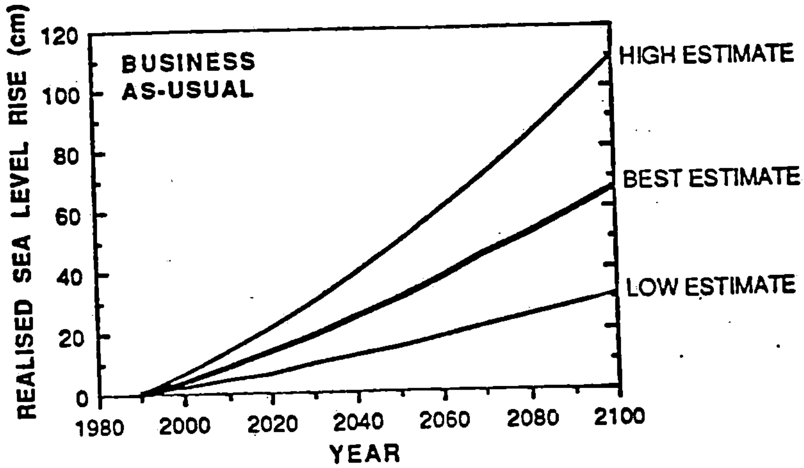
- The size of the warming over the last century is broadly consistent with the predictions of climate models, but is also of the same magnitude as natural climate variability. If the sole cause of the observed warming were the human-made greenhouse effect, then the implied climate sensitivity would be near the lower end of the range inferred from the models. The observed increase could be largely due to natural variability; alternatively this variability and other man-made factors could have offset a still larger man-made greenhouse warming. The unequivocal detection of the enhanced greenhouse effect from observations is not likely for a decade or more, when the committment to future climate change will then be considerably larger than it is today.

Global-mean temperature alone is an inadequate indicator of greenhouse-gas-induced climatic change. Identifying the causes of any global-mean temperature change requires examination of other aspects of the changing climate, particularly its spatial and temporal characteristics - the man-made climate change "signal". Patterns of climate change from models such as the northern hemisphere warming faster than the southern hemisphere, and surface air warming faster over land than over oceans, are not apparent in observations to date. However, we do not yet know what the detailed "signal" looks like because we have limited confidence in our predictions of climate change patterns. Furthermore, any changes to date could be masked by natural variability and other (possibly man-made) factors, and we do not have a clear picture of these.

How much will sea level rise ?

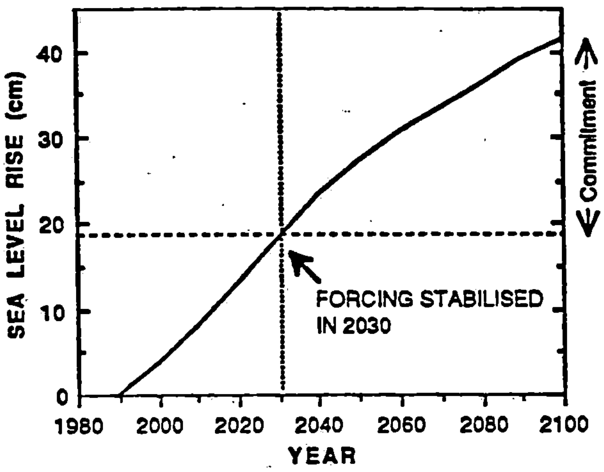
Simple models were used to calculate the rise in sea level to the year 2100; the results are illustrated below. The calculations necessarily ignore any long-term changes, unrelated to greenhouse forcing, that may be occurring but cannot be detected from the present data on land ice and the ocean. The sea-level rise expected from 1990-2100 under the IPCC Business as Usual emissions scenario is shown below. An average rate of global mean sea level rise of about 6cm per decade over the next century (with an uncertainty range of 3 - 10 cm per decade). The predicted rise is about 20cm in global mean sea level by 2030, and 65cm by the end of the next century. There will be significant regional variations.

