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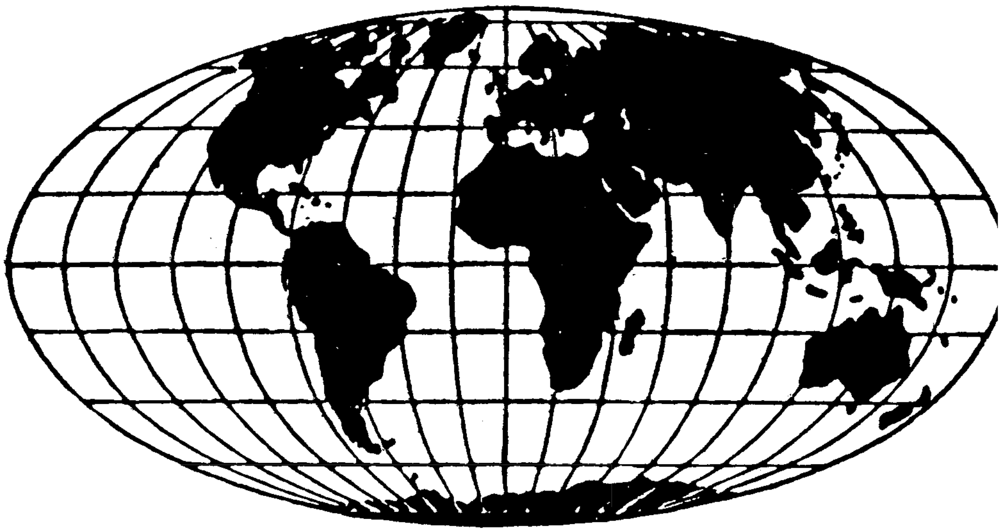

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DETECTING THE CLIMATIC EFFECTS OF INCREASING CARBON DIOXIDE

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FOREWORD

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

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

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

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

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

It is now known that the atmospheric CO_2 concentration in 1900 was approximately 300 parts per million by volume (ppm) (indicated by recent measurements of glacial ice cores by Oeschger). However, it wasn't until 1938 that Callendar presented the first substantive data showing that the concentration of CO_2 in the atmosphere was increasing and suggested that the increase might affect the Earth's climate. After another 20 years Keeling began to monitor the atmospheric CO_2 concentration at the Mauna Loa Observatory in Hawaii. The measurements of atmospheric CO_2 in 1958 showed the annual average concentration was 316 ppm; it was approximately 345 ppm in 1985. Plass outlined theories to explain the relationship between atmospheric CO_2 and climate in 1956, and, soon after, Revelle and Suess described the relationship between CO_2 in the atmosphere and in the oceans, and Kaplan enlarged upon the role of CO_2 in the atmosphere in terms of the global heat balance.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Frederick A. Koomanoff, Director
Carbon Dioxide Research Division
Office of Basic Energy Sciences
U.S. Department of Energy

PREFACE

Virtually all theoretical studies suggest that the increasing atmospheric CO_2 concentration will significantly increase global average temperature. However, even though the CO_2 concentration has risen about 25% since the middle of the last century, such a warming is not yet clearly identifiable in the observational record. Therefore, the problem for the climate research community is to detect the change in climate attributable to the increasing atmospheric CO_2 concentration and to isolate it from all other contributing factors. Being able to verify whether temperature and other climate parameters (such as precipitation, snow and ice amount, and lake freeze-up and break-up dates) are changing as predicted is essential for making accurate estimates of future climate change. It is the objective of this volume to document the state of the climate data base and to assess its consistency with projected CO_2-induced climate changes.

During the last decade, researchers have attempted to correlate the temperature record of the last 130 years with predicted CO_2-induced climate changes. Using different assumptions and data sets, most have found that the record is consistent with the predictions. However, even with differences in the analyses and the simple models that have been used, none have been able to verify a cause and effect relationship regarding the role of CO_2 in climate change. In addition, predictions have only been available for global scale changes because of the the lack of representative regional data and the lack of models that can accurately simulate regional variations.

Important progress has been made. Knowledge of the CO_2 concentration that existed earlier this century and in the 19th century has been markedly improved. Theoretical projections of the potential climate changes resulting from an increased CO_2 concentration are now made with models of greater complexity and that include more realistic representations of the oceans, geography, and the seasonal cycle. There has been substantial expansion of the data base needed to search for the projected CO_2-induced climate changes.

The climate changes due to an increase in the CO_2 concentration will occur slowly and vary regionally. The data base for the assessment of regional change (over both land and ocean) must be extended to include a longer time period and to cover larger geographic regions. The observational effort is an important adjunct to the modeling effort. As multiple climate elements are examined for change, they are not only compared to model estimates of regional rates and magnitudes of climate change, but they are also used to modify and improve models. Only the combined effort of improving data bases and modeling can then establish the cause and effect relationship and provide the scientific basis required for the policy process.

Sincere thanks are due to Michael C. MacCracken and Frederick M. Luther, Lawrence Livermore National Laboratory, for their contributions to this document as editors, authors, and researchers. They have participated in making this volume possible from its inception to its final production. Additional thanks are due to the contributing authors for the hard work, patience, and dedication involved in writing the chapters in this volume.

Michael R. Riches, Program Manager
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Office of Basic Energy Sciences
United States Department of Energy

EDITORS' PREFACE

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

The outline for this SOA was initially proposed in the fall of 1982, where suggestions were offered by those attending a DOE-sponsored CO_2 research conference at Berkeley Springs, West Virginia. After considerable revision, chapter authors were selected in late 1983 and outlines and early drafts of chapters were prepared for most chapters by mid-1984. These initial materials were reviewed at a meeting in Washington, D.C. attended by authors and other scientists active in studying this area.

To assure the authoritativeness of the reviews, DOE contracted with the American Association for the Advancement of Science (AAAS) to arrange for external peer review of the individual chapters. This extensive review started during the second half of 1984 and continued for more than a year. The reviewers selected by AAAS provided many important comments and suggestions that have led to significant improvement of the chapters. Their assistance has been greatly appreciated.

After receiving the reviewers' comments, chapter authors modified and updated their chapters to assure that the SOA report adequately covers the many advances being made in understanding the potential effects of the increasing CO_2 concentration. Final versions of the chapters were submitted for editorial review and a final AAAS review throughout 1985.

The editors gratefully acknowledge the dedicated efforts of all of the chapter authors. They responded with patience and persistence to up to twelve AAAS-sponsored reviews for each chapter, to many comments from fellow authors, to extensive editorial suggestions intended to help better integrate the various chapters, and to requests to aid in review and preparation of summary and recommendation chapters. This has been greatly appreciated.

While these chapters have undergone extensive review, the views finally expressed are those of the authors of each chapter. As in all areas where active research is under way, there are differences of interpretation and emphasis. In reviewing these issues, the editors have attempted to assure thorough presentation and well-reasoned statements, not to coerce uniformity of view. Where differences are evident or uncertainties are presented, further study is recommended.

We want to express special gratitude to the many secretaries who have labored on these chapters at the various institutions. At LLNL, particular and special thanks go to Floy Worden, who has done everything from typing chapters from handwritten scribbles to communicating with authors and preparing camera-ready copy. She has been aided extensively by Nancy Badal, Pam Drumtra, Sandra Eyre, Lonnette Robinson, and Doris Swan, each of whom has willingly helped with each succeeding updating of texts.

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

Michael C. MacCracken
Frederick M. Luther
Editors

This volume was prepared under the auspices of the Carbon Dioxide Research Division of the Office of Basic Energy Sciences, U.S. Department of Energy by the Lawrence Livermore National Laboratory under contract W-7405-ENG-48. Mr. Michael R. Riches served as DOE Project Manager.

AAAS REVIEW

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

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

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

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

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

Roger Revelle, Chairman
Committee on Climate
American Association for the
Advancement of Science (AAAS)

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EXECUTIVE SUMMARY

The potential climatic effects of the increasing atmospheric carbon dioxide (CO_2) concentration, as currently projected by numerical models of the climate system, would constitute a major, extended alteration of the climatic regime that may have far-reaching economic and social consequences. It is, therefore, essential that confirmatory evidence of the projected CO_2-induced climate changes be obtained as soon as possible. The strategy for detecting CO_2-induced climate changes in the observational record consists of the following: (1) determining the climate changes that have occurred, (2) identifying and quantifying the various factors that might have caused or contributed to the observed changes, and (3) isolating those parts of the climate changes that are attributable to the increasing CO_2 concentration, that is, the "CO_2-signal."

The climate research community has devoted a great deal of attention to the problem of isolating and detecting the CO_2-induced climate changes. Recent work has included the assembly and analysis of new geological, historical, and instrumental data. These data are proving useful in developing a better understanding of the sensitivity of the overall climate system to both anthropogenic and natural perturbations, and the data are helpful in verifying and improving present climate models. In turn, the climate modeling studies are helping to identify the climatic variables that appear to be the most promising indicators that climate changes are resulting from the increasing atmospheric CO_2 concentration.

The objective of this volume of the State-of-the-Art series is to document what is known about detecting the CO_2-induced changes in climate and to describe the uncertainties and unknowns associated with this monitoring and analysis effort. This summary follows the order in which material is presented in the rest of this volume. The various approaches for detecting CO_2-induced climate changes are discussed first, followed by a review of applications of these strategies to the various climatic variables that are expected to be changing. Finally, recommendations are presented for research and analysis activities that would contribute to a more definitive identification of the CO_2-induced climate signal.

Climate shows variations on all time scales (monthly, seasonal, annual, decadal, and on up). As a consequence, the appropriate reference climate is not easy to define precisely. A traditional choice is to define climate statistically as the mean state (including the variability) of the atmosphere, ocean, ice, and land surface in a specific region and over a specified time period. Factors such as volcanic emissions, solar variations, and natural fluctuations internal to the climate system may cause variability on time scales of years to decades. On longer time scales, changes in atmospheric composition may also cause climate changes.

Because significant changes in the atmospheric CO_2 concentration started during the second half of the 19th century, it would be desirable to have available a reference climatic data base from prior to that period. Unfortunately, prior to about 1900 the limited accuracy and coverage of observational records of climatic variables pose significant restraints. As a result, a pre-industrial reference climatic state is not available, and the CO_2-induced changes must be sought in data sets that may actually include some early climatic effects, probably relatively small, of rising CO_2 concentrations during the late 19th and early 20th centuries.

Carbon dioxide is not the only atmospheric constituent that may induce climate change. The climatic effects of increasing trace gas concentrations are very likely to be similar to those of the increasing CO_2 concentration, although the magnitude of the changes will depend distinctively on the concentration of each species. Consequently, identification and isolation of the CO_2-induced climate changes from those resulting from other causes are difficult tasks.

THE RADIATIVE SIGNAL OF INCREASING CARBON DIOXIDE

The direct effect of changing CO_2 and trace gas concentrations is alteration of the global radiation balance. This radiative perturbation in turn alters temperatures, which then alter wind fields and other climatic parameters in a continuing sequence. An initial step in isolating CO_2-induced climate changes would, therefore, be identification of the radiative perturbation. To facilitate the detection of these direct changes, the radiative balance of the Earth's atmosphere must be understood. Changes in CO_2 concentration are not expected to result in any significant change in the solar energy absorbed by the atmosphere and Earth's surface. The largest changes are expected to occur in the flux density of longwave (infrared) radiation in the atmosphere. The spectral features (i.e., variations with wavelength) in the longwave flux components are affected by the particular radiative properties of CO_2, water vapor (H_2O), ozone (O_3), and other trace gases, primarily nitrous oxide (N_2O), methane (CH_4), and the chlorocarbons (e.g., $CFCl_3$ and CF_2Cl_2). Changes in the temperature structure of the atmosphere also affect the spectral radiance (distribution of energy over wavelength).

Two viable monitoring approaches for observing this signal are available: to measure the spectral longwave radiance using satellite sensors (looking down from above) or to use ground-based sensors (looking up from below). Both monitoring approaches have their advantages and disadvantages. Satellite sensors could provide global coverage, but calibration of the instruments is more difficult. Ground-based instruments are easily serviced and could be well calibrated, but they cannot easily provide global coverage. In both cases, high-resolution instruments would be required to measure the spectral pattern of change in radiance, which would provide important information about the cause of the change. The relative magnitude of the changes in radiance would be large in certain spectral intervals, but the change in the integrated radiance (called the flux density) would be very small at the top of the atmosphere. Although the radiative signal would be stronger at the Earth's surface than in space, there would be more noise in the signal resulting from the natural variability of temperature and specific humidity in the lower troposphere.

Attempting to measure changes in downward radiance at fixed surface locations does not appear promising in the near future as an approach for detecting CO_2-induced effects on the radiation budget. The primary uncertainty is associated with our limited understanding of the absorption and emission of energy by water vapor. Shifts in the spectral distribution of outgoing radiation at the top of the atmosphere measured by satellites may provide an indication that the CO_2 concentration is increasing, but that is much more easily determined from ground-based sampling. Satellite measurements may also indicate whether radiative fluxes to space are emanating from higher in the atmosphere, as is expected to occur as the CO_2 concentration increases. However, these measurements would not provide detailed information about changes in the radiation budget of the lower atmosphere. Both measurement approaches would benefit from supplemental measurements of the temperature and specific humidity profiles to aid in the interpretation and analysis of the results.

Measurement of changes in downward radiance at fixed surface locations does not appear promising in the near future as a technique for detecting CO_2-induced effects on the radiation budget because natural variations in radiative fluxes are large at the surface and only a very limited areal coverage would be possible.

Shifts in the spectral distribution of the outgoing infrared radiance at the top of the atmosphere measured by satellites may indicate whether radiative fluxes to space are emanating, as expected, from higher in the atmosphere as CO_2 and trace gas concentrations increase. A major source of uncertainty is the lack of understanding about the radiative properties of atmospheric gases, particularly the H_2O absorption continuum, the shape of absorption lines, and the line parameters (and their temperature and pressure dependence). Efforts to reduce instrument noise, provide higher spectral resolution, and provide accurate and stable calibration would contribute to the measurement capabilities.

CLIMATIC DATA BASES

Over the past 5 years, intensive effort has been devoted to improvement of the data bases documenting climatic behavior. The air temperature record over land has, where possible, been extended back from 1880 to 1850 and many more stations have been included. This data set has been carefully scrutinized to identify effects due to changes in the time of recording of temperature, changes in station location, and of urban warming, among other factors. All of these factors can lead to apparent, but false, indications of climate change. A cooperative international effort has made available new records of sea surface temperature and of surface air temperature over the ocean that should greatly improve the spatial representativeness of this record. Records of ocean air temperature and upper atmospheric temperature profiles also have been assembled.

To aid in the analysis and interpretation of temperature changes, data bases also have been developed for volcanic aerosol loading and solar irradiance. The data bases for these quantities vary significantly in the amount and quality of the data, and some of these data sets are inadequate in both respects.

Despite these many efforts to assemble and check the data, there remain limitations in the quality of the data sets. In the case of temperature data, which is the best and most used data set, reliable and standardized instrument techniques were not developed until late in the 19th century. Unfortunately, standardization of measurements did not extend to a policy of a universally adopted observation time so that, even today, significantly different procedures are followed in deriving daily temperature averages. The early data networks were sparse, and, as additional stations were added, they were located primarily over land areas of the Northern Hemisphere. Consequently, data are particularly sparse over large areas of the oceans and the Southern Hemisphere. A significant loss for ocean areas occurred with the reduction in the number of ocean weather stations (located on stationary ships) in the 1970s.

The effects of changes in thermometer type and exposure are considered to be slight, but changes in the technique of measurement of sea surface temperature over the last 100 years have had a significant effect on the readings. Early ship data were taken using the "bucket" method. Over the last 50 years, sea surface temperatures have increasingly been determined using thermometers located in the cooling water intakes. These injection temperatures are estimated to be 0.3 to 0.7°C warmer than the bucket temperatures, which is comparable to the change in temperature estimated to have occurred since the middle of the 19th century as a result of the increasing CO_2 concentration. An important change that can affect a particular station is the growth of towns and cities around the site (the urbanization effect). Increased urbanization around many stations may introduce a warm bias into computed regional temperature trends.

Recent cooperative international efforts have yielded important new data derived from ship records of sea surface temperature and air temperature at sea. Continued expansion of the set of observations of marine air and surface temperatures is essential. Observations at key benchmark stations with long records must be continued.

Data sets documenting changes in volcanic and tropospheric aerosol concentrations and in solar irradiance contain uncertainties that limit their use in interpreting and explaining past climate changes. Continuing and improved monitoring is needed.

ANALYSIS OF CHANGES IN TEMPERATURE

The global mean temperature change is of primary interest because it is the most reliably modeled climatic effect of the increasing CO_2 concentration. Because of limitations in data coverage, there is considerable uncertainty in our knowledge of how global mean temperatures have varied in the past and thus, many workers have used the Northern Hemisphere land-based record as a global proxy. Efforts have been made to fill data-void areas so as to develop hemispheric-mean and global-mean data sets, but such efforts can introduce important uncertainties. In the near future, virtually complete global coverage may be obtainable using satellite data, provided that appropriate calibration techniques can be developed and the satellite and surface records can be related. For the present, however, traditional techniques must suffice.

Five groups have recently published time series of large-scale average surface air temperatures. For land data, the various results are generally highly correlated and show changes of the same magnitude, which is not unexpected because most of the data sources used are common to all analyses. Differences do arise, however, from differences in the number and geographical location of stations employed, from differences in the methods and extent of extrapolation from data-rich to data-poor areas, and from differences in averaging procedures.

Four phases covering the last 130 years can be identified in the time series of Northern Hemisphere temperature: (1) cooling from the mid- to late 1880s, (2) warming from the late 1800s or early 1900s to the 1940s, (3) cooling to the mid-1960s or early 1970s, and (4) warming since the early 1970s. Although the entire globe has shown varying warming and cooling trends, the trends in different regions have tended to differ from each other and from the global and hemispheric mean trends. For example, although there are less vast data for the extensive ocean areas in the Southern Hemisphere, a gradual and more nearly monotonic warming seems to have continued over the period since about 1910. Reconciliation of these different trends is an important problem for detection studies.

Detection of changes in climate that can be unequivocally attributed to the effects of the increasing atmospheric CO_2 concentration involves two steps. The first is a statistical analysis of data directed toward identifying, at a known confidence level, a change in one or more climatic variables. The second is the attribution of at least part of this change to the increasing CO_2 concentration. Statistical attribution is made difficult by the natural variability of the climate, which acts as a noise against which the CO_2, or CO_2 plus trace gas, signal must be detected. In choosing a climatic variable or parameter to monitor for detecting CO_2 effects, it is essential to have both a well-defined signal and a well-defined noise level. The variable that presently best satisfies these criteria is large-scale, area-averaged surface air temperature.

If model results are correct, the rise in the CO_2 concentration since the middle of the 19th century should already have caused an appreciable global surface warming. The amount of the CO_2 warming estimated to have occurred depends on (1) the preindustrial CO_2 concentration, (2) the size of the model-predicted equilibrium temperature increase for a doubling of the CO_2 concentration, and (3) the damping of the response resulting from oceanic thermal inertia effects. Simple energy balance model calculations suggest that temperature changes in response to the increasing CO_2 level may lag a decade to almost a century behind the predicted equilibrium response because of the oceanic effects. Based on land data, the observed increase in global mean surface air temperature since 1850, although apparently oscillatory, is in the range 0.3–0.7°C. (Marine data are in accord with this estimate back to around 1900, but the two data sources diverge prior to this date.) When this temperature range is combined with the probable range of initial CO_2 concentration (260–280 parts per million by volume [ppm]) and the range of ocean lag times mentioned above, then the observations are consistent with an equilibrium temperature increase for a CO_2 doubling in the range of about 1 to 5°C (as calculated using a parameterized ocean-atmosphere model). This result is in broad agreement with the range of climate model estimates of the temperature increase to be expected for a doubling of the CO_2 concentration, which span the range from about 1.5 to 4.5°C.

If more precise values were known for the preindustrial CO_2 concentration and the lag effect of the oceans, a sharper evaluation of climate model results would be possible. For example, if the CO_2 concentration in the mid-1800s was 260 ppm, then the observations would be consistent with a range of equilibrium temperature change from about 1 to 2.5°C, which is not in obvious agreement with the most recent general circulation model results, which project values of about 4°C. Analysis is further complicated by the possible influences that volcanic activity and other forcing factors may have had on the temperature record since 1850. Consequently, although observations of air temperature over the last century are qualitatively in accord with theoretical projections of the climatic effects of the increasing CO_2 concentration, unequivocal identification of a CO_2 signal will require more accurate understanding of the role of the oceans, better model calculations (so that the spatial pattern of projected changes can be sought), and continued improvement of climatically important data bases.

Important progress has been made in assembling climatic data bases, although a major gap still exists in coverage over Southern Hemisphere ocean areas. An essential task now is the integration of different data sets and the explanation of apparent discrepancies between them.

A major problem in detecting the climatic effects of the changing CO_2 concentration is in explaining the decadal and longer time scale fluctuations in the temperature record, particularly the cooling of Northern Hemisphere land areas that occurrred from about 1940 to 1970. Until these medium time scale fluctuations have been adequately explained or overtaken by further warming, claims regarding detection of CO_2 effects must be viewed with caution.

There are major uncertainties in model simulations that make detection efforts on a regional scale difficult at this time. However, such a comparison is needed to gain confidence in the regional scale climate projections made by models for a doubling of CO_2 concentration and as confirmatory evidence of CO_2-induced effects.

ANALYSIS OF CHANGES IN THE OCEANS

The oceans play an influential role in determining the climate of the Earth, moderating temperature excursions, providing a source of heat and moisture to the atmosphere, and affecting temperature and precipitation patterns over land. Changes in the oceans resulting from the increasing CO_2 concentration may in turn affect the climatic patterns over land. To determine the changes that may already have occurred, data bases of the key ocean variables are needed. The key ocean variables for which some data are available include sea level, temperature, and salinity. The oceans are vast areas, however, and the available data suffer from poor temporal and spatial coverage. There are certain specific locations and small areas where relatively long time series of ocean measurements exist, but for some variables these may not be representative of global changes.

The oceanic variable for which the most representative measurements are available is sea level. A substantial increase in sea level would have a dramatic impact on society because a large fraction of the world's population lives near coastal margins. A sea level rise of more than a meter could have serious ramifications. Although data are sparse, analyses of sea level data completed during the past 5 years indicate that sea level has been rising at a rate of 10–25 cm per century since the early 1900s. A few locations, however, have actually experienced a decline in sea level, contrary to the tendency in many regions. Estimates of the rate of past sea level change can vary by a factor of two simply due to the method of analyzing the same data set, so, narrowing the range of estimates of the global rate of sea level rise will require very careful analysis.

Several processes can cause the sea level to rise. An increase in ocean temperature would lead to an increase of sea level through thermal expansion of the oceans. Thermal expansion of the upper ocean over the past 100 years can explain only a part of the observed change, however. Melting of glacial ice in the polar ice sheets of Antarctica, Greenland, and, particularly, midlatitude mountain glaciers may have contributed significantly to the observed rise, but this is by no means proven. Other possible explanations are highly speculative; a very specific pattern of simultaneous spin-up or spin-down of all of the ocean gyres could result in an apparent increase in sea level in many coastal regions, or a simultaneous subsidence of the continental margins and key mid-ocean islands could be contributing to the observed change.

Globally representative estimates of the temperature change of the upper ocean are somewhat uncertain because of problems with the data sets and analysis techniques. There are only a few studies of subsurface temperature change in the ocean over time scales of decades or longer, and the applicability of these studies to the longer term changes associated with possible CO_2-induced effects is uncertain. These studies provide a valuable starting point for future studies. Long-term, coherent basin-wide changes in the temperature of the ocean at depths below the surface layer are not detectable with the current data set. However, significant local changes have been observed on decadal time scales. These changes may represent a noise that must be successfully filtered out if a larger scale, long-term signal is to be detected.

It is not possible yet to estimate the magnitude of any CO_2-induced changes in the oceans' salinity distribution. Significant long-term changes in salinity may be occurring in a few places in the oceans (for example, in the North Atlantic), but no coherent pattern appears to exist. The problem of noise in the signal appears severe and is unlikely to be ameliorated by current hydrographic sampling programs.

Oceanic data indicate that sea level is rising and that the ocean is warming. Both effects are qualitatively consistent with projected CO_2-induced effects, but a quantitative causal coupling is not yet demonstrated.

The major sources of uncertainty in analyses of ocean variables are the lack of global coverage, the lack of long-term records, and the problem of separating a relatively small signal from noisy data.

ANALYSIS OF CHANGES IN SNOW AND ICE

The principal climatic roles of snow and ice relate to their high reflectivity, the insulating effect of sea ice on the ocean beneath, and the thermal buffering provided by their latent heat. Hence, changes in the extent of snow and ice, their thickness, and albedo are of primary significance in studies of climate changes. Thinning of sea ice allows warming of the winter atmosphere because heat can be more easily conducted from the ocean through the ice to the atmosphere. Reductions in snow and ice cover in response to CO_2-induced warming would tend to amplify the warming by allowing increased surface absorption of solar radiation. This amplification process is termed the ice-albedo feedback effect. The amplified warming would lead to additional melting of snow and ice, thereby making snow cover and sea ice extent potentially sensitive indicators of climate change. On the other hand, ground ice (permafrost), glaciers, and ice sheets are slow to respond to changes in climate.

Detailed records of global snow and ice are generally of much shorter duration than those for most other climate system parameters. Consequently, there are many questions about the variability of these parameters and many uncertainties concerning the representativeness of short-term empirical studies and the modeling of climate-cryosphere (snow and ice) interactions. For example, estimates of the expected CO_2-induced changes of snow cover are currently made by simple interpolation between equilibrium climate states having one and two times the present CO_2 concentration, whereas the time-dependent situation may be nonlinear. The actual changes will depend on regionally and seasonally altered atmospheric circulations and the interaction of the increasing radiative perturbation with the changing seasonal dependence of snow and ice coverage. Detection studies to date have focused on searching for potential CO_2-induced changes in snow cover extent over middle- and high-latitude land masses during the spring, when the thin snow cover may be significantly affected. Because of the large variability in the observational record, however, the CO_2 contribution to changing conditions has not been identified.

Recent studies using satellite data indicate a reduction of sea ice in one sector of the Antarctic Ocean compared with ship observations in the 1930s. Although the change is consistent with the postulated CO_2-induced warming, it is still within the range of natural variability. Modeling studies indicate that reductions in snow cover and sea ice extent should be most evident during the spring and fall because the melting would occur earlier and refreezing would occur later. There is also likely to be a reduction in snow and ice thickness, which may be most evident in late summer and winter when extremes in thickness occur.

A major concern involves the possibility of the collapse of the West Antarctic ice sheet, which is grounded on bedrock below sea level, such that its gradual destruction should lead to a rise in world sea level of 6–7 m over as little as several hundred years. The sensitivity of the West Antarctic ice sheet to a warming is quite uncertain because the effects involve air and ocean temperatures, sea ice extent, and changes in accumulation rate. Estimates of the current mass balance for the whole Antarctic ice cap indicate a net accumulation, but recent iceberg monitoring indicates a rate of iceberg calving that is 3 to 4 times greater than previous estimates, possibly in excess of the net accumulation. Retreat of other icecaps and glaciers could also increase sea level. Mountain glacier recession in midlatitudes could have contributed about half of the observed sea level rise during this century.

Other cryospheric data provide some isolated indications that warming may be occurring. For example, measurements of borehole temperatures in permafrost in northern Alaska imply a warming of about 2°C over the past 100 years.

Snow and ice changes that can be directly attributed to the increasing CO_2 concentration are not yet evident, although changes that have occurred are not inconsistent with CO_2 as a cause. Questions about the representativeness of short-term records of cryospheric data at limited geographical locations and the large variability of the data contribute most to the uncertainty in the empirical analyses.

Satellite observations have been providing more homogeneous coverage of global snow and ice extent since the early 1970s.

ANALYSIS OF CHANGES IN PRECIPITATION

Model projections of atmospheric conditions with an increased CO_2 concentration all suggest an increase in the intensity of the evaporation-precipitation cycle. There would be slightly more precipitation on a global average basis, but there could be either an increase or a decrease in precipitation regionally.

To determine whether the increasing concentration of CO_2 is changing the distribution and amount of precipitation, it is necessary to document the past history of the rainfall regime. Although numerous records covering many years exist at many land stations, measurements over the oceans have generally been inadequate. Therefore, estimates of regional and global average values based on past data are highly uncertain. Even though the situation is better for land areas, where it appears feasible to detect regional changes in the record, very important problems must be confronted. For example, instrumentation varies from country to country. Further, individual precipitation systems are relatively small and of short duration, and, accordingly, precipitation records show quite large variabilities on small space scales and short time scales. To generate potentially representative indicators, averaging must be done over large areas and long time frames.

The historical data coverage is not global, so it is not possible to construct a true representation of the global precipitation signal. Further, the data from the land masses do not support the concept of a globally coherent precipitation signal. However, there appears to be a spatially coherent signal in the normalized precipitation anomaly field over a number of continental-scale regions including the United States. The time dependence of these signals is characterized by large decade-to-decade fluctuations on the order of one standard deviation. These variations represent a very large natural noise against which it will be exceedingly difficult to detect a relatively small CO_2-induced signal.

Because CO_2-induced variations may not be uniform (e.g., enhancing tropical and reducing midlatitude precipitation), regional trends and fluctuations must be examined. The recent severe drought in the Sahel has caused numerous investigators to study the past variation of rainfall throughout Africa. The precipitation field over Africa is characterized by large-scale anomaly patterns. The patterns of drought and wet periods persist for years and thus constitute a large background noise. There has been a strong trend toward decreasing rainfall over the continent, which is supported by measurements of the Nile discharge. The cause of this decline in rainfall has not yet been fully determined. These changes are an example of the natural variability on time scales that may obfuscate detection of CO_2 effects and that, on the regional scale, could well be as important as potential CO_2-induced changes, at least over the next few decades.

Limitations in the record of global average precipitation, particularly over the ocean, make near-term identification of any CO_2-induced signal in this record extremely difficult.

When averaged over subcontinental to continental scales, projected CO_2-induced changes in precipitation—even if such changes become better defined as models improve—are likely to be hidden by long-term fluctuations, which are sometimes coherent and perhaps due largely to natural causes.

SUMMARY OF DETECTION STUDIES

The atmospheric CO_2 concentration has increased measurably since the middle of the last century. Northern Hemisphere land temperatures, sea surface temperatures, and sea level have also increased during this period. Model projections of the climatic response to an increased CO_2 concentration indicate that such changes should be expected. The apparent agreement strongly suggests a causal relation.

A critical issue for detection studies is to establish a quantitative relationship that can account for the as-yet unexplained features of the climatic record and is in agreement with model calculations of climatic sensitivity. The non-uniform temporal pattern of the warming on land, particularly the 1940 to 1970 cooling in the Northern Hemisphere, and the conflicting patterns of the changes in the land and ocean records are particularly perplexing, although these difficulties may become less important as the CO_2-induced warming continues to increase over the next few decades. It is particularly important to determine the fraction of the projected equilibrium sensitivity calculated by models that should be evident in current climate records and how this fraction may change with time. This will require improved understanding of oceanic uptake of heat and its transport into the deeper ocean, both theoretically and as portrayed in climate models.

Developing the needed quantitative causal relationship can be accelerated by pursuing signal-to-noise, noise reduction, and multicomponent detection strategies. All three approaches require improved and extended data bases, further analysis, and more accurate modeling studies. Although some studies using each of these approaches have indicated the presence of a CO_2 effect, aspects of these different analyses are not consistent with each other and the derived CO_2 effect is somewhat smaller than suggested by recent climate modeling studies.

Several factors make it impossible to predict precisely when the CO_2-induced changes will be able to be identified with convincing statistical significance. These factors include the uncertainties present in model projections of the induced climate changes, particularly because of uncertainties in representing the oceans, and the possibility that climatic perturbations resulting from changing influences by other causal factors (e.g., volcanic and solar activity, ocean temperatures) could disguise the expected effect of increasing CO_2 and trace gas concentrations. If CO_2 and trace gas concentrations continue to rise as projected and model calculations are essentially correct, the increasing global scale warming should become much more evident over the next few decades. If such changes do not become apparent, our understanding of the uncertainties and completeness of current climate models will require extensive reconsideration.

Trend analysis of long-term records of land and ocean temperatures and sea level are qualitatively consistent with the climate changes projected by modeling studies.

Development of a convincing quantitative cause-effect relationship has been limited by uncertainties in available data sets and in model projections of expected changes, particularly concerning the role of the oceans in delaying the projected climatic warming. Depending on the relative roles of various causal factors, the CO_2 signal should become much more evident over the next few decades.

TASKS FOR THE FUTURE

The overall goal of the CO_2 research program sponsored by the U.S. Department of Energy is to provide a stronger scientific and technical basis for projecting the climatic effects of the increasing CO_2 and trace gas concentrations. Understanding how key climatic variables have changed since the middle of the 19th century and determining the causes of these changes would contribute greatly to the overall objective of the research program. Validation of the model calculations by comparison with observations is a high-priority task. A two-pronged effort must be pursued: (1) to improve the quality of the data bases and (2) to develop and apply a research and detection strategy that can isolate the CO_2 and trace gas effects from the effects of other forcing factors.

Detection studies require data bases of both the factors that may cause climate changes and of the climatic variables that may be changed. In the State-of-the-Art volume in this series on the carbon cycle, recommendations are presented for improving the record of past CO_2 concentrations. Similar efforts are needed to improve records of changes in solar irradiance, volcanic aerosol loading, and trace gas concentrations over the past 100–150 years and to assure that better records are maintained in the future. Understanding of the climate system and our means of detecting changes also require substantial improvement.

The following research tasks are needed to improve our ability to detect climate change, arranged by the variable being investigated:

1. *Changes in the radiative signal*
   - Accurate measurements of the radiative properties of trace gases (spectral line parameters for the absorption bands and their variation with temperature and pressure) are needed to assess the effect of these gases on the spectral radiance. Also, an effort must be made to validate existing radiative transfer models against laboratory and field measurements.
   - Before proceeding with new monitoring efforts, techniques for extracting meaningful radiative signals from the available radiance data (which have a high noise level due to natural atmospheric variability) must be developed and demonstrated.
2. *Changes in temperature*
   - More extensive data coverage is needed. Satellite data would provide the needed coverage, but the accuracy of temperature retrievals near the Earth's surface is not currently adequate for trend analyses. A considerable effort will be required to determine the correspondence between satellite-derived and surface measurements and to improve calibration and temperature retrieval methods.
   - The causes of the medium (decadal) and longer time scale fluctuations in surface air temperature must be adequately explained.
3. *Changes in the oceans*
   - Sea level stations should be established and sea surface temperature should be better monitored in the Southern Hemisphere oceans. Increased sampling of hydrographic data is needed for selected regions where data records already exist. The sampling program should be sufficiently frequent in time to allow an effective filtering of the high-frequency variability that contributes greatly to the noise level in present data sets.

- Historical archives of ship observations need to be thoroughly examined to identify possible global-scale changes in various ocean and over-ocean climatic parameters. The apparent disagreement in land and ocean records prior to 1900 needs to be resolved. The cause of the recent freshening of North Atlantic deep water and its relationship to climate and to the bottom water formation rate must be investigated.
- Better numerical models of the circulation of the oceans, including prediction of the distribution of temperature, salinity, and density, need to be developed. Carefully chosen information from key ocean regions, when put in the dynamical context of such an ocean model, could help identify the data needed to separate possible CO_2-induced effects from other processes that may be causing long-term changes in the oceans.

4. *Changes in the cryosphere*
   - The factors controlling decadal scale fluctuations in sea ice extent and thickness and the stability of pack ice to changes in climate (and vice versa) need to be better understood. Coupled ice-ocean-atmosphere models need to account for the detailed physical processes known to be important so that improved estimates can be made of the changes in sea ice area and thickness expected to occur as the CO_2 concentration increases.
   - Modeling studies, supported by additional field measurements, are needed to ascertain the stability of the West Antarctic ice sheet and adjacent ice shelves to a CO_2 doubling. This question is critical to projecting global sea level on time scales greater than about 100 years.
   - Better observational data must be taken to determine the mass balance and volume changes of the two major ice sheets and a representative coverage of the world's glaciers in order to assess their actual and potential contributions to sea level rise as global temperatures increase.

5. *Changes in precipitation*
   - The precipitation data base must be expanded and homogenized. Although CO_2-induced effects on precipitation are unlikely to be detectable in the next few decades, detailed studies of this data base are needed so that meaningful estimates of regional averages can be developed to compare with model simulations.
   - Possible feedback mechanisms that may amplify the effects of changes in precipitation need to be evaluated. Changes in rainfall, coupled with changes in air temperature, evapotranspiration, and vegetation, for example, may produce a relatively larger effect on runoff and subsequent river flow.

6. *Coupled changes in climatic variables*
   - Improved model calculations are needed that relate, on a regional and seasonal basis, the expected CO_2-induced changes in many climatic variables as a function of time. These changes must be differentiated from the coupled responses that may arise as a result of natural climatic variations and perturbations forced by other causal factors such as volcanic emissions, changes in solar irradiance, and long-term interactions between different components of the climate system.
   - Data sets for individual variables must be improved and compared to assure a continuation of comparable coverage, quality, and length of record.

1. THE CHALLENGE OF DETECTING CLIMATE CHANGE INDUCED BY INCREASING CARBON DIOXIDE

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CONTENTS

1.1 INTRODUCTION

Determination of the potential climatic effects of increasing atmospheric carbon dioxide (CO_2) is an important and challenging problem for the climate research community: important because such effects may represent a major, extended alteration of the climatic regime with far-reaching economic and social consequences; challenging because it tests the ability of observational, theoretical, and modeling studies to provide a satisfactory description of a specific climate change. If we cannot provide a clear explanation, it suggests our inability to answer a wide variety of other questions involving both the potential human impact on the climate and the potential for natural climate change (see National Research Council 1983).

In the past several years the climate research community has devoted a great deal of attention to the problem of the isolation and detection of the climatic changes expected to be induced by the increasing CO_2 concentration (e.g., World Meteorological Organization 1982; National Research Council 1983). Recent work has included the assembly and examination of new geological, historical, and instrumental data; empirical analyses attempting to attribute climatic variations to various astronomical and atmospheric factors; and the design, application, and analysis of new numerical experiments with climate models. Important new insights into the possible climatic response to the increasing concentration of CO_2 have been obtained, together with a better understanding of the sensitivity of the overall climate system to both anthropogenic and natural perturbations. Although this research has yielded neither an unequivocal nor complete identification of the effects of increasing CO_2 on the climate, it has provided many of the scientific tools and much of the evidence required to do so.

The purpose of this introductory chapter is to discuss the framework that provides the basis for efforts to detect the climatic effects of increasing CO_2 concentrations. As background, the second section reviews some of the characteristics of climate, how it has changed, and some of the factors that may play a role in inducing such changes. This overview will also serve to introduce some of the terminology that has developed in CO_2-climate studies. The following sections describe approaches for determining if the climate is changing now (or has changed in the recent past) and, if so, whether it is the increasing CO_2 concentration that is causing these changes.

1.2 CHARACTERISTICS OF CLIMATE AND CLIMATE CHANGE

Climate is in some sense the average of weather, but a clearer statement or definition is required. One such definition (National Academy of Sciences 1975) says that climate is the statistical description of the mean state (including the variability) of the atmosphere, ocean, ice, and land surface in a specific region and over a specified time period. The conventional period of averaging is at least 30 years. For times prior to the instrumental record, the averaging period is often longer because the temporal resolution of the proxy method used to estimate the climatic state is often quite limited.

Climate *change* or *variation* is defined as the difference between climate statistics evaluated over two similar time intervals; for example, a change of January climate from one year to the next or a change of global climate between centuries. All of the atmospheric, oceanic, cryospheric, and land surface statistics required for a complete specification of climate are never available. Because of this, it is common to consider only some of the variables in the climate system. In comparing different climates it is important to remember that changes of only a single element may not completely characterize a change of climate, and that two apparently similar climates may in fact be different because elements omitted may be dissimilar.

1.2.1 Factors Controlling the Climate

The Earth's climate is not random; it is instead an organized condition determined by the interplay of a large number of factors. Some of these factors are independently determined by influences not normally considered part of the climate system; these factors are referred to as *external* forcing factors. But the climate is also the result of interactions within and among various components of the climate system itself; these factors are referred to as *internal* forcing factors.

1.2.1.1 External Forcing Factors

The fundamental factor responsible for the maintenance of the Earth's climate, and thus a potential cause of its variation, is the shortwave radiation received from the Sun. Of the incoming solar radiation, about 50% is absorbed at the Earth's surface and 20% by the atmosphere. The remainder is reflected to space. Were it not for the atmosphere and oceans, however, this solar radiation would only be able to create a climate such as that which exists on the Moon. The radiatively active gases in the atmosphere, including water vapor, CO_2, and other gases present in trace amounts (*trace gases*), act together as an essentially invisible blanket, letting the solar radiation enter but restricting the emission of terrestrial infrared radiation. This effect, often referred to as the *greenhouse* effect, accounts for a net warming of the overall Earth-atmosphere system by about 30°C. Changes in the solar radiation at the top of the atmosphere or changes in the concentration of CO_2 and other radiatively active trace gases are, therefore, possible external factors that may result in climate change. Because the climatic effects of trace gases and CO_2 are so similar, it will be very difficult to distinguish them. In this volume, trace gases will be included implicitly when speaking of increasing CO_2 as a factor in climate change.

Except for the long-term changes in the seasonal distribution of solar radiation associated with variations in the Earth's orbital parameters (the so-called Milankovitch variations) and recent measurements of small transient variations that may relate to sunspots or to changes in the Sun's diameter, it is not known whether the Sun's shortwave radiative emission has changed appreciably over the past few hundred million years. The atmosphere's content of radiatively active gases, however, is believed to have undergone significant variations over this time, partly as a result of interactions with the evolving surface and ocean biomass. Based on measurements of the gases trapped in glacial ice, it has been estimated that the CO_2 content of the atmosphere has varied by more than 20% between the glacial and interglacial periods that have occurred over the last several hundred thousand years. Additional geological evidence indicates that the atmospheric CO_2 concentration may have changed by much larger amounts (i.e., by a factor of several) during the past few hundred million years. Interestingly, numerical models of the Earth's climate suggest that a 2% change in the solar constant and a doubling of the current atmospheric CO_2 content would have generally similar effects on the global surface climate. This similarity in effects seems to result even though the heating due to CO_2 is distributed through the troposphere whereas that due to solar radiation is largely confined to the surface.

The content of water vapor in the atmosphere is controlled by the temperature and by the processes of evaporation and condensation. Not only is water vapor responsible for much of the atmosphere's present greenhouse warming, but its presence serves to effectively amplify the warming that may be caused by other effects (such as an increase in solar radiation or an increase in CO_2 concentration). This is the result of a strong positive feedback between water vapor and temperature, whereby a change in temperature results in a change in the water vapor content, which in turn reinforces the initial temperature change. Although this feedback mechanism may be offset to some extent by clouds, it results in the direct radiative heating of increased CO_2 (or other trace gases) being amplified several fold, and is responsible for most of the warming commonly associated with enhanced atmospheric CO_2. These processes are discussed in more detail in the companion State-of-the-Art Report on *Projecting the Climatic Effects of Increasing Carbon Dioxide* (MacCracken and Luther 1985).

Atmospheric aerosols also may be considered an external forcing for the climate, even though they reside within the atmosphere along with CO_2. The normal or background concentrations of dust, particles, and smoke are actually the result of surface interaction with the atmosphere, although they are not often treated as time-dependent variables. More clearly external to the atmosphere-surface system are the aerosols from volcanic eruptions that may have an influence on the climate for periods of months, seasons, or years.

Another factor generally regarded as external to the climate system is the surface biomass. On time scales of decades to centuries, the surface biomass responds primarily to changes of the surface temperature and precipitation, and the climate may be altered in turn by changes in the vegetation and soil properties. The nature of these biogeophysical feedbacks is not well understood, although there is evidence that they may be particularly important in the maintenance of some subtropical climates.

Other external factors that could affect the climate are the basic physical parameters of the Earth-atmosphere system, such as its mass, the rotation speed, and the distribution of land and ocean at the Earth's surface. Although all of these factors have changed over the course of Earth's history, they may be considered constant on time scales connected with the CO_2 question.

We must consider the possible effects of external forcing factors on the climate in any attempt to isolate or detect the climatic effects of any single factor such as CO_2. Although some information on the effects of the variation of external influences is available from climate models, their effects must be estimated from the observed climate record with the aid of special sensitivity studies and diagnostic analyses. The possible effect of such external forcing provides a range of climate change against which the effect of increased atmospheric CO_2 must be judged.

1.2.1.2 Internal Forcing Factors

In addition to the climatic effects of external forcing, the climate may vary on monthly, seasonal, interannual, decadal, or longer time scales without any change of external factors whatsoever. Such internal or natural climate variability is a result of the interactions among the many physical processes in the atmosphere (and in the ocean and on the land surface, if they are included in the internal climate system). These processes represent the primary feedback linkages among the variables of the system. They include effects such as the horizontal advection and vertical convective transport of temperature, moisture, and momentum; the interaction of clouds, radiation, and temperature; and the interactions of temperature with surface snow, ice, and ground moisture. Because the atmosphere is free to move and supports wavelike phenomena whose influence may be felt at a distance, such effects are ultimately spread throughout the system and give rise to climatic fluctuations on a wide variety of time and space scales. Such fluctuations are likely to be unpredictable to the extent that they may be set in motion by the smallest of perturbations.

The emerging picture of climate change is one in which the (atmospheric) climate displays a spectrum of seemingly erratic fluctuations, because the effects of a temporary local enhancement of one feedback process or another can ultimately be spread to all parts of the system. This internal variability has been called *climatic noise* and takes place in addition to the background of average climatic conditions determined by the external forcing and basic geometry of the climate system (and possibly by the climatic noise itself). The amplitude of this variability has a characteristic value for each climate variable, depending on how the variable is involved in the major feedback processes and how sensitive those processes are to other (internal) influences. In general, on time scales up to hundreds of years, the climatic noise can be comparable in magnitude to climatic variations caused by varying external factors, while on longer time scales external factors have a larger influence. Like the mean climate itself, climatic noise displays a characteristic geographical and seasonal distribution. The determination of this distribution for each climate element is an important part of the diagnosis of climate change. An approximate measure of this variability may be obtained from the observed climate record, with due regard to the uncertainties caused by measurement and sampling limitations and to the possible influence of changing external factors that cannot be readily eliminated in an unambiguous way. Such studies can yield an estimate of the total variability, including the effects of both internal climatic noise and any changing external conditions.

1.2.2 The Baseline Climate

The *baseline* climate acts as the background against which we seek the climate changes projected to occur as CO_2 concentrations change. This baseline is an estimate of the climatic conditions that would prevail in the absence of the significant increase in CO_2 and other trace gas concentrations that started in about 1850 (see Trabalka 1985).

1.2.2.1 Climate Over the Last 10^4 to 10^5 Years

On the time scale of tens of thousands of years, the consensus of paleoclimatic evidence suggests that an interglacial stage started about 10,000 years B.P. (National Academy of Sciences 1975). Because the last several interglacial periods have been only about 10,000 years long, arguments have been made that the return to a much cooler natural glacial climate is to be expected within the next several centuries or millennia (Kukla et al. 1972). However, because the mechanisms causing the interglacial-glacial transitions on this time scale are not yet completely understood, the above argument has little predictive value (Mitchell 1973). Much of the long-term climate variation is widely recognized to be due to the changing orbital parameters (an external forcing factor) as the Earth moves around the Sun, parameters that can be calculated from the laws of celestial mechanics (Berger 1977). On that basis the first glacial episode of moderate intensity is expected to peak 4000–5000 years from now (Imbrie and Imbrie 1980; Kukla et al. 1981). If the baseline for the next 5,000 years is considered to be equivalent to the last interglacial-glacial transition, the volume of the land-locked ice would grow to about 30 to 70% of peak glacial volume, sea level would drop by about 20–40 m, and the average global temperature would decline by about 3°C (judging from proxy data and assuming that temperatures dropped in proportion to the ice volume [Bloom 1971]). If the temperature decrease were linear over the next 5000 years, the cooling rate per century would be very small. However, some proxy paleoclimatic evidence indicates that the transitions may have been more rapid and occurred over decades to centuries (e.g., Bryson et al. 1970; Dansgaard et al. 1972), at least at some locations.

As indicated by MacCracken and Luther (1985), a temperature increase of more than 1°C per century may be possible for several centuries as a result of increasing CO_2 and trace gas concentrations. Therefore, over the next few centuries at least, increasing CO_2 and trace gas effects are likely to be larger than any changes associated with a transition to more glacial conditions.

1.2.2.2 Climate Over the Last 10^2 to 10^4 Years

The warmest period during the last 10,000 years occurred approximately 6000 years B.P. (Webb and Wigley 1985). Since then, climate in at least some regions has become colder and drier. Insufficient information is available to state by precisely how much the climate has changed, but the present global mean temperature is unlikely to be more than about 1°C cooler than during that period and the gross features of the atmospheric and oceanic circulation have remained essentially unchanged during the last 3000 to 4000 years (Williams and Wigley 1983; Webb and Wigley 1985).

Climatic variations that have been noted over the last 1000 years have been mainly regional in scope. The most significant climatic variations occurred in Europe, where the climate was cooler in the 17th, 18th, and 19th centuries (the *Little Ice Age*), principally as a result of the higher frequency of cold winters (Lamb 1977). Larger scale climate variations prior to 1850 cannot be calculated because of the lack of data, although efforts have been made to fill part of the gap through analysis of tree-ring widths (Hughes et al. 1982).

Since 1850, instrumental data have started to become widely available. These data are considered extensively in Chapters 3 and 4 of this volume. Although we cannot be sure that the natural climate would not have changed substantially if unperturbed by deforestation, increasing CO_2 and trace gas concentrations, and other societal activities, available evidence indicates that the global average of the temperature of preceding centuries has been relatively stable (i.e., at most several tenths of a degree) compared with the expected changes resulting from the increasing CO_2 and trace gas concentrations.

1.3 THE DETECTION PROBLEM

The considerations noted above pose several problems for the isolation and identification of the climatic change induced by increasing CO_2 concentrations, referred to here as the *detection problem*. The observed climate record displays changes that are due both to external forcing factors and to the internal variability of the atmosphere itself. Unfortunately, these effects are intertwined in observations of the climate system, and separation of them is a formidable task (World Meteorological Organization 1982). Furthermore, increases in radiatively active trace gases other than CO_2 will induce very similar, if not indistinguishable, climate changes. As a result, the combined climatic effect (*signal*) of increasing CO_2 and trace gas concentrations must be sought; separation of the signals can only be provided theoretically by numerical models.

Three approaches have been suggested as means for attempting to identify the combined effect of increasing CO_2 and trace gas concentrations: *signal-to-noise* analysis; *noise-reduction* analysis, and *fingerprint* analysis. In signal-to-noise analysis, separate records of a single variable (e.g., hemispheric average temperature over different periods) are compared to determine if a statistically significant change has occurred. This approach can identify a statistical relationship between, for example, increasing CO_2 concentrations and warming temperatures or between other changes that are expected as a result of model calculations.

The noise-reduction analysis approach attempts to factor out the variations in the record of a particular variable that may be occurring as a result of changes in forcing factors other than CO_2, thereby making the CO_2 signal more susceptible to signal-to-noise analysis. Accomplishment of this requires that the variations in the other forcing factor (e.g., solar irradiance and volcanic aerosol amount) and the resulting climatic effects be well understood (either based on signal-to-noise analysis or model calculations).

The fingerprint analysis approach relies on a multivariate analysis to reduce the obscuring effect of variations caused by other internal and external forcing factors (Department of Energy 1982; MacCracken and Moses 1983; National Research Council 1983). This approach requires that one have results from theoretical or general circulation models (GCMs) available to generate the fingerprint being sought. Although in their present state of development these models are far from perfect, they provide the only method whereby the climate can be studied under controlled conditions. Aside from their expense, a drawback to the use of GCMs in such research has so far been the difficulty in identifying the reasons for a particular GCM's performance or response (including its limitations in simulating the observed climate); that is, different models have been yielding different fingerprints.

Application of these three approaches requires, to varying degrees, a reliance on the available data records, on our understanding of the climatic effects of various processes and forcing factors, and on the models constructed to make climatic projections. Each of these resources has its particular limitations.

1.3.1 Detection and the Observational Record

Whether from internal and/or external causes, it is characteristic of climate change to yield a "wiggly" record regardless of the resolution with which it is measured or simulated; the climate does not seem to be capable of changing either discontinuously or in a sustained linear fashion (see Chapter 4 of this volume). This feature is partly due to the statistical effects of averaging and partly to the operation of feedback processes that serve to regulate climatic excursions. Climate also has a rich spatial structure that needs further exploration, especially in terms of the relation of local and regional climates to that on the large scale.

The reconstruction and analysis of the space-time structure of climate and climate change raises a number of statistical problems that ultimately bear on the question of how to distinguish between one climatic state and another. The first of these is the problem of sampling the variables of the system to construct the statistics of the climate. Depending on the climatic state being considered (i.e., the length of time and the region of space over which data will be assembled), avoidance of excessive sampling errors generally requires that the climatic variables be known with a resolution finer than the dimensions of the climatic state itself. This means that the average temperature over a year, for example, might be reasonably well constructed from monthly data, whereas the average precipitation in a particular region might require daily data at more closely spaced locations. From a statistical viewpoint, the frequency of sampling should be such that further increases in time and space resolution produce no significant change in the statistics of the climatic state.

In practice the choices in the selection of sampling frequency are limited, as when the climate of a month, season, or year is being examined with conventional synoptic data. In this case the uncertainty is expressed in the statistics of the calculated climate by means of conventional measures of sampling variance. This statistical (or climatic) uncertainty is generally different for different variables, with the largest uncertainty being for those with the highest intrinsic variability in space and time, such as precipitation.

Once the statistics describing the appropriate climatic states have been determined, the next question is that of estimating the statistical significance of changes in the distribution of either the observed or simulated climate. From the structure of the data constituting the climatic states themselves, any of several statistical techniques may be used to estimate the significance of local changes in both the mean and variance of the climatic variables of interest. By applying such tests at each station or at every point of a GCM climate experiment, those regions may be identified where an increase of CO_2, for example, has produced a statistically significant climate change. As normally applied at individual points, such univariate significance tests do not take into account the spatial dependence of the data, although serial correlation (in time) can be considered. A difficulty with this technique is that apparent significance will be indicated by chance at a fraction of the points corresponding to the statistical risk being accepted. Newer multivariate techniques are more satisfactory in this regard and can be used to yield the large-scale pattern of that part of the apparently simulated climate change that is significant at a given level of risk. By using an a priori sequence of estimated patterns characterized by only a few parameters, the spatial dependence of a GCM's results can be taken into account and their dimensionality greatly reduced in a hypothesis-testing strategy (Frankignoul 1985). In this and other statistical tests, the sampled data base should be sufficiently large so that the effects of natural variability in the simulated (and observed) data are minimized. This is usually done by extending simulations over time or by considering an ensemble of shorter simulations, depending on the kind of climate change being studied.

1.3.2 Detection and Our Understanding of Climate Processes

In searching for the climate change induced by a changing forcing factor, our knowledge of the perturbations to be expected depends on the adequacy of our understanding of the behavior of the climate system and participating processes. As explained in MacCracken and Luther (1985), this knowledge is usually assembled from numerical models that simulate the climate and the effect of various perturbations on the climate. Statistical tests can be used to isolate the changes caused by the perturbation from the natural fluctuations generated by the model climate. In principle, the simulated climate changes that are attributable to various changes in external forcing can be assembled. If more than one forcing factor has been changed at the same time, it is generally difficult to isolate the climatic effect of each. This is also usually the case when different models are compared, or when a model is compared with observation. As valuable as such statistical tests are, they cannot be expected to reveal the true dimensions of the climate changes resulting from a specific physical change such as an increased CO_2 concentration. In other words, statistical tests show us how the model responds and the results to which attention should be paid, but not necessarily those to be believed; models are, after all, only approximate descriptions of nature.

The question of determining the *physical* realism of simulated climate change (as compared with statistical significance only) has not received the attention it must have if the isolation problem is to be satisfactorily solved. The heart of this problem lies in the nature of the GCMs themselves. When such models are used in a climate experiment to determine the changes that are due to a specific physical effect or forcing, the results found are not independent of the particular model used, even though the changes may have been found to be statistically significant. In addition to the choice of boundary and initial conditions, each GCM exists in a number of versions or combinations of the resolution and numerical solution schemes available, and offers a variety of choices with respect to the parameterizations of sub-grid-scale processes that may be used. The climate (and therefore the climate change) simulated by a particular GCM version may therefore be detectably different for each possible combination of these choices, and in some cases these differences are probably comparable to the true (and unknown) climate change being sought. Because there is usually no way of knowing beforehand which version (or versions) of the model (or models) is the best for any particular investigation (unless we have a uniformly near-perfect model), a simulated climate change therefore will be influenced to some degree by the model itself.

By denoting the total simulated climate change as Δc, this concept may be represented functionally by the equation $\Delta c = f(\Delta c_r, \Delta c_m)$, where r denotes the true or physically attributable part of Δc, and m denotes the model-dependent part. In general, the relative sizes of Δc_r and Δc_m are not known, and there is no way to separate them in any particular climate change experiment. By recognizing that Δc_r must be independent of the model being used, and considering only those simulated Δc that have proven to be statistically significant (so that we are not dealing with climatic noise), then it may be possible to form an estimate of Δc_r by somehow combining the results of an ensemble of experiments made with alternative (and presumably equally acceptable) versions of a GCM (or GCMs). Although it cannot be assumed that the ensemble average of Δc_m will approach zero because different models may make similar errors, Δc_r might be identified given sufficient knowledge of the dependence of a model's (or models') solutions on the numerical methods and physical parameterizations employed. It also may be possible to estimate Δc_r directly from a single climate change experiment with a superior model by appropriately transforming Δc. The structure of such a statistical function would then represent the variation of Δc over the acceptable parameter space of the GCM. Although in principle the Δc_r determined in this way from versions of one model should converge toward that found from versions of another model (because for this purpose all GCMs could be considered versions of each other), this is not likely to happen unless the models are themselves reasonably accurate in the first place.

When applied to a wide range of simulated variables over as many scales of space and time as possible, such techniques may be a useful way to isolate the climate change that is due to a specific cause, such as an increased atmospheric CO_2 concentration, but proof or confirmation will rest with observations.

1.3.3 Climate Models and Identification of the Carbon Dioxide Signal

The valid or conclusive detection of a climate change resulting from the increasing atmospheric CO_2 and trace gas concentrations will depend not only on the availability of adequate sets of observed climate data and an adequate understanding of climatic processes, but also on the existence of a reliable method for finding the induced signal in the data.

Before observed climate changes can be attributed to increasing CO_2 and trace gas concentrations, we must first know what changes we are looking for. In the noise-reduction and fingerprint approaches, that part of an apparent climate change that is due to external factors must be systematically identified in the record, and the remaining part must be evaluated in light of the expected response to increased CO_2. In each of these steps, the results from climate models could be an important part of the analysis. For all three detection strategies, model projections are indispensable in demonstrating attribution of climate change to increasing CO_2 and trace gas concentrations (World Meteorological Organization 1982; National Research Council 1983).

Of all the methods available for the analysis of climate change (including synoptic, statistical, and analog techniques), climate models (especially GCMs) possess several unique advantages: they address a wide variety of variables and processes in at least a physically consistent manner; they provide a complete data set of relatively high resolution in both space and time; and they offer the opportunity to experiment with the effects of specific changes in the climate system, including the possibility of climate prediction. Although a major disadvantage of models is their possible inaccuracy, all of their attributes should be exploited in the design of a comprehensive detection strategy; this is especially so for the fingerprint strategy.

Assume that, at some future time, a reasonably accurate GCM (preferably one that includes the ocean) has been run for a long time and that the geographical and seasonal dependence of the model's climate and its variability has been determined. The mapping of the noise levels of all variables in the model would document the background statistical uncertainty inherent in the model. This analysis should include, for example, the interannual fluctuations of monthly, seasonal, and yearly averages, which should be reasonably well determined by several decades of simulation. In addition to these characteristic noise levels, it may be assumed that the distributions of the corresponding time means of the climatic variables themselves also have been satisfactorily determined.

With these data available, assume next that the same GCM has been used in an experiment with an increased (or increasing) atmospheric CO_2 concentration, and that the corresponding seasonal and geographical variation of the model's mean climatic state and natural variability also have been determined. The simulated climate changes due to the increased CO_2 concentration is then given by the difference between the statistics in the perturbed and control runs (including possible differences in the variability), and their statistical significance may be estimated by the methods discussed earlier. In practical terms the significant changes are those that would not be significantly modified by further running of the model, and are the only parts of the output of the GCM to which attention needs to be given.

It is now necessary to determine what portion of the statistically significant climate change of the GCM is due to the particular model being used. As noted earlier, this involves the estimation of the model dependence of the results by performing a series or ensemble of similar climate change experiments (and corresponding controls) with alternative versions of the model (such as with different resolutions, alternative numerical algorithms, or alternative physical parameterizations) and with other models. By knowing the sensitivity of the statistically significant simulated climate change (due, say, to a doubled CO_2 concentration) to alternative (and presumably equally acceptable) model versions, one may be able to estimate the essentially model-dependent part of the results, thereby isolating the model-independent or "real" part. As noted above, this may be given by averaging over an appropriate ensemble of climate change experiments, or it may require more subtle statistical techniques. The central role that this isolation step plays in the identification of a CO_2-induced climate signal has not been sufficiently recognized.

It is these common or model-independent climate changes that may be regarded as likely to be physically significant. These changes form the set from which the change in either a single variable or a set of variables (the fingerprint) that is unique for increased CO_2 may be extracted. For each fixed CO_2 level (such as 600 or 1200 parts per million by volume [ppm] or, more generally, for CO_2 levels as a function of time), the CO_2-induced change of a wide variety of climatic variables can be identified. The change a variable displays will depend on its noise level and its role in the climate system; for example, variables like cloudiness and precipitation may show less significant change than temperature. The resultant mapping would show the geographical distribution of each variable's change on a monthly, seasonal, or annual basis and would constitute an atlas of the expected CO_2-induced climate change signal. If such information were available for several CO_2 concentrations (or from an ensemble of experiments with time-dependent CO_2 concentrations), then an effective schedule of expected climate changes might be constructed (or at least envisaged), with due allowance being made for the delaying effect of the ocean if it is not included in the GCM.

This catalog of the expected physically significant climate changes is the unique contribution of climate models to the detection problem; without such information we would not know which climatic variables were the most likely to display a CO_2-induced change, how much of a change should be expected, and in which regions and in which seasons the change might occur. This information apparently cannot be obtained in any other way and is an essential ingredient in the fingerprint strategy.

Even with such information about the expected changes, however, there remains the problem of applying these strategies to the detection of a CO_2-induced climate change in observed data. Part of the difficulty stems from the incompleteness and inaccuracy of the observations themselves; many climatic variables are either not observed or are observed incompletely, and when they are observed they are often not representative of the large-scale distribution addressed by climate models. An additional part of the detection problem is the fact that the observed data contain the effects of other external influences on the climate, such as volcanic eruptions or large-scale environmental or geophysical events, which are not properly accounted for in the model-generated signal. Although these factors eventually may be taken into account in climate models, there always will be a residual uncertainty in the attribution of an observed climate change;

the purpose of each of the detection strategies, especially the multivariate fingerprint approach, is to make this uncertainty as small as possible.

1.4 SEARCHING FOR THE CARBON DIOXIDE SIGNAL

At present, only a few of the elements necessary for the authoritative detection and attribution of a CO_2-induced climate change are available. As already noted, observations are almost always incomplete, and models do not adequately consider many potentially important processes, such as the interaction of the atmosphere with the ocean and with the surface biomass. Despite these limitations, however, it is important that a systematic detection strategy be set in place as soon as possible to coordinate observational and modeling efforts. What is needed is an operational CO_2 effects detection system that can maintain an ongoing evaluation of all available observations in light of the best available results from climate model experiments and that can periodically estimate the likelihood that CO_2-induced effects are occurring. Such a system also should be able to incorporate new information as it becomes available and to upgrade its assessments in a cumulative manner. The possibility that such a *CO_2 climate watch* may not detect a significant effect, at least in its early stages, is not a sufficient reason for not having such a detection strategy in place; its existence would place current efforts in a coherent context and would provide a rationale for further studies.

In addition to new observations, further research in modeling the climate and its response to increasing CO_2 and trace gas concentrations is needed, and new work should be done on the application of statistical techniques to the detection problem. Statistical methods play important roles in the removal of climatic noise from a model's simulation and in the estimation of the model-dependent portion of the results. As important as statistics are in the judgment of whether or not observed climate changes are a significant match to the model-consensus CO_2 signal (or fingerprint), it is the physical reality of the models themselves (as judged by their simulation of observed climates) on which our confidence must ultimately rest.

It should be emphasized that the limitations of numerical models make definition of the CO_2-induced climate signal difficult, and that the limitations in the observational data sets make statistically significant identification of change in any climatic parameter uncertain. Nevertheless, a search using all available methods and data is under way for the climatic effect of increasing CO_2 concentrations, and the remaining chapters in this report review the state of the art in this search.

Perturbation of atmospheric radiative fluxes is the direct effect of increasing CO_2 concentrations, which occurs before any induced climate changes. Chapter 2, therefore, evaluates the prospects for identifying changes in the radiative fluxes at the surface and at the top of the atmosphere. If the radiative fluxes are perturbed, then the temperature will increase. Chapter 3 describes the data sets that are available for an analysis of whether such a temperature change has taken place and why it may have occurred. Chapter 3 also reviews the availability of some other related data sets, including precipitation, volcanic aerosols, and solar irradiance; the cryospheric and oceanic data sets are described in later chapters. Particular attention is given in this chapter to the factors that may make definitive identification of the CO_2 effect difficult.

Recognizing the limitations in the temperature record, Chapter 4 reviews the attempts to identify whether the climate has changed since the middle of the last century. The problem of determining whether any identifiable changes are due to the increase in CO_2, to changes in other factors, or to natural variations in the climate system is considered. Because changes in atmospheric temperatures are also expected to occur over the ocean, a warming of the ocean is projected. This heating would lead to thermal expansion that, along with water from melting glacial ice, would raise the sea level. Changing winds could alter currents, and modifications of evaporation and precipitation could change salinity. Chapter 5 reviews the evidence for these and other changes in oceanic climate.

Climate models generally project that temperature increases will be larger at high latitudes than at low latitudes, in part because of climatic feedbacks involving changes in the extent of snow and sea ice and in part because of the presence of near-surface temperature inversions. Chapter 6 describes the search for changes in these features as possible measures of CO_2-induced climate change. Climate models also generally project that the intensity of the global hydrologic cycle will increase as the temperature rises. Chapter 7, therefore, describes the search to determine if precipitation has increased, using the available records at many observation stations. Based on an analysis of observational evidence, Chapter 8 provides a review of the earlier chapters and returns to the question of whether the ensemble of characteristic changes that is projected to result from increasing CO_2 concentrations (the CO_2 signal) is evident.

Chapter 9 presents recommendations for further assessments and analyses. Central to this effort is the development of a systematic detection strategy to provide a framework for the coordinated evaluation of the findings of observational studies and the projections of climate models.

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2. SEARCHING FOR THE RADIATIVE SIGNAL OF INCREASING CARBON DIOXIDE AND OTHER TRACE GASES

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CONTENTS

2.1 INTRODUCTION

This chapter discusses the spectral distribution of radiative energy emitted in the Earth's atmosphere and changes in the spectral distribution caused by changes in the amount of emitting gas and by changes in the temperature structure of the atmosphere. The specific way in which the changes in spectral distribution serve as an indication of what factors are causing changes in gas amounts and temperature is also discussed. Currently, measured radiances from satellites are used to retrieve such atmospheric variables as the vertical profiles of temperature and gas amounts. These quantities are then used in numerical weather prediction models. This chapter discusses the use of measured radiances for *climate* problems. The strategies discussed in this chapter for detecting the radiative signal of increasing carbon dioxide (CO_2) would employ data that have been and still are being collected from operational satellites. The methods described also rely on the actual radiances. It must be stressed, however, that the problem of detecting the radiative signal due to these gases is similar to most problems in atmospheric science. That is, there are a large number of parameters to the problem which are changing at the same time. Through a comparison of measurements and calculations such as those presented in this chapter, the number of causes for a given change might be narrowed considerably.

2.2 THE SPECTRAL DISTRIBUTION OF LONGWAVE FLUX IN THE ATMOSPHERE

The radiative balance of the Earth's atmosphere and how this radiant energy is distributed in space, as well as in the frequency domain, must be understood to facilitate the detection of any radiative signal. The distribution of outgoing radiative energy per unit area per solid angle per wave number interval for the Earth's atmosphere (radiance) calculated from a detailed narrow band model, which is an extension of Kiehl's (1983) model, is shown in Figure 2.1. The spectral resolution of the model is 5 cm^{-1}. The transmission of longwave radiation through the atmosphere is calculated for each of these spectral intervals with a Malkmus random model. The details of this model and the spectral data that it employs have been discussed by Kiehl and Ramanathan (1981). A description of how radiance is calculated with a narrow band model is given in Chapter 2 of MacCracken and Luther (1985). Vertical profiles of temperature and gas concentrations are needed to calculate atmospheric radiances. Profiles of temperature, ozone (O_3), and water vapor (H_2O) used in the calculations are those given by McClatchey et al. (1971). The vertical mixing ratio of CO_2 is assumed to be constant at a value of 340 parts per million by volume (ppm). All results are for cloud free conditions. Outgoing radiances at the top of the atmosphere are simulated as viewed straight down. The radiance is plotted as a function of wave number, ranging from 0 cm^{-1} (wavelength >2000 μm) to 2500 cm^{-1} (4 μm). The more important features to note about this figure are that (1) the largest emission to space occurs between 800 and 1000 cm^{-1}, (2) there are large reductions in emission centered near 650 and 1000 cm^{-1}, and (3) there is very low emission for wave numbers beyond roughly 1500 cm^{-1}. Each of these features is important in detecting a radiative signal resulting from increased greenhouse gases.

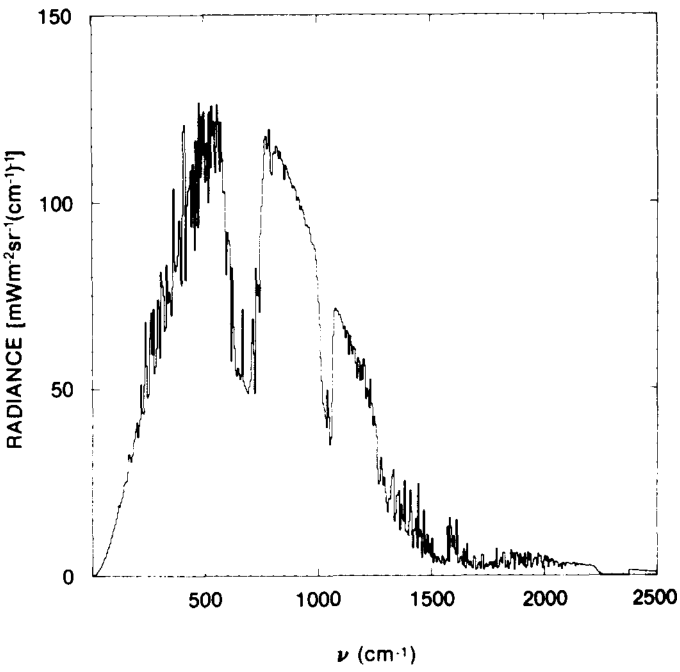

Figure 2.1. Outgoing radiance as a function of wave number for the midlatitude summer profile of McClatchey et al. (1971).

The large emission between 800 and 1000 cm^{-1} originates from the surface of the Earth. The only active absorber of significance in this spectral region is water vapor. This absorption is most important in the lowest 1–3 km of the atmosphere. If increases in greenhouse gases do lead to increases in tropospheric temperatures, accompanying increases in atmospheric water vapor would be expected. Thus, the emission in this spectral region, 800–1000 cm^{-1}, would be affected to some extent. The large decrease in emission around 650 cm^{-1} is caused by absorption by the 15-μm band (bending mode) of CO_2. This radiant energy originates from the colder stratosphere. The spike in radiance located right at 667 cm^{-1} comes from the upper stratosphere. An increase in the atmospheric amount of CO_2 would be expected to alter the emission in this spectral region. The decrease in emission near 1000 cm^{-1} is caused by the absorption resulting from O_3. Changes in O_3 because of anthropogenic activity also will result in changes in the emission in this region of the radiance curve. Finally, the magnitude of the radiance beyond 1500 cm^{-1} is roughly two orders of magnitude smaller than the maximum emission at 800 cm^{-1}. Although relative changes in this region may be large, the absolute change will be very small. However, the instrument noise is much smaller in this spectral region, so that the detection of changes in this region is feasible (Charlock 1984).

Figure 2.2 illustrates the dependence of the outgoing radiance on latitude and season. Outgoing radiances for tropical (A) and subarctic winter (B) conditions are shown. The most apparent difference between these two figures is the overall reduced emission in the subarctic winter case, which is due to the colder subarctic winter temperature profile. These results show that outgoing radiance is strongly affected by temperature. Thus, variations in temperature will, in turn, lead to variations in radiance. The effect of variations in temperature on the measured radiance will be discussed in Section 2.5.

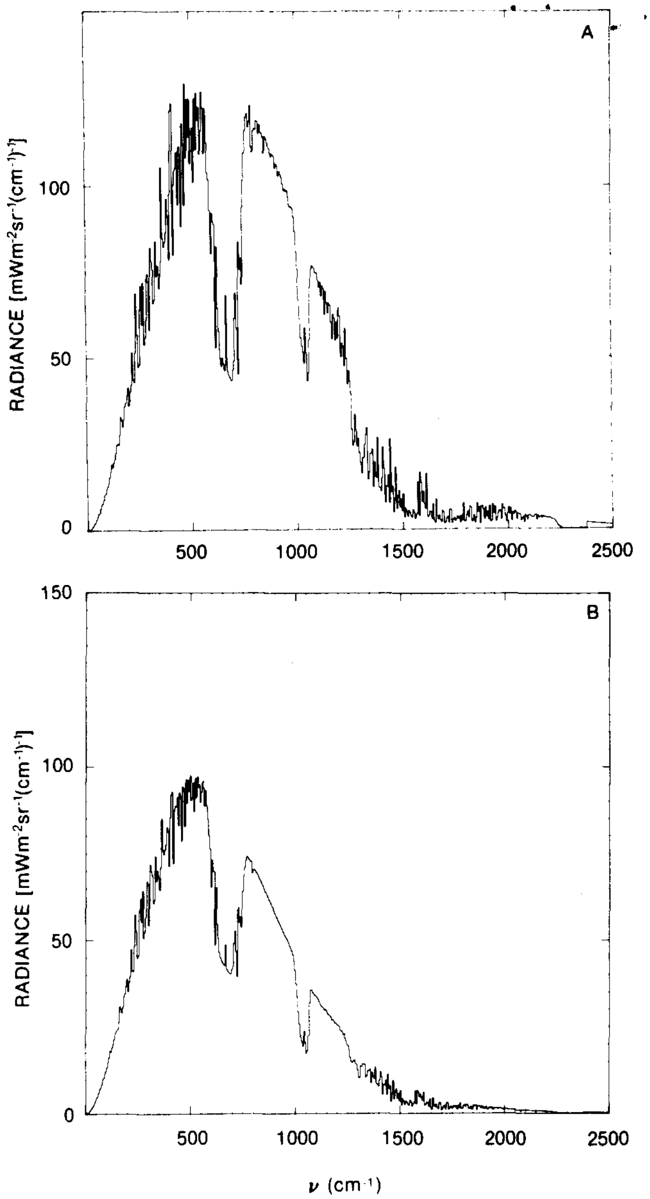

Figure 2.2. Outgoing radiance as a function of wave number for the (A) tropical and (B) subarctic winter profiles of McClatchey et al. (1971).

The longwave radiation emitted by the atmosphere and received at the surface is shown in Figure 2.3 (curve a) for the midlatitude summer condition. The spectral distribution of radiance looks similar to a pure blackbody curve (curve b). The major difference between these two curves is the reduced radiance between 800 and 1300 cm^{-1}. This lower emission is in the window region where there is very little gaseous emission. The small peak in emission near 1050 cm^{-1} is due to O_3. The lower emission in the window region is strongly dependent on the amount of water vapor in the atmosphere. This is illustrated by curve c in Figure 2.3, which shows the downward emission for the midlatitude summer case, but with an increase in water vapor of 50%. The overall magnitude of emission has increased, and the minimum in the window region also has increased. This indicates the radiative importance of the water vapor emission. The importance that changes in water vapor play in detecting a radiative signal will be addressed in Section 2.3.

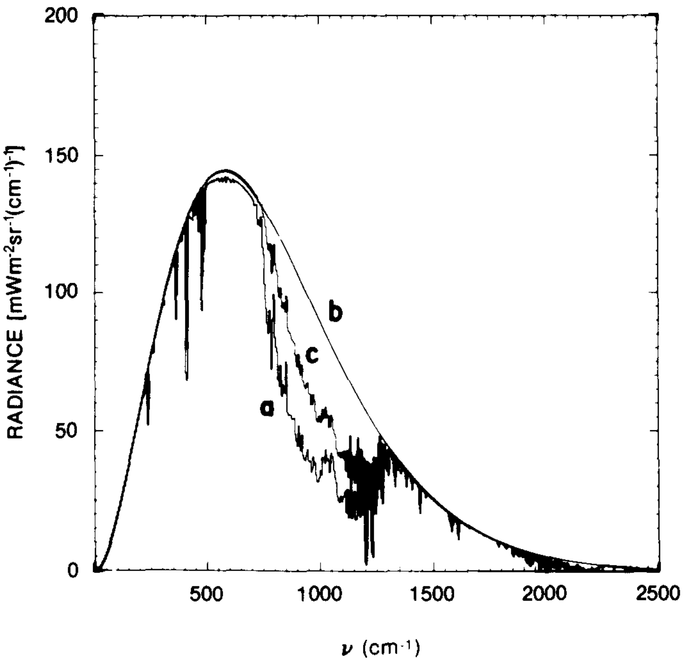

Figure 2.3. Downward radiance at the surface as a function of wave number for the midlatitude summer profile described by McClatchey et al. (1971): (a) standard water vapor amount, (b) pure black body, (c) 50% increase in water vapor.

One of the most important concepts in radiative transfer that is essential for understanding the problem of detecting the radiative signal is the saturation effect of absorption bands. Figure 2.4 illustrates this effect. Shown is the absorption due to CO_2 as a function of the amount of CO_2 (in centimeter atmospheres) for three spectral intervals. The present atmosphere contains about 300 cm atm of CO_2. The absorption for the wave number interval centered at 722.5 cm^{-1} increases linearly with an increase in the amount of CO_2, whereas the relative increases in absorption for the same increase in the amount of CO_2 for the other two spectral intervals are much smaller. These two spectral intervals are nearing saturation. Looking in those spectral regions where the changes will be largest on a relative scale will aid in detecting a radiative signal associated with increasing CO_2 amount. Because the center of the CO_2 band is near saturation, large changes would not be found in this spectral region associated with increases of CO_2. The absorption of longwave radiation, although smaller, is still significant on either side of the band center at 667 cm^{-1}. This region is filled with a number of weaker bands that are not nearly as saturated as the band center. The principle of band saturation applies to all absorbing gases and is important for detecting the radiative signal for other trace gases as well. However, many of these trace gases are not near saturation, because their concentration in the atmosphere is much smaller than that of CO_2.

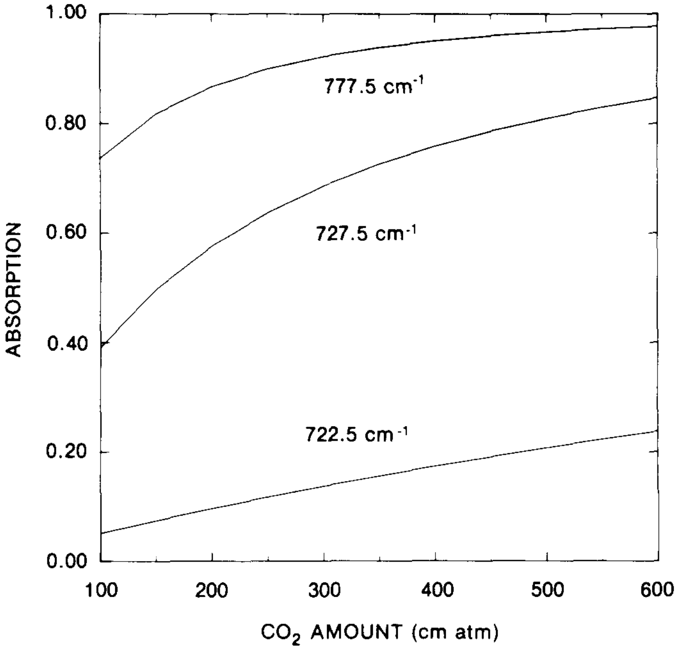

Figure 2.4. Fractional absorption by CO_2 for three spectral intervals as a function of CO_2 amount.

Based on the above discussion, general changes would be expected to occur in the distribution of emitted radiation at the top of the atmosphere and at the Earth's surface because of changes in any gas concentrations. Transmission curves for a number of gases found in the atmosphere are shown in Figure 2.5. Water vapor absorbs over the entire spectral region from 0 to 2500 cm^{-1}. The largest transmission occurs from 800 to 1200 cm^{-1} and for wave numbers greater than 2000 cm^{-1}. CO_2 possesses two strong fundamental bands centered at 667 and 2350 cm^{-1}. There are also a number of weaker bands present between these two strong bands. The strongest band of O_3 is centered at 1042 cm^{-1}, but there are also a number of weaker bands throughout the entire spectral region. Methane (CH_4) has a strong band at 1285 cm^{-1}. Nitrous oxide (N_2O)

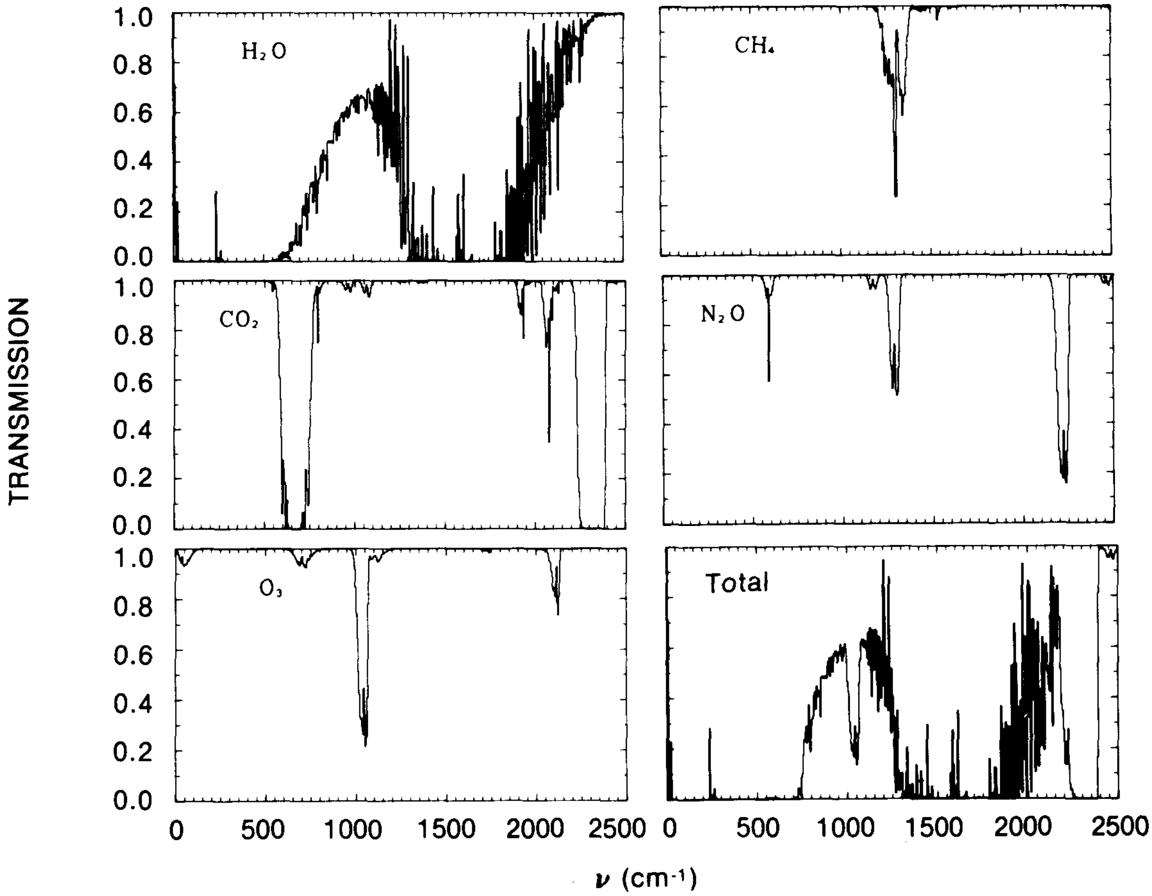

Figure 2.5. Transmission due to various gases present in the Earth's atmosphere as a function of wave number.

possesses three relatively strong bands spread over the entire spectral region. Finally, there are a large number of trace gases present in the Earth's atmosphere which are not shown that possess absorption bands throughout this spectral region. The most important of these gases are the chlorofluorocarbons: $CFCl_3$ (CFC-11), CF_2Cl_2 (CFC-12), and CF_3Cl (CFC-13). The multiplicity of all of these absorption bands for the different gases brings forth another problem for detecting changes in the radiative signal. Figure 2.5 shows many regions where different gases possess absorption bands in nearly the same spectral region (e.g., CO_2 and N_2O, or H_2O and CO_2). If a change in emission is detected in a given spectral region where two or more gases overlap, then this change cannot be attributed with absolute certainty to one gas. This problem implies that the regions that contain only one active absorber should be used for detection purposes.

2.3 CHANGES IN RADIATIVE EMISSION FROM THE ATMOSPHERE

The two spatial locations where changes in the spectral distribution of emitted radiation can be observed are the Earth's surface and outer space. There are definite advantages and disadvantages to both of these locations.

The advantages of a ground-based network to monitor the changes in the radiative spectral distribution are essentially related to the accessibility of the monitoring instrument. High resolution interferometers (<5 cm^{-1}) could be used to look at the downward emission from the atmosphere. This high resolution would enable one to essentially choose specific spectral regions for monitoring. The advantage of this is the elimination of many of the overlap problems that were discussed in Section 2.2. The ground-based instruments also would always be available for calibration checks, leading to an elimination of one source of uncertainty in the signal.

The disadvantages of the ground-based technique are twofold. First, ground-based instruments are limited to land areas. Thus, the spatial coverage would not be global. The area coverage of the available data is important for reducing the sample variance of the data set. Second, the signal may be affected by the large natural variability of temperature and moisture. The magnitude of this variability will be discussed in Section 2.5. It may be very difficult to see changes resulting from the increase in gases because of these effects. However, according to Luther (1983), location of instruments at selectively chosen sites could minimize this problem.

The satellite-based observations would entail use of either the high resolution radiance data as measured by the existing radiometers, such as the High Resolution Infrared Radiation Sounder (HIRS), the Advanced High Resolution Radiometer (AVHRR), the Visible Infrared Spin Scan Radiometer Atmospheric Sounder (VAS), or the use of an interferometer aboard either a satellite or the space shuttle. Each of these instruments has advantages and disadvantages for measuring changes in the radiative signal.

The advantage of using satellites is that they cover a large spatial area. There is almost a decade of National Oceanic and Atmospheric Administration (NOAA) and Tiros N data already available for analysis. These large data sets ensure flexibility in choosing a particular spatial location to minimize noise as a result of natural variability.

The disadvantage of using satellites is that the instruments measuring the radiances are not easily accessible. Accounting for changes in the calibration of the instrument may introduce a source of error into the radiances. Also, the present instruments do not have detailed resolution in many of the spectral regions of interest, and satellite programs are more expensive to implement than ground-based techniques.

2.4 POSSIBLE QUANTITATIVE CHANGES IN THE SPECTRAL DISTRIBUTION OF RADIANCE

To calculate the expected changes in the spectral distribution of radiative emission, some assumptions must be made regarding the increases in the gaseous constituents. The ability to predict the future concentrations of CO_2 and the trace gases depends on many factors, many of which are not yet fully understood. For example, something must be known about the sources and sinks of the gases. This would involve studying the chemical cycles of all important constituents affecting a given gas. Furthermore, it must be known whether these sources and sinks for the constituents are changing with time because of either natural or anthropogenic factors. At present there are many unknowns involved for many of the constituents to be studied. To assess the importance of each of these constituents, scenarios must be chosen that would cover the reasonable future development of the gas concentrations. To avoid this scenario dependence, calculations are performed for a limited number of specified fractional increases in the gas concentrations.

As pointed out in Section 2.1, changes in the amount of atmospheric emission can arise by two distinct means. The emission of radiation to space and to the Earth's surface is governed by the temperature structure of the atmosphere and the amounts and distribution of the absorbing-emitting gases. If any of these quantities change, the atmospheric emission will change. At this point we must ask what are the more important variables to monitor. If the temperature structure does not change when CO_2 is increasing, then the change of the radiative signal is merely an indication of this increase in CO_2. As has been shown previously (Kiehl 1983), the spectral distribution of the change in the radiative signal appears differently for a case in which the atmospheric temperatures do not change while the CO_2 concentration is increased compared with a case in which atmospheric temperatures do change as the CO_2 concentration is increased. More importantly, the change in atmospheric emission occurs at particular locations in the spectral distribution of emission. This spectral dependence of the change in emission allows the difficult problem of causality of climate change to be addressed. By monitoring the changes in many spectral intervals in concert with the type of calculations presented in this chapter, a causal link between an increase in constituents and changes in temperature structure may be established. The three cases to be discussed in detail are as follows: case I, atmospheric temperatures do

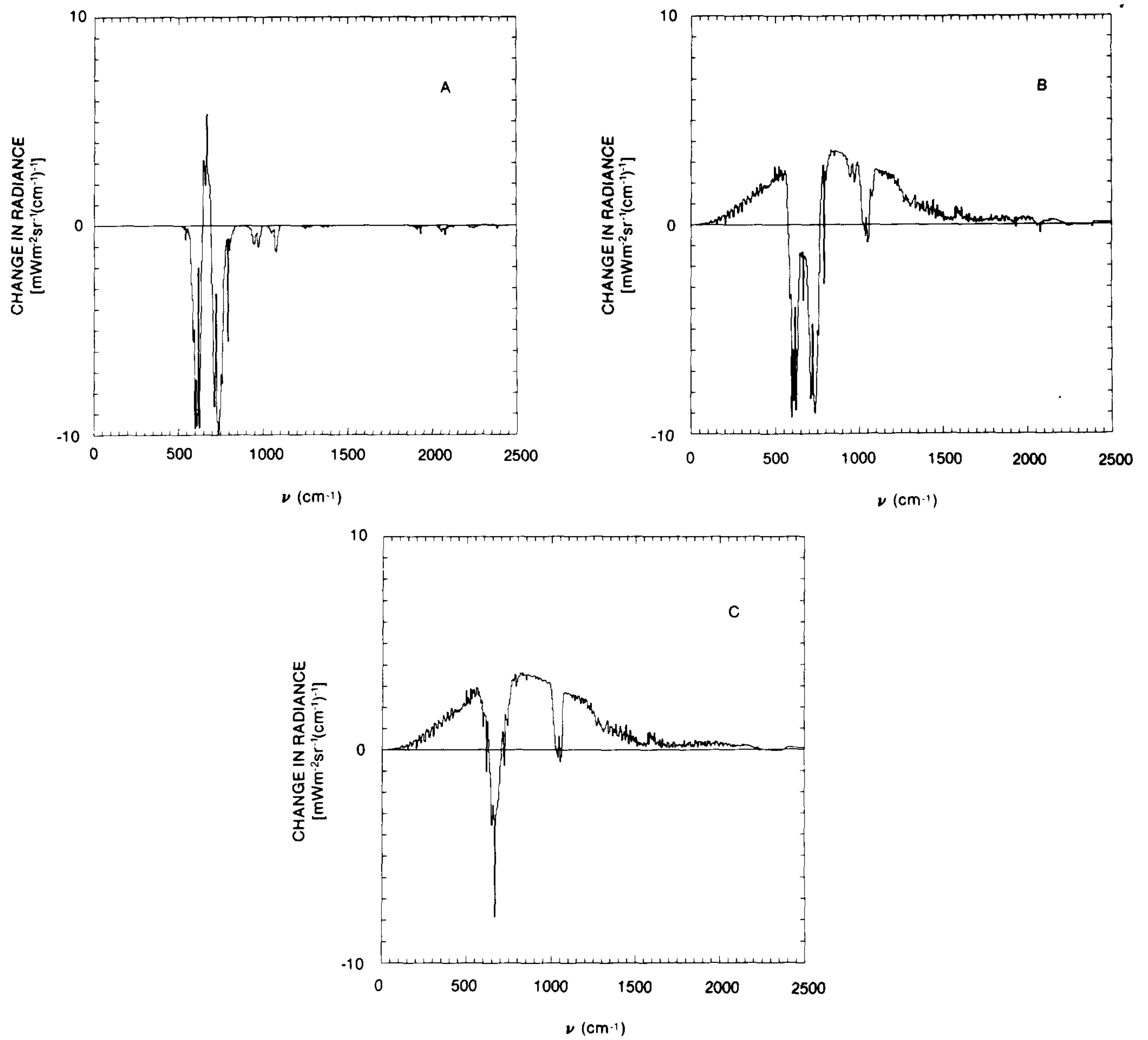

Figure 2.6. Change in outgoing radiance resulting from (A) doubling of CO_2 amount only, case I; (B) doubling CO_2 and changing atmospheric temperatures, case II; and (C) case (II) minus case (I).

not change when the CO_2 concentration is changed; case II, atmospheric temperatures do change when gas concentration is changed; case III, case II minus case I, which illustrates changes in emission caused by changes in gas temperature.

Figures 2.6 and 2.7 illustrate these points. The change in outgoing radiance caused by a doubling of the CO_2 concentration with no change in the atmospheric temperature (case I) is shown in Figure 2.6A. The large change between 500 and 800 cm^{-1} is due to the 667-cm^{-1} bending mode absorption band of CO_2. The smaller changes near 1000 cm^{-1} are caused by the weaker band system. The changes near 2000 cm^{-1} are due to the near infrared bands. As discussed by Kiehl (1983), the increased emission at 667 cm^{-1} arises from the regions of the upper stratosphere. This increased emission leads directly to the large stratospheric cooling that occurs in model calculations of temperature changes that are due to increased CO_2 concentration. The change in outgoing radiance that is due to a doubling of CO_2, along with a change in the atmospheric temperature structure (case II), is shown in Figure 2.6B. The changes in atmospheric temperature structure are determined by computing Equations (2.1) and (2.3), which are defined below. The large decrease centered near 667 cm^{-1} is due to emission from the colder stratosphere. Increased emission on either side of band center is due to emission from the warmer troposphere. The changes shown in Figure 2.6B arise from two sources, increased CO_2 and changes in the temperature structure. The difference between these two results (case III) represents changes due only to the change in thermal structure; this difference is shown in Figure 2.6C. The changes are large between 300 and 1300 cm^{-1} [>1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$]. More importantly, this distribution of the change in radiance is caused by this combination of changes in temperature and CO_2 concentration. This pattern serves as a fingerprint for the assumed change in climate. The same arguments can be applied to changes in downward emission at the surface, which is shown in Figure 2.7A-C for a doubling of the CO_2 concentrations for the midlatitude summer case.

The change in the downward directed radiance caused by an increase in CO_2 alone is quite large (Figure 2.7A). The emission is located at the CO_2 absorption band regions. If the tropospheric temperature is increased along with the CO_2 amount (Figure 2.7B), the increase in downward emission covers nearly all spectral regions. It should be kept in mind that the amount of tropospheric water vapor has not been increased for these cases. If the water vapor amount had been increased, even larger increases in downward emission would result. These calculations indicate that increases in the downward directed radiance resulting from increased CO_2 and tropospheric temperature are quite large.

To assess the relative magnitude of changes in radiance resulting from increased CO_2 and changes in the atmospheric temperature structure, a number of calculations were carried out for various increases in CO_2 concentration for several latitudes and seasons. The increases in CO_2 used for the present study are uniform increases of 10, 25, 50 and 100%. A particular time-dependent scenario for the increase in CO_2 concentration was not used for the calculations because these scenarios depend on many complex factors. The baseline CO_2 amount was chosen to be 340 ppm. The temperature changes were obtained from two sources. The tropospheric temperature changes use the equation given by Madden and Ramanathan (1980). For the tropics and midlatitudes the tropospheric increases in temperature (°C) are given by

$$\Delta T_{trop} = 3 \ln g, \tag{2.1}$$

where g is the fractional increase in CO_2. For high latitudes the tropospheric increase in temperature is given by

$$\Delta T_{trop} = 6 \ln g, \tag{2.2}$$

where the larger increase accounts for albedo feedback processes. Changes in stratospheric temperatures are assumed to be independent of latitude (Fels et al. 1980) and are given by

$$\Delta T_{strat} = -a(z)\frac{(\sqrt{g}-1)}{(\sqrt{2}-1)} \quad \text{for} \quad 1 \le g \le 2, \tag{2.3}$$

where $-a(z)$ is the change in stratospheric temperature due to a doubling of CO_2, and z is altitude. $a(z)$ is latitude dependent only with respect to the change in tropopause height. By using the narrow band radiance model described in Section 2.2, the change in radiance due to increased CO_2 concentrations can be calculated anywhere in the atmosphere.

Figure 2.8A-D illustrates the change in the outgoing radiance that is due to a 10, 25, 50, and 100%, respectively, increase in CO_2 for the tropics. This change in radiance is equivalent to case III discussed above. The change in radiance is due to temperature changes caused by increases in CO_2. The changes for a 10% increase of CO_2 are less than 1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$ at all wave numbers. The magnitude of this change is approximately equal to the magnitude of the natural variation of the daily zonal mean radiance (see Section 2.5). For a 10% increase in CO_2, there is only one place where the change in radiance is large [~3 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$]. This signal is roughly three times as large as the standard deviation of radiance that is due to the natural variability of the atmosphere. For 50 and 100% increases in CO_2, radiance changes are large for many wave number intervals. The change in the downward radiance due to a 10, 25, 50, and 100% increase in CO_2 for subarctic winter conditions is shown in Figure 2.9A-D. The change in radiance is again defined as described above for case III.

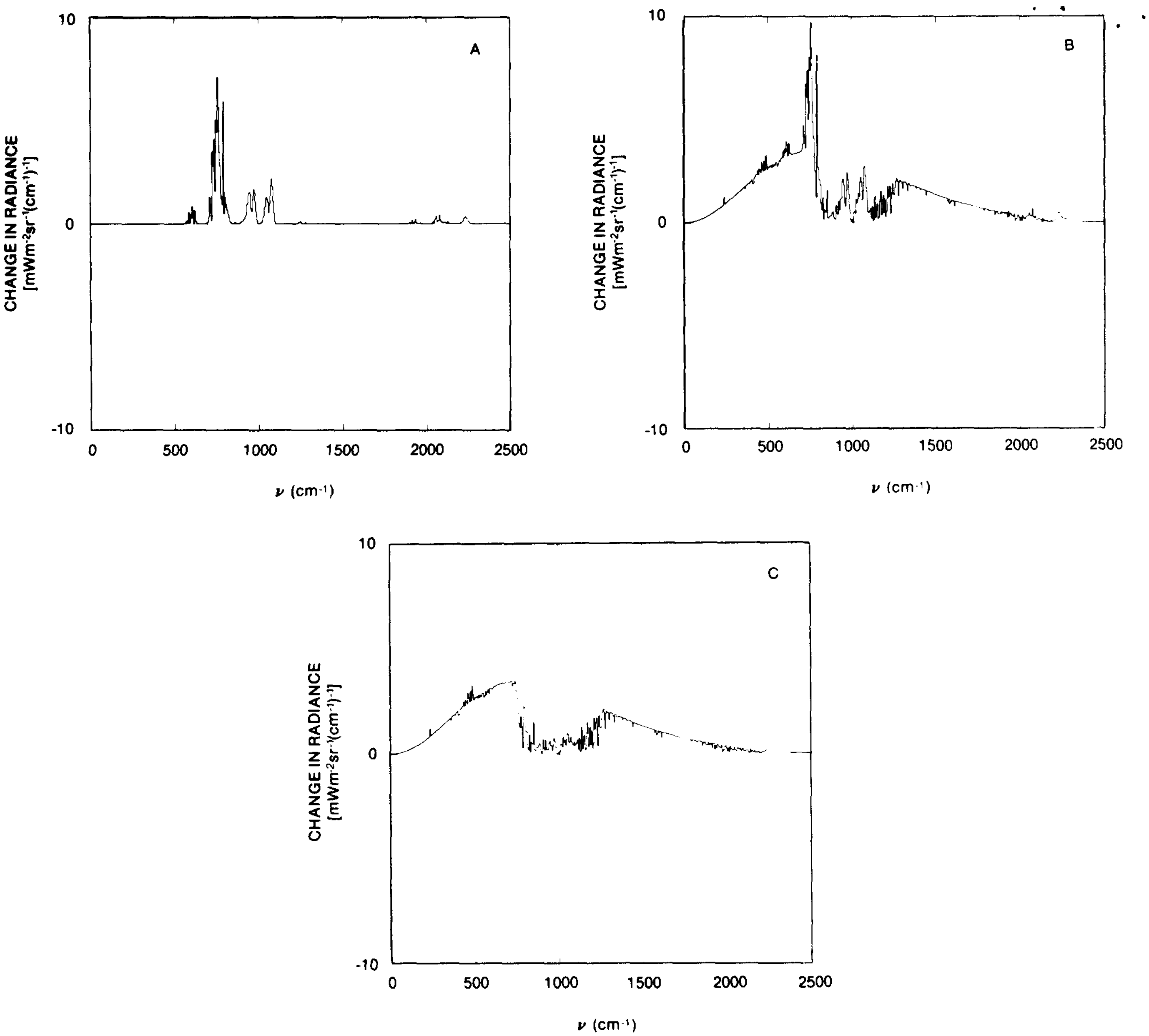

Figure 2.7. Change in downward radiance at the surface resulting from (A) doubling of CO_2 amount only, case I; (B) doubling CO_2 and changing atmospheric temperatures, case II; and (C) case (II) minus case (I).