The best estimate in each case is made up mainly of positive contributions from thermal expansion of the oceans and the melting of glaciers. Although, over the next 100 years, the effect of the Antarctic and Greenland ice sheets is expected to be small, they make a major contribution to the uncertainty in predictions.



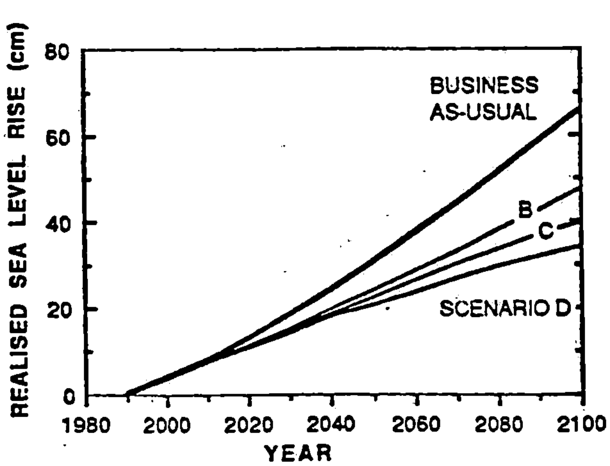
Sea level rise predicted to result from Business-as-Usual emissions, showing the best-estimate and range

Even if greenhouse forcing increased no further, there would still be a commitment to a continuing sea level rise for many decades and even centuries, due to delays in climate, ocean and ice mass responses. As an illustration, if the increases in greenhouse gas concentrations were to suddenly stop in 2030, sea level would go on rising from 2030 to 2100, by as much again as from 1990-2030, as shown in the diagram below.



Commitment to sea level rise in the year 2030. The curve shows the sea level rise due to Business-as-Usual emissions to 2030, with the additional rise that would occur in the remainder of the century even if climate forcing was stabilised in 2030.

Predicted sea level rises due to the other three emissions scenarios are shown below, with the Business-as-Usual case for comparison; only best-estimate calculations are shown.



Model estimates of sea-level rise from 1990-2100 due to all four emissions scenarios.

The West Antarctic Ice Sheet is of special concern. A large portion of it, containing an amount of ice equivalent to about 5m of global sea level, is grounded far below sea level. There have been suggestions that a sudden outflow of ice might result from global warming and raise sea level quickly and substantially. Recent studies have shown that individual ice streams are changing rapidly on a decade-to- century timescale; however this is not necessarily related to climate change. Within the next century, it is not likely that there will be a major outflow of ice from West Antarctica due directly to global warming.

Any rise in sea level is not expected to be uniform over the globe. Thermal expansion, changes in ocean circulation, and surface air pressure will vary from region to region as the world warms, but in an as yet unknown way. Such regional details await further development of more realistic coupled ocean atmosphere models. In addition, vertical land movements can be as large or even larger than changes in global mean sea level; these movements have to be taken into account when predicting local change in sea level relative to land.

The most severe effects of sea-level rise are likely to result from extreme events (for example, storm surges) the incidence of which may be affected by climatic change.

What will be the effect of climate change on ecosystems?

Ecosystem processes such as photosynthesis and respiration are dependent on climatic factors and carbon dioxide concentration in the short term. In the longer term, climate and carbon dioxide are among the factors which control ecosystem structure, i.e., species composition, either directly by increasing mortality in poorly adapted species, or indirectly by mediating the competition between species. Ecosystems will respond to local changes in temperature (including its rate of change), precipitation, soil moisture and extreme events. Current models are unable to make reliable estimates of changes in these parameters on the required local scales.

Photosynthesis captures atmospheric carbon dioxide, water and solar energy and stores them in organic compounds which are then used for subsequent plant growth, the growth of animals or the growth of microbes in the soil. All of these organisms release carbon dioxide via respiration into the atmosphere. Most land plants have a system of photosynthesis which will respond positively to increased atmospheric carbon dioxide ("the carbon dioxide fertilization effect") but the response varies with species. The effect may decrease with time when restricted by other ecological limitations, for example, nutrient availability. It should be emphasized that the carbon content of the terrestrial biosphere will increase only if the forest ecosystems in a state of maturity will be able to store more carbon in a warmer climate and at higher concentrations of carbon dioxide. We do not yet know if this is the case.