The change in downward radiance caused by a 10% increase in CO_2 is no larger than 1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$. This change is small compared with the changes in radiance which would be caused by the natural variability of temperature at these latitudes (see Section 2.5). Thus, it would be impossible to see this change in radiance. A 25% increase in CO_2 causes a change in radiance of approximately 2 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$ for the wave number region between 500 and 800 cm^{-1}. This signal may still be difficult to measure above the variations caused by temperature variability. Increases in CO_2 of 50 and 100% produce changes in radiance that are large [>2 mW m^{-2} $sr^{-1}(cm^{-1})^{-1}$] and therefore would be easier to detect. The changes in radiance for all of these cases for a 50 and 100% increase of CO_2 are fairly large [>5 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$].

CO_2 is not the only greenhouse gas that has the potential to alter the climate (World Meteorological Organization 1982). There are a large number of trace gases whose concentrations are increasing

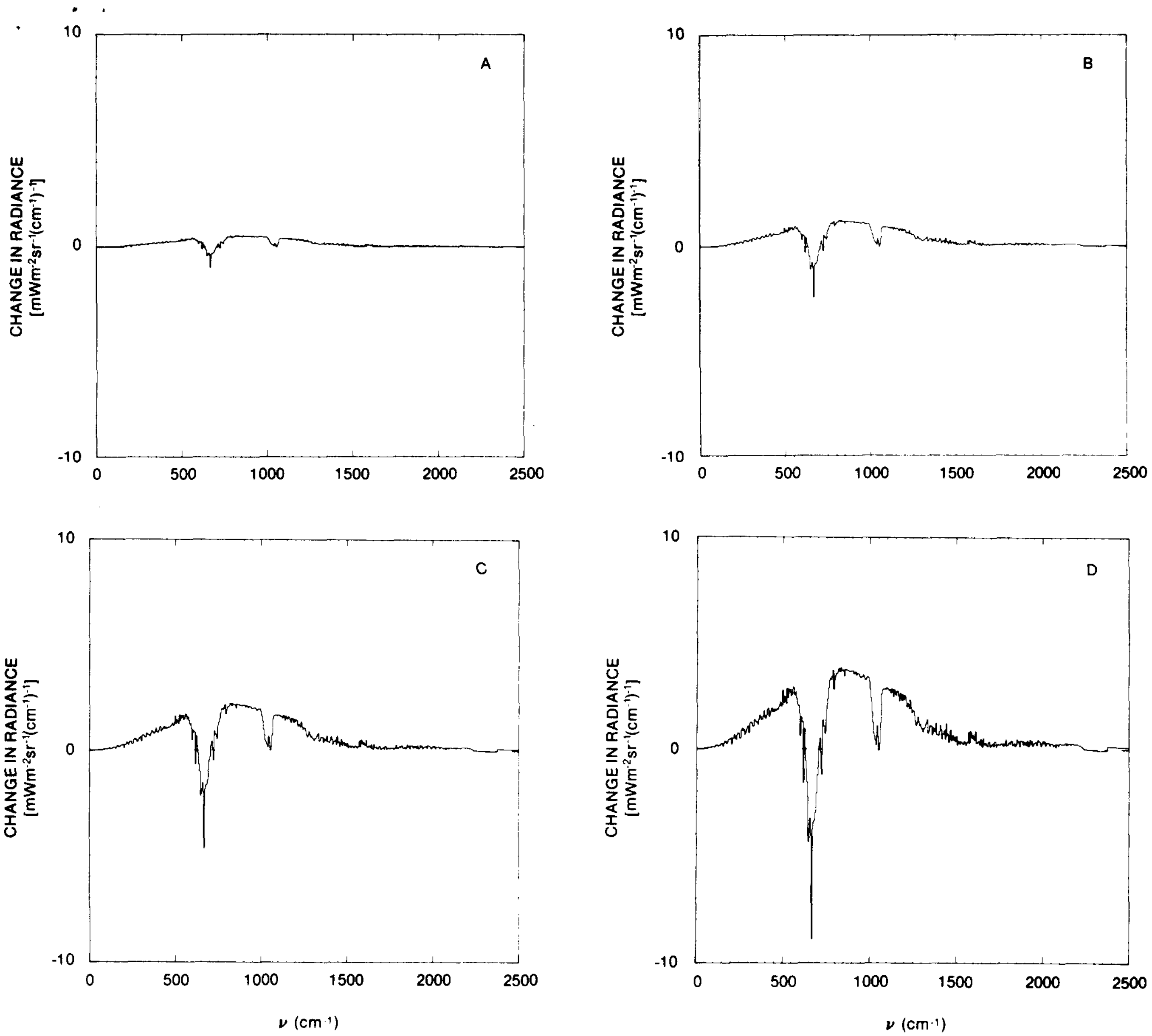

Figure 2.8. Change in outgoing radiance for the tropics profile for case III for (A) 10, (B) 25, (C) 50, and (D) 100% increase in CO_2 amount.

at a rapid rate, even greater than the rate of increase of CO_2. Many of these gases possess strong absorption bands in the longwave spectral region. Unfortunately, detailed narrow band spectral parameters are not available for many of these trace gases. There are data for CH_4 and N_2O, which are included in the calculation. Detailed data also exist for one band of CF_2Cl_2 (CFC-12) (Goldman et al. 1976a) and one band of $CFCl_3$ (CFC-11) (Goldman et al. 1976b). Changes in the concentrations of any of these trace constituents would lead to unique changes in the spectral distribution of radiance. With the help of a detailed radiation model, the gases that were contributing to the changes could be established. For example, decreased stratospheric temperatures could also be caused by decreased O_3 owing to increased CFC-11 and CFC-12. However, these changes would result in different spectral distributions of radiance change. The change in radiance for case I that is due to a doubling of CO_2,

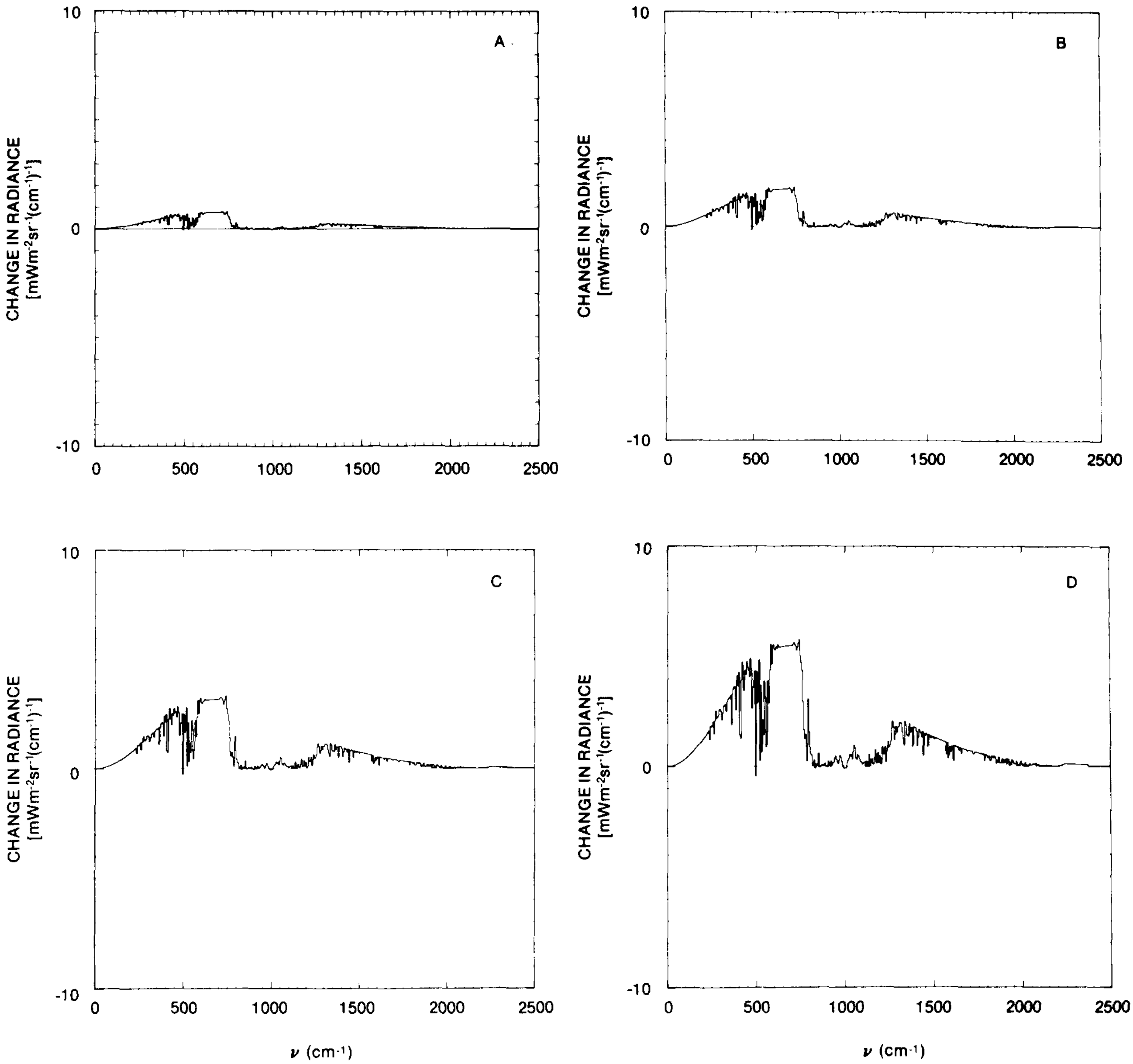

Figure 2.9. Change in downward radiance at the surface for the subarctic winter profile for case III for (A) 10, (B) 25, (C) 50, and (D) 100% increase in CO_2 amount.

CH_4, and N_2O, along with a stratospheric O_3 reduction and tropospheric O_3 increase, has been calculated (Wuebbles 1983). The resultant changes in outgoing radiance and downward radiance for a midlatitude summer profile are shown in Figure 2.10A-B. These results should be compared with Figures 2.6A and 2.7A, respectively. The spectral distribution is significantly different when changes in the concentrations of trace gases are included. The bands of CFC-11 and CFC-12 would also be apparent in these figures if their absorption properties were included in the model.

2.5 SOURCES OF NOISE

Detecting changes in radiance similar to those calculated in Section 2.4 is hindered by inherent random variation in measured radiances. The ultimate problem in detecting climate changes rests in detecting change due to a particular cause (e.g., CO_2, CH_4, O_3) above these sources of noise (Klein 1982).

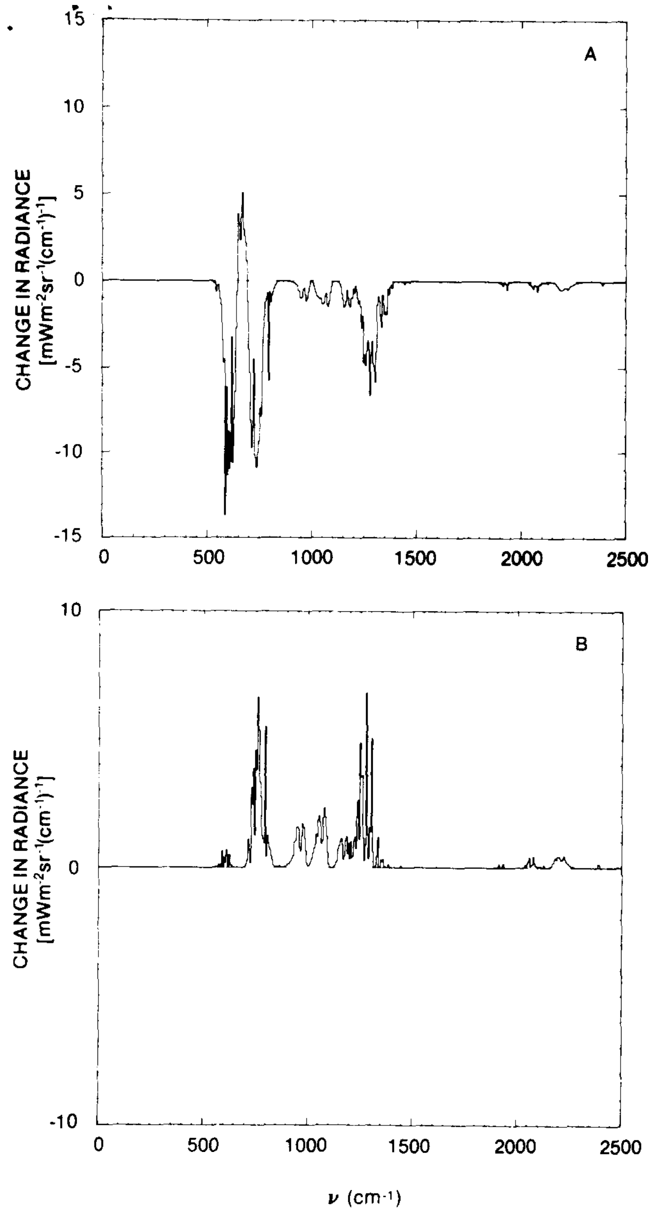

Figure 2.10. Change for the midlatitude summer profile due to a doubling of CO_2, increases in CH_4 and N_2O concentrations, and a change in the O_3 profile for: (A) outgoing radiance and (B) downward radiance at the surface.

These sources of noise can arise from either the instruments that are used to measure the signal (e.g., the change in radiance) or from the natural variability of the Earth's atmosphere.

Noise arising from the instruments used to measure radiances comes from a number of sources. For satellite radiometers, Houghton (1977) lists the most important of these factors as variations caused by temperature fluctuations, detector response, and electronic gain. Barnett (1980) discusses the importance of these factors to satellite measurements. Measurement of radiances from satellites has received much more attention in the last 10 years than have ground-based methods. There is more information regarding the noise characteristics of these instruments. Houghton (1977) and Barnett et al. (1975) discuss the magnitude of noise expected from one instrument, the selective chopper radiometer (SCR). They determined the magnitude of the noise by selecting quiescent areas in the atmosphere from which they had radiosonde data to determine the natural variability. By comparing the variations in their measured radiances with those due to natural variability, they concluded that the noise of the SCR was less than 0.1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$ for most of the radiometer's channels. This is less than most of the signals calculated in the previous section. Thus, at least for the SCR, instrument noise places little limitation on detecting the radiative signal.

Another instrument that would be invaluable for monitoring the radiative signal is the Michelson interferometer. This instrument has been discussed by Hanel et al. (1971). Their interferometer has a spectral resolution of 2.8 cm^{-1}, which is higher than the resolution shown in Figure 2.1. The interferometer is able to look at specific spectral regions to detect changes in the distribution of radiance. The noise equivalent radiance for this instrument was calculated to be approximately 1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$, which is larger than that for the SCR but still would not limit detection of changes in radiance.

One example of a ground-based monitoring technique can be found in the work of Oetjen et al. (1960) and Bell et al. (1960). These investigators used two spectrographs to measure the downward clear and cloudy sky radiances at various geographical locations. The resolution of their instrument was 0.25 μm (or 25 cm^{-1} at 10 μm). Oetjen et al. stated that their measured radiances were accurate to within 1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$ at 10 μm. Thus, instrument noise does not place severe limitations on detecting a radiative signal.

The second source of random variation comes from the natural variability of the atmosphere. This variability arises from the small scale motions present at all times in the atmosphere. The magnitude of this variability is dependent on the geographical location, altitude, and time of year. Most meteorological variables (e.g., temperature, specific humidity, and ozone) exhibit this inherent variability (Oort 1983). This implies that radiances, which depend on the temperature structure and distribution of absorber gas concentrations, will reflect changes in these the variables. The magnitude of the variability is defined with respect to space and time averaging. For example, the day-to-day standard deviation of temperature or humidity from a long-term mean could be considered. This type of deviation for a zonally averaged case is shown in Figure 2.11. The standard deviations could be larger or smaller at any given longitude. These figures point out that near the tropics there is small variability for temperature, but large variability for specific humidity, whereas the reverse situation prevails for higher latitudes. Surface measurements of radiances must be chosen with care to minimize the effects of this variability.

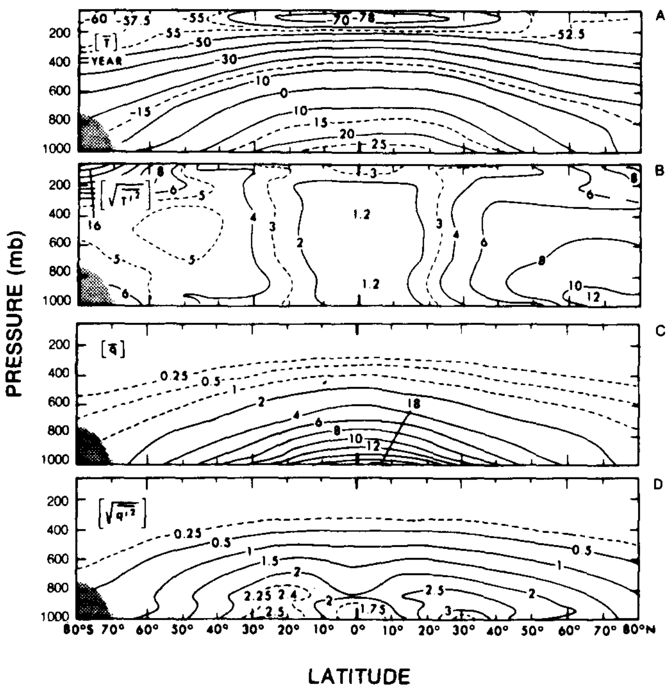

Figure 2.11. (A) Zonal mean temperature (°C), (B) day-to-day standard deviation of temperature (°C), (C) zonal mean specific humidity in g kg^{-1}, and (D) day-to-day standard deviation of specific humidity in g kg^{-1} from Oort and Peixoto (1983). Reprinted by permission of Academic Press.)

The variability of the radiances of the middle atmosphere for an altitude of approximately 45 km and three latitudes has been given by Barnett (1980) and suggests that the best location to look for a change in the radiances is in low latitude regions where the standard deviation of the radiance is nearly constant at 1 mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$.

The task of detecting a signal, in this case a change in radiance, amidst the natural variability of the atmosphere must rely on the application of a number of statistical techniques. These techniques have been reviewed by Klein (1982) for problems relevant to detecting climatic change due to increased CO_2 concentrations.

2.6 FUTURE WORK IN DETECTING THE RADIATIVE SIGNAL

There are three specific areas where future research in detecting the radiative signal should be directed. First, more spectroscopic parameters are needed for the trace gases in order to assess their effect on the spectral distribution of radiances. Currently, there is little narrow spectral interval data on the chlorofluorocarbons. These species are effective greenhouse gases (Ramanathan 1975), and it is important to be able to calculate changes in radiance due to their increase. Also, an effort must be made to validate existing radiative transfer models against laboratory and field measurements. To date, very few measurements relevant to atmospheric situations exist which could be used to test the radiative transfer calculations used to study the radiative balance of the Earth's atmosphere.

Second, data analysis techniques are needed to search for changes in radiances. Data for many spectral regions have been collected over the past decade. Continued collection and analysis of these spectral regions are urged. The revival of interest in ground based measurements should also be investigated.

Finally, an intense effort must be initiated to establish the absolute accuracy of current satellite measurements. Thus, the present instruments (HIRS, AVHRR, etc.) must be checked for absolute calibration. This issue is fundamental for locating changes in the radiative signal (due to increases in gas concentration, temperature change, or most likely, both). The chances of identifying a trend in radiances would be greatly enhanced if absolute calibration could be achieved. This goal could be achieved if a high resolution interferometer could be flown on the space shuttle. This instrument could then be used to look at the same field of view as a particular satellite instrument. Comparison of these two measurements could be used to check the accuracy of the satellite instrument. The high resolution instrument, either on a satellite or on board the space shuttle, would also be able to monitor virtually the entire longwave spectral region with very high resolution (<2 cm^{-1}).

2.7 SUMMARY

This chapter has addressed the problem of how the distribution of longwave radiation changes when CO_2 or trace gases increase in the Earth's atmosphere. The major atmospheric parameters that determine the distribution of longwave radiation were reviewed. The major determinants of the spectral distribution of longwave radiation were shown to be the temperature distribution, gas composition, and gas concentration. The importance of looking into different spectral intervals was also discussed. Changes in the distribution of longwave radiation were calculated from a high spectral resolution radiation model. Changes in radiance at the top of the atmosphere and at the Earth's surface were evaluated for fractional increases in CO_2 amount. Changes in radiance for increases in CO_2 greater than 50% were sufficiently large to be detectable by present day instrumentation. The various sources of noise were considered. It was argued that the greatest noise source arises from the natural variability in atmospheric temperature and water vapor amount. This natural variability constrains detection to particular regions of the atmosphere. The need for more spectroscopic data for the trace gases was also stressed. The need for a long-term monitoring program employing a high resolution interferometer was also discussed.

ACKNOWLEDGMENTS

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3. DATA BASES FOR ISOLATING THE EFFECTS OF THE INCREASING CARBON DIOXIDE CONCENTRATION

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CONTENTS

3.1 INTRODUCTION

The debate over the climatic significance of anthropogenically produced carbon dioxide (CO_2) began with Callendar's (1938, 1949) studies of worldwide temperature increases in the 1930s and 1940s. Empirical studies of long-term instrumental climatic data have continued to play an important role in attempts to identify causative factors in climatic variations. Instrumentally recorded data have been used in five major types of studies:

1. To produce spatially averaged time series of temperature for large areas (e.g., Northern Hemisphere averages, which are generally derived for continental areas only, or zonal averages for latitudinal strips around a hemisphere, again generally continent based) (e.g. Callendar 1961; Mitchell 1963; Vinnikov and Groisman 1981; Yamamoto 1980; Jones et al. 1982);
2. To identify modes of climatic anomaly, often using principal components analysis of temperature or precipitation data, and to relate these to changes in circulation pattern and determine if changes in anomaly patterns have occurred through time (van Loon and Williams 1976; Gray 1981; Tabony 1981; Jones and Kelly 1983);
3. To verify computer simulations of hemispheric or zonal temperature trends in relation to various forcing factors including CO_2 increase, solar output variations, volcanic aerosol loading of the atmosphere (Reitan 1974; Schneider and Mass 1975; Robock 1979; Hansen et al. 1981; Gilliland and Schneider 1984).
4. To examine empirically the effects of forcing factors (particularly volcanic aerosols) on the instrumental record of climate (Miles and Gildersleeves 1978; Yamamoto and Hoshiai 1980; Kelly and Sear 1984);
5. To derive scenarios of the possible climatic effects of CO_2-induced warming by identifying warm-year or warm-season anomalies in the long-term instrumental data set and using these as analogs for future climate (Wigley et al. 1980; Williams 1980; Pittock and Salinger 1982; Jäger and Kellogg 1984).

In almost all of these types of studies, the focus has been on mean monthly temperature data from Northern Hemisphere continental stations. Only recently have efforts been made to incorporate long-term air temperature data from over the oceans to derive more representative hemispheric averages (Folland et al. 1984; see also Chapter 4 of this volume). Here, we discuss briefly the history of instrumentally recorded temperature data and review some of the limitations of the data and the methods used in deriving large-scale averages. This is followed by a brief discussion of long-term precipitation and pressure data sets and the limitations of these data sets. A final section discusses data on volcanic aerosol loading of the atmosphere and solar irradiance variations, both of which may have played an important role in recent temperature variations.

Chapter 4 describes analyses of the surface and free air temperature data bases. Later chapters in this volume extend the scope of the search for the CO_2-induced signal to include other climatic variables, including changes in the ocean (Chapter 5), changes in snow and ice extent and other cryospheric variables (Chapter 6), and changes in precipitation (Chapter 7). The data bases needed for these analyses are described in those chapters.

3.2 TEMPERATURE DATA

3.2.1 History of Worldwide Instrumentation Network

Although various experiments to record temperature had been made earlier, the first reliable thermometers were not developed until the mid-18th century (Gerland 1896). Only careful analyses have been able to extend records back further into the past, and these require a great deal of faith in the early instruments and their calibration scales (e.g., Manley (1974) compiled a central England temperature series commencing in 1659) (also see Dettwiller 1981; Schaake 1982). Some attempts to organize regional observations in a systematic way began in the late 18th century, the best-known example of this being the network coordinated from Mannheim, West Germany (the measurements from which were published as the *Mannheim Ephemerides* for 1781–1785 [Kington 1974]). Similarly, the Société Royale de Medicine de France coordinated another network during the same interval (1780–1790) (Kington 1970). However, there was no major expansion of temperature recording networks until a century later, following the Vienna Meteorological Congress of 1873. This meeting provided the impetus for the expansion of recording networks worldwide and for the international exchange of data that continues today under the auspices of the World Meteorological Organization (WMO). While national meteorological agencies were established in the late 19th century, observing networks expanded, instruments and instrument shelters were standardized, and uniform instructions to observers were issued. As a result, a reasonably comprehensive global network of temperature-recording stations emerged, many of which have continued in operation to the present. Unfortunately, standardization of measurements did not extend to a universally adopted observation time policy so that, even today, vastly different protocols are followed in deriving daily temperature averages. This problem is discussed in more detail below. Further discussion on the history of instrumentation has been presented by Lamb and Johnson (1966), von Rudloff (1967) and Middleton (1966, 1969).

3.2.2 Data Inhomogeneities

> A numerical series representing the variations of a climatological element is called homogeneous if the variations are caused only by variations of weather or climate (Conrad and Pollak 1962, p. 223).

Although observers may take readings with meticulous care, nonclimatic influences can easily affect the readings. Some factors, such as the type of instrument, its exposure, and the method of measurement, may be under the control of the observer; other factors, such as observation times and the station environment, may not. Figure 3.1 illustrates the relative magnitude of error introduced by these various factors according to Mitchell (1953). The following sections discuss the most important causes of inhomogeneity:

- Changes in instrumentation, exposure, and measurement technique;
- Changes in station location (both position and elevation);
- Changes in observation times and the methods used to calculate monthly averages;
- Changes in the environment of the station, particularly with reference to urbanization.

After this discussion, methods of data homogenization are considered.

3.2.3 Changes in Instrumentation, Exposure, and Measuring Techniques

For land-based temperature data, the effects of changes in instrumentation are slight (at least after the early 19th century). More important is the change in exposure of the instruments. Thermometers are now located in louvered screened enclosures, which are generally painted white. Earlier readings often were from shaded wall locations and need correction. The effects of changes in thermometer type and screen, however, are considered to be slight (Mitchell 1953).

Changes in the techniques for measuring sea surface temperatures (SSTs) over the last 100 years have significantly affected readings. From the early 19th century, an increasing number of ships took daily readings of SST, air temperature and sometimes pressure, wind speed and direction, and other weather observations (Fletcher 1984). Until about 1940, SST was measured using the so-called bucket method. The values were recorded in log books together with the ship's position. With the increasing size of ships, SSTs began to be measured with thermometers located in the ship's cooling water intakes. These injection temperatures are estimated to be 0.3 to 0.7°C higher than bucket temperatures because of the internal warmth of the ship and elimination of evaporative cooling effects (Saur 1963; James and Fox 1972; Tabata 1978; Folland et al. 1984; see also Chapter 5 of this volume).

Many observations in the world data bank are supposed to be either all bucket or all injection SST data, but the analysis of Barnett (1984) shows that many contain both types of observations. Barnett concluded that better estimates of the temperature conditions over the oceans may be gained using the air temperatures measured on ships. Although these may contain the effects of increasing ship size, specific location of the instruments on board the ship, and the color of the louvered screen, these errors are more likely to be reduced by averaging

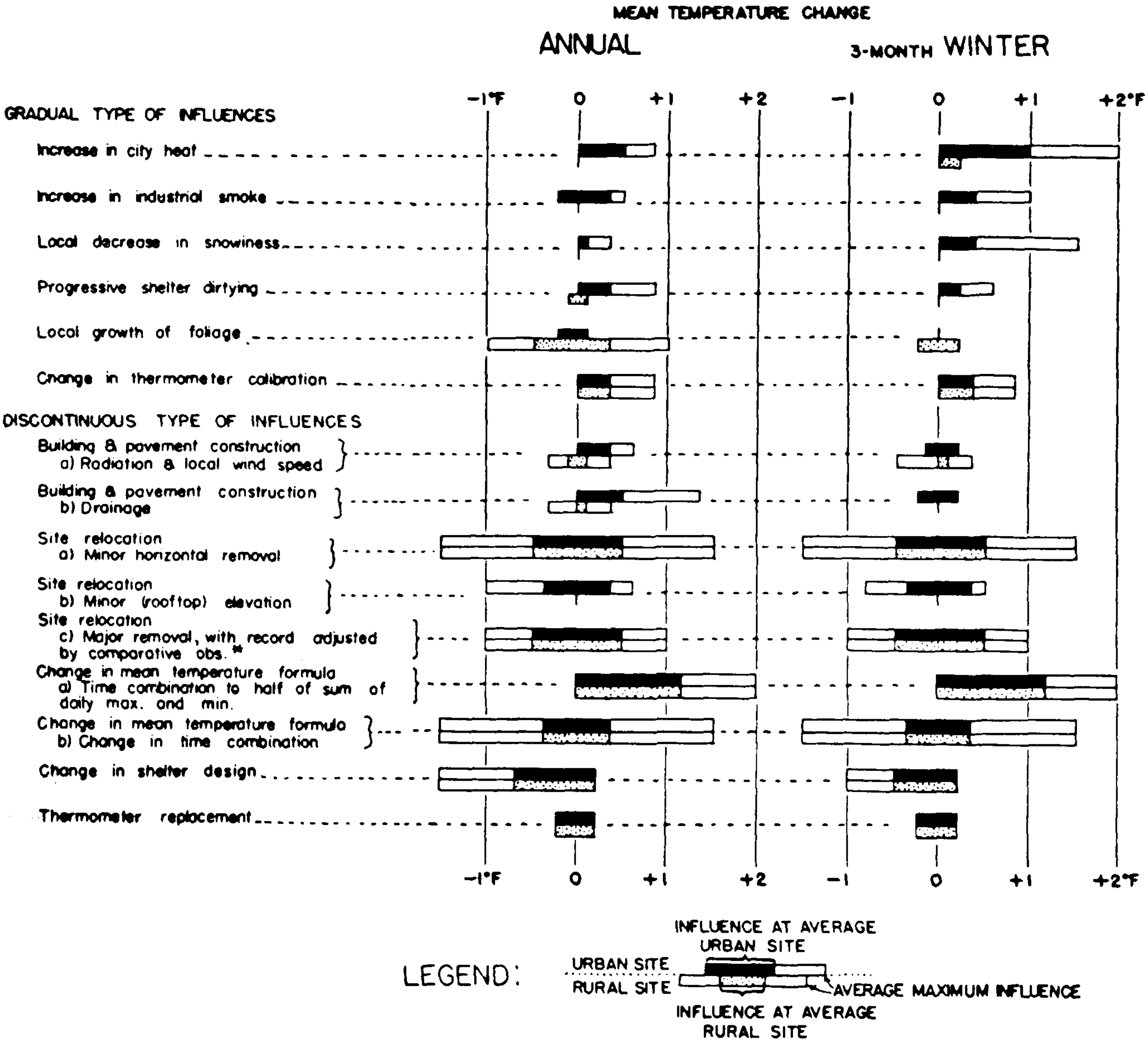

Figure 3.1. Estimated worldwide average and average extreme magnitudes of instrumental and local environmental influences on secular records of annual and winter mean temperatures over a 100-year period, based on observational and environmental changes typical of the past century. Source: Mitchell (1953).

than are systematic errors apparent in SST measurements (cf. Ramage 1984). One such systematic bias (toward higher temperatures in recent years) may be present as a result of improved weather forecasts, which have enabled ships to avoid bad weather, thereby increasing the number of observations taken under fair conditions.

3.2.4 Changes in Station Location

In any data bank, the specified station location and height probably will be that in current use. Earlier locations of the station, if any, can be found by consulting the meteorological archives or the written information available with a data set, such as the volumes of World Weather Records (WWR). Information concerning station moves is of primary importance to homogenization. The significance of moves can only be assessed station by station in the homogenization process (see Section 3.2.7).

3.2.5 Changes in Observation Times and Methods Used to Calculate Monthly Averages

In the late 19th and early 20th centuries, there was considerable discussion in the climatological literature concerning the best method of calculating true daily and, hence, monthly average temperatures (e.g., Ellis 1890; Donnell 1912; Hartzell 1919; Rumbaugh 1934). Many different schools of thought existed, and unfortunately, no one system prevailed in all regions.

The publication *Reseau Mondial,* which issued monthly mean data in yearly volumes from 1910 to 1934, used correction factors to reduce readings to means based on observations every hour (a 24-hour mean). In most cases, however, the corrections used are not given. In the United States, many sets of observation times were devised by the various supervising authorities and it was not until 1890 that a widespread policy of computing daily means—(maximum + minimum)/ 2—was introduced. However, despite this use in locally published material in the United States, data published in WWR for the United States are corrected to 24-hour means using factors derived by Bigelow (1909). In WWR these adjustments continued until 1940 or 1950. After this time, adjustments were not made and the measurements were simply calculated by (maximum + minimum)/2.

In the United States it is possible to unravel the effects of such adjustments and to correct the data sets accordingly. In other countries the situation is more chaotic with numerous changes to (maximum + minimum)/2 from fixed hours or vice versa. Examples for almost all countries have been described by Bradley et al. (1985). Only in 15 countries has a consistent methodology been followed since the 19th century. Of these, the most important (spatially) are those from Canada and India.

At present, there is no uniform system, although the (maximum + minimum)/2 method is used by over half of the members of WMO. The remainder calculate monthly means from observations at fixed hours or use complex formulae that employ station constants. Norway, for example, bases mean daily temperatures on the formula:

$$T = \frac{1}{3}(08 + 14 + 19) + C$$

where 08, 14, and 19 refer to the temperatures at these local times (in hours), and C is an empirically derived station constant which varies from month to month. For countries that use the maximum/minimum formula, the time of observation during the day can affect monthly mean values (Baker 1975; Blackburn 1983). In the United States the time of maximum/minimum observation at cooperative stations has changed from morning to evening, and this may result in a spurious cooling trend in temperature data (Schaal and Dale 1977).

To correct all readings to a common standard (e.g., [maximum + minimum]/2) would be extremely difficult. Such an effort is not necessary however if monthly temperature values and hemispheric estimates are calculated as *anomalies* from a selected reference period. Using this approach, one can assume that departures from differently derived mean values are comparable, provided that the observing system has remained internally consistent.

3.2.6 Changes in the Station Environment

The most important change that can affect a particular station is the growth of towns and cities around the site (the urbanization effect). Other changes at a site resulting from deforestation and irrigation are important only at single sites and generally not over entire regions.

Increasing urbanization around many stations may introduce a warm bias into computed regional temperature trends. Dronia (1967) calculated urban/rural differences by using 67 station pairs and by classifying the stations used by Mitchell (1963) into urban/rural categories based on population. His results showed an average urban warming trend of about 0.08°C per decade. Applying this correction to Mitchell's curve, he concluded that the mean hemispheric temperature in the 1960s was lower, not higher, than during the 1880s and 1890s. However, the spatial distribution of the stations used, the large distances of up to 2000 km separating the paired stations, and the paucity of data from the late 19th century make his conclusions questionable.

Other estimates of the magnitude of urban warming are made either on a regional basis or for a single city (e.g., Kukla et al. 1985). Most results suggest an urban effect from about 0.10 to 0.30°C per decade and refer to cities with populations greater than 100,000. However, because it is very difficult to separate any urban bias from other station inhomogeneities, these results should be viewed as a combined impact of both processes.

Recent studies of gridded temperature data from Northern Hemisphere land areas indicate that in studies of large-scale temperature trends, urbanization effects are relatively insignificant (Jones et al. 1985, 1986a). Indeed, those regions that are most important in contributing to hemispheric temperature trends are in high latitudes, north of 55°N, where urbanization effects are probably small (Jones and Kelly 1983).

3.2.7 Data Homogenization Techniques

Evaluation of the homogeneity of a data set can be accomplished in two stages. Data errors, generally from mistakes made in keypunching of handwritten records for computer entry, can be found by checking on outliers and flagging values that are at least three standard deviations from monthly means. (Alternatively, outliers could be flagged from 30-year filtered series.) These can be checked with original documents and with neighboring stations to decide whether the observation is correct. After such checking, the data can be considered "clean," although the data set may still not be homogeneous.

Many methods have been proposed for testing the homogeneity of station records relative to those at adjacent stations (Kohler 1949; Conrad and Pollak 1962; WMO 1966). Generally, tests of homogeneity involve the null hypothesis that a time series of differences between adjacent observation sets will exhibit the characteristics of a random series (Mitchell 1961; Bradley 1976; Craddock 1977; Jones 1983; Jones et al. 1986a).

These methods use climatological data (long-term series of monthly mean values) and make the assumption that non-climatic factors influencing one record will become apparent when the record is compared with a similar record (or group of records) from neighboring sites. If differences between observations are plotted against time, an abrupt nonlinearity in the plot may reveal where a station moved or a change in instrumentation or observation time at one of the stations occurred (Figure 3.2). When stations are analyzed in a small area it may become apparent which station or stations are in error. The station history information concerning the errant station can be checked for confirmations of a change. With this type of abrupt error, corrections can be made easily. However, if the graphs reveal that the changes at one station are gradual, increasing urbanization may be causing a gradual rise in monthly mean temperatures at one of the sites. With these types of subtle, time-dependent errors, corrections are difficult to apply, and part or all of the record should be removed from subsequent analyses.

This method of homogenization is only possible for monthly mean data. The corrections made on a monthly basis are not applicable to the daily data, such as for a time series of daily maximum and minimum air temperatures; such series are still nonhomogeneous. The method can be easily computerized, but the final deletion/correction/acceptance of a record can only be made by a careful subjective analysis.

This approach also requires that station records be from a dense network so that the stations are close together, although the appropriate distances depend on the region and on meteorological variables. For precipitation, which is generally more spatially variable, a denser network is required. For many parts of the world and many early periods, an adequate station density is not available. However, station networks since the 1880s are dense enough over the midlatitude regions of North America and Europe. In these regions homogeneity tests will be particularly important, because here the effects of urbanization may be large (Dronia 1967; Kukla et al. 1985).

Although further and more detailed analyses of temperature records are possible, such detail is considered impractical and unnecessary for the calculation of averages over large areas. The tests outlined above are considered to be adequate for this task. They are not, however, adequate for the analysis of individual station records.

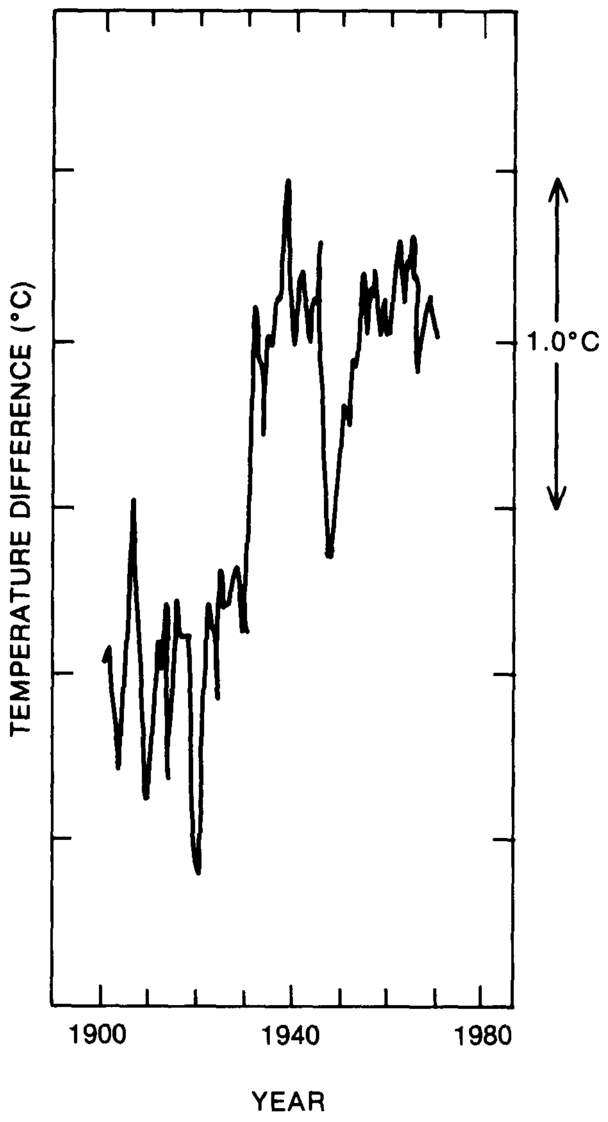

Figure 3.2. Station temperature difference time series: Reykjavik (64.0°N, 22.0°W) minus Vestmannaeyjar (63.4°N, 20.3°W), 1901–1970. The plot identifies Vestmannaeyjar as the errant station because a similar jump also occurs in 1931 when the station is compared with Stykkisholmur (65.0°N, 22.8°W). WWR station history data reveal that the station was moved in 1931.

In certain instances the analysis of long and isolated records from particular parts of the world is important. An example is the record from Sitka, Alaska (57.1°N, 135.0°W), where temperature, precipitation, and pressure were measured between 1832 and 1887. The unique position of this site (no records are available from the western half of North America until the 1850s in California and the 1870s in Washington State) make it ideal for further study. Detailed analyses of the record by Parker (1981, 1984) may allow the record to be used in hemispheric and regional analyses. The examination of station history information is very important for such single-site analyses.

Similar detailed studies of air temperature records have led to the production of many long temperature series; for example, in central England (Manley 1974), Paris (Dettwiller 1970), Berlin (Schaake 1982), and Philadelphia (Landsberg et al. 1968). There is the potential for similar research on long temperature records from other areas of the world.

3.2.8 Data Compilation

The best-known set of climatological data is generally referred to as World Weather Records (WWR). The project to produce this compilation initially was led by H. H. Clayton and funded by the Smithsonian Institution, Washington, D.C. Data were published by the Smithsonian Institution in 1927, 1934, and 1947, and updates containing 10 additional years of data were published by the U.S. Weather Bureau (various volumes from 1959–1982). These records have been recently digitized, and copies can be obtained from the National Center for Atmospheric Research (NCAR) (Jenne 1975). The individual station records that make up this data set are not homogeneous and contain many basic errors, but this data set has been used widely to study temperature variations over the last 100 years (Yamamoto and Hoshiai 1979; Hansen et al. 1981; Jones et al. 1982).

The selection of data for inclusion in WWR, however, was somewhat arbitrary. Before 1920 data were requested through personal contacts and not in any exhaustive fashion. This resulted in rather unrepresentative spatial coverage, with certain countries dominating the records in early years. For example, in the 1880s, half the temperature records came from three countries (the United States, U.S.S.R., and India). It is apparent that many long-term records were not included in these compilations.

Initial efforts to homogenize the records from before 1950 that comprise WWR were undertaken in cases for which the editors could gain sufficient information from the country concerned. The disproportionate information given by some countries can be seen in the station information section that

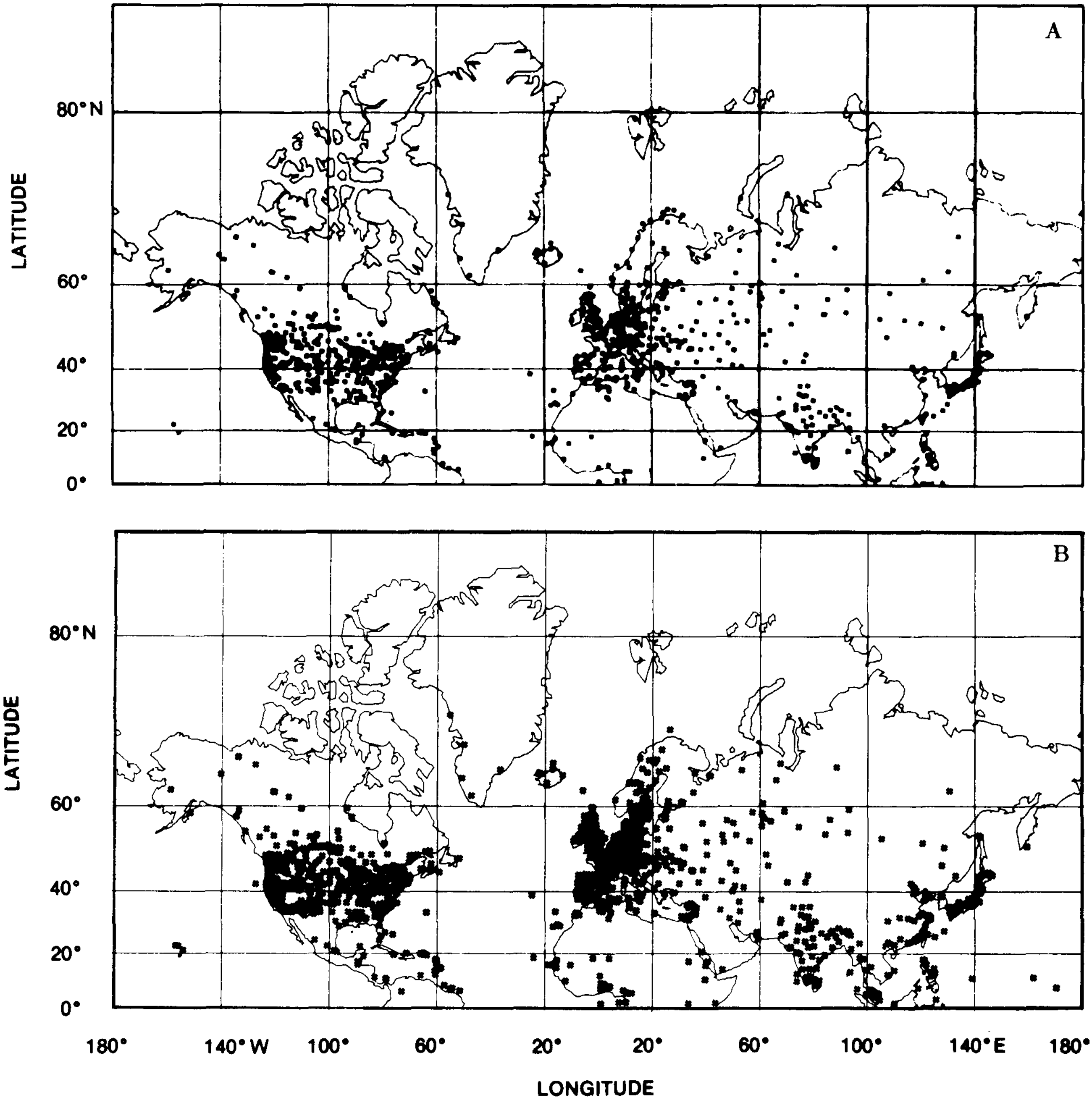

Figure 3.3. Locations of 19th century station records of 10 years or more in the Department of Energy Data Set. All stations shown have at least 10 years of data before 1900 and are generally continuous through at least 1960. (A) Temperature stations; (B) precipitation stations.

precedes each volume (see Bradley et al. [1985] for further details). After 1951, the data are simply those submitted through WMO by member countries. Any homogenization performed has been done in the individual country, and it must be assumed, in some cases, that no checks have been made.

The WWR data set formed the basis for the Department of Energy (DOE) temperature and precipitation data set produced by Bradley et al. (1985). Improvements were made for the period 1850 to 1900 by searching national meteorological archives for both published and unpublished materials. These data were key punched and added to the data set, extending existing records or forming new ones (Figure 3.3).

Many published compilations of early records were found in the course of this work. Two are particularly important, and while these were available to Clayton, neither was included in WWR. Between 1881 and 1890 the Central Physical Observatory at St. Petersburg (now Leningrad) published a set of volumes entitled *Repertorium für Meteorologie* (Wild 1887). These gave all climatological data (temperature, precipitation, and air pressure) collected for Russian territory and some adjacent areas, particularly China. The source extends instrumental data for the U.S.S.R. back to about the 1830s and significantly earlier in European parts of the U.S.S.R. Because no data for the U.S.S.R. are available in WWR before 1881, all analyses of hemispheric temperature trends have been constrained to begin in 1881 (surprisingly even those of the Russians themselves; Gruza and Ran'kova 1979; Vinnikov et al. 1980).

The most significant source of early temperature records for the world (pre-1860 to 1870) were published by Döve between 1838 and 1868. By correspondence, he built up a network of stations throughout the world. His network contained more than 1700 stations (although not all were operating at the same time); this number of stations has only been exceeded since 1951. However, most records were from Europe and eastern North America. He diligently gave his sources for every site (sometimes these are in very obscure journals), observation times, and units. No check was made on station homogeneity. This data set formed the basis of the first attempt to produce hemispheric temperature averages (Köppen 1873).

3.2.9 Methods of Producing Hemispheric Temperature Estimates

A world map of the locations of station temperature records shows that there are areas where station density is high (e.g., Europe, United States), areas where density is extremely low (Tibetan China, Antarctica, and the Amazon Basin), and ocean areas where there are no stations at all. Therefore, to produce hemispheric or regional estimates of mean temperatures, adjustments need to be made so that each area for which temperature records exist receives proper weighting. Another problem facing analysts is that temperature stations are at different altitudes and therefore may have quite different mean temperatures. Ideally, one would like to factor out the altitude effect, perhaps by extrapolating all data to a common level such as mean sea level (although topoclimatic factors would still be important). This, unfortunately, is extremely difficult to accomplish. Almost all analysts, therefore, subtract a reference period mean from station data. These anomaly values are comparable and contourable (see, for instance, *Die Grosswetterlagen Europas*, which publishes monthly a Northern Hemisphere map of temperature anomalies from the 1931 to 1960 normal reference period). The choice of reference period is especially important because a station can be used only if it has a record that is long enough to produce a reference period mean. Many workers (e.g., Yamamoto and Hoshiai 1980) select either the period 1881 to 1980 or 1881 to 1975. With a long reference period, one is immediately faced with deciding how many years need to have measurements before a reliable reference period mean can be produced. The choice of the almost 100-year period omits much station data and allows very few low-latitude (<30°N) data records to enter any subsequent analysis. To use most of the available data, it is easiest to select a reference period that has the best data coverage. Jones et al. (1982) selected the period 1946 to 1960, from which the maximum number of station records are available.

Many techniques have been used for forming hemispheric average temperature values.[1] Mitchell (1961, 1963) divided the world into 10° latitude zones, averaging all the zones (with area weighting). Hansen et al. (1981) divided the Northern Hemisphere into 40 boxes of equal area. Yearly temperature estimates were formed for each box, with the average of all available box values being the hemispheric estimate. Yamamoto and Hoshiai (1979) organized the data onto a regular but extremely coarse grid network, using the method of optimum interpolation, which was introduced into the field of meteorology by Gandin (1963). They then averaged the grid point values, with cosine latitude weights, using zero for all grid points with no data. This last step is

[1] Analysis of time series produced by various authors is discussed in Chapter 4 of this volume.

of dubious validity and causes a considerable damping of the variability. Jones et al. (1982) organized the data onto a 5° latitude by 10° longitude grid using an inverse distance-weighted, best-fit plane (for stations within 300 km) and then produced a hemispheric average using the weighted (by cosine latitude weighting) average of all the grid point values.

The last technique discussed here is that used by a team at the State Hydrological Institute, Leningrad, under Vinnikov et al. (1980). All available station data are plotted on maps and contoured by trained analysts. This is done routinely in Leningrad for forecasting purposes. The Soviet scientists have digitized the temperature anomaly charts by reading off interpolated values at 5° latitude by 10° longitude on the 1200 monthly charts. They have managed to use most of the available data (even short records) by changing the reference period mean three times over the period 1881 to 1980. Their method has been discussed at length by Jones et al. (1982) and Robock (1982).

Which method is best? Is there a best method? Undoubtedly the Russian method (Vinnikov et al. 1980) has certain advantages. The maps highlight data errors, but the contouring is subjective, and there may be differences among analysts in the extrapolation technique over areas with poor data. The Russians extrapolate their analyses over the ocean areas of the Northern Hemisphere, even though only isolated island data are available. This procedure is dubious and gives a false impression of the true data coverage. Tables and figures of the gridded area, as given by Jones et al. (1982), are more informative. The significance of the missing areas can be assessed, but this is fraught with danger, as highlighted by the analysis of Jones and Kelly (1983).

All the analyses discussed so far purport to be for the Northern Hemisphere (Jones et al. 1982). However, the area for which hard data are available is, at best, only 57% of the surface area of the Northern Hemisphere (Figure 3.4), dropping to less than 8% in the 1850s. The problem is worse when the Southern Hemisphere is considered; Hansen et al. (1981) produced a "global average," but when the number of stations in the Southern Hemisphere is considered, it is clearly inadequate. Global estimates cannot be produced until temperature trends over the oceans of the world are considered. Recently, estimates that include the ocean areas as well have been produced by Paltridge and Woodruff (1981), Folland and Kates (1984) and Folland et al. (1984), and by Barnett (1984); see also Chapter 4 of this volume. The results of these analyses show trends similar to those derived from the land series records, although there is still considerable controversy over the quality and homogeneity of ocean area air temperature data (Barnett 1984; Ramage 1984).

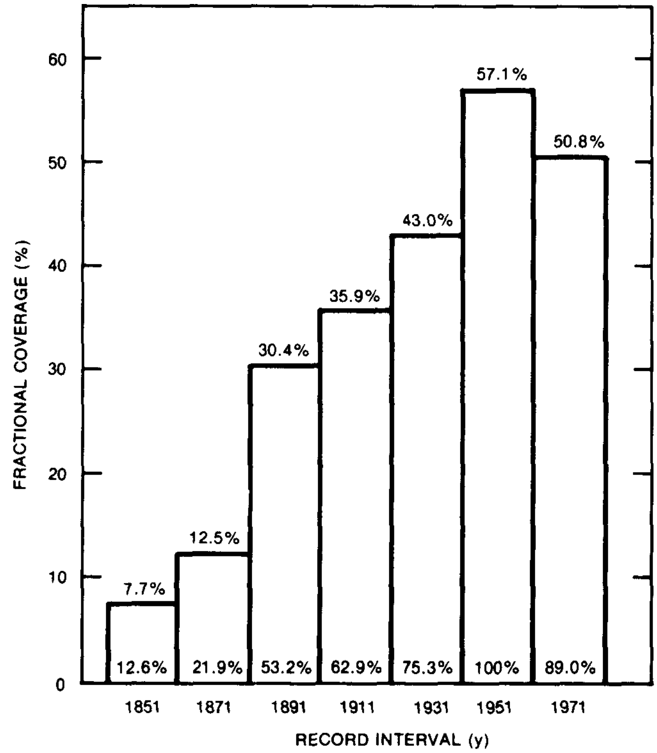

Figure 3.4. Percentage of the surface area of the Northern Hemisphere represented by gridded land-based temperature data, at 20-year intervals, in the Department of Energy Data Set. The numbers within the vertical bars indicate the relative coverage (in percent) compared to the maximum coverage achieved during the period centered on 1951.

Improvements are under way to rectify many of the shortcomings of recent analyses. Work by Bradley et al. (1985) has considerably increased the digitized land-based station data for the last century and increased the areas of the Northern Hemisphere where gridded estimates of air temperature can be made. Analyses of the global SST and air temperature data from ocean areas are being undertaken at the National Oceanic and Atmospheric Administration (NOAA)/University of Colorado Cooperative Institute for Research in Environmental

Sciences (Fletcher et al. 1984; Slutz et al. 1985; Woodruff 1985), at the U.K. Meteorological Office (Shearman 1983) and at the Scripps Institution of Oceanography (Barnett 1984). The combination of the new land estimates and ocean estimates from ship-based air temperatures will produce the best possible hemispheric estimates. The inclusion of Southern Hemisphere ocean area temperature data in time will improve global estimates (see Chapter 4 of this volume). An important development in this field is the new estimates of hemispheric and global surface temperatures that recently have been calculated from information from polar-orbiting satellites (Chahine et al. 1983). Such estimates may give an improved picture, but only after they have been checked and calibrated against dense arrays of surface observations. This approach will not provide any historical information, but it may enable the representativeness of limited area averages (such as those from continental land areas) to be assessed.

3.3 OTHER INSTRUMENTALLY RECORDED DATA BASES

Three other data bases are considered here. The precipitation and pressure data sets are comparable in length to that for temperature. However, upper air analyses, which give tropospheric and stratospheric measures of change, are only available since about 1950. All three data bases are discussed with respect to their errors, homogeneity, and potential significance in evaluating the climatic effect of increasing CO_2 concentrations.

3.3.1 Precipitation

Precipitation stations throughout the world far outnumber those for temperature, but the distribution is similar, with a plethora of records from Europe and North America. As with temperature, many precipitation site records are not homogeneous. There are problems because of changes in precipitation gauge size, gauge shielding, the height that the gauge is located above the ground, the growth of vegetation near the gauge, or the construction of buildings. All these factors can impair the performance of the gauge or alter the efficiency of the catch. Furthermore, various methods have been devised to deal with the difficulties of accurately measuring snowfall. This is a particularly acute problem at high latitudes and high elevations. Rodda (1969) has reviewed the possible errors and inhomogeneities that can result in precipitation records.

The total of these problems, coupled with the greater spatial variability of precipitation data, makes the problem of precipitation record homogenization significantly more difficult than that for temperature. Precipitation networks are not dense enough to allow the homogenization checks proposed in Section 3.2.6 to be easily undertaken. Analysis of precipitation records should not be undertaken on single-site records because of the high spatial variability of precipitation. Instead, analyses should be based on regional precipitation averages, formed by aggregation of many records in an area (Bradley et al. 1983; see also Chapter 7 of this volume).

The basic data source is again World Weather Records (WWR); improvements to the data set have been made by Bradley et al. (1985) by incorporating many new records, including 350 station records from the western United States (from Bradley et al. 1983), 180 homogenized station records for western Europe (Tabony 1980), many early records for the U.S.S.R. (Wild 1887), and the approximately 1000 records collected for Africa by Nicholson (1976).

The changes in precipitation patterns that are likely to accompany CO_2-induced climate changes are extremely important but have not been the focus of much research to date. Most CO_2 experiments suggest that global total precipitation will rise (National Research Council 1983), although precipitation may decrease in some areas. It is these regional-scale changes, especially on seasonal and shorter time scales, that are of most importance (Revelle and Waggoner 1983). Unfortunately, such resolution is not yet possible from available models, nor is it likely to be so in the near future. The only means of assessing future precipitation levels on these scales at present is the use of scenario studies (Lough et al. 1983; Webb and Wigley 1985).

3.3.2 Surface Pressure

Pressure data have an advantage over the temperature and precipitation data bases because they are routinely analyzed onto regular grid networks for weather forecasting purposes. Monthly mean data for the Northern Hemisphere extend back to 1873, but for the Southern Hemisphere routine analysis only began in the late 1940s.[2]

The Northern Hemisphere data, however, are not truly homogeneous. There are several different data sets (produced by the United Kingdom Meteorological Office [UKMO], the National Oceanic and Atmospheric Administration [NOAA], Deutscher Wetterdienst, and Soviet sources). Most data sets are the same for 1899 to 1939, but since 1945 they are generally different because of different interpolation algorithms. Analyses of the UKMO and NOAA data sets by Williams and van Loon (1976) and of the NOAA data set by Trenberth and Paolino (1980) have revealed that in many areas the gridded data do not agree with monthly mean station data. Further comparison of the different data sources has been undertaken by Parker (1980). Differences are most apparent over highland areas such as the Rockies and the Himalayas. Furthermore, it is suspected that early analysts overemphasized the "Arctic High," particularly over North America before 1920. Pressure values here are thought to be 2–4 mb (0.2–0.4 kPa) too high compared with mean values from recent periods (Namias 1958).

Jones et al. (1983, 1986b) derived a method to extend pressure data further back in time and to estimate values in data-poor regions. The method uses multiple regression techniques to transform between principal components of the gridded sea level pressure and principal components of station pressure data. The regression equations developed with data from the 20th century can then be applied to earlier periods, depending on the availability of station pressure data (back to the 1850s over Europe and North America). Extension back to even earlier times (~1780) is possible over Europe using a mixture of station pressure, temperature, and precipitation data instead of station pressure data alone. For Europe, continuous monthly mean pressure series have been produced for Paris, Edinburgh, Trondheim, and Milan back to the late 18th century (see Jones et al. [1986b] for details).

Station pressure data are also available from WWR.[3] The pressure data have all the problems of temperature data with station moves, observation time changes, correction to true means, and instrumentation problems. Further difficulties arise because all pressure readings should be corrected for temperature and for gravity at 45°N; the height of the instrument must also be known unless the observation is reported as mean sea level pressure. Jones et al. (1983) analyzed 32 pressure sites over Europe, and although WWR authors noted that corrections had been made in all cases, only five records were found to be error free. The most common problem was because of a station move, with subsequent data being corrected to a new height.

3.3.3 Upper Air Analyses: Heights, Temperature, Thickness

Since 1945, worldwide measurements of upper level pressures, winds, and temperatures have been made with radiosondes, rocketsondes, and, more recently, satellite-derived data. There are many different types of sonde manufactured, and their use differs from country to country. The data obtained from these instruments are extremely important meteorologically for weather forecasting purposes. As a result, the data come to climatologists secondhand after their operational use. Interpolated daily grid-point values of heights and temperatures at certain fixed pressures are averaged to form monthly mean values. The thickness of the atmosphere between any two levels is directly related to the average temperature of the layer. Analysis of free air temperature measurements is discussed in Chapter 4 of this volume.

Because they are affected by changes in instrumentation and changes in correction procedures, upper air data are far from being error free. Over data-sparse oceanic regions, the grid point analyses are dependent on the dynamic model that provides the "first guess" pressure field (Parker 1980). Spurious trends therefore may be introduced into grid point time series by changes in analysis procedures.

[2] The routine synoptic analysis of the Southern Hemisphere was started at the end of the 1940s in South Africa and was taken up in the early 1960s by the International Antarctic Analysis Center in Melbourne, Australia. The U.S. Weather Bureau prepared its own analysis in the 1960s and 1970s in support of the Apollo project.

[3] In numerous cases, station and sea level pressure data sets published in WWR appear to have been reversed.

Although changes related to sonde type may not be apparent from day to day, climatological analysis has unearthed some slight, but important, changes. For example, Parker (1980) has analyzed sonde data from Arctic regions and found that data from U.S.S.R. stations bordering Norway were abruptly different at the same time as similar changes were noted on the eastern border of the U.S.S.R. with Japan. It was concluded that the change in sonde manufacture had affected the U.S.S.R. sonde data. Such checks between stations in adjacent countries need to be conducted on a continuing basis to ensure the homogeneity of the data set in the future. Parker (1980) also compared the routinely analyzed charts of surface pressure and upper level analyses performed by national meteorological agencies in the U.S.S.R., the United States, West Germany, and the United Kingdom. Some important differences were revealed, but by using the station data it was possible to select the most correct chart for any particular case. However, there was no consistent pattern in selecting the optimum analysis.

3.4 VOLCANIC AEROSOLS

Several studies of the effects of a CO_2 buildup on the climate during the last 100 years have pointed to the importance of isolating the significance of volcanic aerosol loading of the upper atmosphere (Schneider and Mass 1975; Miles and Gildersleeves 1978; Hansen et al. 1981). Because this factor is possibly the most important variable affecting temperature variations on the time scale of 1 to 10^2 years, the following section discusses various data sets concerned with stratospheric aerosol loading. A similar review has been provided by MacCracken (1983).

Although the idea that volcanic eruptions might influence climate far from the eruption site was proposed over 200 years ago (Franklin 1789), serious attention was not directed to the question until the early 20th century (Abbott and Fowle 1913; Humphreys 1913; Arctowski 1915). The first attempt at a comprehensive catalog of explosive eruptions was that of Lamb (1970), who compiled historical data on known eruptions (tephra volume) and records of unusual post-sunset sky coloration as a proxy of eruptions that had injected aerosols into the stratosphere. Using the fairly comprehensive studies of Krakatoa (Symons 1888) as a reference standard (to which he assigned a dust veil index [DVI] of 1000), Lamb made estimates of the magnitude of other, less well-known eruptions, and even of unknown eruptions observed via the sky coloration records in historical literature. This approach recently has been applied to years earliler than 630 A.D. using European (principally Roman and Greek) literature sources (Stothers and Rampino 1983). Lamb's DVI often has been criticized because of the occasional use of temperature records to assess eruption magnitude (by reference to the extent of cooling after an eruption) (Robock 1981). However, this is only true for a few cases; according to Newhall and Self (1982), of Lamb's 250 DVI estimates, only 5% are solely based on temperature estimates. The majority (48%) are based on qualitative descriptions of eruptions or post-sunset sky coloration and another 42% are based either on Sapper's (1927) semiquantitative estimates of tephra volume or on later, more detailed calculations of tephra volume. A small percentage of values (5%) are based on actual radiation data. Lamb recognized the dangers of circular reasoning in using the DVI to assess the effect of explosive eruptions on temperature and was careful to isolate any data derived from climatological reasoning (Figure 3.5). He also pointed out explicitly that his estimates were probably only good to within one order of magnitude. This is discussed further by Kelly and Sear (1982).

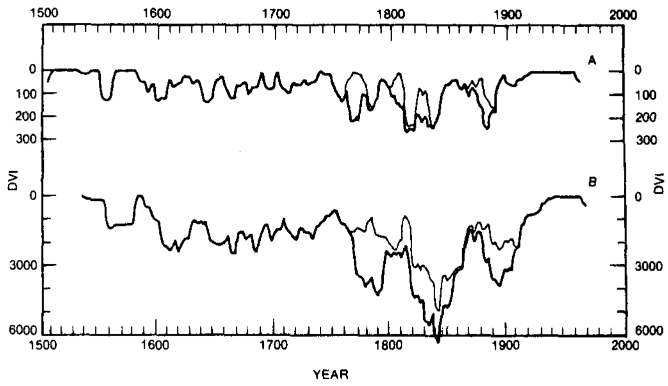

Figure 3.5. (A) Ten-year running means of Lamb's DVI plotted at the middle of each decade. (B) Twenty-five-year cumulative DVI. In (A) and (B) the finer lines indicate values obtained by ignoring cases of dust veils assessed solely on evidence of temperature anomalies. Source: Lamb (1970).

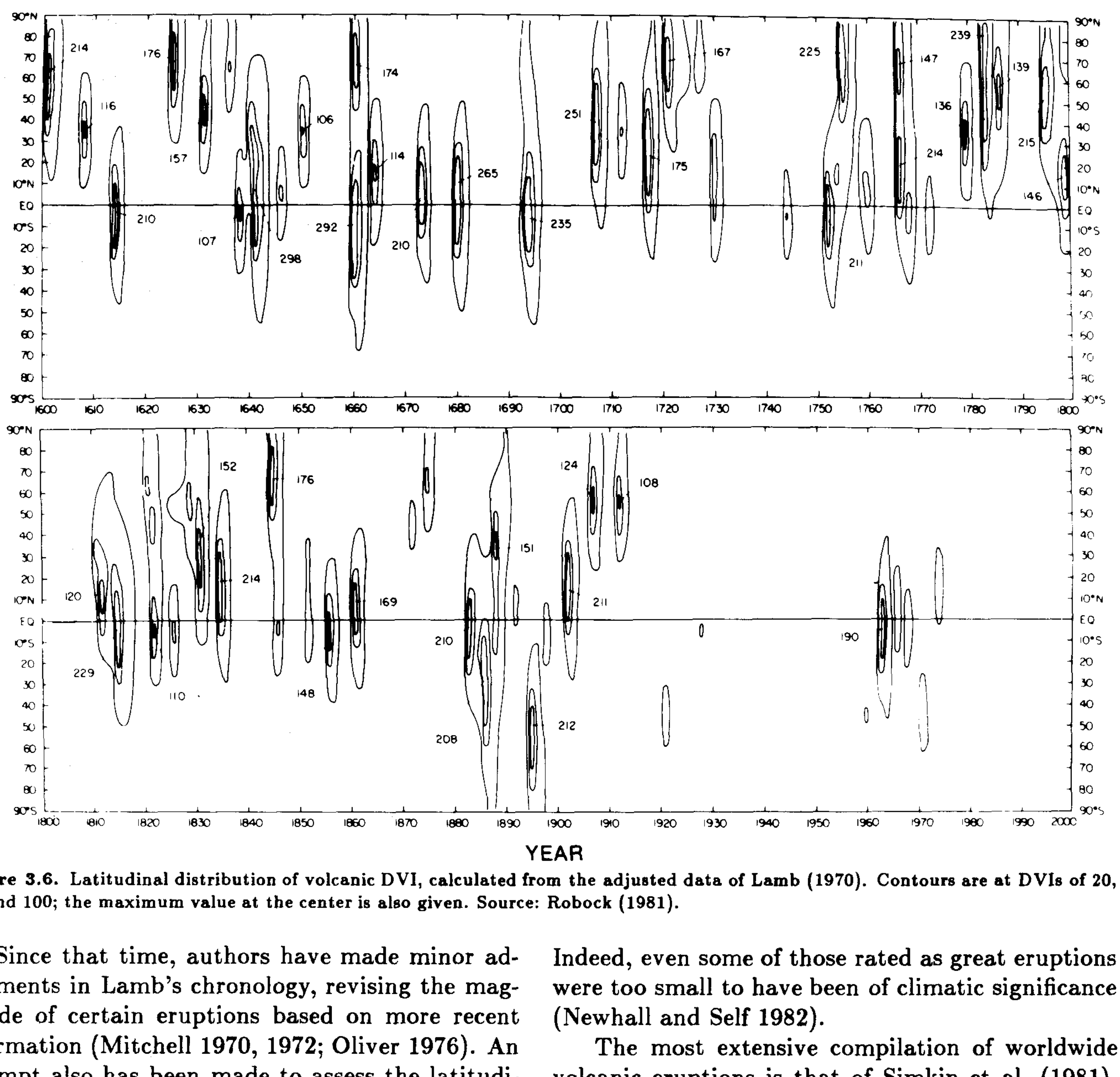

Figure 3.6. Latitudinal distribution of volcanic DVI, calculated from the adjusted data of Lamb (1970). Contours are at DVIs of 20, 60, and 100; the maximum value at the center is also given. Source: Robock (1981).

Since that time, authors have made minor adjustments in Lamb's chronology, revising the magnitude of certain eruptions based on more recent information (Mitchell 1970, 1972; Oliver 1976). An attempt also has been made to assess the latitudinal and temporal dimensions of aerosol clouds, recognizing that the location of eruptions is important in subsequent aerosol dispersion history, and that aerosols may remain dispersed for several years following an eruption (Robock 1981). Robock's model of DVI is shown in Figure 3.6.

A more detailed chronology of volcanic eruptions has been compiled by Hirschboeck (1980), with data spanning the interval from 2227 B.C. to A.D. 1969. Hirschboeck's chronology contains an order of magnitude more eruption events, but the vast majority are rated as minor or moderate, which are unlikely to have been of climatic significance. Indeed, even some of those rated as great eruptions were too small to have been of climatic significance (Newhall and Self 1982).

The most extensive compilation of worldwide volcanic eruptions is that of Simkin et al. (1981), who used geological criteria to rank eruptions on a scale of 0 (small) to 8 (massive) to form a volcanic explosivity index (VEI). Although this provides an objective chronology of eruption magnitude, the ratings were not designed to indicate potential climatic significance. The 1980 eruption of Mt. St. Helens, for example, was given a rating of 5, whereas the 1963 eruption of Mt. Agung, known to be of far greater climatic significance, was rated only as a 4. It would be useful to develop an index of climatically important explosive eruptions that would take into account eruption magnitude (volume and size of material, and injection height) as well as tephra composition, particularly sulfate content (Devine et al. 1984; Rampino and Self 1984). Other factors such as season of eruption, latitude, and vent elevation are also of significance.

All these factors notwithstanding, analysis of the VEI chronology provides some insight into the major characteristics of all eruption chronologies:

1. A total of 75% of major eruptions (VEI > 5) are recorded as occurring in the Northern Hemisphere (Table 3.1).

Table 3.1
Major Volcanic Eruptions (with VEI $\geq$ 5) Since 1600.

Volcano	Latitude	Date of Eruption	VEI
Awu	3.67°N	January 1641	5
Usu	45.5°N	August 1663	5
Tarumai	42.7°N	August 1667	5
Long Island	5.4°N	1700±100	6
Tarumai	42.7°N	August 1739	5
Katla	63.6°N	October 1755	5
Tambora	8.3°S	April 1815	7
Galunggung	7.3°S	October 1822	5[a]
Cosiguina	12.98°N	June 1835	5
Sheveluch	56.78°N	February 1854	5
Askja	65.0°N	March 1875	5
Krakatoa	6.1°S	August 1883	6
Tarawera	38.2°S	June 1886	5
Santa Maria	14.75°N	October 1902	6
Ksudach	51.8°N	March 1907	5
Novarupta (Katmai)	58.28°N	June 1911	6
Quizapu (Cerro Azul)	35.67°S	April 1932	5
Bezymianny	56.07°N	March 1956	5
St. Helens	46.2°N	May 1980	5

[a] Considered questionable.
Source: Simkin et al. (1981).

2. The record is incomplete and biased toward areas with good historical records. Prior to 1800, most reports of large, high-latitude eruptions were from Iceland and Japan, whereas in the late 19th and 20th centuries more high-latitude eruptions were reported from Alaska, the Aleutian Islands, and Kamchatka. This strongly suggests that many other eruptions occurred in these areas during historical times, but were never reported. Even today, eruptions in the Aleutian Islands may not be noticed for weeks to months. High latitude Southern Hemisphere eruptions are virtually unknown; only two major eruptions poleward of 30°S have been recorded, in 1886 and 1932.
3. Because of the probable bias in reporting, the maximum eruption density (eruptions per unit area) for eruptions with a VEI $\geq$ 4 occurs in latitude zone 60°–65°N (1.4 eruptions per 10^6 km^2 for the period 1500–1980). In view of other potentially unrecorded eruptions in this zone, this in fact may be true, but only further volcanological studies will be able to confirm it.
4. Although reports of eruptions have increased toward the present, it is, nevertheless, clear that the interval 1923–1950 was relatively free of major eruptions. The number of large eruptions (VEI $\geq$ 4) has doubled since the early 1950s, with an average of two major eruptions every 3 years since then (Simkin et al. 1981).

Because of the limited historical and volcanological data available and the inherent difficulties of assessing eruption magnitude ex post facto, the value of a precise proxy for volcanic eruptions is clear. Recently, the potential for using ice core electrolytic conductivity data as a proxy of explosive sulfate-rich eruptions has been demonstrated. Gases such as hydrogen sulfide and sulfur dioxide injected into the stratosphere are rapidly dispersed around the hemisphere. They are photochemically oxidized en route and combine with water molecules to form sulfuric acid, which is eventually washed out in precipitation, thereby raising the conductivity of snow (and eventually ice layers) in polar ice caps. Using Lamb's dust veil index (DVI) as a check, Hammer (1977) and Hammer et al. (1978, 1980) have demonstrated that highly acidic layers, significantly above background acidity values, reveal a record that closely matches eruptions of known ages. However, the ice core record is biased toward high-latitude eruptions because the washout of acidic precipitation is greater near the source. For example, in the Greenland Crête ice core, the highest acidity levels in the last 2000 years resulted from the eruptions of Lakagigar, Iceland, in 1783 and Eldgja, Iceland, in 934 A.D. $\pm$ 2. However, the most explosive eruption in the last millennium (and also perhaps the eruption of greatest climatic significance) was that of Tambora in 1815, a low-latitude event (8°S) rated 7 in the VEI (Stothers 1984). Perhaps a Southern Hemisphere ice core acidity record will provide a better perspective of the magnitude of these equatorial eruptions, even though they appear to have been of climatic significance in both hemispheres (Dunwiddie and LaMarche 1980; LaMarche and Hirschboeck 1984). Perhaps, the greatest value of ice core acidity records is the prospect of constructing a unique chronology of explosive eruptions back into prehistoric times.

The principal value of all these chronologies is as an estimate of the potential climatic effect of eruptions through their effect of reducing the solar radiation that reaches the lower atmosphere. A more direct index of volcanic aerosol loading may thus be long-term actinometric measurements of solar radiation (Pivovarova 1977). Although there are uncertainties in the quality of these records over time (World Meteorological Organization 1984a) they have been used as an index of stratospheric aerosol loading by Bryson and Goodman (1980), who have analyzed actinometric data for 42 Northern Hemisphere stations between 20° and 60°N on cloud-free days. These data were converted into mean aerosol optical depth as a measure of turbidity in the overlying atmospheric column (Figure 3.7). The early data in the record show relatively sharp year-to-year changes that appear to be correlated with the volcanic record. However, the later data seem to be much less related to volcanic eruptions. Because the data cannot be used to distinguish between tropospheric and stratospheric aerosol loading, the effect of urbanization and industrial and agricultural activity may be important factors in this time series. The data also include any effects of solar variability.

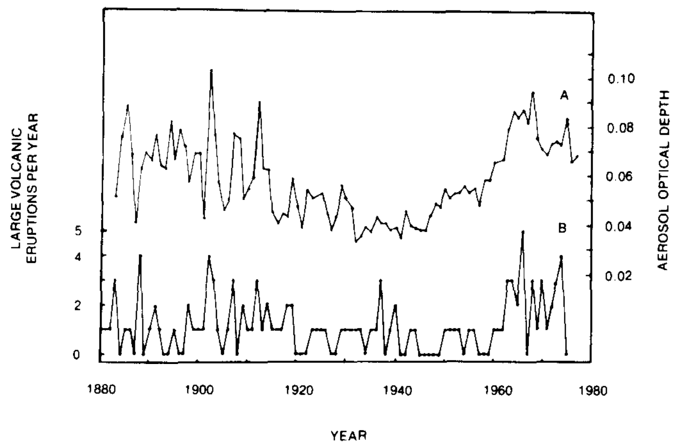

Figure 3.7. (A) Mean annual aerosol optical depth, based on 42 stations between 20° and 65°N, and (B) the number of Northern Hemisphere volcanic eruptions of large magnitude per year. Source: Bryson and Goodman (1980).

In view of the wide variety of volcanic aerosol indices, it is interesting to compare them for the period of overlap. Correlation coefficients are shown in Table 3.2.[4] Correlations between most series are statistically significant at a level <5 percent after accounting for autocorrelation. Pivovarova'sactinometric index is inversely correlated with the other indices because it is a direct measurement of solar radiation received at the surface rather than of aerosol loading. Also shown are correlations between the various indices and the Northern Hemisphere continental temperature data set of Jones et al. (1982). All series (except Pivovarova's) show statistically significant negative correlations with the temperature records. This analysis suggests that a better understanding of volcanic aerosol effects on climate will be needed to adequately assess CO_2-climate relationships in the future.

3.5 SOLAR IRRADIANCE

Until recently (i.e., since the availabiltiy of satellites) there have been no accurate measurements of solar irradiance. However, a number of proxy indicators have been used in solar climate studies: sunspots, atmospheric carbon-14 concentrations and solar diameter (see Wigley [1981] for a thorough discussion of these proxy indicators of solar irradiance and also Repin et al. [1980]; papers concerning solar-climate relationships are the focus of Centre National d'Etudes Spatiales [1978, 1980]).

Sunspots were rediscovered by Galileo in 1611, shortly after the invention of the telescope, even though the dark spots or blemishes had been seen on the face of the Sun with the naked eye for many centuries. Long records of sunspot activity can be gleaned from descriptions of the Sun and aurorae in Chinese records (Youji et al. 1983), but these are imprecise. Sunspots vary greatly in size, with some being as large as one-twentieth of the visible area of the Sun, and they occasionally occur in groups.

Since the 17th century, the number of sunspots has been counted by a regular network of observatories around the world. Wolf began a compilation of the sunspot counts on a monthly basis (known as the Zurich relative sunspot number), which has been continually updated. The sunspot numbers

[4] S. Clegg and T.M.L. Wigley; personal communication.

Table 3.2.
Correlations Between Different Series of Volcanic Forcing[a]

Source[b]	1	2	3	4	5	6	7	8
9	−0.36	−0.44	−0.54	−0.29	−0.48	−0.66	(0.31)	−0.47
8	0.28	0.37	0.44	0.30	−0.36	0.52	(−0.50)	
7	−0.48	−0.32	−0.46	(−0.15)	−0.47	−0.43		
6	(0.26)	0.43	0.63	−0.25	−0.56			
5	0.65	(0.16)	0.81	0.65				
4	0.61	0.30	0.63					
3	0.56	0.41						
2	0.23							
9	(−0.48)	(−0.67)	(−0.67)	(−0.35)	(−0.61)	(−0.75)	(0.33)	(−0.57)
8	(0.25)	(0.45)	(0.58)	(0.48)	(0.53)	(0.62)	(−0.40)	
7	(−0.51)	(−0.33)	−0.68	−0.51	(−0.72)	(−0.47)		
6	(0.44)	(0.63)	(0.72)	(0.49)	(0.60)			
5	0.70	(0.18)	(0.84)	0.66				
4	0.80	(0.33)	(0.54)					
3	(0.62)	(0.51)						
2	(0.39)							

[a] In the lower half of the table, the volcanic data have been subjected to a low pass (3 year) binomial filter. Values in parentheses are not significant at the 5% level after accounting for autocorrelation.

[b] Sources are as follows:

1. Simkin et al. (1981); VEI.
2. Hammer et al. (1980); ice core acidity.
3. Lamb (1970); DVI.
4. Mitchell (1970); based on Lamb's (1970) DVI.
5. Oliver (1976); based on data from Lamb (1970) and Mitchell (1970).
6. Bryson and Dittberner (1976); alternative DVI.
7. Pivovarova (1977); strength of direct solar beam (actinometric).
8. Bryson and Goodman (1980); mean annual aerosol optical depth (cf. Pivovarova 1977).
9. Northern Hemisphere temperatures from Jones et al. (1982).

vary in a periodic way with roughly 11 years between each sunspot maxima, although the cycle has had lengths ranging from 7.3 to 17.1 years over the last 300 years (see Sonett [1983] for a discussion of the sunspot index spectrum). The sunspot record extends back to 1749 with high reliability (Eddy 1976), but is less reliable before and extremely sketchy prior to 1700. Part of the reason for this sketchiness is due to the belief that there were no sunspots for certain periods during the 17th century. This period has come to be known as the Maunder Minimum (1654 to 1714). It is now generally believed that there were some sunspots, although the numbers were reduced compared with those counted after 1800. Earlier sunspot minimum periods are also believed to have occurred (e.g., the Spörer [1416 to 1534] and Wolf [around 1280 to 1350] minima).

The area of the Sun covered by sunspots has been measured since 1874. The umbral/penumbral ratio (the ratio of the size of the inner to the outer ring of the spot) has also been measured since that time because it was thought that it might be a more useful irradiance proxy than the sunspot number (Hoyt 1979). The area time series has been used by Eddy et al. (1982) to reconstruct values of the solar irradiance. Such a series may be a more reliable indicator of variations of solar irradiance received by the Earth than is the number of sunspots.

Another indicator of solar output variations is the amount of radioactive carbon-14 in the atmosphere. Such variations can be measured by comparing carbon-14 dates with accurate calendrical dates of tree rings. As carbon-14 production is influenced by solar flare activity and the strength of the solar wind (Stuiver and Quay 1980), it is a valid proxy measurement of solar activity, but its relationship to irradiance is uncertain. Major variations in atmospheric carbon-14 are associated with modulations in the amplitude of the 11- and 22-year solar sunspot cycle and may allow periods similar to the Maunder Minimum to be identified back many thousands of years. The variability of atmospheric carbon-14 provides firm evidence of solar variability on the century time scale, but as noted above, the link with irradiance has not yet been determined.

The relationships between variations in sunspot activity, sunspot area, solar flare activity, and solar irradiance are still a matter of debate (Newkirk 1983). The issues cannot be resolved until long and accurate time series of satellite-measured irradiance become available. Such direct measurements of solar irradiance have been made with the aid of satellites only since 1975. Earlier rocket-based or ground-based data are considered inadequate for estimating decadal time scale changes (Willson 1984). The results from the many satellite missions since 1975, in particular the Solar Maximum Mission, have been summarized by Smith et al. (1983) and Willson (1984). Correlations on daily to annual time scales between solar irradiance and sunspots have been demonstrated (Willson et al. 1981; Eddy et al. 1982; Smith et al. 1983). These data suggest that there may be irradiance changes of ± 0.05 percent associated with the sunspot cycle. The associated change in global mean temperature would almost certainly be less than ± 0.1°C if due account is taken of transient response effects (Hoffert and Flannery 1985).

Extrapolation of the satellite record to longer periods suggests that solar variability may vary by up to 0.2% over a 40-year period. Such variations may be more important for climatic studies, partly because of the amplitude of such changes and partly because the transient response damping would be less for lower frequency forcing. These variations appear to be in accord with astronomical evidence that the solar diameter varies on the decadal to century time scale. Parkinson et al. (1980) and Gilliland (1981) suggest that the diameter of the solar disk varies with an approximately 80-year cycle. There is also evidence of such a periodicity in the modulations of the 11- and 22-year sunspot cycle over the last 250 years. Although the quantitative link is still uncertain (to within two orders of magnitude according to Gilliland 1981), changes in solar diameter should be reflected by changes in solar irradiance. Smith et al. (1983) have suggested that some of the observed changes in irradiance from satellite data might be caused by such changes.

Global mean temperature changes of up to $\pm 0.2°C$ appear to be possible as a result of solar irradiance changes, consistent with recent satellite observations (Willson 1984). However, the shortness of the satellite record precludes any definitive statement at present. Changes of $\pm 0.2°C$ would be extremely difficult to detect in the observational record. Nevertheless, the magnitude is consistent with the empirical estimates of Gilliland (1982) and Gilliland and Schneider (1984), even though their results are not statistically significant. Almost all proposed solar-climate relationships have been found to be statistically unsound when analyzed critically (Pittock 1978, 1983), but the possibility of solar influences on the climate cannot be dismissed. The effects are almost certainly small on the interannual time scale, but they may still be significant on the 10 to 100 year time scale.

Determination of the significance of solar irradiance change on climate and the relationship between solar irradiance and the proxy measurements discussed above require a relatively long and continuing record of accurate measurements from satellites. The present Earth Radiation Budget satellites will probably not be operative after 1987–1988. After 1993, a new generation of satellites will become operational (World Meteorological Organization 1984b). There will therefore be a gap in the hitherto continuous record of broad-spectrum radiation measurements from satellites between 1988 and 1993 (World Meteorological Organization 1984b). To avoid such a gap, immediate action is required. The contingency measures suggested by the World Climate Research Programme (World Meteorological Organization 1984a) to bridge this gap may not provide the required accuracy.

3.6 SUMMARY AND RESEARCH RECOMMENDATIONS

Almost all data used in studies of long-term climatic variations have significant limitations arising from the fact that the data were never collected for this purpose. Table 3.3 summarizes the most important data sets that have been assembled and that are in machine-readable form. Users of these data sets who wish to isolate the effect of increasing CO_2 concentrations or to understand climatic variations of the past must carefully consider the data quality and spatial and temporal resolution of the data, as discussed above. Only with such a realistic appraisal of these data can meaningful conclusions be reached.

Compilation of a detailed land-based temperature and precipitation data set extending back to 1851 has recently been accomplished. It is unlikely that a significant improvement in the spatial coverage of this long-term set can be achieved. Studies of the homogeneity of the data sets, particularly of precipitation and pressure data, are needed, however. Both sea surface and marine air temperatures need to be carefully homogenized and merged with land-area data to provide a more comprehensive view of hemispheric and global temperature variations. The record of large explosive volcanic eruptions (i.e., those likely to be of climatic significance) is poor and improvements are needed to more fully understand the effects of these events on the long-term climatic record. In particular, the chemical characteristics of past volcanic emissions and their relationship to atmospheric optical depth through time are areas requiring further study. Satellite measurements of solar irradiance are also extremely

Table 3.3
Sources of Data Sets Useful for Detecting CO_2-Induced Climate Change.

Variable	Region	Source and Availability of Data Set	References and Remarks
Surface Air Temperature (land stations only)	Global	World Weather Records (WWR)—NCAR, Boulder, CO	Jenne (1975) and updated current NCAR documentation
Surface Air Temperature (land stations only)	Northern Hemisphere	U.S. Dept. of Energy Data Bank—Carbon Dioxide Information Center (CDIC), Oak Ridge, TN	Bradley et al. (1985). 19^{th} century records in WWR are a subset
Gridded Air Temperature (5° latitude by 10° longitude for land areas	Northern Hemisphere	U.S. Dept. of Energy Data Bank—CDIC, Oak Ridge, TN	Jones et al. (1986a). Gridden version of station data in Bradley et al. (1985) and WWR
Marine Data (including sea surface and ocean air temperature)	Global	Comprehensive Ocean-Atmosphere Data Set (COADS)—NCAR or CIRES, Boulder, CO	Slutz et al. (1985). Data set contains many other variables (e.g., windspeed, sea level pressure, cloudiness)
Marine Data (sea surface and ocean air temperature only)	Global	U.K. Meteorological Office	Folland et al. (1984)
Station *Precipitation* (land areas only)	Global	WWR-NCAR, Boulder, CO	Jenne (1975) and updated current NCAR documentation
Station Precipitation (land areas only)	Northern Hemisphere	U.S. Dept. of Energy Data Bank—CDIC, Oak Ridge, TN	Bradley et al. (1985) (19^{th} century records in WWR are a subset)
Sea Level Pressure (gridded)	Northern Hemisphere	NCAR (1899-), Boulder, CO	Jenne (1975). See remarks concerning homogeneity in Williams and van Loon (1976) and Trenberth and Paolino (1980)
Sea Level Pressure (gridded)	Northern Hemisphere	U.K. Meteorological Office (1873-)	See Williams and van Loon (1976)
Upper Air Analyses (various levels [1000-100 mb], heights, and temperatures)	Northern Hemisphere	NCAR, Boulder, CO	Jenne (1975). Most standard levels available although 500 and 700 mb are of longer duration (since ~1946) and are probably most reliable
Sea Level Pressure (gridded), Upper Air Analyses (various levels (1000-100 mb), and heights)	Southern Hemisphere	Australian Bureau of Meteorology, Melbourne Australia	Only available since 1972

important and it is critical that these observtions be continued, particularly over the next decade.

Improvements in large area and long-term data sets have been and continue to be made. Analysis of these data should provide new insights into the variability of past climate and the importance of various factors in producing the variability. This will be a major step towards isolating the effects of the increasing CO_2 concentration on climate.

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4. ANALYSIS OF THE TEMPERATURE RECORD

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CONTENTS

4.1 INTRODUCTION

This chapter reviews current knowledge of the atmospheric temperature record and summarizes attempts to explain past temperature variations. Because the effects of carbon dioxide (CO_2) are our main concern, the discussion is restricted to relatively large spatial scales and to medium (on the order of decades) and longer time scales.

Empirical detection of the effects of CO_2 on climate is a necessary element of the CO_2 research program because of inherent uncertainties in model predictions. These uncertainties become larger as the spatial scale diminishes from global mean values downward. Because of this, the emphasis of this chapter is on hemispheric-scale changes. Local and regional variations in temperature undoubtedly are important in assessing the impacts of climate and climate change on society, but it is large-scale averages that are most likely to provide detectable changes in climate because of changes in the atmospheric carbon dioxide concentration. The possibility of using regional-scale data is discussed elsewhere in this volume in reference to changes near snow and ice margins where feedback mechanisms may amplify the CO_2 signal (Chapter 6 of this volume), but such efforts are hampered by the large natural variability in these regions.

This chapter is also restricted to discussions of data that span periods of decades or more. Because detection requires a reliable definition of the background level of natural variability on time scales relevant to the possible effects of CO_2, long record lengths are required to establish these noise levels.

4.2 SURFACE AIR TEMPERATURE

4.2.1 Review of Past Work

We begin with a brief review of previous compilations and analyses of surface air temperature data. Other reviews have been given by Chen (1982), Jones et al. (1982), Wallén (1984), and Ellsaesser et al. (1986). The first attempt to estimate global mean surface air temperatures was made in 1873 by Köppen (1873, 1914) using data extending back to 1750 collected almost entirely by Döve (see Chapter 3 of this volume). This series was later updated by Humphreys (1940). Although the original data are of considerable value, these early results do not give reliable large area averages because of limited data coverage and possible inhomogeneities in the early data (see Chapter 3). Some years later Willett (1950) produced a "global mean" time series of 5-year nonoverlapping averages spanning the period 1845-1940. Willett's station network was relatively uniform, but too sparse for one to be confident of his estimates of changes in global mean temperature. Nevertheless, his results showed a considerable warming trend during the early 20th century that was later confirmed, using a denser data network, by Callendar (1961). Mitchell (1961, 1963) extended Willett's analysis beyond 1940, improved the method of area averaging, and found that the warming prior to 1940 had subsequently become a cooling trend (as suggested earlier by Kincer [1946]). Whether or not this cooling was common to both hemispheres was uncertain, because data coverage over the Southern Hemisphere was, and still is, sparse. However, as later discussions in this chapter will show, any Southern Hemisphere cooling was almost certainly less marked than that in the Northern Hemisphere. The Willett-Mitchell curve was later updated by Mitchell (1970), Reitan (1974), and Brinkmann (1976), with the latter work demonstrating a leveling-off of the post-1940 cooling, at least in the Northern Hemisphere.

In recent years, five groups have published time series of large-scale average surface air temperatures: a Soviet group, including Budyko, Vinnikov, Gruza, Borzenkova and others; a Japanese group, Yamamoto and co-workers; Jones, Wigley, Kelly, and others at the Climatic Research Unit in the United Kingdom; Hansen and collaborators in the United States; and Folland, Parker, and Kates of the United Kingdom Meteorological Office (UKMO). The first four groups have used only land-based data and data from a few fixed-position weather ships. The UKMO workers have analyzed marine data from the vast number of ships that report information to the global meteorological network (so-called ships of opportunity) and have examined both marine air temperatures (MAT) and sea surface temperatures (SST). An extensive discussion of SST data is given in Chapter 5. In addition to these five groups, valuable contributions have been made by a number of individuals including Chen (1982), who combined both land and marine data from the period 1949 to 1972, Barnett (1978), who also considered both land and marine data (1950–1976), Paltridge and Woodruff (1981), who used mainly SST data, and Barnett (1984), who has critically examined MAT and SST data in selected areas.

We will begin here by considering the land-based data and then discuss the MAT and SST data in light of the land results. For land data, the results of various researchers are generally highly correlated and show changes of the same magnitude. This is not surprising because most of the data sources used are common to all analyses. Differences arise from differences in spatial coverage and the precise number of stations employed, in the methods and extent of extrapolation over data-poor areas, in averaging procedures (see Table 4.1), and in the treatment of the raw data with regard to possible data inhomogeneities (see Chapter 3).

TABLE 4.1
Averaging Procedures Used in Calculating Spatial Mean Surface Air Temperature

Procedure	Investigators
Inverse-distance weighted best-fit plane onto a 5° latitude by 10° longitude grid, followed by areally weighted averaging	Jones et al. (1982), Kelly et al. (1982), Raper et al. (1983, 1984)
Inverse-distance weighted average onto a 5° latitude by 10° longitude grid, followed by areally weighted averaging	Jones et al. (1986)
Forty equal-area boxes in each hemisphere with box size determined by correlation distance for primary dynamical transports; unweighted averaging of box values	Hansen et al. (1981)
Cubic splines under tension (1), or optimum interpolation (2), onto a 10° latitude by 30° longitude grid (45° longitude at 80°N), followed by areally weighted averaging with missing grid values set equal to zero	(1) Yamamoto et al. (1975) (2) Yamamoto and Hoshiai (1979, 1980), Yamamoto (1980, 1981), Yamamoto and Iwashima (1981)
Subjective contouring of maps with values interpolated onto a 5° latitude by 10° longitude grid, followed by areally weighted averaging	Borzenkova et al. (1976), Gruza and Ran'kova (1979), Vinnikov et al. (1980), Vinnikov and Lugina (1982)

Table 4.2
Areas of Coverage for Spatial Mean Surface Air Temperatures

Jones et al. (1982, 1986)	0–90°N
Kelly et al. (1982)	65–90°N
Raper et al. (1983, 1984)	65–90°S
Hansen et al. (1981)	23.6–90°S, 23.6°S–23.6°N, 23.6°S–90°N, 90°S–90°N
Yamamoto et al. (1975)	0–90°N
Yamamoto and Hoshiai (1979)	0–30°N, 30–60°N, 60–90°N
Yamamoto (1980, 1981)	0–30°N, 30–60°N, 60–90°N, 0–90°N
Yamamoto and Iwashima (1981)	60–90°N
Budyko (1969)	0–90°N
Borzenkova et al. (1976)	17.5–37.5°N, 37.5–57.5°N, 57.5–72.5°N, 72.5–87.5°N, 17.5–87.5°N
Gruza and Ran'kova (1979)	17.5–87.5°N and many others
Vinnikov et al. (1980)	As for Borzenkova et al. (1976)
Vinnikov and Lugina (1982)	72.5–87.5°N

Summaries of the published time series are given in Tables 4.2 and 4.3, which show the claimed areas of coverage (Table 4.2) and the time intervals covered and reference periods (Table 4.3). What is not clear from Table 4.2 is the true station coverage, or how this coverage varies with time. These issues will be addressed later in this chapter. It should be noted, however, that all of the results of studies summarized in Tables 4.2 and 4.3 are strongly biased toward the continental areas, and coverage over the oceans is invariably poor, coming only from ocean islands and, in recent years, a few weather ships. Note also the limited amount of work on Southern Hemisphere data. Because roughly 80% of the Southern Hemisphere is ocean, analyses of land-based data for this hemisphere cannot confidently be expected to represent the hemisphere as a whole.

Of the available time series spanning more than a few decades, the Soviet group, the Climatic Research Unit, and Yamamoto (1980, 1981) have included tabulated data (annual, monthly, and seasonal values, respectively). Very few researchers have published their methods of data collection and analysis in any detail.

In their initial work, the Japanese group used cubic splines under tension to extend data to regions (particularly oceans) where data were sparse or nonexistent (Yamamoto et al. 1975). Later Yamamoto and Hoshiai (1979, 1980) and Yamamoto

Table 4.3
Time Periods Covered by Large-Scale Spatial Mean Surface Air Temperature Series and Corresponding Reference Periods

Investigators	Period covered	Reference period
Jones et al. (1982)	1881–1981	1946–1960
Jones et al. (1986)	1851–1984	1951–1970
Kelly et al. (1982)	1881–1981	1946–1960
Raper et al. (1983, 1984)	1957–1983	1957–1975
Hansen et al. (1981)	1881–1980	1881–1980
Yamamoto et al. (1975)	1951–1972	1951–1972
Yamamoto and Hoshiai (1979)	1951–1977	1951–1975
Yamamoto (1980, 1981)	1876–1975	1931–1960
Yamamoto and Iwashima (1981)	1951–1977	1951–1975
Budyko (1969)	1881–1960	1881–1960
Borzenkova et al. (1976)	1881–1975	1881–1975
Gruza and Ran'kova (1979)	1891–1976	Absolute values given
Vinnikov et al. (1980)	1881–1978	1881–1975
Vinnikov and Lugina (1982)	1881–1981	Absolute values given
Chen (1982)	1949–1972	1949–1972
Paltridge and Woodruff (1981)	1880–1977	1880–1977
Folland et al. (1984a)	1856–1981	1951–1960

(1980, 1981) used an optimum interpolation technique. Their results show changes that are similar to those of other workers, but with a much smaller range of variation, because they set anomaly values to zero where no data were available. This is equivalent to assuming no climate change over a substantial portion of the study region, an assumption that is not supported by other work (e.g., Jones et al. 1982, their Figure 4a and accompanying discussion). The results of Hansen et al. (1981; see their Figure 3) are of considerable interest, but they give little information about their data sources or analysis methods. Some of their results are probably unrepresentative of the claimed region of coverage (a point that will be discussed further below), namely their low latitude (23.6°N–23.6°S) and southern latitude (23.6°S–90°S) time series.