The response to increased carbon dioxide results in greater efficiencies of water, light and nitrogen use. These increased efficiencies may be particularly important during drought and in arid/semi-arid and infertile areas.

Because species respond differently to climatic change, some will increase in abundance and/or range while others will decrease. Ecosystems will therefore change in structure and composition. Some species may be displaced to higher latitudes and altitudes, and may be more prone to local, and possibly even global, extinction; other species may thrive.

As stated above, ecosystem structure and species distribution are particularly sensitive to the rate of change of climate. We can deduce something about how quickly global temperature has changed in the past from paleoclimatological records. As an example, at the end of the last glaciation, within about a century, temperature increased by up to 5°C in the North Atlantic region, mainly in Western Europe. Although during the increase from the glacial to the current interglacial temperature simple tundra ecosystems responded positively, a similar rapid temperature increase applied to more developed ecosystems could result in their instability.

Deforestation and Reforestation

Man has been deforesting the Earth for millennia. Until the early part of the century, this was mainly in temperate regions, more recently it has been concentrated in the tropics. Deforestation has several potential impacts on climate: through the carbon and nitrogen cycles (where it can lead to changes in atmospheric carbon dioxide concentrations), through the change in reflectivity of terrain when forests are cleared, through its effect on the hydrological cycle (precipitation, evaporation and runoff) and surface roughness and thus atmospheric circulation which can produce remote effects on climate.

It is estimated that each year about 2 Gt of carbon (GtC) is released to the atmosphere due to tropical deforestation. The rate of forest clearing is difficult to estimate; probably until the mid-20th century, temperate deforestation and the loss of organic matter from soils was a more important contributor to atmospheric carbon dioxide than was the burning of fossil fuels. Since then, fossil fuels have become dominant; one estimate is that around 1980, 1.6 GtC was being released annually from the clearing of tropical forests, compared with about 5 GtC from the burning of fossil fuels. If all the tropical forests were removed, the input is variously estimated at from 150 to 240 GtC; this would increase atmospheric carbon dioxide by 35 to 60 ppmv.

To analyse the effect of reforestation we assume that 10 million hectares of forests are planted each year for a period of 40 years, ie 4 million km2 would then have been planted by 2030, at which time 1GtC would be absorbed annually until these forests reach maturity. This would happen in 40-100 years for most forests. The above scenario implies an accumulated uptake of about 20GtC by the year 2030 and up to 80GtC after 100 years. This accumulation of carbon in forests is equivalent to some 5-10% of the emission due to fossil fuel burning in the Business-as-Usual scenario.

Deforestation can also alter climate directly by increasing reflectivity and decreasing evapotranspiration. Experiments with climate models predict that replacing all the forests of the Amazon Basin by grassland would reduce the rainfall over the basin by about 20%, and increase mean temperature by several degrees.

What should be done to reduce uncertainties, and how long will this take?

Although we can say that some climate change is unavoidable, much uncertainty exists in the prediction of global climate properties such as the temperature and rainfall. Even greater uncertainty exists in predictions of regional climate change, and the subsequent consequences for sea level and ecosystems. The key areas of scientific uncertainty are:

- **clouds:** primarily cloud formation, dissipation, and radiative properties, which influence the response of the atmosphere to greenhouse forcing;

- **oceans:** the exchange of energy between the ocean and the atmosphere, between the upper layers of the ocean and the deep ocean, and transport within the ocean, all of which control the rate of global climate change and the patterns of regional change;

- **greenhouse gases:** quantification of the uptake and release of the greenhouse gases, their chemical reactions in the atmosphere, and how these may be influenced by climate change.

- **polar ice sheets:** which affect predictions of sea level rise

Studies of land surface hydrology, and of impact on ecosystems, are also important.