The data presented by Jones et al. (1982) were constructed by interpolating monthly mean station data onto a 5° latitude by 10° longitude grid and then averaging the areally weighted grid point values. As the stations have different altitudes, it is not possible to interpolate raw station data. Instead, it is necessary to use temperature anomalies from a reference period. The period 1946–1960, which had the best data coverage, was used as the reference period for calculating anomalies. Many stations began recording temperatures in 1951, and a number of long-term stations ceased or were moved to airport locations during the 1950s. For a station to be included in their analysis, they required at least 10 years of data between 1946 and 1960. The method of gridding used an inverse-distance-weighted best-fit plane and has been described in detail by Jones et al. (1982). This work has recently been updated (Jones et al. 1986). A considerable amount of previously unused data has been incorporated into the analysis (see Bradley et al. [1985a], for details), and the time series of hemispheric mean temperature has been extended back to 1851. In their reanalysis, Jones et al. (1986) carefully screened all data for inhomogeneities (see below), and used a slightly different (and simpler) method of gridding and a different reference period (1951–1970).

The Soviet data are areally weighted averages of 5° latitude by 10° longitude grid point values extracted from carefully (but subjectively) analyzed monthly mean charts. Three series of charts were used. There are differences of opinion on the homogeneity of the data. These differences are discussed in detail by Vinnikov (1977) and also by Jones et al. (1982), Robock (1982), and Ellsaesser et al. (1986). The basic data charts do not represent a homogeneous time series, but it is clear that some effort has gone into producing homogeneous area-averaged series. The data published by Borzenkova et al. (1976) and Vinnikov et al. (1980) are essentially updates of data published earlier by Budyko (1969), but a comparison of these series shows a data discontinuity around 1940 of about 0.09°C, and there are other differences on the interannual time scale. These differences probably arise because adjustments made by Borzenkova et al. and Vinnikov et al. were not all incorporated in the earlier data of Budyko.

4.2.2 Time Series

The three most easily compared time series are those representing Northern Hemisphere annual mean temperatures given by Vinnikov et al. (1980), Hansen et al. (1981), and Jones et al. (1982). Although Hansen et al. (1981) have only published 5-year running means of their data, we have obtained the individual annual values to facilitate this comparison.[1] The data of Jones et al. (1982) have recently been updated and revised, but we use the original data for comparison here, because the revised version employs many sources not used by either Hansen et al. or Vinnikov et al. The three time series compared here are based on essentially the same data sources. (For further information on sources see Chapter 3.) Although the basic sources are similar, there are differences in the methods used for producing area averages (see Table 4.1) and slight, but possibly important, differences in claimed area of coverage (Table 4.2). Vinnikov et al. cover the region 17.5°–87.5°N, Hansen et al. cover 23.6°–90°N, and Jones et al. claim their series represents the whole Northern Hemisphere, but that coverage is far from complete. The three series are shown in Figure 4.1. Correlation coefficients between the series are $r_{JV} = 0.955$, $r_{JH} = 0.965$, and $r_{VH} = 0.938$ over the maximum periods of overlap, with the subscript letters representing the first letters of the names of the primary authors. Because of these strong correlations and the overlap of data sources, more detailed analyses of Northern Hemisphere data (below) will be restricted to the updated version of the data set of Jones et al. (1986).

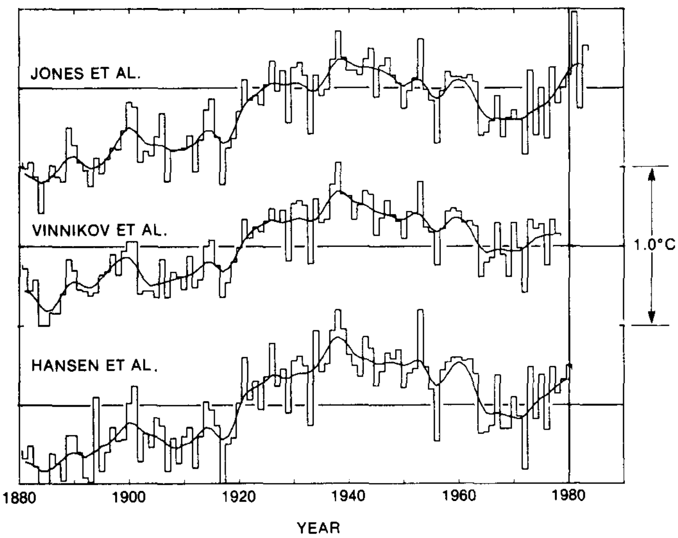

Figure 4.1. Surface air temperature changes representative of the Northern Hemisphere land masses from Vinnikov et al. (1980), Hansen et al. (1981), and Jones et al. (1982). Smooth curves were obtained by using a 10-year Gaussian filter.

[1] J. E. Hansen, personal communication.

Figure 4.2 shows the Northern Hemisphere data of Jones et al. (1986) by season. All seasons show the same long-term trends, trends that are also common to all other land-based data sets: a warming from the 1880s to around 1940, cooling to the mid-1960s/early-1970s (less obvious in winter), and subsequent warming, beginning later in summer and autumn than in spring and winter. On the monthly time scale, all months show a long-term, statistically significant warming trend, although the magnitudes of the trends and the levels of significance vary noticeably (Figure 4.3). The most significant trends occur in late spring.

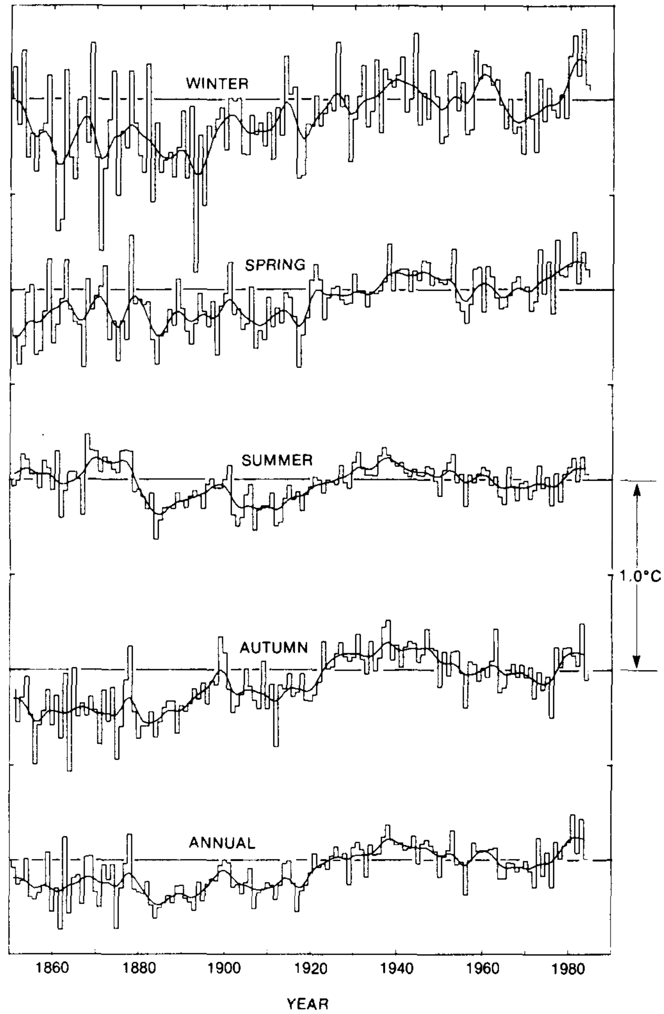

Figure 4.2. Northern Hemisphere surface air temperature fluctuations by season. Data are from Jones et al. (1986). Smooth curves were obtained by using a 10-year Gaussian filter.

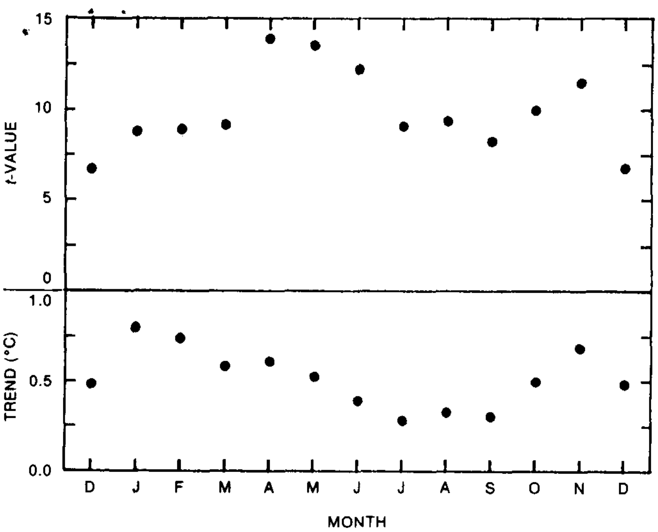

Figure 4.3. Total trends and corresponding t statistics month-by-month for the Northern Hemisphere surface air temperature data of Jones et al. (1986). The trends shown here are the temperature changes over the period 1881–1984 inclusive obtained by fitting a straight line to the appropriate monthly data. All t values are highly significant. The largest warming trends occur in the late winter months of January and February, but the most significant trends (highest t values) occur in late spring (April, May).

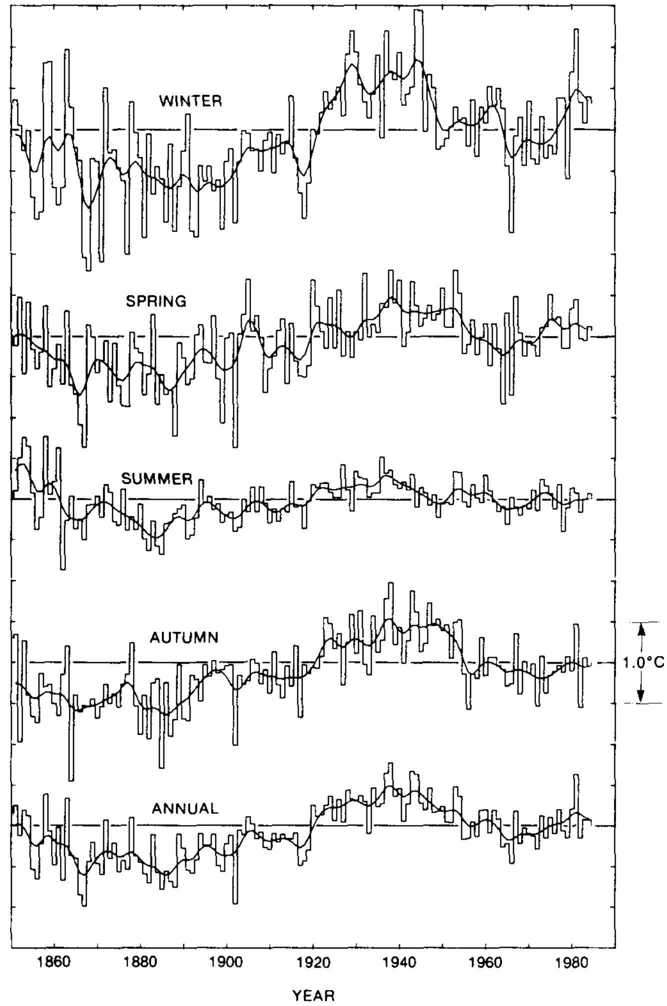

Figure 4.4. Arctic (65°–90°N) surface air temperature fluctuations by season. The data are essentially those of Kelly et al. (1982), as revised and updated by Jones et al. (1986). Smooth curves were obtained by using a 10-year Gaussian filter.

In Figure 4.4 we show the seasonal time series for the Arctic, 65°–90°N (from Kelly et al. 1982, updated). Other Arctic data have been presented by Yamamoto (1980, 1981), Yamamoto and Iwashima (1981), Borzenkova et al. (1976), and Vinnikov and Lugina (1982). The data in Figure 4.4 show similar trends to those from the Northern Hemisphere. Hemispheric variations tend to be dominated by those in high latitudes, a point made earlier by van Loon and Williams (1976). Variability is clearly higher in higher latitudes (see also Figure 4.8).

Hansen et al. (1981) give one of the few long time series of Southern Hemisphere data. Their published results are for the region south of 23.6°S. In Figure 4.5 we show their results for the entire Southern Hemisphere.[2] For the early part of the record, data coverage probably amounts to less than 10% of the total area, while even in recent decades coverage is poor because so much of the hemisphere is ocean. We therefore have doubts about the representativeness of these data as a true indicator of changes in Southern Hemisphere annual mean temperatures.

Figure 4.6 shows data for the Antarctic (65°–90°S) from Raper et al. (1983, 1984). These data are based on what is still rather limited coverage (see Raper et al. [1984], their Figure 2). No reliable area-average values can be estimated prior to 1957, the time when most Antarctic stations were established. In recent years some key stations have ceased regular recording (e.g., the Byrd station).

All of the data series mentioned above have restricted coverage. The main gaps are over ocean areas, and these can only be incorporated into the network by using ship-based observations. Recently, Folland et al. (1984a; see also Folland and Kates 1984 and Folland et al. 1984b) have published global averages of seasonal mean sea surface temperatures (SST) and nighttime marine air temperatures (NMAT) for the period 1856–1981, and annual NMAT data for the Northern Hemisphere back to

[2] Ibid.

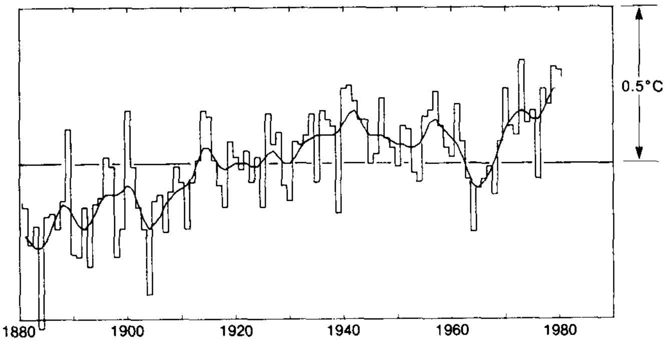

Figure 4.5. Southern Hemisphere annual mean temperature fluctuations (land-based stations) from Hansen et al. (1981). The data shown are for 0°–90°S supplied by J.E. Hansen (personal communication). Hansen et al. (1981) only give the 23.6°–90°S average. A smooth curve was obtained by using a 10-year Gaussian filter.

1881, which help to fill this gap. Their annual mean NMAT data for both hemispheres back to 1856 are shown in Figure 4.7.[3]

The main difficulty in using ship-based data is that spurious temperature fluctuations can arise because of changes in instrumentation (for an extensive discussion, see Barnett [1984] and Chapter 5). For SST data, there was a general change from uninsulated bucket measurements to engine intake measurements around 1940, although insulated buckets still have been widely used since that time (Folland et al. 1984a). For MAT, data have been influenced by the thermal inertia of the ship (which has changed with time as ship sizes have changed) and have been affected by the changing height of observations as ships have become larger. The adjustments applied to correct for these nonclimatic effects are of similar magnitude to the observed, long time scale changes. Corrections for daytime MAT data are so uncertain that only nighttime MAT data are considered by Folland et al. (1984a) to be of value, and even here the corrections are open to debate (see Barnett 1984).

Folland et al. have justified the corrections made to their NMAT and SST data, but there are still some doubts about the homogeneity of the series, especially prior to 1900. What inspires confidence in their work is the strong similarity between their annual NMAT data for the Northern Hemisphere oceans and the totally independent land-based data

[3] D.E. Parker and C. K. Folland, personal communication.

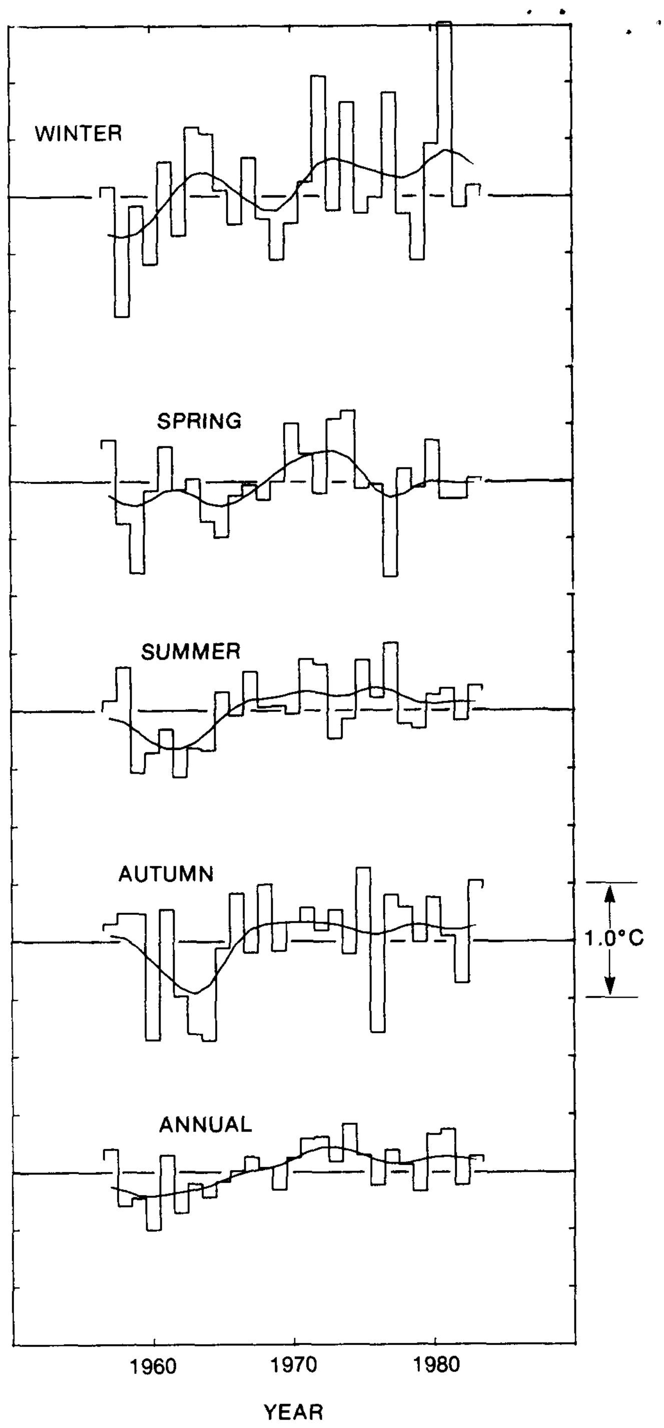

Figure 4.6. Antarctic (65°–90°S) surface air temperature fluctuations by season. Data from Raper et al. (1984, updated). Smooth curves were obtained by using a 10-year Gaussian filter.

of Jones et al. (1982, 1986). Folland et al. (1984a, see their Figure 2) compared their results with the data published by Jones et al. (1982). We compare the results of Folland et al. with those of Jones et al. (1986) here (Figure 4.7) and in Section 4.5. The

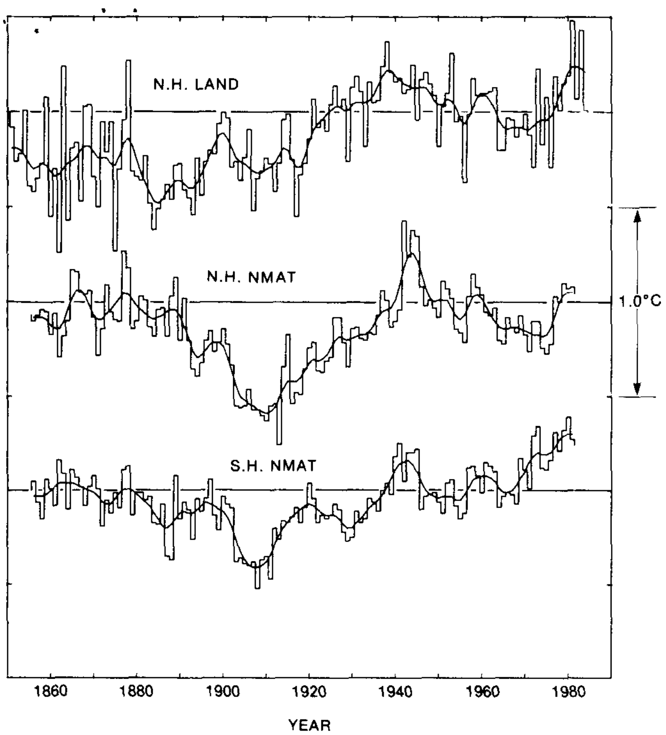

Figure 4.7. Comparison of Northern and Southern Hemisphere surface air temperature fluctuations. Data are from Jones et al. (1986) for the Northern Hemisphere land areas and from Folland et al. (1984a) for Northern and Southern Hemisphere nighttime marine air temperatures (NMAT). The unpublished Southern Hemisphere data were supplied by D. E. Parker (personal communication). Smooth curves were obtained by using a 10-year Gaussian filter.

two time series are highly correlated back to around 1895, and both series show long-term changes and interannual variability of similar magnitude (the latter being slightly less for the marine data). This indicates that the corrections applied to the NMAT data are reasonable and, because the same corrections have been applied globally, that the corrected Southern Hemisphere data are also reliable. Because the SST corrections have been calculated on the basis of correlations with MAT data, the SST data are probably equally reliable. There are, however, large variations in coverage that particularly may affect the representativeness of the Southern Hemisphere data and all data prior to about 1900. This is discussed in more detail in Section 4.5.

In mentioning sources of inhomogeneity in the marine data, we do not mean to imply that land-based data are perfect. This is far from the truth, and individual station records often show changes or trends that are not climatic in origin. The main sources of inhomogeneity in these records are station relocations, changes in observation times or in the methods used to calculate daily means, and urban warming (see, for example, Mitchell [1953, 1958], Dronia [1967], and Chapter 3). In producing area averages, data must be carefully selected to avoid those records which are obviously not homogeneous or records that must be homogenized to remove nonclimatic effects. It is likely that, in all analyses of land-based data, some poor-quality station records have been included. However, as shown by Jones et al. (1986) and by the similarities between land and marine time series, any resulting errors in large-scale area averages are almost certainly small.

4.2.3 Spatial Patterns

Although the globe as a whole has shown significant warming and cooling trends over the past century, the trends in different regions tend to differ markedly from each other and from the global or hemispheric mean trends. This has been known for many years and is well documented in the works of Mitchell (1961, 1963), van Loon and Williams (1976), and Brinkmann (1979). More recent analyses have been carried out by Jones and Kelly (1983), who considered changes over three periods between 1917 and 1980, and by Vinnikov and Kovyneva (1983), who considered overall trend patterns for the period 1891–1979.

Brinkmann, in particular, notes that the relationships between trends in different regions change with time as a result of decadal time scale shifts in the longitudinal positions of the long waves and the main atmospheric centers of action. Similar results have been obtained by Jones and Kelly. They began by noting that three phases can be identified in the land-based time series for the Northern Hemisphere covering the last 100 years, a warming to around 1940, cooling to the mid 1960s/early 1970s, and warming since then. They then examined the spatial patterns of changes in temperature during these three periods, specifically for 1917–1939, 1940–1964, and 1965–1980. During the first period, most of the land areas showed strong warming, especially in high northern latitudes. However, cooling occurred in central northern Canada and over part of central Asia. In the subsequent cooling phase, cooling was strongest in high latitudes. There were, once again, large areas where the changes were in the opposite direction to the general trend in the Northern Hemisphere. However, these areas of contrary trends were *not* the same as in the earlier warming, and the cooling pattern was *not* the opposite of the previous warming period pattern. The pattern of most recent warming is different again and shows little similarity with the earlier warming pattern. Since the mid-1960s, in fact, there has been a significant part of the Arctic in which the trend has been one of cooling.

These results have two important implications. First, even if models could reliably predict the spatial patterns of changes in surface air temperature resulting from an increasing CO_2 concentration, these patterns are unlikely to be easily identified in the pre-1980 record simply because the data for this period show no consistent spatial signal. Second, it is unlikely that there is any shortcut to obtaining indices of hemispheric average temperatures based solely on information from small-scale "key" regions on the continents or oceans. This second point is illustrated forcefully by the maps of the correlations between the hemispheric mean and individual grid point values given by Jones and Kelly (1983). The correlation pattern for the period 1901–1940 differs markedly from that for 1941–1980. In general, at any individual grid point, the correlation with the hemispheric mean is not a constant, but varies significantly with time. As noted above, this is a consequence of medium time scale shifts in the general circulation. If the spatial scale is sufficiently large, however (i.e., larger than the scale of long waves, but smaller than hemispheric), the above conclusion may not apply. Indeed, Bradley et al. (1985b) have found that decadal time scale fluctuations in the spatial mean temperature over China (spanning longitudes from 80°–130°E) correlate highly with the hemispheric mean. Similar high correlations are found for other regions (e.g., North America).

Some authors have used proxy climate records or long instrumental data series from specific small regions as indicators for the whole Northern Hemisphere: examples include the Iceland drift ice (Koch 1945; Bergthórsson 1969; Ogilvie 1984), temperatures from central England (which extend back to 1659; Manley 1974), oxygen isotope data from Greenland ice cores (Dansgaard et al. 1971; Hammer et al. 1980), and the long Bristlecone pine tree ring record from the southwestern United States (LaMarche 1974). (The references here are to the original data sources, in which hemispheric representativeness was not necessarily claimed.) The study by Jones and Kelly (1983) showed that in none of the areas cited above was the correlation between local temperature and the Northern Hemisphere average consistent before and after 1940.

These results also throw doubt on the analog forecasts of CO_2-induced climate change produced by Groisman (1981), Vinnikov and Groisman (1979), Kovyneva (1982), and Vinnikov and Kovyneva (1983), and on the empirically determined patterns of sensitivity to carbon dioxide and volcanic forcing given by Vinnikov and Groisman (1982). All of these studies are based on relationships between local climate and the Northern Hemisphere average temperature. Analysis of such relationships shows that many appear to be statistically significant when examined over the full period of the record. However, if the data were considered in two halves, it is likely, based on information presented by Jones and Kelly (1983), that most local relationships would be unstable. Thus, empirically based *forecasts* of the detailed patterns of future climate change are of doubtful reliability. This does not, however, preclude the use of such data for the construction of instrumental climate *scenarios* (Namias 1980; Wigley et al. 1980; Williams 1980; Jäger and Kellogg 1983; Lough et al. 1983; Palutikof et al. 1984). A more detailed discussion of climate scenarios is given by Webb and Wigley (1985).

To conclude this section, we show, in Figure 4.8, the spatial pattern of the variability of annual mean temperature (1951–1970). The climate state is not defined solely by mean values, but requires information about higher order moments. Variability (i.e., standard deviation) is important because it is a measure of the background natural noise level against which a CO_2 (or CO_2 plus other trace gases) signal must be detected. This noise may include variations due to forcing factors other than CO_2 (e.g., volcanic or solar effects). For the Northern Hemisphere as a whole, the standard deviation of year-to-year temperature variations is approximately 0.2°C. Both MAT and SST data are less variable than the land-based data on the interannual time scale (Barnett 1978), but all data show similar magnitudes of long time scale variations. Uncertainty in defining the variability arises for two reasons, partly from the incomplete data coverage and also because standard deviation is not always a stable property of the available data series. Times of strong secular trends (e.g., 1910–1940) tend to have greater standard deviation than times when conditions have shown few trends (e.g., 1940–1960); and even when the lower frequency changes are filtered out, the variability changes noticeably with time. The data presented in Figure 4.7 show that there has been a considerable increase in the interannual variability of the land-based Northern Hemisphere mean temperature since 1970 that is not evident in the NMAT data. Figure 4.8 confirms the frequently stated result that variability of surface air temperatures is highest in high latitudes. In terms of latitudinal dependence, the changes in variability roughly parallel the latitudinal variations in the expected equilibrium temperature change resulting from a CO_2 doubling based on, for example, the work of Manabe and Stouffer (1979, 1980). (See Schlesinger and Mitchell [1985] for a discussion of latitudinal variations in recent model studies.) Thus, even though the signal may be highest in high latitudes, the high-latitude noise level is larger than elsewhere (at least in the Northern Hemisphere), and there would appear to be no strong advantage in choosing a particular latitude zone for detection, given the uncertainties in both the predicted signal and the observed noise. This point is further discussed below (Section 4.6.3).

4.3 SPATIAL REPRESENTATIVENESS

Most of the discussion above was concerned with Northern Hemisphere data. However, it is a global mean temperature change that is the most reliably modeled effect of CO_2 on climate. The emphasis on the Northern Hemisphere is because this is the region with best data coverage. Even if land and marine data sets were combined, there would still be gaps in coverage in the Northern Hemisphere and much larger gaps in the Southern Hemisphere. There is thus considerable uncertainty in our knowledge of how global mean temperatures have varied in the past. Because of this, many workers have used the Northern Hemisphere land-based record as a global proxy. In a few studies Southern Hemisphere land data have been included (as in Hansen et al. 1981), but even when this is done, overall coverage is far from complete. With the recent publication of marine data sets (Folland et al. 1984a, 1984b; Fletcher 1984; Slutz et al. 1985; Woodruff 1985), it should be possible to produce better estimates of global mean temperature change, but to date no thorough global synthesis of marine and land data has been published. It is, essential, therefore, to examine the global representativeness of hemispheric or smaller-area data.

How representative is the available data network of the true hemispheric or global average? How much are changes in estimated hemispheric or global mean temperatures affected by changes in data coverage? These questions have received relatively little attention in the literature, with the most recent relevant work being that of Barnett (1978), Oort (1978), Parker (1981), Chen (1982), and Jones et al. (1982, 1986). Of all published land-based temperature records spanning more than a few decades, only Jones et al. (1982, 1986) have attempted to quantify overall coverage and changes in coverage. Their work gives some idea of the scale of the coverage problem. For the Northern Hemisphere, Jones et al. (1986) show that, even when coverage is best, temperature anomaly values can be calculated for only about 58% of the area of the Northern Hemisphere. In 1900 this percentage was only about 27%. These values are slightly less than those in Jones et al. (1982) because, despite a large increase in the amount of data used, the number of gridded points was reduced through the use of more stringent extrapolation and interpolation methods. The total area of coverage changed only slightly because of the inclusion of additional data in low latitudes.

Barnett (1978) used both sea surface and land-based air temperature data for the period 1950–1976 to estimate the representativeness of a spatially limited network. The data he used covered about 50% of the area between 20° and 80°N, or approximately 30% of the Northern Hemisphere. His analysis showed that the interannual variability of temperature over land is four to six times greater than that over the oceans for winter averages, and two to three times greater for annual averages. From this one might conclude that quite large

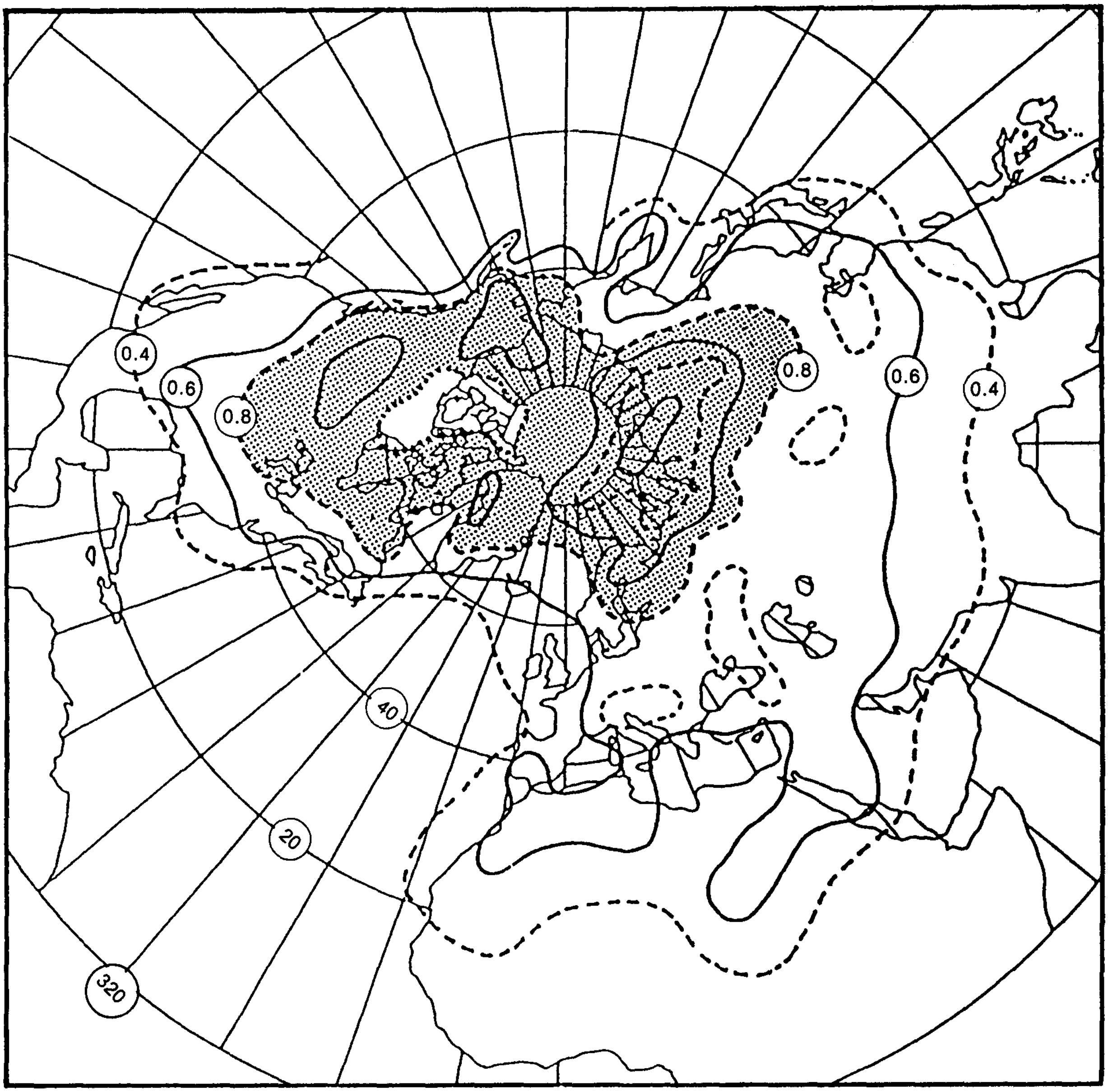

Figure 4.8. Variability of annual mean surface air temperature. The values are standard deviations calculated at individual grid points over the period 1951–1970 from the grid-point data set of Jones et al. (1986).

data gaps over the oceans could be tolerated. However, interannual variability is not directly relevant to CO_2-induced changes in climate. As noted above, long time scale changes over the land and ocean are of similar magnitude, so ocean coverage would seem to be essential to identify changes on time scales relevant to the projected CO_2 increase. Barnett reached a similar conclusion using principal components analysis, a method that isolates the medium to long time scale fluctuations in the first component. He concluded that for any incomplete Northern Hemisphere temperature network, the spatial average will not represent the hemispheric average well unless the ocean areas form a substantial part of the network.

This conclusion may be too pessimistic. Jones et al. (1982) compared averages using their full data set with averages based on data coverage available in the year 1900 and found the two series to be strongly correlated ($r = 0.92$ over the period 1946–1980). This indicates that even the limited coverage available at around 1900 may give good estimates of much larger-scale averages over the land. More recently Jones et al. (1986) have shown that with their expanded station data base, reasonable estimates of the land-based Northern Hemisphere mean temperature can be made back to around 1880. Work by Chen (1982) indicates that inclusion of data from ocean areas in developing an estimate of the Northern Hemisphere average has little effect on medium time scale fluctuations in the estimated large-scale average. On the other hand, Parker (1981) has suggested that results may depend crucially on the positions of stations relative to the (changeable) positions of long-waves in the circulation. Clearly, some caution must be used in interpreting published time series of hemispheric or near-hemispheric mean temperatures. Nevertheless, Northern Hemisphere land-based trends do appear to be representative of the whole hemisphere, as evidenced by the strongly parallel trends in the land and marine time series (Figure 4.7). This similarity will be further discussed below (Section 4.5).

In the Southern Hemisphere the situation is more extreme. Approximately 80% of the hemisphere is ocean, with the region 40°–65°S being more than 95% ocean. No land-based Southern Hemisphere time series can be considered representative of the whole hemisphere, and even over the land there are substantial gaps. For example, for the Antarctic region (south of 65°S) there are practically no data prior to 1957. Ship-based data are essential to obtain information about hemispheric scale temperature changes, but even with ship-based data, there are coverage problems away from shipping routes, particularly in the Pacific and south of 40°S. The spatial representativeness of the available Southern Hemisphere data has yet to be assessed. In this regard, assessments guided by general circulation model results (e.g., Oort 1978) may be necessary.

Completely adequate coverage using ground-based observations will never be achieved. However, in the near future, virtually complete global coverage might be obtained using satellite data, provided that ground truth can be established and appropriate calibration equations can be developed. In the meantime, although there are still some problems to be resolved, coverage does appear to be adequate to assess medium to long time scale changes back to around 1900. Further discussion is given in Section 4.5.

4.4 TROPOSPHERIC AND STRATOSPHERIC TEMPERATURES

4.4.1 Introduction

Roughly 1000 stations over the globe make radiosonde measurements of free atmospheric parameters on a regular basis. They record temperature and moisture content as a function of pressure, and also produce information on wind velocities and pressure as a function of geopotential height. The upper air network began to develop in the 1940s, but balloons seldom went above the 100-mb[4] level (approximately 16 km) until the early 1950s. Today, balloons usually reach the 30-mb level (approximately 24 km) or higher. Over the past few decades, small rockets (rocketsondes) have been launched regularly, usually weekly, from a network of about 20 sites to heights of more than 50 km (mainly in the western half of the Northern Hemisphere). Unfortunately, the rocketsonde program was reduced recently, and even those data that are available now have homogeneity problems that limit their use (see Section 4.4.4).

Radiosonde and rocketsonde data may be analyzed for trends by using two types of data, the temperature at mandatory pressure surfaces or the difference in height (thickness) between mandatory pressure surfaces, with the latter data being proportional to the mean virtual temperature within the layer bounded by the pressure surfaces.[5] Two different types of thickness data have been used for determining temperature changes, raw station data, and data that have been interpolated and extrapolated onto a regular grid. Analyses have been based either on the daily or twice daily gridded data (e.g., Harley 1978, 1980; Boer and Higuchi 1980) or on monthly mean station data from *Monthly Climatic Data for the World* (as in the work by Angell and Korshover that forms the basis for this section; Angell and Korshover 1977, 1978a, 1978b, 1983a).

[4] 100 mb = 10 kPa. Because millibars are still used extensively in operational meteorology, we retain these units in further discussion.

[5] Virtual temperature depends on both temperature and atmospheric moisture content, but fluctuations in virtual temperature (T_v) are essentially identical to those in temperature (T). If relative humidity is constant, then $\Delta T_v \approx \alpha \Delta T$ when α differs from 1 by less than 1%, except close to the surface where the difference may reach 10–20%.

Raw temperature data have been examined by only a few workers (e.g., Starr and Oort 1973; Oort 1983). These analyses, which span the period 1958–1973 only, show quasi-periodic temperature fluctuations that can be partly attributed to the influence of sea surface temperature changes (Newell and Weare 1976a, 1976b; Pan and Oort 1983). There may, however, be data homogeneity problems (see below). In any event, the time series available are short, and trend estimation for possible detection of CO_2 effects is premature.

A number of authors have examined grid point thickness data. Dronia (1974) analyzed 1000–500 mb data averaged over middle and high latitudes of the Northern Hemisphere back to 1949 and found an overall cooling trend, largely due to an almost step function drop in 1963–1964. Some authors have attributed this drop to the possible cooling effect of the eruption of the volcano Mt. Agung (March 1963), but this conclusion may be suspect because of data problems (see below). Namias (1980) used 1000–700 mb data and found a general cooling trend over the period 1951–1978, with substantial year-to-year fluctuations. Harley (1980) analyzed daily 1000–500 mb thickness data from 25–90°N for 30 years (1949–1978) to determine mean temperature changes in the lower troposphere between successive 5-year periods. He found four small areas with an unbroken record of 25 years of cooling and one (northern Iran) with an unbroken record of continuous warming. He concluded that there was no evidence of any large-scale warming. Boer and Higuchi (1980) also found no evidence of large-scale warming in their analysis of twice daily 1000–500 mb grid point thickness data from the U.K. Meteorological Office. Their study covered the region 25°–90°N and the period 1949–1975. Over this period there was a small cooling trend, but little change after 1964.

The results cited above must be viewed with caution, since both raw station data and grid point data may be influenced by nonclimatic effects. Inhomogeneities in station data may arise from changes in instrumentation, changes in the standard corrections that are made to compensate for radiative effects, and changes in observation time. Grid point data may also be affected by changes in analysis procedures. In this connection, a major problem with grid point data has been exposed by Parker (1980). He found that values determined by different meteorological agencies can show considerable disagreement, particularly over the oceans and in the subtropics. Similar data discrepancies were found by Higuchi (1985) in repeating the earlier analysis by Boer and Higuchi (1980) using National Center for Atmospheric Research (NCAR) data. Drastically different results were found prior to 1963. Large differences between estimates of spatially averaged upper air temperatures on seasonal to decadal time scales have also been reported by Arpe (1980) and Lau and Oort (1981). In the latter work, 1958–1972 means of zonally averaged winter temperatures at 850 mb and 30°N based on two different grid point analyses, those of the Geophysical Fluid Dynamics Laboratory (GFDL) and the U.S. National Meteorological Center (NMC) were found to differ by 1°C, even though both analyses used the same basic radiosonde data network.

Despite these reservations regarding their use, grid point data can produce results that correlate well with estimates based on station data. For example, Dronia's (1974) 1000–500 mb results for 25°–90°N and Angell and Korshover's station data-based Northern Hemisphere mean for 850–300 mb (see Section 4.4.3) are strongly correlated over the period 1958–1973, although the strength of this result is partly due to the temperature drop around 1963–1964 that is common to both data series. Furthermore, although the differences between data sets noted above imply considerable uncertainty in *absolute* temperature levels, the same uncertainty may not apply to estimates of year-to-year differences, annual anomalies from a long-term mean, or trends obtained from a single, self-consistent data set.

In addition to the problems stated above, both station-based and grid point upper air analyses may be influenced by the restricted network of observation sites. The network is much less extensive than the surface data network. Because large-scale spatial averages must inevitably involve averaging over areas in which little or no data are available, spatially averaged data may not be representative of true area averages. Such averages must have an inherent uncertainty, the magnitude of which is, in general, unknown.

4.4.2 Station Data Network

The locations of the 63 radiosonde stations used by Angell and Korshover and in the present analysis of tropospheric and stratospheric temperatures are shown in Figure 4.9. The basic pressure-height data have been obtained from *Monthly Climatic Data for the World,* a publication of the National Climatic Data Center, National Oceanic and Atmospheric Administration (NOAA), as well as from real-time teletype data in recent years. The record was begun in 1958 because of the improved data coverage provided during the International Geophysical Year (IGY) and because of possible data homogeneity problems arising from a 3-hour change in upper air observation times in 1957. Extension back to around 1950 would be possible, and this has been done for the tropical region (20°N to 20°S) by Parker (1985a). In choosing radiosonde locations, every effort has been made to produce a spatially uniform network. The globe has been divided into seven latitudinal zones; north and south polar (60°–90°), north and south temperate (30°–60°), north and south subtropical (10°–30°), and equatorial (10°S–10°N). In general, longitudinal station spacing is about 30° except in the zones south of 30°S, where the spacing is larger because of lack of data over the ocean areas and the relatively poor network in the Antarctic. Of the 63 stations, 8 are located in the north polar zone, 12 in the north temperate, 12 in the north subtropical, 9 in the north equatorial, 10 in the south subtropical, 6 in the south temperate, and 6 in the south polar zone. San Diego, although just north of 30°N, is included in the north subtropical zone.

As a consequence of the relatively uniform longitudinal spacing of the radiosonde stations, irrespective of continental and oceanic areas, and because column-mean temperatures tend to be spatially more uniform than surface temperatures (Angell and Korshover, 1983a), this relatively sparse radiosonde network probably samples the large-scale domains (climatic zones, hemispheres, and globe) satisfactorily. However, that there are shortcomings in the radiosonde data cannot be denied. In some countries the data are erratic, particularly those from the stratosphere, and care has had to be taken with regard to time of observation to ensure that daytime and nighttime soundings are not being compared. Because of the greater complexity of the measurement, the overall quality of the radiosonde data is probably not as good as that of the surface data. Nonetheless, as will be shown below, correlations with surface data provide strong support for the reliability of the area averages described in the next section.

4.4.3 Tropospheric Temperature Variations

As indicators of tropospheric column mean temperatures, we have used thickness data for the layers 850–300 mb (1.5–9 km) and 300–100 mb (9–16 km). Area averages for each climatic zone were calculated simply by averaging the station data with no weighting. Except in the tropics, the 850–300 mb layer includes most of the troposphere, and these data generally avoid the natural and anthropogenic vagaries (surface inversions, urban heat islands, etc.) associated with the planetary boundary layer.

In this context it must be emphasized that in polar zones and in the north temperate zone in winter the trends in surface and tropospheric temperature need not be the same, because the surface may become almost decoupled from the troposphere when strong inversions are present. Furthermore, any breakdown of this inversion by strong winds could cause a surface warming that is not reflected in a tropospheric warming.

Figure 4.10 shows annual temperature anomalies (temperature deviations from the 1958–1975 mean) during 1958–1983 for the 850–300 mb layer. The anomalies are presented for the seven climatic zones, both hemispheres, and the globe. The hemispheric average is based on a 1,2,2,1 weighting of polar, temperate, subtropical, and equatorial zones (approximately their areal extent), respectively, and the global average is an average of the two hemispheres.

Figure 4.10 shows that the Northern Hemisphere troposphere cooled during the early part of the interval 1958–1983 and warmed during the later part, with temperatures after 1980 being comparable to those before 1960. In the Southern Hemisphere, the tropospheric temperature trend is similar, with the warming beginning about a decade earlier and the temperatures after 1980 being higher

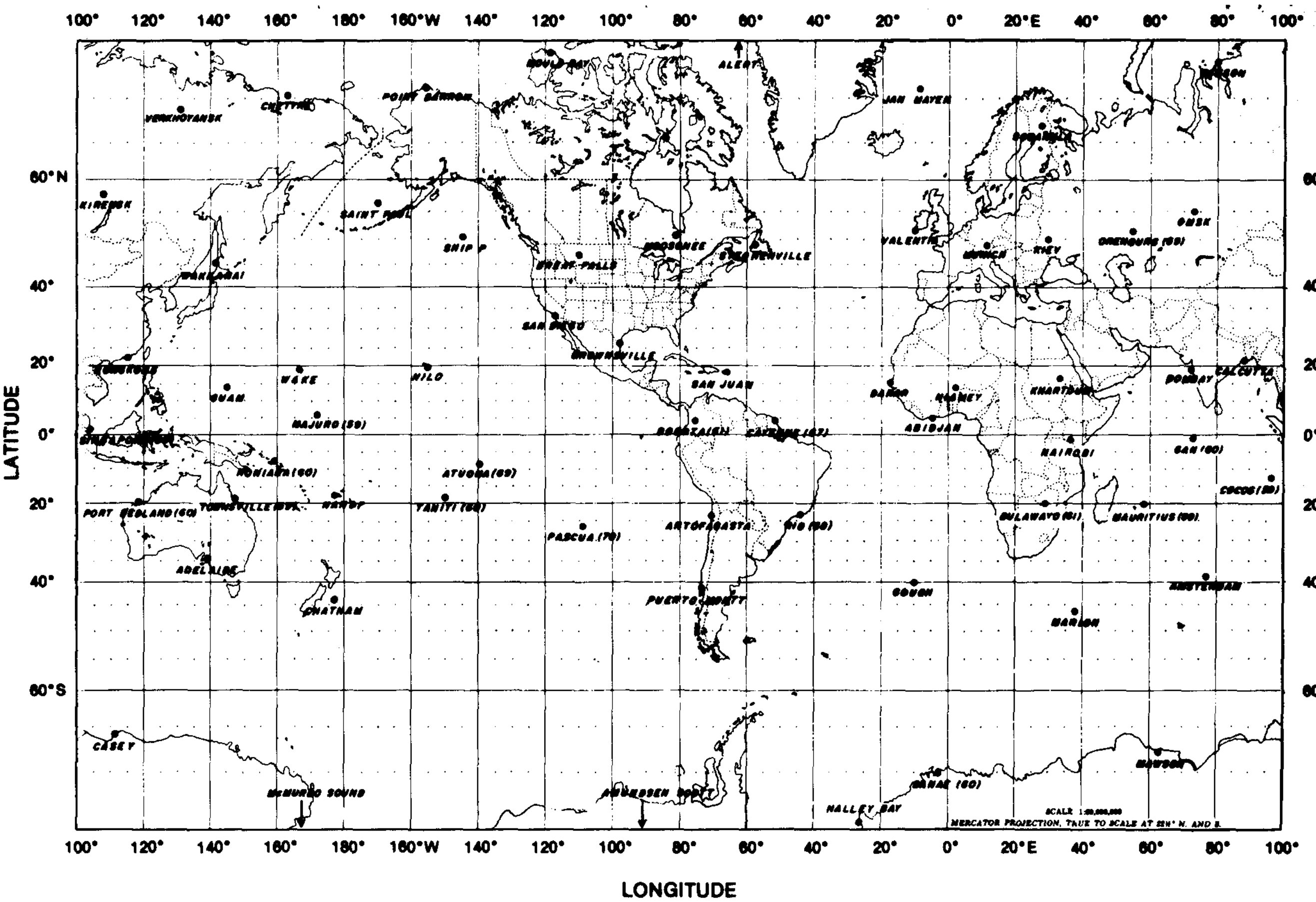

Figure 4.9. Station network for upper air data. The number in parentheses shows the first year of record if later than 1958.

than those before 1960. There is considerable variability both on interannual and interdecadal time scales. The latter variations are in both directions, except for the equatorial and south polar zones, which show steady warming. The steady equatorial warming also shows up in Parker's (1985a) analysis of 850–200 mb data for the zone 20°N–20°S, but it is slightly less evident when the strong correlation with the El Niño/Southern Oscillation Index is factored out.

Figure 4.10 also shows that, in the north temperate zone, the equatorial zone, the Southern Hemisphere, and for the world as a whole, the warmest tropospheric temperatures of the 26-year record were observed during 1983. The year 1983 was also an exceptionally warm year in the record of Jones et al. (1986) (see Figure 4.4), second only to that in 1981. This warmth is despite the El Chichón (Mexico) eruption at 17°N in the spring of 1982. At least in tropical regions, part of the warmth of 1983 appears to be due to the strong El Niño in 1982/1983 (cf. Angell and Korshover 1984; Parker 1985a).

It is useful to examine tropospheric temperature variations by season, both for the purpose of determining which seasons are mainly responsible for the annual temperature variation and for comparison with seasonally specific model predictions of the effects of a CO_2 increase. The earlier models indicated that, for the case of an abrupt doubling of the CO_2 concentration, the equilibrium atmospheric warming should be greater in winter than in summer in extratropical regions; but more recent model studies of the transient case suggest that, initially, the largest warmings might be in the transition seasons of spring and autumn.

Figure 4.11 shows tropospheric temperature variations by season in the Northern Hemisphere from which the data are best. The differences between seasons are not large, although in the north temperate zone the trends have been larger in winter than in summer, and in the north polar zone

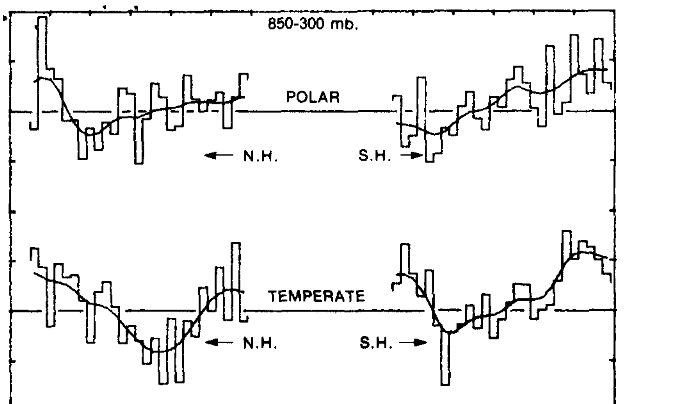

Figure 4.10. The 850–300 mb annual mean temperature fluctuations. Smooth curves were obtained by using a 10-year Gaussian filter.

there is little evidence of warming in winter or autumn. The absence of any strong seasonal character may be because the tropospheric data record is not yet long enough to adequately define such differences. Certainly we cannot infer a CO_2 effect based on seasonal differences in warming, especially when there are still doubts about the seasonal details of the response to increasing CO_2. Because of this latter point, no attempts have been made to assess relative signal-to-noise ratios for different seasons. However, the low interannual variability in the temperate zone in summer (see Figure 4.11) suggests that it might be useful to examine these data separately when searching for a CO_2 effect.

It is doubtful whether the sparse 63-station radiosonde network used here can detect the subtle changes in tropospheric lapse rate that might be associated with an increase in the CO_2 concentration. To examine this matter, Figure 4.12 presents a comparison of annual hemispheric temperature variations at the surface and in the 850–300 mb and 300–100 mb layers. The latter layer is in the troposphere in the tropics, but in the stratosphere in the polar zone. Lapse rate changes in the lower troposphere could be estimated by taking the difference of the surface and 850–300 mb anomalies, but the value of such a calculation is doubtful because, as will be shown in Section 4.5, the 63-station surface network is too sparse to give a reliable indication of surface air temperature changes. As can be seen from the data in Figure 4.12 there are no consistent trends in the lower troposphere lapse rate. For the upper troposphere-lower stratosphere, lapse rate changes can be inferred from differences between the 850–300 mb and 300–100 mb anomalies. For the Northern Hemisphere there is a trend toward increasing lapse rates over the past 10 years, whereas in the Southern Hemisphere there is a strong increase in lapse rates over the past 20 years. The decreases in 300–100 mb temperatures that contribute to these lapse rate changes arise mainly from changes in the tropics.

Whether or not these changes are in accord with the anticipated effects of increasing CO_2 is difficult to judge, because lapse rate changes may differ significantly with latitude (e.g., J.F.B. Mitchell 1983). The station network used here is probably too sparse to identify latitudinal differences in lapse rate trends. Furthermore, different models give different results for CO_2-induced lapse rate changes. Because of this, a detailed comparison between observations and models is unjustified at present, but the hemispheric mean changes described above are worthy of further study (see also Section 4.6.4).

As a final point in this discussion of tropospheric temperature changes, the relationship between sea surface temperatures in the eastern equatorial Pacific (which correlate strongly with the El Niño/Southern Oscillation phenomenon) and tropical tropospheric temperatures should be kept in mind (e.g., Newell and Weare 1976a, 1976b; Navato et al. 1981; Angell 1981; Pan and Oort 1983). This relationship may obscure not only the effects of CO_2, but also the effects of other climate forcing factors such as volcanic activity (see Section

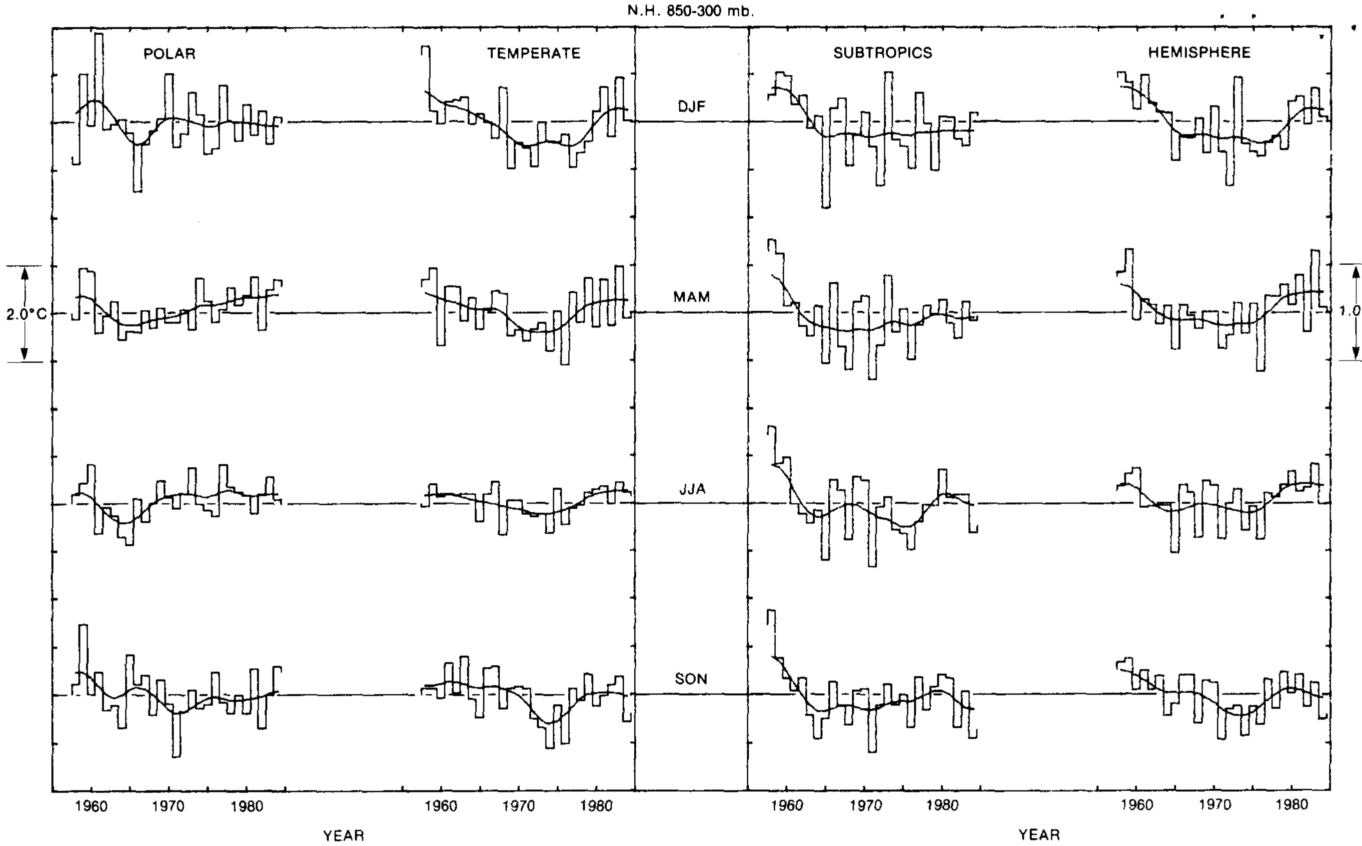

Figure 4.11. The 850–300 mb mean temperature fluctuations for the Northern Hemisphere by season. Smooth curves were obtained by using a 10-year Gaussian filter. (Note the difference in scale between the left and right halves of the figure.)

4.6.5). The SST values in the region 0°–10°S, 180°–80°W are significantly correlated with tropospheric temperatures in the equatorial and subtropical zones, with maximum correlations occurring when the temperatures lag two seasons behind the SST data (Pan and Oort 1983). Similar correlations exist with the Southern Oscillation Index; a low index value, implying low pressure in the southern tropical mid-Pacific, tends to be followed (with a lag of about two seasons) by a warm tropical troposphere. Up to 50% of the variance in tropical tropospheric temperatures can be explained by this relationship, with the exact amount depending on the season (Parker 1985a).

4.4.4 Stratospheric Temperatures

Atmospheric models show that an increase in the CO_2 concentration should lead to stratospheric cooling, with the amount of cooling increasing with height to near the stratopause at 50 km. The magnitude of this cooling is uncertain. J.F.B. Mitchell (1983) finds a temperature decrease at 50 mb of about 10°C for a CO_2 doubling, while Manabe and Stouffer (1980) find that only a CO_2 quadrupling would bring about such a large temperature decrease.

Temperature variations in the lower stratosphere have been estimated from radiosonde-derived differences in the height of the 100 and 50 mb pressure surfaces, yielding mean temperatures for the 16–20 km layer. The mean temperature for the 100–30 mb (16–24 km) layer has also been evaluated from radiosonde data, although in some areas of the world the radiosondes do not reach the 30-mb pressure surface with regularity. Mean temperatures for the layers from 26–35, 36–45, and 46–55 km in the western half of the Northern Hemisphere have been obtained by averaging rocketsonde temperatures at 1-km intervals. Table 4.4 gives the location of the rocketsonde stations and the period that they kept records. Three rocketsonde stations have closed since 1979, making it even more difficult

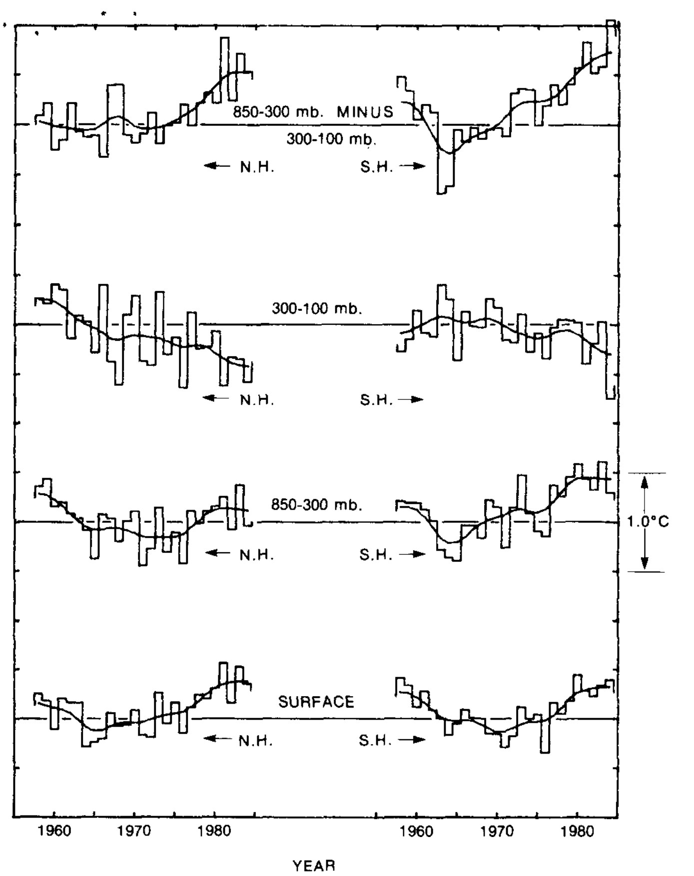

Figure 4.12. Comparison of hemispheric annual mean temperature fluctuations at the surface, in the lower to middle troposphere (850–300 mb), and in the upper troposphere-lower stratosphere (300–100 mb). The differences between the 850–300 mb and 300–100 mb data (upper curves) provide an indication of changes in upper tropospheric mean stability. An upward trend indicates decreasing stability (i.e., an increase in lapse rate). Smooth curves were obtained by using a 10-year Gaussian filter.

to delineate temperature trends in the middle and high stratosphere.

Figure 4.13 shows annual temperature anomalies for the 100–50 mb (16–20 km) layer of the lower stratosphere. In some zones, data prior to (approximately) 1965 show quite different characteristics compared with the more recent data; for example, interannual variability increases dramatically for the equatorial zone and decreases noticeably for the global mean. These changes, and the relatively high interannual variability in general, make it difficult to detect any trends. The perception of trends is also influenced by short-term warming events such as the warming in the lower stratosphere probably associated with the Agung eruption in 1963 (Newell 1970; McInturff et al. 1971), and

Table 4.4
Rocketsonde Stations and Their Period of Record

Station	Latitude	Longitude	Period of record
Thule, Greenland	78°N	71°W	1975–1980
Poker Flat, Alaska	70°N	148°W	1965–1974
Fort Churchill, Canada	59°N	94°W	1971–1981
Primrose Lake, Canada	55°N	110°W	1982–present
Shemya, Alaska	53°N	173°E	1976–present
Wallops Island, Virginia	38°N	75°W	1965–present
Point Mugu, California	34°N	119°W	1965–present
White Sands, New Mexico	33°N	106°W	1965–present
Cape Canaveral, Florida	28°N	80°W	1965–present
Barking Sands, Hawaii	22°N	165°W	1971–present
Antigua, West Indies	17°N	62°W	1972–present
Fort Sherman, Canal Zone	9°N	80°W	1967–1979
Kodaikanal, Marianas	7°N	171°E	1972–present
Ascension Island	8°S	14°W	1965–present

the weaker warming after the El Chichón eruption in 1982 (Parker and Brownscombe 1983; Labitzke et al. 1983; Quiroz 1983). (Note that the bulk of the El Chichón dust cloud was initially above the lower stratosphere [Angell and Korshover 1983b].) Nevertheless, there is evidence of a strong cooling in the Southern Hemisphere subtropical zone. For the hemispheric means and the global mean, however, no trends are apparent over the last 20 years.

Figure 4.14 shows trends in hemispheric temperature for atmospheric layers extending from just below the tropopause to just above the stratopause in the Northern Hemisphere, and through the low stratosphere in the Southern Hemisphere. The data up to 30 mb (24 km) are from radiosondes, and data above this are from rocketsondes. In the Southern Hemisphere, from which the data are relatively poor, there is evidence of an increase in lapse rate in the 9–24 km layer, but this is not so in the Northern Hemisphere, from which the data are relatively good.

Although the rocketsonde data show strong trends, these cannot be accepted as evidence of climate change because of the sparseness of the network and instrumentation problems. The representativeness of the network must be affected by their location in the western half of the Northern Hemisphere, and the data are biased by an inhomogeneity of several degrees Celsius (recently uncovered) caused by an instrumental alteration during the early 1970s.[6]

[6] K. Johnson, personal communication.

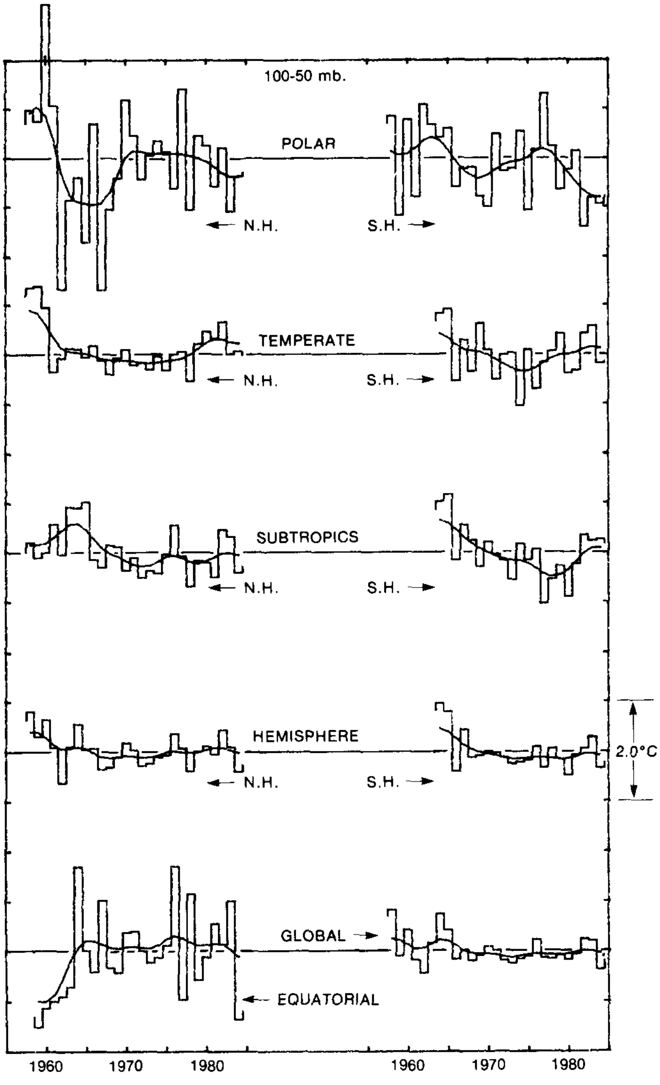

Figure 4.13. The 100–50 mb annual mean temperature fluctuations. Smooth curves were obtained by using a 10-year Gaussian filter.

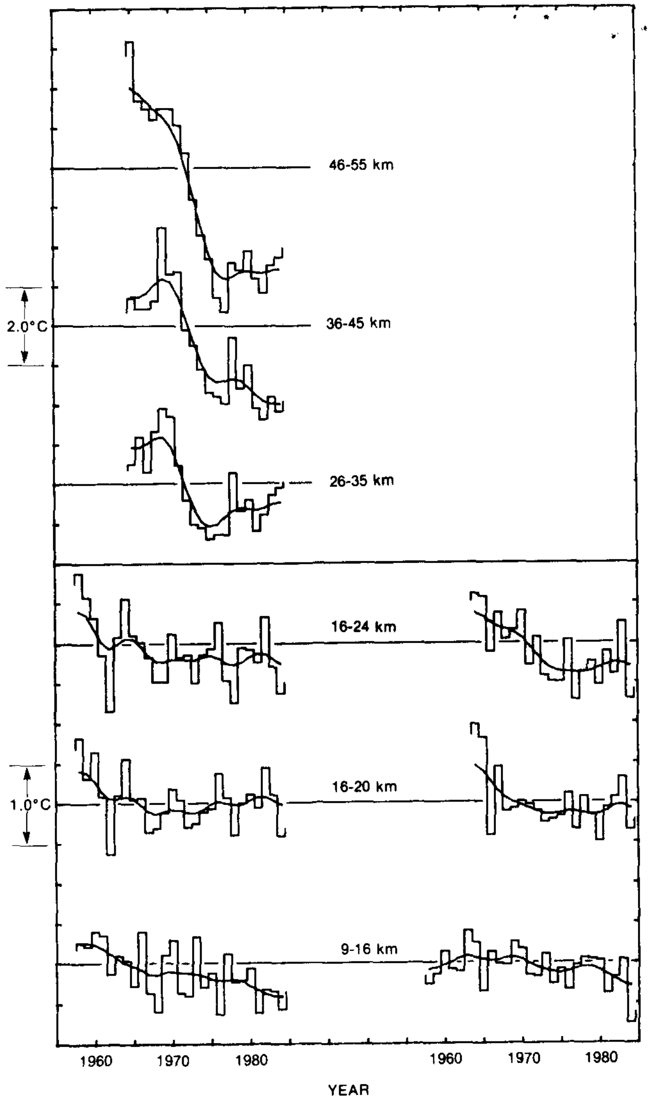

Figure 4.14. Stratospheric annual mean temperature fluctuations. Smooth curves were obtained using a 10-year Gaussian filter. (Note the difference in scale between the top and bottom halves of the figure.)

4.5 COMPARISON OF MARINE, LAND-BASED, AND FREE ATMOSPHERE TEMPERATURE CHANGES

No systematic comparison of the three independent temperature data sets (marine, land-based, and free atmosphere data) has been published, nor have the marine and land-based data yet been combined to give an optimum estimate of global mean surface air temperature variations (although a preliminary attempt has been made by Wigley et al. 1986). Although the latter exercise is possible with currently available data, there are problems present in the individual data sets that need to be resolved. Questions about data reliability, homogeneity, and representativeness have been raised and discussed in the preceding sections. Some aspects of these questions can be partly answered by comparison of the different data sets.

Before doing so, we present some simple heat budget arguments to provide a basis for discussing these comparisons. We consider the global climate system as four boxes: lower atmosphere boxes (up to 300 mb) over the land (29% of the global area) and the oceans (71% of the global area), an ocean mixed-layer box, and a land surface box. The total heat content of the boxes is

$$Q = \sum_{i=1}^{4} A_i C_i T_i, \tag{4.1}$$

where A denotes area, C denotes bulk heat capacity per unit area, and T is temperature. As a rough approximation the bulk heat capacities per unit area for the land and for the atmosphere boxes are equal and about 1/40th of the value for the mixed layer. Equation (4.1) therefore can be written as

$$Q = \frac{29}{100}\frac{AC}{40}\left[T_L + T_{AL} + \frac{71}{29}\left(40T_O + T_{AO}\right)\right], \tag{4.2}$$

where A is the total area of the globe, C is the mixed-layer bulk heat capacity per unit area, and the subscripts on the temperature terms are as follows: L, land; AL, atmosphere over land; AO, atmosphere over ocean; and O, ocean (i.e., mixed layer). The coefficient of the mixed-layer term far exceeds those of the other terms, so mixed-layer temperature changes effectively reflect changes in the heat content of the climate system below the tropopause. It is this heat content that we expect to be affected by an increasing atmospheric CO_2 concentration. Therefore, to detect CO_2 effects, we should look at changes in ocean mixed-layer temperatures. However, because of the large heat capacity of this element of the climate system, and because heat exchanges between the various boxes are relatively rapid on time scales relevant to CO_2 changes, we should expect temperature fluctuations on the CO_2 time scale to be similar in all boxes. In a sense, CO_2 forcing can be pictured as driving the ocean mixed-layer flywheel whose inertia drives the other large-scale components of the climate system. Quite large interannual time scale fluctuations can be expected to occur in the lower heat capacity boxes over and above this slower driving force, and these fluctuations constitute an element of the noise that makes CO_2 detection difficult (other forcing factors may contribute additional noise).

We can now relate the terms in Equation (4.2) to the available data. The term T_L can be identified with the land-based surface air temperature, T_{AL} and T_{AO} correspond to free atmosphere temperatures (850–300 mb data in this case), and T_O is reflected by both marine data sets (SST and NMAT). The available time series do not distinguish between T_{AL} and T_{AO}. In comparing these data sets, therefore, we expect to find quite high correlations reflecting the underlying common, slower (i.e., decadal or longer) changes, with these correlations being somewhat obscured by uncorrelated changes occurring on interannual time scales. The strength of these interrelationships will depend on the lengths of the data records and on the magnitude of slower time scale forcing effects relative to the interannual variability. These arguments apply not only to CO_2 forcing, but to any forcing that operates on time scales longer than those for exchange between the various boxes in our simple conceptual model.

We compare here four sets of annual data, hemispheric averages and high latitude averages for both the Northern and Southern Hemispheres. For the Northern Hemisphere as a whole we compare surface air temperature averages from the Angell and Korshover network, 850–300 mb averages from the same network, the land-based surface air temperature record of Jones et al. (1986), and the SST and NMAT data from Folland et al. (1984a). For the Southern Hemisphere, we compare Angell and Korshover's surface and 850–300 mb data with the land-based data of Hansen et al. (1981),[7] and NMAT and SST data from Folland et al. (1984a).[8] For the Arctic and Antarctic we compare surface and 850–300 mb data of Angell and Korshover with surface air temperature data from Kelly et al. (1982, updated) and Raper et al. (1984). For all these comparisons, product-moment correlation coefficients are given in Table 4.5, and visual comparisons are made in Figures 4.15–4.18.

Table 4.5 shows the strong correlations between the appropriate surface data of Angell and Korshover and data for the Northern Hemisphere land-based record ($r = 0.90$), NMAT ($r = 0.61$), the Arctic ($r = 0.69$), and the Antarctic ($r = 0.79$). Corresponding correlations for the 850–300 mb data are also quite high: Northern Hemisphere land, $r = 0.80$; Northern Hemisphere NMAT, $r = 0.80$; Arctic, $r = 0.48$; Antarctic, $r = 0.70$. For the Southern Hemisphere as a whole, however, the surface data of Angell and Korshover are poorly correlated with the data from Hansen et al. ($r = 0.42$)

[7] Ibid. Hansen.

[8] Additional unpublished data from D. E. Parker and C. K. Folland, personal communication.

Table 4.5
Correlation Coefficients Between Various Independent Annual Mean Temperature Records

	AK (trop)	AK (surface)	NMAT	SST
Northern Hemisphere				
J et al.	0.80(26)	0.90(26)	0.73(78)	0.62(80)
AK (trop)	1	0.71(26)	0.80(26)	0.69(26)
AK (surface)		1	0.61(26)	0.26(26)
NMAT			1	0.90(78)
Southern Hemisphere				
H et al.	0.82(26)	0.42(26)	0.65(77)	0.61(77)
AK (trop)	1	0.66(26)	0.81(26)	0.78(26)
AK (surface)		1	0.36(26)	0.41(26)
NMAT			1	0.91(78)
Arctic				
K et al.	0.48(26)	0.69(26)		
AK (trop)	1	0.49(26)		
Antarctic				
R et al.	0.70(26)	0.79(26)		
AK (trop)	1	0.77(26)		

Note: Sample sizes are given in parentheses. *Abbreviations:* J et al., Jones et al. (1986); AK, Angell and Korshover data as given in this chapter and in references cited in the text; trop, 850–300 mb data; NMAT, nighttime marine air temperature data from Folland et al. (1984a); SST, sea surface temperature from the same source; H et al., Hansen et al.'s (1981) Southern Hemisphere mean; K et al., Kelly et al. (1982) as revised and updated by Jones et al. (1986); R et al., Raper et al. (1984).

and the NMAT ($r = 0.36$) and SST data ($r = 0.41$); but the corresponding correlations with the 850–300 mb data are much higher ($r = 0.82$, 0.81, and 0.78, respectively). In general, marine data in both hemispheres are more strongly correlated with 850–300 mb data than with surface data, particularly the surface data of Angell and Korshover. For the 850–300 mb data and land-based surface data, correlations with NMAT data are generally higher than those with SST data. Correlations in the Northern Hemisphere are not noticeably different from those in the Southern Hemisphere.

These results are in accord with our expectations and provide support for the reliability and representativeness of all data sets. The lower correlations with SST data probably reflect the fact that SSTs are further removed (in both a physical process and direct spatial sense) from the air temperature records, although it may indicate some additional uncertainty in the SST data compared with the NMAT data. Larger corrections have been applied in producing the homogenized SST time series than is the case for the NMAT series (Folland et al. 1984a). The lower correlations with the surface data of Angell and Korshover imply that the existing network is inadequate for establishing large-area averages of surface air temperature, but the network appears to be satisfactory for the free atmosphere. This is in accord with the observation of Angell and Korshover (1983a) that column mean temperatures are more spatially coherent than surface temperatures.

We have not attempted to assess the statistical significance of these correlations because any significance tests would be strongly influenced by trends in the data. As the results are in accord with our a priori expectations, they provide good evidence of the reliability of the basic temperature data sets, at least back to the early years of the 20th century. Prior to about 1900 there are discrepancies between the marine and land-based data that require explanation (see Figure 4.7). Jones et al. (1986) have attempted to assess the homogeneity of their land-based data series. They conclude that there is little likelihood of a substantial trend going undetected as a result of their limited 19th century data coverage. This suggests that either there are inhomogeneities in the marine data (SST and NMAT) or, if the land-marine difference is real, that the late 19th century climatic regime was radically different from that of the 20th century. However, the possibility of an undetected trend in the land-based data cannot yet be entirely eliminated. If the cooling trend evident in the marine data is real, this would imply the existence of some strong (non-CO_2) forcing factor whose effects are sufficiently large that, if they occurred in the future, they might obscure the effects of increasing CO_2 well into the 21st century.

The similarity between marine and land-based data in the 20th century does not confirm the earlier, qualitative comparison of marine and land data made by Paltridge and Woodruff (1981). These authors produced a global mean temperature time series based largely on SST data and compared it with the land-based record of Mitchell (1963). The SST data showed a wider range of variation than the land data and appear to lag some 20 years behind the land data. These results must be viewed with considerable skepticism for several reasons. First, the large amplitude of the SST fluctuations in part may be an artifact of the way Paltridge and Woodruff produced their annual averages; they averaged only summer and winter data. Second, their data were not systematically corrected for instrumental changes. As noted earlier in this chapter, these corrections are of a similar magnitude to the observed interdecadal temperature fluctuations and cannot be ignored. Third, there is no physical reason to expect variations in large-scale average SSTs to lag decades behind land-based surface air temperature.

The lag idea can be confusing. It is true that the transient response to changing CO_2 level lags decades behind the instantaneous equilibrium response (Hoffert and Flannery 1985), but this is merely a way of looking at the degree of disequilibrium of the climate system. Exchange between the various components of the climate system occurs on a relatively short time scale. Energy balance model calculations show that ocean-midcontinent lags can be at most a few years, a result that is also apparent in the general circulation model (GCM) calculations of Bryan et al. (1982, their Figure 2). Because all land-based temperature records contain data from the large coastal fringe zones of the continents, where land-sea interchanges are rapid, one would not expect to see any appreciable lag between marine and land-based temperature fluctuations. This fact is apparent in Figures 4.15–4.18.

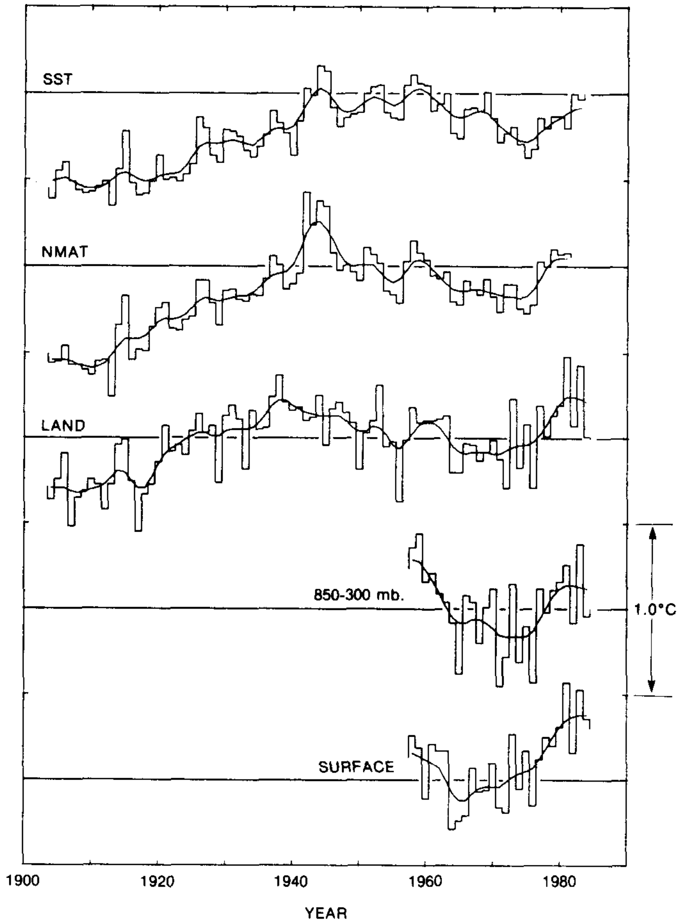

Figure 4.15. Comparison of independent estimates of temperature fluctuations for the Northern Hemisphere back to 1904. For details of sources see Table 4.5. Smooth curves were obtained by using a 10-year Gaussian filter.

4.6 THE USE OF TEMPERATURE DATA IN DETECTION STUDIES

4.6.1 Introduction

Early detection studies are directed toward isolating, as soon as possible, those parts of observed climate fluctuations that can be unequivocally attributed to the effects of increasing atmospheric CO_2, that is, the CO_2 signal. (Note that the effects of other trace gases are not readily separated from those of CO_2, so CO_2 and these trace gases probably need to be considered together.) Detection of the CO_2 signal involves two steps. The first is a statistical analysis of data directed toward identifying, at a known confidence level, a change in one or more climatic variables. The second is the attribution of at least part of this change to increasing CO_2 concentrations. The first step has been expressed by some authors in terms of the signal-to-noise ratio. All climate variables show considerable natural variability on all time scales, and this variability constitutes the background noise against which a potential CO_2 signal is to be detected. When an *observed* signal (i.e., a change in some climate variable) becomes sufficiently large relative to the noise (i.e., the signal-to-noise ratio is sufficiently high), then it might be claimed that the signal has been detected at some stated level of significance. This approach has been applied by Madden and Ramanathan (1980) and Wigley and Jones (1981, 1982) to surface air temperature data to determine the most appropriate data to use in detection by comparing the predicted CO_2 signal to the observed noise.

In choosing a climate variable or parameter to monitor for detecting CO_2 effects, it is essential to

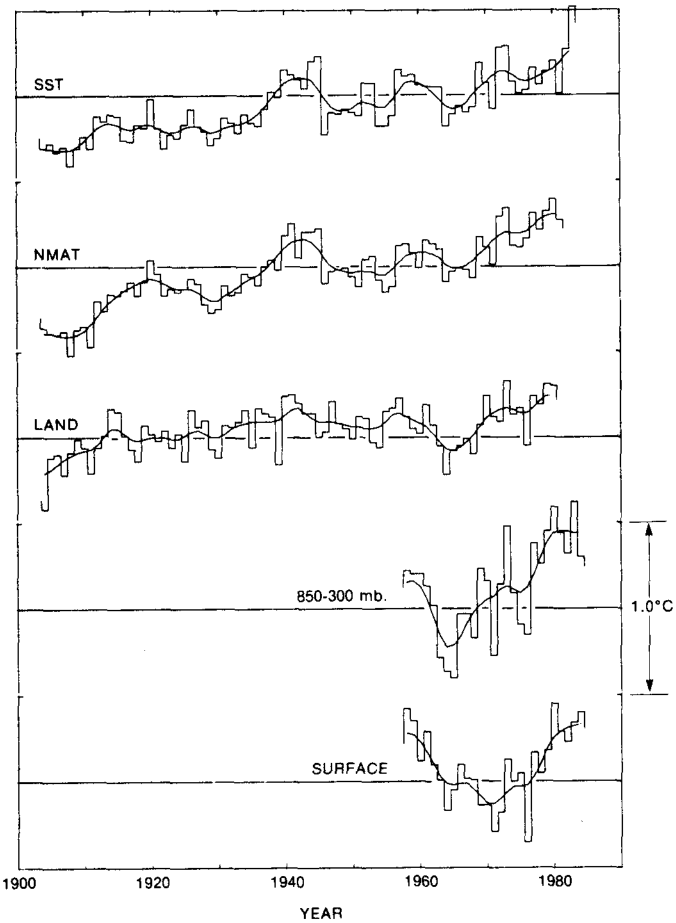

Figure 4.16. Comparison of independent estimates of temperature fluctuations for the Southern Hemisphere back to 1904. For details of sources see Table 4.5. Smooth curves were obtained by using a 10-year Gaussian filter.

have both a well-defined predicted signal and a well-defined noise level. The present state of the art of numerical modeling controls the choice of possible signals. Thus, climate parameters such as global mean annual surface air temperature and annual mean stratospheric temperature are possible signal choices, whereas parameters that depend on the regional details of climate or on poorly understood physical links with the climate system are inappropriate choices. Definition of the noise level is primarily dependent on the quality and amount of data available. Data series must be of sufficient length to allow appropriate statistical tests to be applied and to ensure that noise levels are well defined on time scales relevant to the possible effects of CO_2. Very few climate variables have a record long enough to be able to establish the lower frequency characteristics of their natural variability.

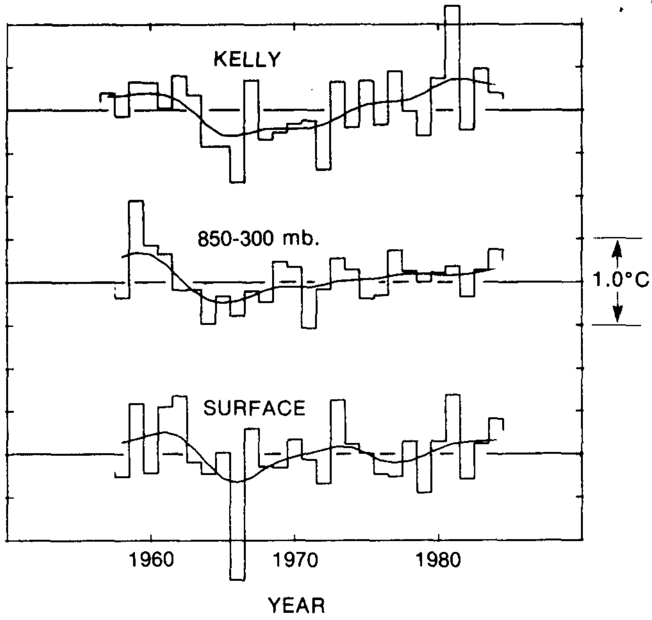

Figure 4.17. Comparison of independent estimates of temperature fluctuations for the Arctic. For details of sources see Table 4.5. Smooth curves were obtained by using a 10-year Gaussian filter.

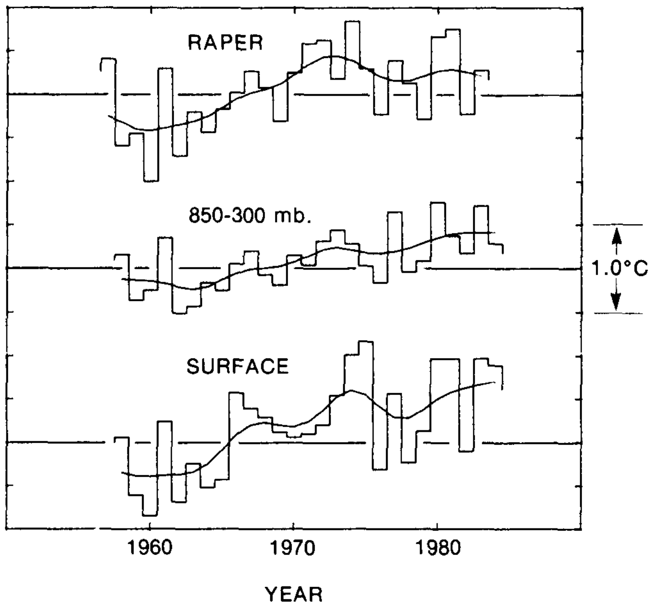

Figure 4.18. Comparison of independent estimates of temperature fluctuations for the Antarctic. For details of sources see Table 4.5. Smooth curves were obtained by using a 10-year Gaussian filter.

The only variable that satisfies both of these criteria at present is large-scale area average surface air temperature. (SST data convey much the same information over the oceans.) Lower tropospheric temperatures may be more representative of CO_2-induced changes, being less subject to local, near-surface anomalous effects (urban heat islands, surface inversions, etc.), but these data are less useful for detection purposes because of their limited length of record (less than 40 years) and because of poorer spatial coverage compared with surface air temperature data. These difficulties also apply to stratospheric temperature data and to derived parameters such as the stratosphere-lower troposphere temperature difference (cf. our discussion below in Section 4.6.4 of Epstein's [1982] suggestion that the stratosphere-troposphere temperature difference would be a more sensitive detection variable than would either temperature separately).

If model results are correct, the rise in CO_2 concentration since the middle of the 19th century should already have caused an appreciable global mean surface warming. The amount of CO_2-induced warming depends on the preindustrial initial CO_2 level, the size of the model-predicted equilibrium signal for a CO_2 doubling (ΔT_{2x}), and the damping of the response resulting from oceanic thermal inertia effects. Simple energy balance model calculations (Siegenthaler and Oeschger 1984; Wigley and Schlesinger 1985a; see also Hoffert and Flannery 1985) suggest that temperature changes in response to increasing CO_2 levels may lag some 10–60 years behind the instantaneous equilibrium response, that is, that the present climate may be far from equilibrium with respect to the prevailing CO_2 concentration. Weller et al. (1983, their Figure 5.5) tried to account for this lag effect in a rather ad hoc way and concluded that, if the globe had warmed by 0.5°C since 1881 (which is the best estimate based on current observational evidence), then this implied that there would be a CO_2-doubling temperature change in the lower end of the range 1.5–4.5°C. This conclusion is at odds with the latest GCM results, which give values of about 4°C (see Hansen et al. 1984; Washington and Meehl 1984; Schlesinger et al. 1985; and discussion in Schlesinger and Mitchell 1985). However, if the lag effect is accounted for more rigorously, it may be possible to resolve this conflict. This subject is an area of active research interest (see Hoffert and Flannery 1985), and the results of various groups still differ.

As an example of how this difference between recent model results and observations may be resolved, Figure 4.19 shows the estimated 1850–1980 temperature change due to increasing CO_2 as a function of the initial CO_2 level and ΔT_{2x}. This figure is based on a transient response analysis using the simple box-diffusion climate model of Wigley and Schlesinger (1985a). The two diagrams span the range of uncertainty arising from uncertainties in the degree of disequilibrium, which in turn arise from uncertainties in the mixed-layer depth and in the mixing processes below the mixed layer. For an initial CO_2 concentration in the range 260–280 parts per million by volume (ppm) (World Meteorological Organization 1983; Wigley 1983; Siegenthaler 1984) and ΔT_{2x} in the range 1.5–4.5°C, the expected temperature change, ΔT(1850–1980), lies in the range 0.26–0.96°C. Other trace gas effects could raise this range by one or two tenths of a degree Celsius (see Wigley and Schlesinger 1985a). Given that other forcing factors (such as volcanic eruptions and changes in solar irradiance) might cause temperature changes of at least a few tenths of a degree Celsius, and given recent ice core evidence suggesting that the 1850 CO_2 level may be higher than the upper limit assumed here (Neftal et al. 1985), observed temperature changes must be judged compatible with the probable range of equilibrium doubling temperature changes recommended by the Carbon Dioxide Assessment Committee (National Research Council 1983). However, to say that observed temperature changes may be compatible with the CO_2 hypothesis does *not* mean that we can claim to have detected the CO_2 signal and therefore to have verified model predictions.

The difficulty in unequivocally detecting a CO_2 signal arises mainly because of the medium to long time scale (on the order of decades or more) noisiness of the past temperature record and because of our inability to explain these fluctuations adequately. Marked hemispheric-scale temperature fluctuations in both directions on all time scales have been observed since the data first allowed reliable estimates to be made. Perhaps the most striking of these fluctuations is the post-1940 cooling episode. Between about 1940 and 1970, the Northern Hemisphere cooled by approximately 0.2°C (see Figure 4.15), contrary to the predicted CO_2-induced warming effect. In some data sets this cooling is less

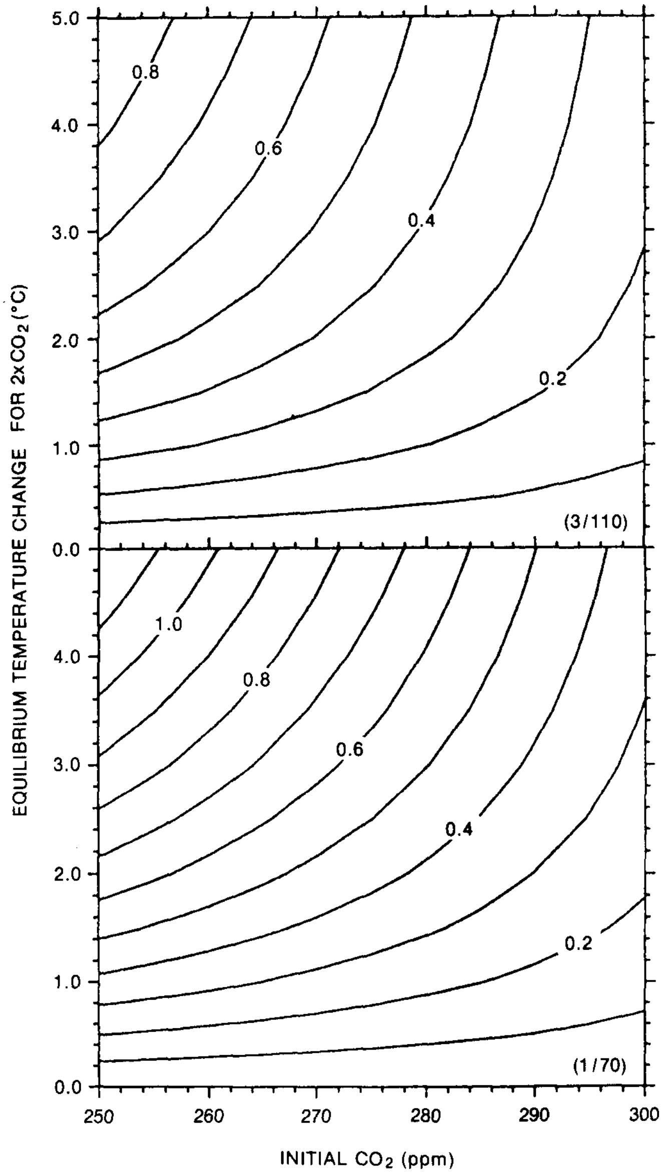

Figure 4.19. Surface temperature changes between 1850 and 1980 allowing for oceanic thermal inertia effects (from Wigley and Schlesinger 1985a). The upper diagram shows values based on a sub-mixed-layer diffusivity of 3 $cm^2\ s^{-1}$ and a mixed layer depth of 110 m. The lower diagram gives values for diffusivity of 1 $cm^2\ s^{-1}$ and a mixed layer depth of 70 m. These two diagrams therefore span the likely range of uncertainty due to uncertainties in these two ocean parameters. The 1980 CO_2 concentration was taken as 338 ppm. Trace gas effects (CH_4, N_2O, etc.) were not included.

pronounced than in others; for example, in the recent reanalysis of land-based data for the Northern Hemisphere by Jones et al. (1986) (cf. Figures 4.1 and 4.2). This cooling episode is also less obvious in Northern Hemisphere SST data, with these data showing a cooling beginning around 1960 and ending in the mid-1970s (Figure 4.15; a result noted earlier by Chen 1982); and it shows up only as a pause in a longer time scale warming trend in the Southern Hemisphere (Figure 4.16). Fluctuations like this are examples of the type and magnitude of unexplained medium time scale noise against which a CO_2 signal must be detected.

4.6.2 Detection Strategies

Two different approaches to detection may be distinguished. First, we may seek to detect either a significant change in the mean value of a given parameter or a significant trend in a given parameter over a specified time interval (and then attempt to attribute some part of this change to the effects of increasing CO_2 concentration). Alternatively, regression techniques or simple models may be used to associate a statistically significant part of past variations of a parameter with past variations in atmospheric CO_2. The first method is that employed by Madden and Ramanathan (1980) and has been explained in simpler terms by Wigley and Jones (1981, 1982). From a purely statistical viewpoint, detecting a trend is closely related to detecting a change in the mean, although the appropriate statistical test (a *t*-test) is slightly more efficient in the former case. The second method is equivalent to reducing the noise level, therefore enhancing the signal-to-noise ratio. A number of authors (see Section 4.6.5) have attempted to do this in works which were not necessarily aimed at detection of CO_2 effects per se, but toward explaining recent changes in global mean temperature. The technique is to identify part of the record with other causal factors, therefore reducing the noise by factoring out these effects. The method is, however, fraught with statistical difficulties and is considerably hampered by our current poor understanding of the processes that cause decadal and longer time scale climate fluctuations, by the relatively poor quality of the records of past forcing factors, and by uncertainties in the large-scale observational temperature record.

4.6.3 Simple Signal-to-Noise Ratio Studies

The analyses of Madden and Ramanathan (1980) and Wigley and Jones (1981, 1982) essentially assume that all observed variations are noise and that this noise is random with a simple first-order autoregressive structure. The signal to be detected is a change in mean temperature, and the observed noise level is taken to be the square root of the variance of the sampling distribution of the mean. For data with no serial correlation, this is given by $s(\overline{X}) = \hat{s}/\sqrt{N}$, where $\hat{s}$ is the observed standard deviation and $\overline{X}$ the mean temperature over a time period with a length of N years. However, since the temperature data are autocorrelated, allowance must be made for this, with $s(\overline{X})$ being appropriately inflated. Madden and Ramanathan (1980) have achieved this using a frequency-domain approach, whereas Wigley and Jones (1981, 1982) have used a time domain approach.

Madden and Ramanathan were primarily concerned with identifying a significant signal in the observational temperature record around 60°N. They concluded that the signal cannot yet (i.e., in 1980) be detected. Wigley and Jones were concerned with choosing the right combination of season and latitude band to maximize the ratio of theoretical signal to observed noise. They found the highest ratio for summer, midlatitude temperatures; but, given the uncertainties in both signal and noise levels, all seasons and latitude bands were similar. They concluded that the best detection variable is probably the mean annual temperature averaged over as large an area as possible. Wigley and Jones also concluded that the CO_2 signal cannot yet be unequivocally detected in the available data.

There are a number of problems with this approach which these authors do not address. First, the assumption of a first-order autoregressive (AR1) noise structure is difficult to justify. Although the lag-1 autocorrelation (r_1) can easily be calculated, its value can only be partly the result of the autoregressive nature of internal climatic fluctuations. A considerable contribution to r_1 may come from medium to long time scale trends that are not stochastic but that have deterministic causes. Second, and closely related to this, the lag-1 autocorrelation coefficient itself is nonstationary and varies considerably, depending on which part of the time series is examined. This is a frequent property of climatological time series and is particularly evident in the Northern Hemisphere surface air temperature series (see, for example, Jones et al. 1982, their Table 2).