To reduce the current scientific uncertainties in each of these areas will require internationally coordinated research, the goal of which is to improve our capability to observe, model and understand the global climate system. Such a program of research will reduce the scientific uncertainties and assist in the formulation of sound national and international response strategies.

Systematic long-term **observations** of the system are of vital importance for understanding the natural variability of the Earth's climate system, detecting whether man's activities are changing it, parametrising key processes for models, and verifying model simulations. Increased accuracy and coverage in many observations are required. Associated with expanded observations is the need to develop appropriate comprehensive global information bases for the rapid and efficient dissemination and utilization of data. The main observational requirements are:

i) the maintenance and improvement of observations (such as those from satellites) provided by the World Weather Watch Programme of WMO

ii) the maintenance and enhancement of a programme of monitoring, both from satellite-based and surface-based instruments, of key climate elements for which accurate observations on a continuous basis are required, such as the distribution of important atmospheric constituents, clouds, the earth's radiation budget, precipitation, winds, sea surface temperatures and terrestrial ecosystem extent, type and productivity.

iii) the establishment of a global ocean observing system to measure changes in such variables as ocean surface topography, circulation, transport of heat and chemicals, and sea-ice extent and thickness.

iv) the development of major new systems to obtain data on the oceans, atmosphere and terrestrial ecosystems using both satellite-based instruments and instruments based on the surface, on automated instrumented vehicles in the ocean, on floating and deep sea buoys, and on aircraft and balloons.

v) the use of paleoclimatological and historical instrumental records to document natural variability and changes in the climate system, and subsequent environmental response.

The **modelling** of climate change requires the development of global models which couple together atmosphere, land, ocean and ice models and which incorporate more realistic formulations of the relevant processes and the interactions between the different components. Processes in the biosphere (both on land and in the ocean) also need to be included. Higher spatial resolution than is currently generally used is required if regional patterns are to be predicted. These models will require the largest computers which are planned to be available during the next decades.

Understanding of the climate system will be developed from analyses of observations and of the results from model simulations. In addition, detailed studies of particular processes will be required through targetted observational campaigns. Examples of such field campaigns include combined observational and small scale modelling studies for different regions, of the formation, dissipation, radiative, dynamical and microphysical properties of clouds, and ground-based (ocean and land) and aircraft measurements of the fluxes of greenhouse gases from specific ecosystems. In particular, emphasis must be placed on field experiments that will assist in the development and improvement of sub-grid-scale parametrizations for models.

The required program of research will require unprecedented international cooperation, with the World Climate Research Programme (WCRP) of the World Meteorological Organization and International Council of Scientific Unions (ICSU), and the International Geosphere-Biosphere Programme (IGBP) of ICSU both playing vital roles. These are large and complex endeavours that will require the involvement of all nations, particularly the developing countries. Implementation of existing and planned projects will require increased financial and human resources; the latter requirement has immediate implications at all levels of education, and the international community of scientists needs to be widened to include more members from developing countries.

The WCRP and IGBP have a number of ongoing or planned research programs, that address each of the three key areas of scientific uncertainty. Examples include:

- **clouds:**
  International Satellite Cloud Climatology Project (ISCCP);
  Global Energy and Water Cycle Experiment (GEWEX).

- **oceans:**
  World Ocean Circulation Experiment (WOCE);
  Tropical Oceans and Global Atmosphere (TOGA).

- **trace gases:**
  Joint Global Ocean Flux Study (JGOFS);
  International Global Atmospheric Chemistry (IGAC);
  Past Global Changes (PAGES).

As research advances, increased understanding and improved observations will lead to progressively more reliable climate predictions. However considering the complex nature of the problem and the scale of the scientific programmes to be undertaken we know that rapid results cannot be expected. Indeed further scientific advances may expose unforeseen problems and areas of ignorance.

Timescales for narrowing the uncertainties will be dictated by progress over the next 10-15 years in two main areas:

- Use of the fastest possible computers, to take into account coupling of the atmosphere and the oceans in models, and to provide sufficient resolution for regional predictions.