Rather than look at changes in mean temperature, an alternative (but basically equivalent) approach would be to examine trends in large-scale, area average temperatures. As noted above, the statistical test for a trend is effectively the same as testing for a significant change in the mean, although it is slightly more efficient (in the sense that, all other factors being equal, less data are required to establish a significant trend than to establish a significant change in the mean). No comprehensive exploration of this approach has been published. Detection of a trend that could be definitely related to increasing CO_2 would depend on the time interval over which the trend was measured and on the chosen upper limit for the natural (non-CO_2) rate of warming. This approach also has problems. Consider, for example, the CO_2-induced trends expected around the year 2000. For ΔT_{2x} in the range 1.5–4.5°C, these range between 0.12–0.29°C per decade, based on the transient response model of Wigley and Schlesinger (1985b, their Figure 2; see also Hansen et al. 1984, their Figure 18). (Note that the magnitude of the expected trend does not change much with time, although this result is sensitive to the assumed future CO_2 scenario.) Natural trends in the past have been as much as 0.2°C per decade over 20-year periods, so it would still be possible to deny the existence of a CO_2 effect even if the observed trend were as much as this by resorting to arguments about data representativeness or by invoking other causal factors. To eliminate the former requires better data coverage on a global basis. To eliminate the latter requires better understanding of the causes of climate change in general and a better knowledge of past and future variations in the forcing parameters. Alternatively, the warming trend would have to be observed over a period well in excess of 20 years.

4.6.4 Comparison of Tropospheric and Stratospheric Temperatures

Epstein (1982) has shown statistically that, if the data were reliable and representative, the joint behavior of the temperature in the troposphere and the stratosphere may be a more sensitive indicator of climate change (natural or anthropogenically induced) than is the surface temperature by itself. Unfortunately, both surface and upper atmosphere data have quality problems, and there are noninstrumental errors or uncertainties arising from limited spatial coverage that differ between data sets. Furthermore, the record length for tropospheric and stratospheric data is too short to be able to define the level of natural variability on decadal and longer time scales. Epstein (1982) does not consider decadal time scale noise, yet it is clear from Figure 4.12 that there are fluctuations on these time scales that are inconsistent with the hypothesized effects of an increasing CO_2 concentration. Because of these data quality and record length difficulties, application of the relatively sophisticated technique proposed by Epstein does not yet appear practical. Separate analyses of surface, tropospheric, and stratospheric data are more informative and provide more detailed insights into the complexities of climate change.

Nevertheless, Parker (1985b) has followed Epstein's lead in presenting a careful analysis of data from different levels in the free atmosphere. On the basis of model indications that there should be a relatively large CO_2-induced warming at 300 mb and a cooling at 50 mb, he examined 50 mb minus 300 mb temperatures at 27 extratropical stations in a search for these differences (cf. our discussion of lapse rate changes in Section 4.4.3). Extratropical stations were used to minimize the influence of the quasi-biennial oscillation (see Quiroz 1983). Parker found little evidence of a significant negative trend at these stations over the past 30 years and concluded that there was no evidence of a CO_2 effect.

Both of these studies (Epstein 1982; Parker 1985b) may be considered as simple examples of the fingerprint approach to detection (see Chapters 1 and 8 of this volume). It has been suggested that detection would be facilitated by developing "a unique CO_2-specific 'fingerprint' for the CO_2 response involving a set of parameters, distinctive from responses that would be caused by all other known influences, and to search for this correlated pattern of changes, not just for a change in one isolated parameter" (MacCracken and Moses 1982, p. 1172). The tropospheric-stratospheric temperature combination is a simple two-element fingerprint. Apart from the potential statistical advantages pointed out by Epstein, such a fingerprint could help in the attribution phase of detection since it may allow us to distinguish between solar and CO_2 forcing, although it would not allow CO_2 effects to be readily isolated from, for example, the effects of a changing stratospheric aerosol concentration. Unfortunately, the difficulties in detecting the effects of changing CO_2 using a single climate variable also apply to the fingerprint method, but they are additionally magnified because the uncertainty inherent in a multivariate fingerprint must be at least as great as that of its most poorly defined part.

4.6.5 Noise Reduction Studies

It is clear from signal-to-noise ratio arguments that detection will be hastened if the noise level can be reduced. The primary need is to explain past decadal time scale fluctuations in large-scale, area-average temperatures convincingly. The basic noise reduction approach, therefore, is to attempt to relate part of the past variations to specific forcing factors, remove these effects, and consider that which remains as residual noise. This approach has been used indirectly by many authors in works that were not always directed toward detecting the effects of CO_2 on climate (Bryson and Dittberner 1976; Miles and Gildersleeves 1977, 1978; Hoyt 1979a; Robock 1979; Bryson and Goodman 1980; Hansen et al. 1981; Gilliland 1982; J.M. Mitchell 1983; Gilliland and Schneider 1984). These studies range from the purely empirical to strongly model-oriented work. Most of them support the realism of theoretical estimates of the magnitude of CO_2-induced warming, but none provides statistically convincing evidence that such a warming has already occurred. Gilliland (1982) and Gilliland and Schneider (1984), in particular, stress the lack of statistical significance in their results.

In contrast, Vinnikov and Groisman (1981, 1982) are more directly concerned with detecting CO_2 effects, and they state that this exercise has been successful, not only in detecting CO_2-induced global warming, but even in identifying the spatial and seasonal patterns of this warming. These conclusions are statistically unconvincing for reasons that will be outlined below.

To illustrate the statistical problems involved, consider a direct regression approach. Suppose global or hemispheric mean surface air temperature change can be expressed in the form

$$T_i = aV_i + bS_i + cC_i + \beta_i + e_i \qquad (4.3)$$

where V, S, and C are past variations in volcanic aerosol loading, solar irradiance, and atmospheric CO_2 level (in year i), respectively; β_i is an autoregressive term [i.e., $\beta_i = f(T_{i-1})$]; e_i is a residual error term; and a, b, and c are regression coefficients. (Note that there may be forcing factors other than those included in this equation. Internal factors related to changes in the atmospheric or, more importantly, the oceanic circulation could be significant. Ocean circulation effects represent a particularly glaring gap in our present knowledge.) Fitting an equation like this to observed temperatures produces estimates of a, b, and c with associated confidence limits. If the CO_2 coefficient c is significantly greater than zero, this might be taken as proof of a significant CO_2 effect. It is, however, a difficult task to assign confidence limits to these regression coefficients, and standard formulae are inappropriate for two reasons.

First, the forcing functions V, S, and C are not well defined (nor, as has been noted earlier, is the response variable T a completely reliable measure of global or hemispheric temperature change). For example, the volcanic forcing functions used by Vinnikov and Groisman (1981, 1982), Hansen et al. (1981), and Gilliland (1982) are quite different. Vinnikov and Groisman use Pivovarova's (1977) actinometric data on atmospheric transmissivity (P), Hansen et al. use Lamb's (1970) dust veil index (L), and Gilliland uses the ice core acidity record of Hammer et al. (1980) (H). The correlations between these three volcanic forcing functions are quite small: $r_{PL} = -0.46$, $r_{PH} = -0.32$, and $r_{LH} = 0.41$, with all being statistically significant at the 5% level, but with little variance in common. The solar and CO_2 forcing functions used by these authors are also quite different (Vinnikov and Groisman have no solar term, although a solar effect may be implicit in their transmissivity data). Not only do the assumed CO_2 forcing functions differ from author to author, but all three studies used an initial value at the high end (or above) the currently accepted range of possible values. Finally, in none of these modeling studies is the transient response of the system adequately modeled.

All three studies obtain good fits between modeled and observed temperature variations (results from Hansen et al. [1981] and Gilliland [1982] are shown in Figure 4.20 as examples). However, as pointed out by MacCracken (1983), it is impossible for essentially the same temperature fluctuations to be correctly explained by three different sets of forcing function records. The only way that the results can be compatible is for all to be subject to considerable statistical uncertainty, that is, to have wide confidence bands for the regression coefficients and for the total explained variance.

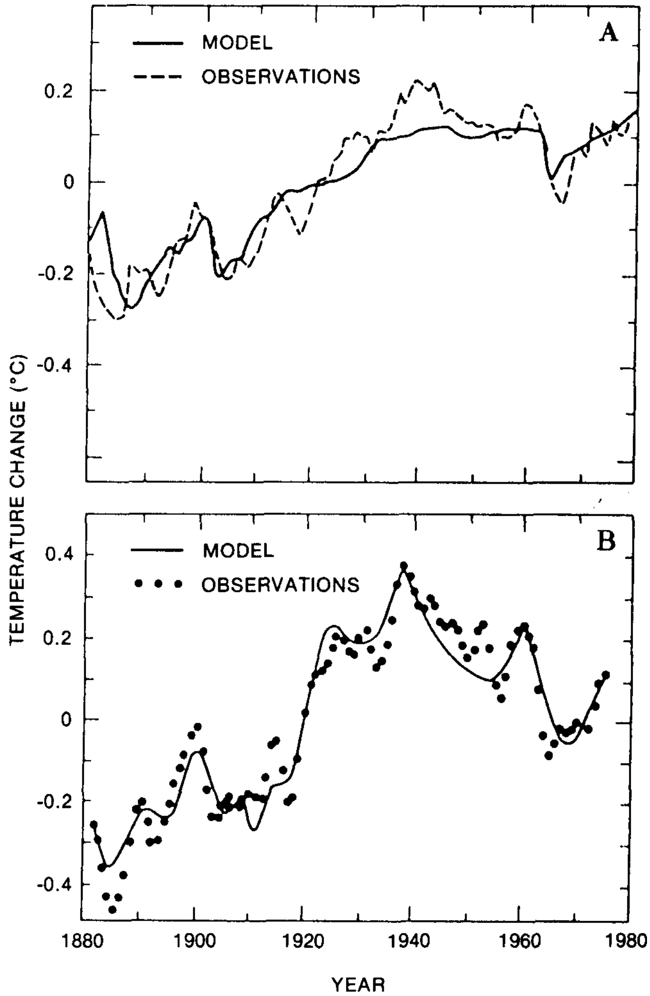

Figure 4.20. Comparison of the observed change in surface air temperature and model predictions of the change in temperature due to the increase of CO_2 concentration and changes in solar irradiance and stratospheric aerosol loading. (A) from Hansen et al. (1981); (B) from Gilliland (1982).

The second reason why confidence levels are difficult to assign is purely statistical. Because both the response and predictor variables show medium to long time scale trends, a good fit could be obtained if some of these trends happened to match by chance. For example, the umbral/penumbral solar index ratio (Hoyt 1979a, 1979b) used by Hansen et al. (1981) has rising and falling trends either side of (approximately) 1940 which are similar to those in the temperature record used by these authors. This could be a chance match, but it is sufficient to ensure a good correlation between T_i and S_i. Building an arbitrary and variable time lag into any relationship (as in Gilliland's analysis) introduces multiplicity[9] and further reduces the statistical significance of any results.

The central problem here is that the number of degrees of freedom that should be used in testing significance cannot easily be determined. Standard methods for accounting for autocorrelation (e.g., Bartlett's method; see Quenouille 1952) may not be applicable. To compound the statistical problems, on medium to long time scales all three forcing variables can show similar low-frequency fluctuations, making it impossible to assign a causal role to any particular variable.

Given these difficulties, the noise reduction method can, at present, only give supporting evidence in the detection problem. In principle, part of the low-frequency variance could be removed if it could be ascribed confidently (and with *known* confidence) to other causes, but, because present knowledge of non-CO_2 climate forcing mechanisms is as poor as or poorer than our understanding of CO_2 effects, a statistically rigorous application of this method would be extremely difficult.

While it is clearly a difficult task to reduce noise levels by factoring out the effects of external forcing factors other than those of CO_2, reduction of internally generated noise is possible in some cases. For example, Wigley and Jones (1981) suggested that the influence of the El Niño/Southern Oscillation phenomenon could be removed from some data series, and we have noted already that this factor can be an important contributor to climate change. Parker (1985a) has examined tropical tropospheric temperatures (20°S–20°N, 850–200 mb) over the period 1950–1983 in an attempt to isolate the effects of the volcanic eruptions of Agung and El Chichón. When the influence of the Southern Oscillation is removed, he finds no statistically significant cooling that can be ascribed unambiguously to either event. (There is, however, convincing evidence that volcanic eruptions cause tropospheric cooling in extratropical regions; see Kelly and Sear 1984.) A similar study of stratospheric temperatures (30 mb station data near the Equator and zonally averaged data for 15°–45°N) has been made by Quiroz (1983). To isolate the effects of El Chichón, Quiroz removed the effects of the Southern Oscillation, the quasibiennial oscillation, and dynamically produced temperature changes associated with sudden warming episodes in the high-latitude stratosphere. He found a substantial residual 30 mb warming of 1°–3°C between 0°–35°N attributable to the eruption. These techniques are applicable to the detection of the climatic effects of an increasing CO_2 concentration, but they may be of secondary importance because they relate to limited areas and to relatively short time scales.

[9] If a large number of comparisons is carried out (i.e., for different lags), then the probability of obtaining a significant match by chance must increase.

4.7 SUMMARY AND RESEARCH RECOMMENDATIONS

Unequivocal, statistically rigorous detection of the effects of changing CO_2 levels on atmospheric temperatures is not yet possible. There is a dichotomy between our ability to model temperature changes and our knowledge of past changes. Model results are, on the one hand, most reliable for large-scale averages, while the observational record, on the other hand, is most securely established only over relatively small regions. Furthermore, even on the largest spatial scales (global mean values) our understanding of the causes of climate change is still rudimentary. In this chapter we have tried to give both an objective and critical account of the present state of knowledge of the observational record and some insight into the confidence with which available data can be used to evaluate model simulations of climatic change. The various points discussed lead us to the following conclusions and recommendations.

1. More extensive data coverage is required. While it is possible to estimate global mean surface air

temperatures, it must be stressed that such estimates incorporate uncertainties in the available data and uncertainties arising from substantial gaps in coverage (especially in the Southern Hemisphere). The estimated global mean warming of 0.5° over the last 80–100 years is probably only reliable to within ± 0.2°C. Uncertainties still exist, and it should be a high priority item to make full use of satellite data to extend data coverage. However, this will require considerable effort in establishing and maintaining ground truth and in calibrating the satellite observations against surface instrumental data.

2. Further examinations and analyses of existing surface and upper air data are required to improve the quality of these data and to establish more carefully the large-scale spatial representativeness (or otherwise) of limited-coverage data sets. Assessments of spatial representativeness may be aided by the use of general circulation model simulations of present-day climate. Intercomparisons of different data sets, perhaps guided by relatively simple climate models, can help in assessing data reliability and in understanding the causes of past climate change.
3. Explaining the decadal and longer time scale fluctuations in the past record is a major problem in detecting the climatic effects of changing CO_2 and other trace gas concentrations. An example is the unexplained 1940–1970 (approximately) cooling. Attempts to empirically model such changes have, to date, suffered from a number of deficiencies, as discussed in detail in Section 4.6.5. Until such fluctuations have been adequately explained, claims regarding detection of CO_2 effects can be easily criticized.
4. The influence of changes in ocean circulation (vertical and horizontal) on climate is poorly understood and has not been considered in model simulations of past climate change. Interhemispheric differences in decadal and longer time scale temperature trends (see Figure 4.7) may provide clues to the relative importance of this regionally specific internal forcing factor, and further research in this area must be of high priority.
5. Improved theoretical estimates of the regional details of CO_2-induced climate change are required. Uncertainties in model predictions of global mean temperature changes due to transient response effects were mentioned briefly in section 4.6.1. New and more taxing problems are introduced when attempts are made to account for transient response effects on a regional scale.

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5. LONG-TERM CLIMATE CHANGE IN OBSERVED PHYSICAL PROPERTIES OF THE OCEANS

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CONTENTS

5.1. INTRODUCTION

After the Sun, the oceans play, perhaps, the key role in the climate system of the Earth. This is so because they directly or indirectly supply most of the energy (heat) needed to keep the atmosphere running. They also provide the moisture that subsequently falls over the land masses and makes agriculture and life as we know it possible. The long time constant of the oceans and their moderating effects ensure a climate system that is more or less stable and therefore conducive to the development of highly organized social and industrial systems. If any of these three factors (energy, moisture, and stability) were not provided by the oceans, our world would be strikingly different.

It is no wonder that people are concerned with the projected changes in the physical properties of the ocean that may be induced by an increasing concentration of carbon dioxide (CO_2). Will they change the way in which the oceans force the atmosphere? Will they change the distribution of rainfall over the planet? Will they cause a shift to another climate regime that may not be favorable to society? For instance, a substantial atmospheric warming would warm the upper ocean, and melt landlocked sea ice, glaciers, and so forth. The effect would be a rise in global sea level that, if large enough, would cause havoc in the coastal margins of the world.

To determine whether increasing amounts of CO_2 are changing key ocean properties, it is necessary to know two things: (1) There must be a priori theoretical predictions of the oceanic changes expected from increased CO_2 levels. Aside from obvious, qualitative statements about warming of the near-surface waters, such predictions have yet to be made. Thus, one is left with a discussion of the background noise levels in the ocean data set over which any signal must be detected. (2) With this latter point in mind one must know the past history of key ocean variables. Unfortunately, the oceans have been, and still are, imperfectly monitored. The average global distribution of surface temperature, salinity, and sea level are known, but in many areas nothing is known of the secular and shorter term temporal variations of these parameters. Yet, it is just such information that is needed to detect any CO_2-induced changes.

In addition, far less is known about changes in the temperature and salinity of the oceans' interior. The first major expedition designed expressly for scientific study of the oceans was aboard the H.M.S. Challenger which explored all the world's oceans between 1872 and 1876. However, it was not until the 1920s and 1930s that physical measurements of the oceans' interior began in a systematic manner. Even then the number of observations was limited. Only after World War II did relatively extensive hydrographic surveys begin in earnest. These latter surveys constitute the majority of the pitifully small set of interior ocean observations that we have today.

It is true that there are certain specific locations or small areas where relatively long time series of ocean measurements exist. This is especially true for sea surface temperature (SST) and sea level (SL). Unfortunately, local effects generally tend to dominate the variability at these locations. Yet, for many purposes they are all we have and so provide the bulk of the information on long-term changes in the oceans. In summary, changes in key ocean variables (temperature, salinity, sea level) have been poorly documented over space and time. The best we can do is to select data from the few stations and regions where they are most voluminous and use that information in a CO_2 detection and model verification program.

The following sections of this chapter discuss each of the key variables noted above, examining in turn their importance to the climate system and society, their measurement problems and the availability of data, and their time variability over recent decades. It is with such information that schemes aimed at detecting possible CO_2-induced effects must be developed. The focus of this chapter is intended to be rather narrow. Therefore, the ensuing discussion does *not* touch on the possibility of using geological or chemical data to trace and describe climate change in the oceans. Each of these subjects is covered extensively in the literature and offers promising avenues for detecting climate change. Note also that this chapter does *not* deal with theoretically expected changes in the oceans that are projected to result from an increasing CO_2 concentration. Nor does it cover the status of global ocean models or other expressions of theoretical understanding that might be applied to the

CO_2 detection problem. Again, each of these items is worthy of a chapter in its own right.

5.2 MEAN SEA LEVEL

5.2.1 The Importance of Sea Level Variations

Of any ocean variable, a substantial increase in relative sea level (RSL) would have the most dramatic impact on society. (e.g., National Research Council [NRC] 1983; Hoffman et al. 1983). A large fraction of the human race lives near coastal margins. Furthermore, the industrial structure of the world depends on ship-borne transport and the worldwide system of sea ports. A substantial rise in RSL, therefore, would cause serious displacements of populations plus, if large enough, drown most of the world's harbors. The serious ramifications of even a 50–70 cm rise in RSL have been documented (NRC 1983; Titus et al. 1983). This change is only a small fraction of the approximate 100-m change in sea level that has occurred since the last glacial epoch. By contrast, it appears that sea level has risen around much of the world by between 10 and 25 cm since the early 1900s (Barnett 1983a). Changes of RSL of this size, and occasionally much larger, occur regularly because of storms, changes in temperature, pressure, and so forth. For example, the 1982–1983 El Niño event caused the sea level along the west coast of North America to be raised by 20–30 cm above its normal level for many months. Such variability represents a "noise" with which one must reckon in any attempt to detect a "CO_2-signal."

5.2.2 Measurement Methods

Sea level has been measured by a variety of methods over the years. Before the 1900s, observations were sometimes taken visually from a graduated staff. In modern times, the methods are more rigorous, but are still subject to many potential problems (cf. Barnett 1983a, 1983b). In general, the position of a float, or other sea surface sensing device, is recorded and referenced to a geodetic standard, hence the term "relative sea level." In many cases the SL data are recorded in graphical form and later digitized. It is generally supposed that current methods of measurement have not introduced a systematic bias to the SL data set relative to measurements circa 1900.

5.2.3 Data Quality and Distribution

Sea level is subject to variation because of changes in many geophysical factors, including wind, temperature, salinity, precipitation runoff, atmospheric pressure, and so forth (cf. Fairbridge and Krebs 1962). Thus, climate variations or secular trends in any of these variables can induce changes in RSL.

Perhaps more important, any changes in geodetic reference level because of movement of the SL instrument or natural vertical tectonic motion of the coastal margins will also induce an apparent change in RSL. Modernization of harbors has often required displacement of SL instrumentation. Furthermore, rates of actual vertical motions of the coastal land surface can be as large or larger than most current estimates of RSL change (Barnett 1982, 1983a). One should not underestimate the potential biasing effects of these two reference level problems to the SL data sets.

The geographic distribution of useful SL stations was given by Emery (1980). A notable absence of stations in the central gyres and equatorial regions is obvious from this work, and even more clear is the nearly total lack of measurement stations in the Southern Hemisphere. This distribution of data makes it abundantly clear that reliable estimates of change in *global* SL have yet to be made, despite suggestions to the contrary.

In summary, all estimates of RSL change are apt to be contaminated by factors as large as, but having nothing to do with, any CO_2-induced effects. Furthermore, the distribution of stations is not global, leaving serious doubts as to RSL changes in the central gyres of the Northern Hemisphere, the equatorial regions, and particularly the Southern Hemisphere oceans.

5.2.4 Recent Analyses and Variations

Estimating changes in global RSL has been a matter of scientific study since at least 1940 (see Lisitzin [1974] for a summary of older work; and Table 5.1). Fairbridge and Krebs (1962), Barnett (1983a), and others have used representative stations around the continents to estimate rates of change of RSL between 7 and 17 cm per century. Gornitz et al. (1982), Barnett (1984b), and others have used regional averages of SL to estimate change of RSL of 10–25 cm per century. Barnett (1984b) showed that estimates of RSL change could vary by a factor of 2 due simply to the method of analyzing the *same* data set. Thus, estimates of RSL change in the range of 10 to 25 cm per century must be considered as essentially equivalent.

Table 5.1
Estimates of Mean Global Sea Level Increase.

Source	Rate (cm/century)	Method
Thorarinsson (1940)	>5	Cryologic aspects
Gutenberg (1941)	11.8	Sea level (many stations)
Kuenen (1950)	12–14	Different methods combined
Lisitzin (1958)[a]	11.2± 3.6	Sea level (six stations)
Wexler (1961)	11.8	Cryologic estimates
Fairbridge and Krebs (1962)	12	Selected sea level
Emery (1980)	30	Sea level (many stations) and selected stations
Gornitz et al. (1982)[b]	12	Sea level (many stations)
Barnett (1983a)	15.1±1.5	Selected sea level stations
Barnett (1984b)	14–23	Sea level (many stations)

[a] Cited in Lisitzin (1974).
[b] The authors attempt a correction for crustal motion and find 10 cm per century. The value of 12 cm per century is without this correction.

It has also been pointed out by Barnett (1983a, 1984b) that even the existing SL data set does not support the idea of an increase in SL over the domain of present data. The southwest and western Pacific has been experiencing a *decrease* in RSL, a decrease that is independently supported in recent years by expected changes in the ocean's density structure (White and Hasanuma 1980). The causes for these changes are unknown. It may be that local factors are obscuring a global CO_2 signal, if such a signal exists. At any rate, the existence of a global rise in RSL has not been proven.

Typical time series of RSL change obtained from the existing data are shown in Figure 5.1. The results show the variation between different analyses methods noted above. As a general rule, the curves are well fit by a simple linear trend, particularly after the 1920s. The length of record has much to do with the estimated trend. After trend removal from the 1930–1980 period, the residual noise was indistinguishable from white noise in at least two cases (Barnett 1983a, 1984b). Thus, claims of an accelerating increase in the rate of change of RSL are not in agreement with the observations. However, the latter work showed that even modest changes (e.g., 10%) in the current rate of RSL change should be readily detectable on time scales of a few decades.

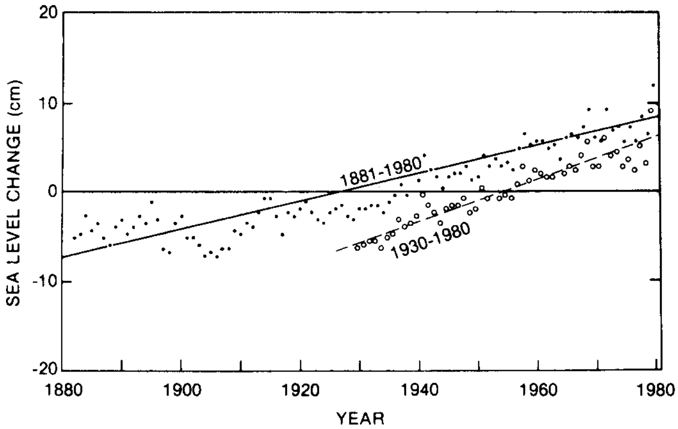

Figure 5.1. Estimates of near global sea level change based on the analysis of the two periods of record shown. The estimated sea level values for the 1930–1980 period (open circles) are different than the corresponding values for that period in the 1881–1980 data set (dots) because the longer period is represented by a different set of (and fewer) stations. Source: Barnett (1984b).

5.2.5 Causes of Relative Sea Level Change

Much work has been done over recent decades to explain possible causes for RSL changes. The summaries of Munk and Revelle (1952), Lambeck (1980), and Barnett (1983a) serve as beginning background references for the following discussion. A number of processes could cause RSL to increase. Some of these are considered in the following sections.

5.2.5.1 Ice Melt

Melting of glacial ice on land in the polar and mid-latitude regions could easily give the observed changes in RSL. The amount of ice needed (360 km^3 of ice is equivalent to 1 mm increase in RSL) would be within the error of estimating the volume of the landlocked polar ice. In a sense, the ice would not be missed. Indeed, even melting of mid- to high-latitude glaciers could account for a significant fraction of the observed RSL rise (Fairbridge 1961; Meier 1984). Melting at both poles of the magnitude needed to give the RSL increase would lead to a drift of the Earth's axis of rotation at almost exactly the same velocity observed over the last 80 years. This drift is associated with the redistribution of the polar ice mass more or less uniformly over the oceans and subsequent change in the Earth's moment of inertia. The increase in length of the day would also be in the sense observed. Furthermore, the melt water would change the ocean's salinity by a value that would be too small to be observed, particularly away from the polar regions. Indeed, no such salinity change is observed except very near the polar caps where it might be expected anyway (see Section 5.4). It should be noted, however, that attractive alternative explanations for the motion of the poles, the change in the length of the day, and the decrease in high-latitude salinity are all available (e.g., Peltier 1985). Attribution of RSL change to ice melt that may or may not have been influenced by CO_2-induced climatic effects has not yet been proven.

5.2.5.2 Temperature Changes

An increase in ocean temperature would lead to an increase of RSL through thermal expansion (steric effect). Conservative estimates of temperature change (Section 5.3) are not large enough to give the observed RSL change. Furthermore, the resultant increase in dynamic height (change in the upper ocean's density) needed to give the RSL is not observed (Section 5.5). Thermal expansion of the upper ocean can explain, at most, a small part of the observed RSL change.

5.2.5.3 Circulation Changes

A very specific pattern of simultaneous spin up and spin down of *all* the ocean's gyres could result in an increase in RSL. The probability of getting just the right changes in the ocean circulation systems to give the observed RSL changes appears small. However, the subject does not appear to have received serious study and so remains an open question.

5.2.5.4 Crustal Motion

A simultaneous subsidence of the continental margins and key mid-ocean islands could give the observed RSL change. Although this effect no doubt contaminates the RSL data, it is unlikely that it would explain the existence of the observed spatially coherent RSL signal that covers most of the coastal margins of the continents. This is because the spatial scale of crustal movements is considerably smaller than the spatial scale of the RSL signal (Newman et al. 1980). Furthermore, it seems unlikely that the amount of subsidence is equal, to within a factor of 2, for the vast majority of ocean margins and mid-ocean islands (Section 5.2.4).

In summary, sea level has apparently risen along many, but not all, of the coastal regions of the world since 1900. The most likely cause of this change is the melting of land-locked ice in the high latitudes of both hemispheres, but this supposition has not been proven.

5.3 TEMPERATURE

5.3.1 Importance

During approximately the winter half of the year the oceans act as a heat source that helps to maintain the atmospheric general circulation and moderate wintertime cooling. During the summer half of the year the oceans absorb solar radiation that is later given off as heat in the winter. The oceans also help in transferring the excess heat from the tropics to high latitudes where it is transferred to the atmosphere. Furthermore, because of their higher density and heat capacity, the oceans have a far larger thermal inertia than the atmosphere. For all of these reasons the oceans act as the prime stabilizing influence on the global climate. Without this influence, the climate of Earth would be as hostile to life as that of Mars or Venus.

The oceans' roles of providing thermal inertia and heat to the atmosphere are often represented by the proxy measurement of sea surface temperature (SST). Although it is the ocean surface that is in direct contact with the atmosphere and although the magnitude of SST affects air-sea heat exchange, it is really the heat transport and content of the ocean and heat flux to and from the atmosphere that are crucial to understanding the oceans' role in climate. Unfortunately, knowledge of these three variables is sparse at best, and so we are left with SST and, occasionally, the distribution of temperature with depth (heat content) as the only measured variables available with which to describe fluctuations in the oceans' thermal properties.

An increase in the CO_2 concentration will warm the atmosphere and the oceans. Except in polar regions, one might expect the SST to increase by about the same amount as the surface air temperature. Normal ocean processes would mix this surface signature to greater depths. The resulting changes in the oceans' density structure would increase the SL and perhaps change some features of the ocean circulation. These changes would probably have a strong regional signature (e.g., marginal seas, regions of water mass formation) because the processes which would redistribute the heat are known to have such characteristics.

5.3.2 Measurement Methods

Estimates of SST have been obtained principally by two methods. In the early decades of the 1900s, measurements were obtained by tossing a bucket overboard to collect a water sample and then using a thermometer to measure the water's temperature. By the time World War II had ended, the use of injection thermometer readings from ships' engine rooms are thought to have substantially replaced the bucket data, but this is not known with certainty.

This situation represents a near disaster for those interested in studying long term, but small (0.2–0.5°C) changes in the SST field. This is due to the fact that many studies have shown bucket values are biased *low* relative to injection values by 0.3–0.7°C (cf. Barnett 1984a for additional references). Thus much of the perceived increase in SST since about 1900 may be due to instrumental effects alone (see below).

Measurements of deep ocean temperature since the 1900s come largely from hydrographic data, that is, precision measurements of temperature and salinity from oceanographic research vessels. Here the instrumentation is generally consistent and calibration techniques are excellent. The bathythermograph[1] data, which began in substantial numbers in the 1950s, are somewhat less accurate, but still probably good enough to look for changes in temperature of the order of 0.5°C once the record has been extended for 1 to 2 more decades to improve the signal and noise characteristics of the data set.

5.3.3 Data Quality and Distribution

The distribution of historical observations of SST from the world's fleet of merchant and military vessels (30–50 million observations) is concentrated principally along the traditional shipping lanes. Thus coverage is excellent in the North Atlantic and Indian Oceans, as well as the coastal routes of the Pacific. The South Pacific is especially data poor, as are all the Southern Hemisphere oceans below about 40–50°S. Recently, there have been two large efforts to consolidate these ship observations into useable, well-ordered data sets. In the United States, Slutz et al. (1984) describe the COADS (Comprehensive Ocean-Atmosphere Data Set) while a similar effort in the United Kingdom is described by Parker (1984).

The ship observations began to be gathered in an organized fashion after 1854. Nevertheless, it was not until after World War I that observations became relatively plentiful in the most frequently traveled ocean areas. Even today only about 100,000 SST observations (on average, one for every few thousand square kilometers of ocean) are available each month to define the surface temperature over all the oceans (72% of the Earth's surface). Modern satellite techniques offer hope that eventually global SST fields will be available. Until that time, it is only possible to trace the history of SST change over selected regions of the oceans since roughly 1900.

Observations of the oceans' interior thermal structure total some one million hydrographic casts since the late 1800s and perhaps several million bathythermograph (BT) measurements. The vast majority of these observations were made after 1950. Furthermore, there are few places in the open ocean, far from the contaminating influences of land, where

[1] A device that is lowered overboard and returns information on temperature versus depth to the ship.

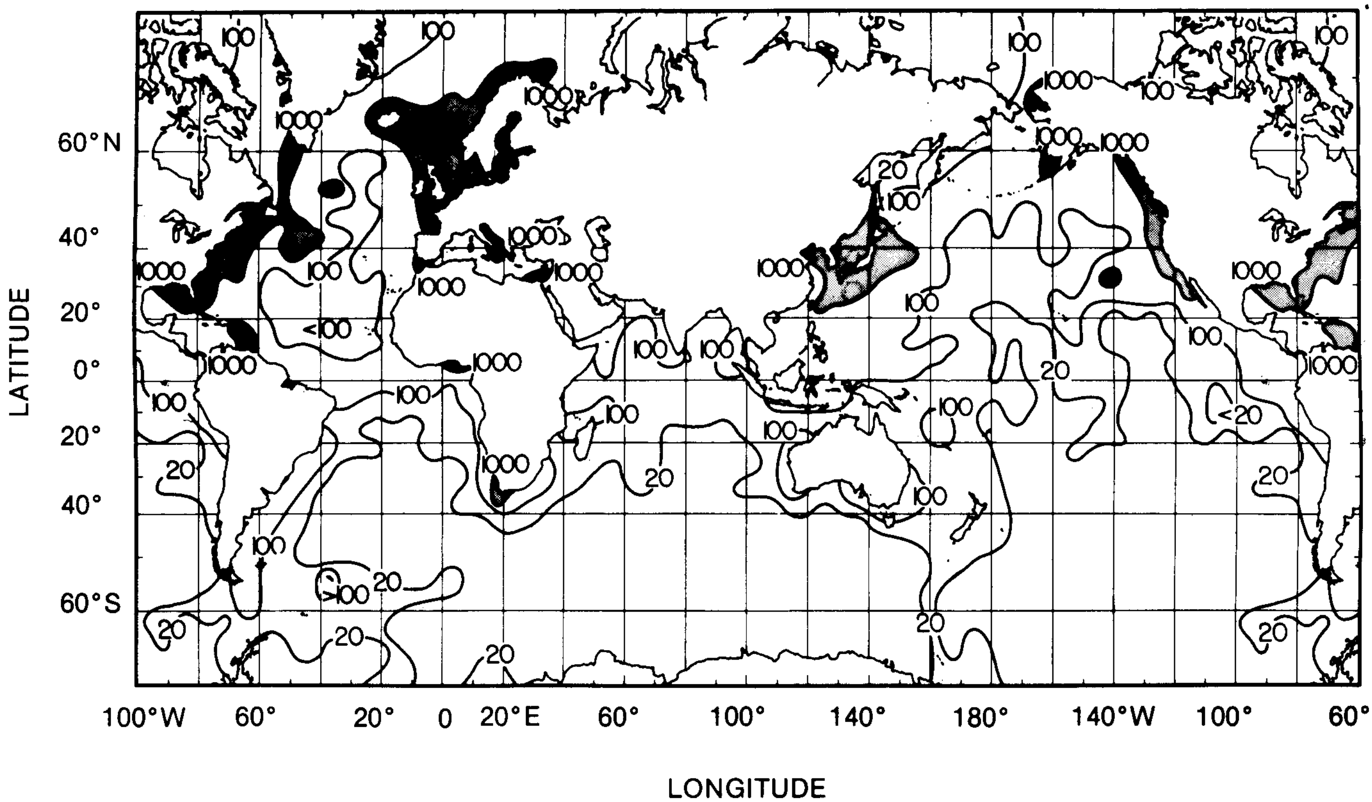

Figure 5.2. Distribution of hydrographic casts by 5° square through March 1978. Areas with over 1000 casts are shaded.

a reasonable time series of hydrographic observations can be constructed (Figure 5.2). These few sites generally coincide with ocean weather ship (OWS) stations. But nearly all the OWS stations were discontinued or relocated in the 1970s so that even these relatively short, 20–30 year records have effectively ended. The BT data continue today but have neither precision nor depth of penetration of the hydrographic casts. However, for CO_2 signal detection studies, the BTs might suffice *if* they are continued at the OWS sites with sufficient frequency.

In summary, ocean temperature data are not globally distributed. The most voluminous set (OWS station observations) is potentially contaminated by instrumental effects so that estimation of relatively small SST changes since about 1900 will be very difficult, at best (see below). Hydrographic and BT temperature data that have the required accuracy and measurement consistency are not plentiful prior to 1950. Background noise levels, withdrawal of OWS stations, and so on, constitute serious problems in attempting to estimate how the temperature in the oceans' interior has changed with time.

5.3.4 Recent Analyses and Variations: Surface

It has been the vogue in recent years to estimate changes in the surface air temperature (SAT) of the Earth and, in some cases, compare the results with those expected from the effects of an increased CO_2 level. The number of investigators attempting this estimate has multiplied in recent years in response to the CO_2 problem (cf. NRC 1983). However, most of the estimates of global temperature have had poor or no information over the oceans. Because over 70% of the Earth is ocean-covered, it is fair to wonder whether the many estimates of SAT change, since about 1900, are particularly meaningful. The keys to answering this question are twofold: (1) Has the SST change since 1900 been quantitatively similar to the SAT change over land? (2) Are SST and marine air temperature (MAT) really the same? The answers to these two questions are crucial to detection of a possible CO_2 signature in the surface temperature field. Finally, if the oceans really have warmed, then the possible impact on the atmosphere and RSL might be estimated if we knew to what depth the temperature perturbation signal has penetrated. These three questions are discussed below.

Estimates of temperature change at the ocean's surface have only recently been made on a global basis. Paltridge and Woodruff (1981) made a first attempt at a near-global estimate using data from regions where they exist. Their record of global SST change is shown in Figure 5.3 along with an estimate of global temperature derived by Mitchell (1961, 1963). Despite potentially serious problems with the Paltridge and Woodruff analysis, the difference between SAT and SST shown on Figure 5.3 seems real. Similar estimates of SST changes over large regions of the Northern Hemisphere ocean recently have been made by Barnett (1984a) and compared with Northern Hemisphere SAT from Jones et al. (1982). Significant differences were again found between SAT and SST. Based on these limited analyses of ocean data, it appears that SST and SAT from land data show increasing trends but differ from each other in several important respects. However, results of a recent study by Folland et al. (1984) did not find this discrepancy (see Figure 4.15). Those authors performed numerous corrections of the original data before arriving at that conclusion, partially given in Folland et al. (1984), Folland and Hsiung (1985), and Parker and Folland (1984). The approach appears promising, but the critical reader may be cautious of corrections that can be as large as the signal itself, especially given assumptions regarding instruments, data distributions, and other aspects that are not possible to document well.

Simultaneous estimates of SST and MAT over the ocean regions have been made by Barnett (1984a). These are shown in Figure 5.4 along with the Northern Hemisphere estimates of SAT from Jones et al. (1982). The SST and MAT over the oceans track reasonably well, although an offset between them is apparent (see below). Apparently, the two near-surface ocean temperatures vary in unison, a result anticipated by Cayan (1980). They do not vary with the SAT estimated from land-only summaries as noted above.

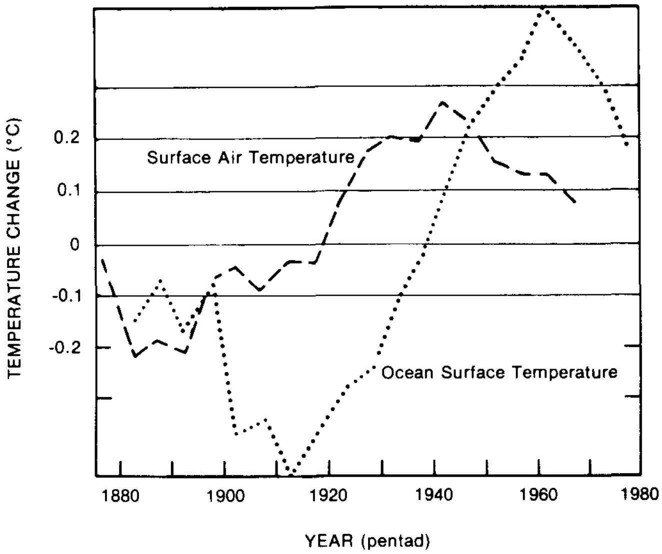

Figure 5.3. Estimates of global temperature change by pentad (five year average). Surface air temperatures (dashed line) are from Mitchell (1961) and ocean surface temperatures (dots) from Paltridge and Woodruff (1981).

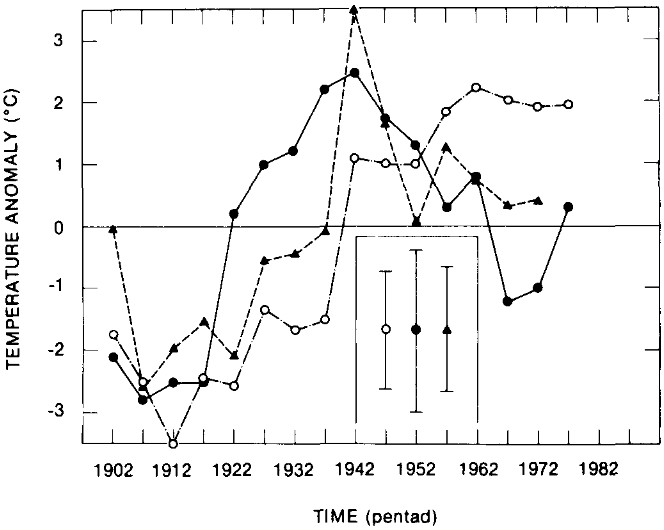

Figure 5.4. Pentad average typical of the Northern Hemisphere sea surface temperature (open circles) and air temperature over the oceans (triangles) versus air temperature over the continents (solid circles) by Jones et al. (1982). The inset shows the one standard deviation associated with each estimate. Source: Barnett (1984a).

The nature of the offset between MAT and SST is shown in Figure 5.5 where the two sets of simultaneous data are subtracted. The bias and offset is what would be expected by a gradual replacement of bucket with injection readings (assuming no bias associated with the MAT data[2]). The unlikely alternative is that the Northern Hemisphere oceans

[2] Folland et al. (1984) and Parker (1984) have suggested that the MAT may have a bias of the order of 0.2°C because of different instrumental measurement heights.

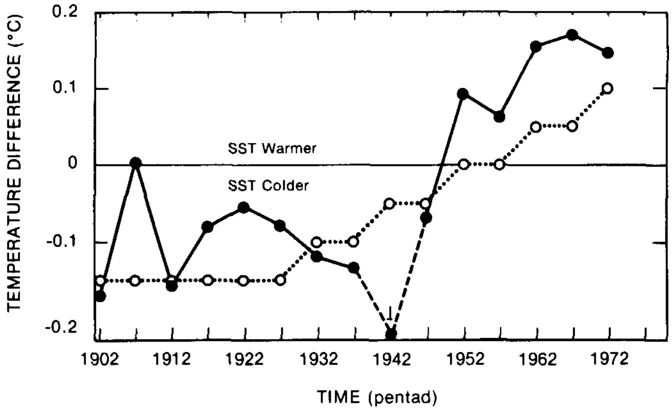

Figure 5.5. Pentad averages of sea minus air temperature differences over the Northern Hemisphere oceans (solid circles) present in the Barnett (1984a) data set. An SST correction term proposed by Folland and Kates (1982) to account for the change from bucket to engine injection methods for measurement of SST is also shown (open circles) with sign reversed for (for comparison purposes) and normalized to 1951–1960.

warmed relative to the atmosphere by 0.3°C in a 10-year period from 1942 to 1952. Note that the value of 0.3°C is roughly one-half the change in any of the temperature records—land or sea—since about 1900. This seems to be the level of uncertainty associated with any estimate of total temperature change since about 1900. It appears that conservative estimates of SST and oceanic MAT changes since 1900 are substantially lower for the Northern Hemisphere oceans than those normally quoted. On the other hand, the corrected estimates of Folland et al. (1984) are roughly the same order of magnitude as the SAT but the corrections, which are arguable, are nearly as large as the signal. The situation regarding the data sets and their analyses are not satisfactory at this time. However, it is hoped that additional work along lines described by Folland et al. (1984) will resolve this situation in the near future (see references to Folland and others noted above).

5.3.5 Recent Analyses and Variations: Subsurface

Studies of subsurface temperature changes in the oceans over time scales of decades or longer are few in number. The applicability of these studies to the longer term changes associated with a possible increase in CO_2 concentration is uncertain. However, it is clear that they will provide a valuable starting point for *future* studies, and so they are summarized below. A new study aimed at determining deep temperature changes over the last 60–70 years is also summarized.

Temperatures in the depth ranges of 200–400 m off Bermuda appear to have decreased by order of 3°C per century, based on studies by Pocklington (1972, 1978) and Wunsch (1972) on data for the period 1954 to 1972. Using a slightly longer version of the *same* data set, Roemmich (1985) comes to the opposite conclusion, that is, warming of the water column. From results of a study carried out in the nearby Caribbean, Worthington (1956) has suggested marginal increases in temperature of the deep water at a rate of roughly 0.05°C per century, although instrumental accuracy makes several of these estimates questionable.

Other studies carried out in the North Pacific (Robinson 1960; Tabata 1981) and the North Atlantic (Gammelsrod and Holm 1981; Ellett 1981) show a wide disparity of results covering the range from increasing to decreasing temperature to no significant change in temperature at all. There is no consensus in the results, so a general statement regarding temperature changes at depths in the oceans cannot be made.

To gain a long-term, large-scale view of possible changes in temperature at depth, a systematic regional analysis of the existing hydrographic data was made. The basic idea was to find oceanic regions that had reasonable hydrographic data densities in both the 1920–1930s and 1960–1970s and then to compare average temperatures at depths over these regions and periods to see whether a significant change occurred over the intervening 40–50 years. The methods and details are identical to those described by Barnett (1983b). The study depths were 200 and 1000 m. The temperature changes at 200 m [T_{200}] for selected regions of the Northern Hemisphere ocean are shown in Figure 5.6 (hundredths of °C per century) and are qualitatively representative of the results from 1000 m. No systematic change in T_{200} was found. Furthermore, the 95% confidence limits in the rate of change of temperature (dT_{200}/dt) were 3–20 times the estimated changes shown on Figure 5.6. None of the values of dT/dt at 200 or 1000 m were statistically distinguishable from zero. This is so for even the largest values in the equatorial region (where there are few estimates) and in the tropical western Pacific (where there is large scatter about the mean). The current results for the Pacific are in general agreement with those of Robinson (1960).

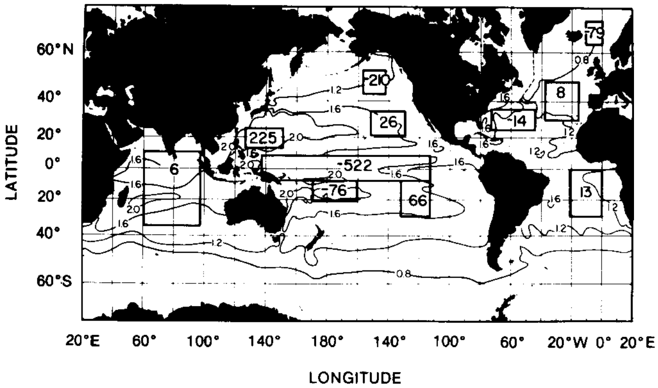

Figure 5.6. Change in temperature at 200-m depths in large averaging areas (boxes) in units of hundredths of °C per century. The background contours refer to the dynamic height relative to the 1000 dbar surface (1 dbar = 1 kPa).

In summary, long-term, coherent, basin-wide changes in the temperatures of the oceans at depths are not detectable with the current data set. However, significant local changes have been documented on decadal time scales. These latter signals may well represent a noise that must be successfully filtered out if a larger scale, long-term signal is to be detected. It is doubtful that current ocean sampling densities are adequate for this task.

5.4 SALINITY

5.4.1 Importance

The salinity, or salt content, of the oceans is one of the major factors in determining the oceans' density field. It is small space and time differences in this density field that help drive the general circulation of the world's oceans. These current systems, in turn, help redistribute heat, oxygen, and other materials necessary to the oceanic ecosystem. Significant, long-term alteration of the global salinity could have serious physical and biological repercussions.

It is not possible at this time to estimate the magnitude of a CO_2-induced change in the oceans' salinity. However, one might guess in advance that such changes, if they occurred, would be regional in nature. For instance, we know qualitatively that a warming will most likely melt high latitude ice, thereby freshening the oceans. Any appreciable change in the rates of evaporation and precipitation over the globe would also change salinity. Regional change can be expected if the oceans must adjust to major change in atmospheric forcing, that is, displacement of the major centers of action. Unfortunately, no qualitative estimates of such changes are currently available. In many situations one cannot even guess the sign of the change.

5.4.2 Measurement Methods

The direct determination of salinity used since the beginning of the century until the 1960s involved chemical titration to estimate the amount of chloride in the sea water. This, in turn, was related to the salinity by an empirically determined relationship. In careful hands, this method gave accuracies of about 0.02 $^{0}/_{00}$ (0.02 parts per thousand). This constitutes a measurement error of about 0.06%.

In the last several decades, the electrical conductivity of sea water has been measured, and that value then has been related directly to salinity. This method was not without its problems in the beginning. However, accuracies of $>$0.01% ($>$0.003 $^{0}/_{00}$ or better), are now quoted for accuracy of salinity determination. For comparison purposes, this measurement error can be compared with a typical value for seasonal cycle variations of $<$0.1 $^{0}/_{00}$ and the interannual signal of 0.5 $^{0}/_{00}$ seen off western Europe.