- Development of improved representation of small scale processes within climate models, as a result of the analysis of data from observational programmes to be conducted on a continuing basis well into the next century.

Annex

EMISSIONS SCENARIOS FROM WORKING GROUP III OF THE INTERGOVERNMENTAL PANEL ON CLIMATE CHANGE

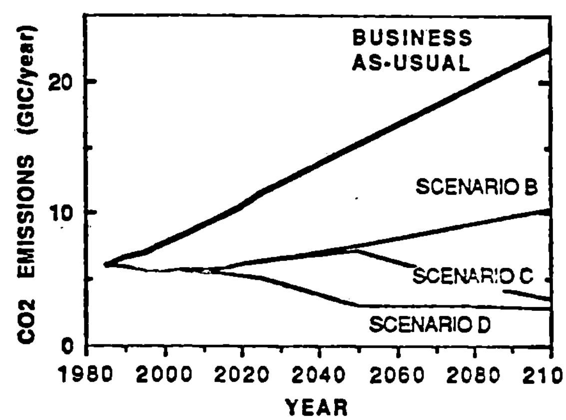
The Steering Group of the Response Strategies Working Group requested the USA and the Netherlands to develop emissions scenarios for evaluation by the IPCC Working Group I. The scenarios cover the emissions of carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), chlorofluorocarbons (CFCs), carbon monoxide (CO) and nitrogen oxides (NO_x) from the present up to the year 2100. Growth of the economy and population was taken common for all scenarios. Population was assumed to approach 10.5 billion in the second half of the next century. Economic growth was assumed to be 2-3% annually in the coming decade in the OECD countries and 3-5 % in the Eastern European and developing countries. The economic growth levels were assumed to decrease thereafter. In order to reach the required targets, levels of technological development and environmental controls were varied.

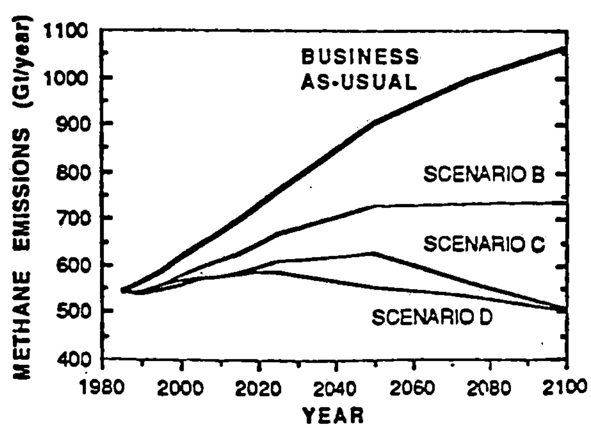
In the **Business-as-Usual scenario** (Scenario A) the energy supply is coal intensive and on the demand side only modest efficiency increases are achieved. Carbon monoxide controls are modest, deforestation continues until the tropical forests are depleted and agricultural emissions of methane and nitrous oxide are uncontrolled. For CFCs the Montreal Protocol is implemented albeit with only partial participation. Note that the aggregation of national projections by IPCC Working Group III gives higher emissions (10 - 20%) of carbon dioxide and methane by 2025.

In **Scenario B** the energy supply mix shifts towards lower carbon fuels, notably natural gas. Large efficiency increases are achieved. Carbon monoxide controls are stringent, deforestation is reversed and the Montreal Protocol implemented with full participation.

In **Scenario C** a shift towards renewables and nuclear energy takes place in the second half of next century. CFCs are now phased out and agricultural emissions limited.

For **Scenario D** a shift to renewables and nuclear in the first half of the next century reduces the emissions of carbon dioxide, initially more or less stabilizing emissions in the industrialized countries. The scenario shows that stringent controls in industrialized countries combined with moderated growth of emissions in developing countries could stabilize atmospheric concentrations. Carbon dioxide emissions are reduced to 50% of 1985 levels by the middle of the next century.





Emissions of carbon dioxide and methane (as examples) to the year 2100, in the four scenarios developed by IPCC Working Group III.