5.4.3 Data Quality and Distribution

Most salinity measurements available in the hydrographic archives are thought to be reasonably reliable. There are some notable exceptions where data from a particular expedition are known to be biased or where a particular country produces poor quality measurements. These are generally the exceptions.

The amount of salinity data available for analysis is distributed as are the hydrographic casts from which they come (Figure 5.2). Only a few limited regions of the oceans have enough information to attempt to construct a time series of salinity changes since the early 1900s. Longer records come from some coastal stations, but these data are susceptible to a myriad of problems (e.g., runoff effects).

5.4.4 Recent Analyses and Variations

Few studies of long-term salinity change in the oceans are available. Perhaps the most comprehensive study was that of Dickson (1971) who studied variations in the *surface* salinity in the coastal regions off western Europe. He found large, quasi-periodic changes in salinity to occur over the entire region. Anomalies of order $>0.5\ ^0/_{00}$ were common and seemed to be associated with a recurrence interval of 3–4 years. The anomalous changes in salinity were related to large-scale changes in the circulation of the North Atlantic Ocean and overlying atmosphere. This variation will constitute a large "noise" against which to detect longer term trends in salinity, for example, the 0.1 $^0/_{00}$ per 50 years suggested to occur, at least locally, in the eastern Atlantic (Ellett 1981).

Changes of salinity at depths seem to fit no overall coherent pattern based on previous work. Robinson (1960) has found no significant change in the Pacific, whereas Pocklington (1972) has found a slow increase of salinity with time for the upper few hundred meters of the ocean off Bermuda. Swift (1983) and Brewer et al. (1983) have shown a recent, significant decrease in salinity in the deep water north of 50°N in the Atlantic which they tentatively attribute to changes in the overlying atmosphere. Etkins and Epstein (1983) might argue that meltwater from Greenland is causing the observed salinity reduction. At any rate, other works that address changes occurring at decadal time scales show the potentially difficult signal-to-noise problem associated with detection of long-term (50+ years) changes in salinity.

In an effort to clarify matters, a study of a selected set of hydrographic data was undertaken for this volume. The approach and methods are identical to those described in Section 5.3.5 and Barnett (1983b). The resulting estimates of salinity change at 200 m in selected Northern Hemisphere oceans are shown in Figure 5.7.

The illustration shows estimated changes in salinity ($^0/_{00} \times 100$ per century) of the order of 0.1 $^0/_{00}$ per century, a value well above the measurement accuracy. Unfortunately, none of the values

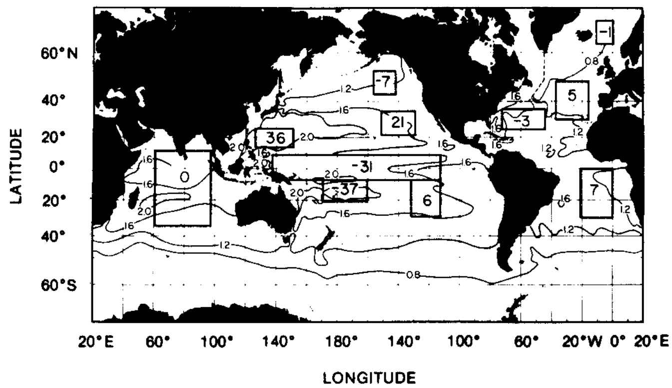

Figure 5.7. Changes in salinity at 200-m depths in large averaging areas (boxes) in units of $^0/_{00} \times 100$ per century. The background contours refer to the dynamic height relative to the 1000 dbar surface.

shown on Figure 5.7 is significant at the 95% confidence level. Again the confidence levels are 3–30 times the values shown. There is also no coherent pattern of change, with increases (+) and decreases (−) being about equally balanced.

In summary, significant long-term changes in salinity may be occurring in a few places in the oceans, but no coherent pattern appears to exist with which one can characterize the Northern Hemisphere oceans. As with temperature, the signal-to-noise problem appears to be severe and not likely to be ameliorated by current hydrographic sampling programs.

5.5 DENSITY

5.5.1 Importance

Spatial and temporal changes in the density field lead to changes in the general circulation of the oceans.[3] This is important for the reasons already mentioned above and will not be elaborated further here. Additional changes in density field will affect sea level and hence relate to the discussion in Section 5.2. Indeed, reasons for changes in RSL, under the right circumstances, could be deduced from the study of changes in the density field. The magnitude and even the direction of expected changes in

[3] Wind-induced changes in the general circulation of the oceans will also require adjustments in the density field. Consideration of any CO_2-induced changes in the global wind stress distribution and subsequent oceanic response will be vital in attempts to detect CO_2 changes in the oceans.

the density field of the oceans that would be due to an increased level of CO_2 are unknown.

5.5.2 Measurement Methods

Density is normally not measured directly in the oceans. Instead, it is derived from knowledge of the temperature and salinity and the depths at which they were measured. The measurement problems associated with these variables were discussed in Sections 5.3.2 and 5.4.2.

The relationship between temperature, salinity, depth, and density is mathematically complicated, and, as a result, the relation is generally obtained from tables derived from laboratory measurements. It has been estimated that the density determination is accurate to >0.02%.

5.5.3 Data Quality and Distribution

All of the problems associated with measurements of the temperature or salinity of the oceans' interior carry over to the density determination. Estimation of the depth of measurement also can add error. Nevertheless, the historical files of hydrographic data are thought to be reliable, in general (see Sections 5.3.3 and 5.4.3 also). The distribution of density observations is generally poor (Figure 5.2), as one would expect from the earlier discussion.

5.5.4 Recent Analyses and Variations

Several studies of local changes in the density structure of the upper ocean suggest significant changes in that quantity (e.g., Pocklington 1972, 1978; Tabata 1981; Gammelsrod and Holm 1981). However, the time series producing these results are relatively short (25–30 years) and appear to contain a large random component, making it difficult to quantify the apparent density changes. Longer records from the eastern Atlantic (for example, Rockall Channel) show a lowering of the surface density that is due to an increase in temperature (2°C) and decrease in salinity (0.1%) in the last 50 years (Ellett 1981). The representativeness of these data, however, are open to question, as changes this large are not observed over most of the world's oceans (Robinson 1960; Barnett 1983b). Furthermore, Ellett's results apply only to surface data, so the effects on the water column, even locally, are unknown.

In a recent study, Barnett (1983b) attempted to determine whether changes in relative sea level measured by tide gauges can be explained in terms of changes in the density structure of the upper ocean. The idea was to compute dynamic height relative to a given pressure surface using hydrographic data from the early 1900s and then to compare similar estimates of dynamic height relative to the same pressure surface with data from recent times, that is, the same tactic described in Sections 5.3.4 and 5.4.4 for temperature and salinity, respectively. The principal results of the study (Figure 5.8) were as follows:

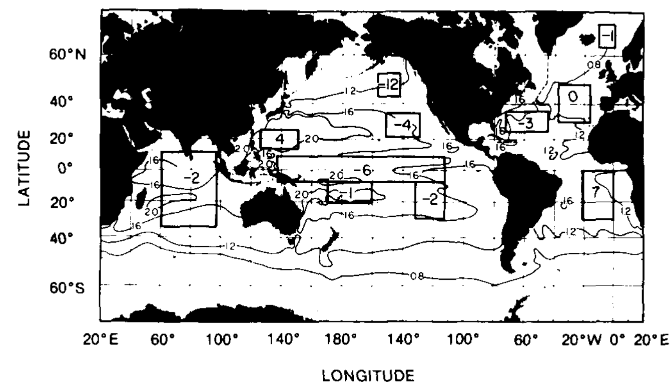

Figure 5.8. Changes in dynamic height (0–1000 dbar) in large averaging areas (boxes) in units of dynamic cm per century. The background contours refer to the dynamic height relative to the 1000 dbar surface.

1. Averaging over all Northern Hemisphere ocean areas where analyses were conducted gives a relative rate of change in dynamic height of approximately −0.75 dynamic cm per century, that is, dynamic height was *falling*. This value is approximately 20 times less than the observed rate of change of relative sea level and of the wrong sign. It is also 10–20 times less than the standard deviation involved in making the estimate. It is a number that for all intents and purposes is statistically zero.
2. The data going into the grand average are noisy. However, the noise levels that are due to instrumental effects are at least a factor of 10 less than those observed. It appears that much of the noise is due to natural variability that extends to

the lower study depth (~1000 m) (e.g., internal waves and eddies). It has been suggested that other measures of ocean properties, for example, temperature-salinity relations, would help reduce this noise level.

3. Several competing hypotheses that might explain the suggested changes in relative sea level were compared with the hydrographic results. A relatively large-scale warming of the upper ocean, as suggested by the data of Paltridge and Woodruff (1981), would give changes in density structure that should have been detectable in this study but were not. A more modest warming, in line with more conservative estimates of the actual changes in sea surface temperature, would give changes in the density structure that might marginally have been detected. Freshening of the surface waters from the melting of the polar ice caps would change the dynamic height by an amount that would be undetectable in the presence of the large, natural background noise. This conclusion depends on the assumed homogeneous mixing of the melt-water on a global scale. The combined effect of modest warming of the upper ocean plus freshening due to ice cap melting would give changes in dynamic height that should have been marginally detectable, but were not.

5.6 SUMMARY AND RESEARCH RECOMMENDATIONS

Climate changes in key ocean variables over the last 80–100 years are rather poorly documented. One of the best observed variables is sea level. However, the data coverage for this variable is poor in the central ocean regions and almost nonexistent in the Southern Hemisphere, making estimation of global sea level change impossible. *Analyses of existing limited data suggest changes in relative sea level of 10–25 cm per century along many, but not all, of the continental margins.*

Another well-observed variable is sea surface temperature (SST). Global coverage is again a problem, but more serious is the apparent, but fictitious, increase in SST due to instrumental bias. Despite this problem, it appears that changes in SST and air temperature over the ocean may differ significantly from air temperature trends measured over land. *Temperature at, or immediately over, the ocean surface must be included proportionately in any estimate of global temperature change.*

Information on changes in the oceans' interior temperature, salinity, and density are meager at best. Local changes in some of these variables have apparently been found in the few high data density areas. However, these changes fit no apparent systematic pattern. Furthermore, small-scale features (eddies, internal waves) in the oceans introduce such high noise levels in attempts to study ocean-wide changes in temperature, salinity and density, that *inferred trends over the last 50 years are generally a factor of 10 less than the uncertainty of their estimates.* Use of different measurements of ocean property change might reduce these noise levels.

The bleak picture described above is not without hope. The following actions, if taken now, will make possible the detection of a CO_2 signal in the ocean, if it exists, much more likely.

1. Quantitative estimates of expected CO_2-induced changes in the ocean must be made. Assuming that an appreciable signal can be expected in the oceanic properties discussed above, then the following steps should be followed.
2. Sea level stations should be established in the southern oceans and on mid-ocean islands as soon as possible. Very long baseline interferometry methods, which are proven, should be used to "shoot in" a widely scattered set of sea level stations, thereby factoring out contaminating influences of vertical crustal motion.
3. The historical archives of ships' observations (for example, the COADS data set) should be examined carefully in an effort to uncover simultaneous, global-scale changes in various ocean or over-ocean parameters (e.g., SST, air temperature, wind velocity, pressure, cloud cover). At the same time a diligent effort should be undertaken to see whether correction schemes can be devised to eliminate the instrumental bias effects that may exist in this data set (e.g., Folland et al. 1984). This huge (50+ million observations) data set is virtually our only hope of studying past surface changes in ocean climate, and we must take full advantage of that resource.
4. There is a clear need for a long-term series of hydrographic data from selected regions of the

ocean. These time series should be planned so that spatial and temporal biasing do not influence the results. Furthermore, the sampling program should be dense in time to allow an effective filtering of the high-frequency variability that appears to contribute noise levels that are almost an order of magnitude greater than expected in the density field because of secular changes in global climate. There appear to be few places in the world's oceans where the required data density may exist for recent times, and these might be the first regions where an increased sampling problem should be undertaken.

Detection of long-term climate changes in the oceans, whether they are due to CO_2 transients or other mechanisms, will be difficult even if the recommendations presented above are invoked. A brute force sampling program would be extremely expensive and may not even be feasible in the next several decades. However, a strategy that will make use of the fragments of information on ocean climate change that we now have, plus those to be collected in the future, can be envisioned.

An attractive and feasible strategy to detect oceanic climate change would be based on a good numerical model of the ocean's circulation. The natural parameters of such a model include the distribution of temperature, salinity, and density. Sea level is a derived variable and is available. Because all numerical models have a limited number of degrees of freedom, we do not need to measure all parameters at every location to verify accurate performance by such a model. This also means that carefully chosen information from key ocean regions, when put in the dynamical context of a good ocean model, can provide the measurements of long-term climate change needed to separate possible CO_2-induced effects from other processes that cause long-term changes in ocean climate.

The intelligent development of ocean modeling and data collection strategies directed toward the detection and explanation of climate change in the oceans will be a substantial task requiring decades of effort, stable support, and purposeful coordination. These latter two items appear to be substantially more difficult problems than the scientific problem that must be addressed.

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6. THE CRYOSPHERE AND CLIMATE CHANGE

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CONTENTS

6.1 INTRODUCTION

6.1.1 Components of the Cryosphere

Snow and ice features occur on the Earth's surface in a wide variety of forms, and these are collectively referred to as the *cryosphere*.[1] The major components are snow cover, freshwater ice in lakes and rivers, sea ice, glaciers, ice sheets, and ground ice or permafrost. Solid precipitation originates in the nucleation of individual ice crystals in the atmosphere, whereas floating ice (except for ice shelves and icebergs) forms primarily from the freezing of water in bulk. Ground at subzero temperatures need not involve any water at all in the case of bedrock. The residence time of water in each of these cryospheric phenomena varies widely. Snow cover and freshwater ice are essentially seasonal, and most sea ice lasts only a few years if it is not seasonal. A given water particle in glaciers, ice sheets, or ground ice, however, may remain frozen for 10^1–10^5 years or longer, and deep ice in parts of East Antarctica may have an age approaching 10^6 years.

It will be helpful to consider first the dimensions of the major cryospheric components and the characteristics that are most commonly determined in each case. Guides to standardized data collection on snow and ice were developed during the International Hydrological Decade, 1965–1974 (United Nations Educational, Scientific, and Cultural Organization [UNESCO], International Association of Scientific Hydrology [IASH] 1970a, 1970b, 1970c, 1970d, 1973), and more recently inventories of specific snow and ice data sets have been compiled (Crane 1979; MacKinnon 1980; Barry 1984a, 1984b). Table 6.1 shows that the majority of the world's ice volume is held in Antarctica, principally in the East Antarctic Ice Sheet, but in terms of areal extent it is the Northern Hemisphere winter snow and ice cover that is largest—amounting on the average to 23% of the hemispheric surface area in January (Table 6.2).

[1] "Cryo-" is a combination form of "κριοσ" (Greek) which means icy cold or frost (Neilson 1958; Klein 1966). It is used in several compound words to refer specifically to the freezing of water or of liquids in general. The term cryosphere has been widely adopted in the climate research community (Untersteiner 1975, 1984).

Table 6.1
Snow and Ice Components
(modified from Hollin and Barry, 1979)

	Area (10^6 km^2)	Ice Volume (10^6 km^3)	Sea Level Equivalent[a] (m)
Land Ice:			
East Antarctica[b]	9.86	25.92	64.8
West Antarctica[c]	2.34	3.40	8.5
Greenland	1.7	3.0	7.6
Small Ice Caps and Mountain Glaciers	0.54	0.12	0.3
Permafrost (excluding Antarctica)			
Continuous	7.6	0.03 to	0.08 to
Discontinuous	17.3	0.07	0.18
Sea Ice			
Arctic[d]			
late February	14.0	0.05	
late August	7.0	0.02	
Antarctic[e]			
September	18.4	0.06	
February	3.6	0.01	
Land Snow Cover[f]			
Northern Hemisphere			
Early February	46.3	0.002	
Late August	3.7		
Southern Hemisphere			
Late July	0.85		
Early May	0.07		

[a] 400,000 km^3 of ice is equivalent to 1 m global sea level.
[b] Grounded ice sheet, excluding peripheral, floating ice shelves (which do not affect sea level). The shelves have a total area of 0.62×10^6 km^2 and a volume of 0.79×10^6 km^3 (Drewry 1983).
[c] Including the Antarctic Peninsula.
[d] Excluding the Sea of Okhotsk, the Baltic Sea, and the Gulf of St. Lawrence (Walsh and Johnson 1979a). Maximum ice extents in these areas are 0.7, 0.4, and 0.2 million km^2, respectively. New estimates from 1973–76 ESMR data (C. Parkinson, personal communication) are March extent 14.7 million km^2, September extent 7.6 million km^2. Actual ice areas, excluding open water, are 7.9 and 4.6 million km^2, respectively.
[e] Actual ice area excluding open water (Zwally et al. 1983a). Ice extent ranges between 4 and 20 million km^2.
[f] Snow cover includes that on land ice, but excludes snow-covered sea ice (Dewey and Heim 1981, 1983).

6.1.2 The Role of the Cryosphere in the Climate System

The principal climatic roles of snow and ice relate to their high reflectivity, low thermal conductivity, and thermal inertia effect (Kukla 1981; Miyakoda 1982). Hence, the extent of snow and ice, their thickness, and their albedo are of primary significance for climate research.

Table 6.2
Seasonal Extremes of Hemispheric Snow and Ice Area, 10^6 km^2

	Northern Hemisphere Land	Northern Hemisphere Ocean		Southern Hemisphere Land	Southern Hemisphere Ocean
Total Area	100	154		49	206
Min.	3.7	7.0	Max.[a]	13.4	18.4
Max.[b]	46.3	15.4	Min.	12.5	3.6
	Percent of Hemisphere				
Min.	1.5	2.8	Max.	5.3	7.2
Max.	18.2	6.0	Min.	4.9	1.4

[a] Maximum sea ice extent in the Southern Hemisphere is about 2–4 weeks later than the minimum in the Northern Hemisphere.
[b] Snow cover occurs about 2–4 weeks earlier than maximum sea ice extent.

The ratio of reflected to incident solar radiation is termed the *albedo*. For climatological purposes we are primarily interested in the integrated value for the spectral range (approximately 0.3–3.5 μm) of solar radiation. Typical albedo values are as follows: 0.80–0.90 for fresh, dry snow; 0.60 for old, melting snow and bare ice; and 0.30–0.40 for melting sea ice with puddles. The areal albedo of snow-covered forests, however, is only 0.25–0.40. Rapid shifts in surface reflectivity occur in autumn and spring, and surface albedo is high in high latitudes (Robock 1980), but the climatic significance of these effects is diminished in autumn because the Earth-atmosphere (or *planetary*) albedo is determined principally by cloud cover and because the total solar radiation received in high latitudes is small as a result of the large zenith angle of the incident radiation. There is high average cloudiness over the Arctic Ocean in summer and autumn.

Snow on land or sea ice builds up a cold reserve during the winter season that acts to depress temperatures in spring and summer when energy is required to warm the snow pack (or ice) to 0°C and then to melt it (2.8×10^6 J kg^{-1}). However, the strong static stability of the atmosphere over areas of extensive snow or ice tends to confine the immediate cooling effect to a relatively shallow layer. Hence, the associated atmospheric anomalies are usually local to regional in scale (Walsh 1984).

Snow cover has an insulating role for the ground surface and likewise sea ice insulates the underlying ocean. The effect is most pronounced in the latter case because a sea ice cover decouples the ocean-atmosphere interface with respect to both heat and moisture fluxes. The flux of moisture from a water surface is eliminated even by a thin skin of ice, whereas the flux of heat through thin ice continues to be substantial until it attains a thickness in excess of 30–40 cm (Maykut, in press).

The global snow and ice cover undergoes major changes in extent and water storage on a seasonal time scale (in the cases of snow cover and sea ice), as well as on a scale of hundreds to thousands of years. The continental ice sheets that now cover most of Greenland and Antarctica, and formerly covered much of Europe and North America, have lifetimes of tens of thousands to tens of millions of years (in the case of East Antarctica). On these time scales, the changes in water storage can lead to substantial changes in global sea level. For example, the sea level fell some 120 to 160 m during the last glacial maximum about 18,000 years ago (Chappell 1981; Hughes et al. 1981). Consequently, the cryosphere-climate interactions involve positive feedbacks such as the temperature response to changing snow and ice albedo (Kellogg 1975), and negative feedbacks such as ice sheet buildup, which eventually causes a lack of accumulation because of an absence of storms penetrating to the interior of the ice mass and glacio-isostasy, which causes a relative rise in snow line (Budd 1981; Ghil 1981). More complex are the short-term interactions between snow and ice phenomena and the atmospheric circulation.

An extensive snow cover over the northern continents is usually produced by cyclonic precipitation in an outbreak of cold Arctic air. Subsequently this snow area may serve to deflect cyclonic activity southward along its margin, further expanding the snow cover through the local cooling effect of the snow on the air crossing it (Lamb 1955). Despite some empirical analyses and modeling studies with general circulation models (GCMs), these relationships are not well defined in quantitative terms (Namias 1962; Williams 1975).

Sea ice limits may be extended as a result of advection of ice by the wind and by ocean currents or by the cooling of surface water below the appropriate freezing temperature (approximately −1.8°C). Both processes operate on synoptic and seasonal time scales (Pease 1980; Ackley 1981).The opposite mechanism of atmospheric forcing by sea ice is less well documented. On the seasonal time scale, it appears that in summer the ice responds to forcing by the atmosphere, whereas during freeze-up there are equivalent tendencies for ice anomalies to precede or lag atmospheric anomalies (Walsh and Johnson 1979b; Walsh 1983). On the synoptic scale, Ackley and Keliher (1976) have illustrated possible links between sea ice divergence causing enhanced oceanic heat flux to the atmosphere and thereby increasing cyclonic activity. However, the synoptic interactions appear to operate in both directions but with differences according to season and geographical location (Crane 1983).

6.1.3 Climate Models and Cryosphere-Climate Interactions

The cryosphere has been represented in climate models by a wide range of parameterizations. Accordingly, it is essential to assess carefully the parameterizations used in particular model experiments to interpret properly any predicted changes in snow and ice conditions.

Simple one-dimensional energy budget models (EBMs) typically feature a zonally averaged ice line represented by a step function albedo parameter that varies latitudinally with temperature (Budyko 1969; Sellers 1969). Such models, however, show an oversensitivity to small changes in external conditions, primarily because of the high value chosen for albedo north of the ice line (Robock 1983). Improved parameterizations of snow and ice for zonally averaged EBMs of this type are being examined by Robock. He uses a seasonal EBM with a surface albedo for each 10° latitude band weighted for the fractional coverage of land, ocean, snow, and sea ice. The seasonal and latitudinal pattern of surface temperature sensitivity to changes of $\pm 1\%$ in solar constant is examined for a series of experiments incorporating various snow and ice feedback effects. Robock notes that experiments with the same model with a doubled atmospheric carbon dioxide (CO_2) concentration give identical response patterns of surface temperature (see Figure 6.1a). He concludes that the determining factor is a sea ice-thermal inertia feedback expressed via the extent of ice. Changes in ice area affect both the ice albedo and the thermal inertia, but the latter effect is shown to be responsible for producing enhanced temperature sensitivity in the polar regions in winter, as found by Manabe and Stouffer (1980), and lesser sensitivity there in summer

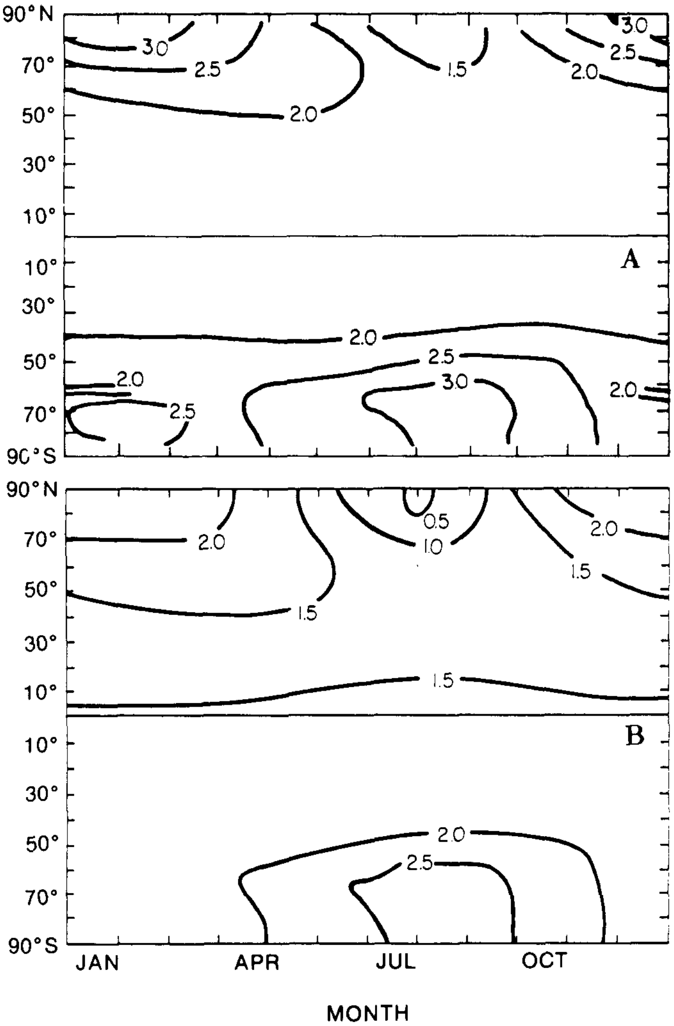

Figure 6.1. Change of zonally averaged surface temperature in an EBM simulation following a 1% increase in solar constant for total run (A) and allowing ice area feedback only (B). The patterns and magnitudes of the changes are virtually identical for a CO_2 doubling. Source: Robock (1983) and personal communication (1984).

(Figure 6.1b). Albedo effects in Robock's EBM experiments are weak, although they add to the net sensitivity without altering the pattern. These results should be contrasted with earlier suggestions (Lian and Cess 1977; Kukla 1982) that the maximum sensitivity is in the summer season. However, EBM experiments must drastically simplify surface processes—such as oceanic heat transport and heat flux through openings in the ice in winter—and they cannot provide regionally specific information on possible CO_2-induced cryospheric changes. Hartmann (1984) demonstrated that longitudinal asymmetries in the snow/ice line (neglected in the zonally symmetric EBMs) enhance the albedo effect on climate sensitivity through the dynamic role of planetary-scale atmospheric waves. High plateaus and mountain ranges are a major cause of longitudinal asymmetries in the snow line; their potential role for augmenting snow/ice albedo feedbacks has been illustrated by Birchfield et al. (1982) for ice caps in the eastern Canadian Arctic.

In higher order GCMs, the cryosphere may be represented through specified sea ice and snow cover. Increasingly, however, the surface hydrological cycle is modeled to include a seasonally varying snow cover (Manabe and Holloway 1975; Manabe et al. 1979). A thermodynamic model of sea ice growth, such as that of Parkinson and Washington (1979), also may be coupled to the GCM (Washington et al. 1980) or incorporated into it (Manabe and Wetherald 1975; Hansen et al. 1983). Using the Goddard Institute for Space Studies (GISS) three-dimensional global climate model, with computed ocean surface temperatures and ice cover, Hansen et al. (1984) calculated a global mean warming of 4°C. Because the direct effect on the radiative balance would cause the surface temperature to rise 1.2°C, a net positive feedback factor of 3–4 is indicated. The feedback components are estimated by inserting the changes in global average surface and planetary albedo, cloud amount and its vertical distribution, atmospheric water vapor, and lapse rate, calculated by the three-dimensional model, into a one-dimensional radiative convective model. The principal feedbacks, which combine nonlinearly, are water vapor (~1.6) and clouds (~1.3). Snow and ice effects are less important (~1.1) and seem to be caused mainly by sea ice changes. Reduced low and middle cloud amounts in low latitudes (positive feedback) offset the effect of increased low clouds in high latitudes (smaller negative feedback).

Climate anomaly experiments involving the cryosphere have been mainly concerned with the equilibrium response of the atmosphere to a change in the extent of snow and ice (Williams 1975; Herman and Johnson 1978). A number of the GCMs used to examine the climatic response to a simulated doubling or quadrupling of atmospheric CO_2 provide some indication of possible effects of such a change on global snow and ice (Parkinson and Kellogg 1979; Manabe and Stouffer 1980; Hansen et al. 1984; Washington and Meehl 1983). A simulation of CO_2 doubling and quadrupling effects with the Geophysical Fluid Dynamics Laboratory (GFDL) GCM (Manabe and Wetherald 1980) incorporating an interactive lapse rate parameterization, an ocean "swamp" surface,[2] a hydrologic cycle, and cloud/radiation feedback shows a high latitude increase in precipitation and runoff, as well as the retreat of the snow and ice limit. The latter gives rise to a 30–40% decrease in albedo at about 75–80°N for a quadrupling of CO_2 concentrations. Cloud feedback appears to have little effect on overall model sensitivity, but no seasonal variation was included in the model. This result differs from the finding of Hansen et al. (1984) with the GISS model discussed above. It reflects the present uncertainties associated with modeling studies and the need for caution in interpreting such simulations.

Simulations with the National Center for Atmospheric Research (NCAR) GCM (sometimes referred to as the Community Climate Model [CCM]) using a swamp ocean formation but no seasonal cycle, for doubled and quadrupled CO_2 (Washington and Meehl 1983), show some differences in predicted changes in snow and ice cover compared with results of earlier work. The snow albedo is set to 0.80 for solar radiation wavelengths less than 0.9 μm and to 0.55 for longer wavelengths. The albedo of bare sea ice is set to 0.70. The albedo changes only for complete melting of snow or ice, and the authors note that this reduces the sensitivity compared with CCM experiments in which the albedo changes in response to a temperature threshold, as discussed above. The model albedos, for a fixed cloud control experiment, compare well with data on annual mean zonally averaged values. In the doubled CO_2 experiments, in which the globally averaged surface air temperature increases by 1.3°C (for both fixed and computed clouds), there are decreases in snow cover over polar land areas, central Asia, and atop Arctic sea ice and smaller areas of increased snow cover over the ice of the Kara-Laptev and Beaufort seas. Further work with the CCM coupled to a mixed-layer ocean model (Washington and Meehl 1984) gives a mean surface air temperature increase for doubled CO_2 of 3.5°C in a seasonal cycle experiment. However, the modeled sea ice in the control case is overextensive, as a result of the omission of ocean heat transport, and this leads to an amplified

[2] This formulation represents an ocean with zero heat capacity and no heat transport, but with a wet surface.

sensitivity to increased CO_2 through surface albedo effects. They also note that the inclusion of ocean heat storage and a seasonal cycle appears to initiate a strong snow- and ice-albedo feedback.

In comparing models that have parameterizations of the cryosphere, particular attention needs to be given to the treatment of surface albedo (see Henderson-Sellers and Wilson 1983; Barry et al. 1984) and thermodynamic processes involved with sea ice (Semtner 1984). Although most models incorporate, in some way, the change between premelt and postmelt ice, there is a large variation in the method by which this is done. For example, the choice of the threshold temperature at which the albedo of a melting ice surface changes is different among GCMs. In the United Kingdom Meteorological Office (UKMO) GCM it is set at 271.2 K (a representative freezing point for sea water), whereas in the GFDL model a threshold value of 263 K is used, well below the temperature at which observed changes in albedo usually occur. As a result of such differences, the UKMO and GFDL models calculate surface albedos of 0.8 and 0.35, respectively, for an identical ice situation. In models in which specified ice cover is used, the influence of albedo differences will be minimal because anomalies in surface energy balance will not be able to feed back on the ice amount or thickness. On the other hand, such specifications limit the use of the GCM to near present-day conditions. Other smaller inter-model differences also exist in the threshold temperature for which sea ice is specified to form.

Ice sheet models have not yet been linked interactively with climate models because of the problem of widely different time scales. The specific response of schematic ice sheets to external (astronomical) forcing has been examined by Weertman (1976), Birchfield (1977), and Pollard et al. (1980), whereas Oerlemans (1979) has considered stochastic forcing (associated with short-term internal fluctuations). Ice sheet dynamics alone can generate glacial-interglacial cycles, particularly when bedrock adjustments are included (Oerlemans 1982b; Peltier 1981). In more complete representations of surface type, such as that used by Ghil and LeTreut (1981), the Earth's surface is segmented to include a land ice cover and an ocean ice cover. The primary interest in most of these studies lies in the fluctuations of ice extent over time, in response either to external forcing of the climate system or to internal stochastic variability. The glaciological and climatic controls of ice sheet growth and decay have also been simulated, incorporating the effects of ice dynamics. Work of this type has been performed by Budd and Smith (1981), who have examined the response of the North American (Laurentide) ice sheet to forcing by the (Milankovitch) astronomical radiation changes. The timing of the growth and decay of the ice sheet is found to be quite realistic if the effects of ice albedo feedback, isostatic crustal response, and decreasing precipitation in the central area of the ice sheet are incorporated. A high resolution, time-dependent ice sheet model incorporating the effects of fast ice streams flow has recently been developed and tested for Antarctica by Budd et al. (1984), whereas Pollard (1983) has improved the treatment of calving processes. A recent book by Oerlemans and van der Veen (1984) summarizes work on modeling ice sheet processes in the context of the climate system.

Possible atmosphere-ice sheet interactions involving the recently identified long-term changes of atmospheric CO_2 during the Quaternary glacial and interglacial intervals remain to be investigated (Oeschger et al. 1984) to determine whether CO_2 operates as a minor feedback, as a major climatic forcing factor, or whether it is primarily a response to other forcing (Stauffer et al. 1984). As Lorius (1984) points out, the glacial-postglacial and preindustrial-1980 changes in CO_2 both amounted to about 30%, but the latter has not yet been accompanied by any identified temperature trend.

6.2 SNOW COVER

6.2.1 Significance

Snow is the most spatially extensive component of the cryosphere (Figure 6.2). In winter, snow at its maximum extent covers between about 37×10^6 and 45×10^6 km^2 of the land area of the Northern Hemisphere (Matson and Wiesnet 1981), and the total for snow-covered land and sea ice amounts to approximately 24% of the hemisphere (Table 6.2). The water equivalent of the seasonal snow pack is a factor of major hydrologic significance. Spring

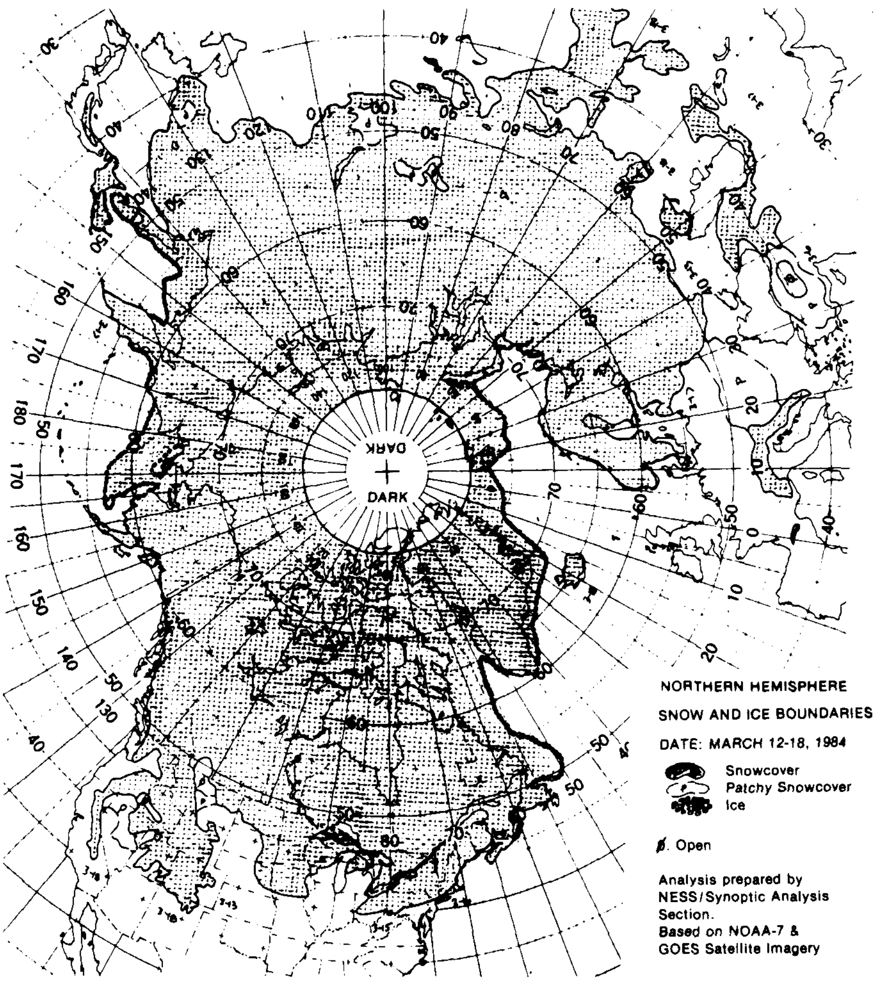

Figure 6.2. Example of Northern Hemisphere weekly snow and ice chart. Source: NOAA-NESDIS.

snow melt provides runoff for irrigation, hydroelectric power generation, and drinking water in many midlatitude and subtropical river basins. Steppuhn (1981) estimates that one-third of global irrigation water comes from snow sources. Excessive snow packs and rapid melt, however, contribute to downstream flooding on uncontrolled rivers. Snow cover provides the basis of winter recreational activities that are major revenue earners in many mountain regions, and, at the same time, very great costs are associated with snow clearance or control on transportation arteries and in population centers (Colbeck et al. 1979). Prolonged snow cover can delay planting in areas with short growing seasons, whereas an absence of snow can expose winter crops to severe cold and dessication. Comprehensive assessments of these effects do not exist (McKay and Adams 1981).

6.2.2 Snow Cover Observations

Four characteristics of snow cover are important fo the present discussion: extent, depth, water con tent, and surface albedo. The first and last are c primary importance for large-scale climate, wherea the other two are principally of hydrological signif cance. Variables that have mainly local significanc (in terms of snow stability, for example) includ grain size, density, free water content, snowpac structure, and other mechanical and thermal prop erties. Electromagnetic and optical properties ar important in terms of the remote sensing of larg scale snow conditions (Warren 1982).

Standard meteorological observations includ the 6- or 12-hourly synoptic weather reports c snowfall and once-daily determinations of sno depth on the ground. Surveys of snow depth an water content at snow courses are also conducte at about monthly intervals by other organizations; in the United States these are carried out by the Soil Conservation Service and cooperating groups on a state-by-state basis. Nearly all of these records begin in this century, and there are no standardized annual global summaries of snow cover duration and average maximum depths. Various national summaries or maps exist, although the time intervals for which these data are available differ greatly and the criteria used to define particular parameters are seldom identical. A little recognized problem is the incompatibility of accumulated snowfall water equivalents on the ground (at open or forested sites) with those measured by precipitation gauges at climatological stations (Goodison 1981). Measurement of the density of new snow is not standard procedure at the majority of Canadian stations, for example. Precipitation gauge design, shielding devices used to minimize wind effects, and measuring techniques also may differ considerably between countries (Goodison et al. 1981).

The value of satellite remote sensing of snow and ice was recognized early in the development of weather satellites, and such systems now provide the primary tool for routine global observations of the cryosphere. Satellites afford frequent repetitive mapping of global snow cover, data collection in the absence of illumination and in the presence of clouds (in certain wavelengths), and a relay and control system for data collection platforms installed in remote locations.

Snow-cover mapping from satellite imagery began operationally in 1966. The National Environmental Satellite (now Data and Information) Service (NESDIS) prepares hemispheric maps of snow and ice approximately weekly. Until April 1983, these maps depicted three reflectivity classes corresponding approximately to highly reflective snow/ice (category 3; albedo ~60%), moderately reflective (category 2; albedo ~45%), and least reflective—such as patchy mountain snow cover (category 1; albedo ~30%) (Matson and Wiesnet 1981). The maps have been digitized for November 1966 through December 1980 by Dewey and Heim (1982) for 7921 grid boxes covering the Northern Hemisphere, but sea ice with and without snow cover and the reflectivity categories are not separately identified. Indices of reflectivity for geographical sectors have also been compiled from the same data source by Kukla and Gavin (1979).

Some deficiencies in this data set should be noted: (1) the charts from 1966 through 1974 did not consistently map Himalayan snow cover; (2) the occasional extension of snow cover beyond the southern limits of the map; (3) the seasonal northern limit of illumination for the satellite sensors in the visible wavelengths; (4) the scattered mountain snows were omitted because of the coarse grid resolution; and (5) 10 individual weekly maps from the series were unavailable due to lost charts. A more detailed assessment of the charts has been provided by Kukla and Robinson (1981).

In April 1983, the maps issued by National Oceanic and Atmospheric Administration (NOAA) NESDIS stopped showing the three reflectivity categories and now indicate only snow-covered land and areas partly covered (Figure 6.2). It is not clear how the two new categories relate to the previous three. The digitized data (Dewey and Heim 1981) designate only snow and no snow and accordingly are not directly applicable to albedo determinations. The overall problem of obtaining realistic monthly global albedo maps remains (Kukla and Robinson 1980; Robock 1980).

The use of satellite passive microwave data to monitor global snow cover is currently being explored (Rango 1980; Kunzi et al. 1982; Foster et al. 1984), although at present it is still only a research technique.

6.2.3 Space-Time Coverage of Data

One of the most extensive snow data sets is provided by the weekly maps of winter snow depth for the United States (U.S. Department of Commerce 1983) that have been recently digitized for 120–70°W by Walsh et al. (1982) for the period 1947–1982. Tabular station records are more widely available for many European countries, and in several instances digital files are being prepared. Published data summaries exist for Austria (Hydrographischer Dienst in Österreich 1962), Finland (University of Oulu 1972), Sweden (Pershagen 1969), and

Switzerland (Schüepp et al. 1980), for example. Station data have also been published for Canada (Atmospheric Environment Service, Canada 1961) and Japan (Japan Meteorological Agency 1964). A tabulation of major sources of charts and surveys has been given in Crane (1979).

While numerous records of snow cover exist, there are several deficiencies that affect their usefulness. The global (satellite-based) record is relatively brief, and intercomparisons are needed with conventional surface observations so that these longer records can be used to assess the spatio-temporal variability more comprehensively. Up to now, snow cover variability has been examined either for specific locations (Uttinger 1963) or in a statistical sense (Karapet'iants 1978). A further problem is caused by the absence in many cases of compact digital files of snow cover data. Synoptic reports of snowfall and snow depth are difficult and expensive to retrieve because they are included within large volume data bases of station synoptic reports.

A more fundamental question concerns the relationship between snow cover variables observed at the surface, those inferred from satellite data, and parameters needed for climatic modeling or climate assessment purposes. The modeler is primarily interested in surface albedo which depends on snow depth, age, water content, and type of vegetation cover. This question is examined below. Neither surface nor space observations of snow cover can yet provide the information desired in a simple, straightforward manner. More work on this problem is required, although there may be no single best solution.

6.2.4 Snow Cover-Climate Interactions

Several physical properties of snow contribute to its great significance in the climate system. Most important is the fact that a snow cover reflects much of the solar radiation incident upon its surface. The reflectance is usually ≥ 0.8 in the wavelength range 0.3–1.1 μm, where most solar radiation is concentrated, although in the near infrared this decreases sharply, and at about 1.5–1.6 μm the reflectance falls below 0.1 (Warren 1982). For clean new snow, the integrated spectral albedo under clear skies is between about 0.8 and 0.9; the figure is increased in the presence of cloud cover as a result of multiple reflections between the snow surface and the cloud base. Average albedos of snow-covered terrain are dependent on the snow depth and the type of vegetation cover. Shallow snow packs expose vegetation and rocks, whereas a snow-covered forest generally has quite a low albedo unless snow is temporarily retained on the tree canopy. In the first few days after a snowfall, snow albedo decreases by about 0.1 as a result of increasing grain size in the snow pack; the deposition of impurities such as soil particles, soot, and plant litter on the surface; and the exposure of vegetation or rocks by wind transport of the snow. As a result of these various effects, snow-covered land surfaces in the Northern Hemisphere have *maximum* albedo values (estimated from satellite image brightness) averaging 0.59 (Robinson 1982). Maximum values over tundra range from 0.35 to 0.80. In individual $1° \times 1°$ boxes with dense coniferous forest the estimates are as low as 0.21.

A snow pack 30–50 cm or more thick insulates the underlying ground surface from the cold air above, by virtue of its low thermal conductivity (Berry 1981; Kukla 1981). Fresh snow is much more effective in this respect than older, more dense snow in which the conductivity may be up to an order of magnitude greater. Another effect of winter snow cover is to eliminate virtually all evaporative moisture flux into the atmosphere. Evapo-sublimation from a snow cover is generally almost negligible, although in dry sunny climates sublimation may account for average losses of between 1 and 6 mm of water per day during February-April (Rylov 1969; Barry 1981). In windy environments like the High Plains or Rocky Mountains, blowing snow particles will undergo substantial sublimation losses (Tabler 1975).

Snow melt is primarily accomplished by the absorption of solar and atmospheric (infrared) radiation, supplemented by downward sensible heat transfer during episodes of warm air advection. However, there are no consistent relationships between the various energy flux processes that contribute to snow melt and characteristic environmental properties (Male and Granger 1981). Rather, the fluxes depend on air masses and large-scale circulation patterns as well as topographic conditions and the altitude of a particular site. Hence, the dominant processes determining snow melt may also vary significantly from year to year.

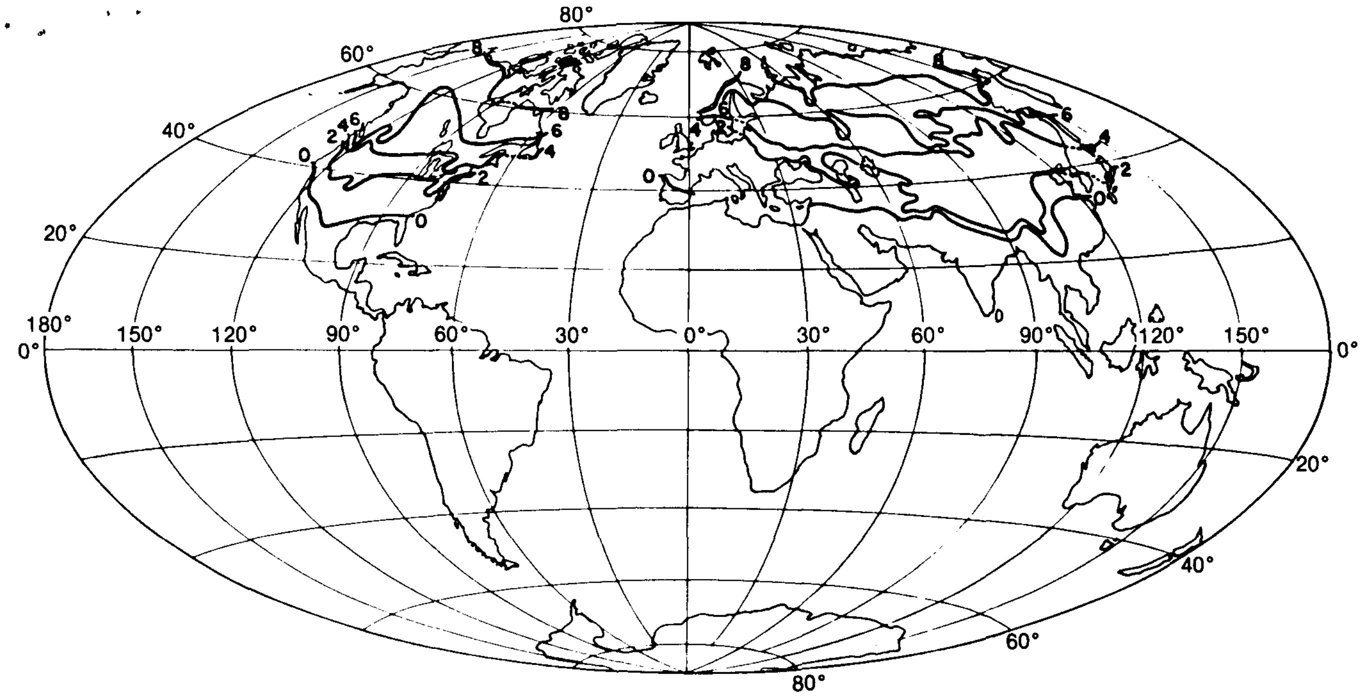

Figure 6.3. World distribution of snow cover. Average duration in months is shown (modified after Rikhter and redrawn from Inland Waters Directorate, Environment Canada). The projection is equal area.

Visible wavelength radiation penetrates several to tens of centimeters into the pack, whereas infrared radiation is absorbed close to the surface. Initially, the pack is warmed to an isothermal state at 0°C, and liquid water percolation and refreezing then complicates the vertical distribution of melting. Melt is the principal process of snow cover removal because the phase change requires only 3.35×10^5 J kg^{-1} compared with 2.5×10^6 J kg^{-1} for vaporization. Energy balance studies indicate that evaporation seldom exceeds 5–6% of the total melt.

The geographical distribution and duration of global snow cover is summarized in Figure 6.3. This conceals the temporal variability. Figure 6.4 illustrates the fluctuations of Northern Hemisphere snow cover for 1967–1984, smoothed by a 52-week running mean. There is an approximately 10% interannual variability, although variations are considerably greater in Eurasia than in North America. It should be noted that the data for 1967–1972 over Eurasia contain some bias toward lower values because of inconsistent mapping of snow cover over the Tibetan Plateau and Himalayas.

Monthly probability maps show that snow covers extensive areas of midlatitude North America and Eurasia in 1 to 5 years out of 10, particularly in the transition seasons (Dickson and Posey 1967). The standard deviation of snow extent reaches an annual maximum in October (Hahn 1981). CO_2-induced warming effects might be expected to be especially important in higher latitudes during these transition seasons. Early snow removal from the tundra in spring, for example, has a major effect in raising summer temperatures according to energy budget model calculations by Lettau and Lettau (1975).

The effects of a general warming on snowfall and snow cover will differ according to latitude. In low and middle latitudes, where the occurrence of snow (rather than rain) is frequently marginal, warming will decrease the frequency of snowfall and the duration of snow cover on the ground, thereby shifting the wide zone of low frequency of snow cover depicted by Dickson and Posey well to the north. This zone has also been displayed in a somewhat different way by Kukla (1981, 1982). He refers to the Snow Transition Zone (zonally averaged around the Earth) delimited by the highest latitude without snow cover and the lowest latitude with snow cover on land. For Antarctica, a log-linear relationship exists between annual accumulation and mean temperature (Limbert 1984). In the subarctic of Canada there is a statistical tendency for warmer winter months to be more snowy as a result of more

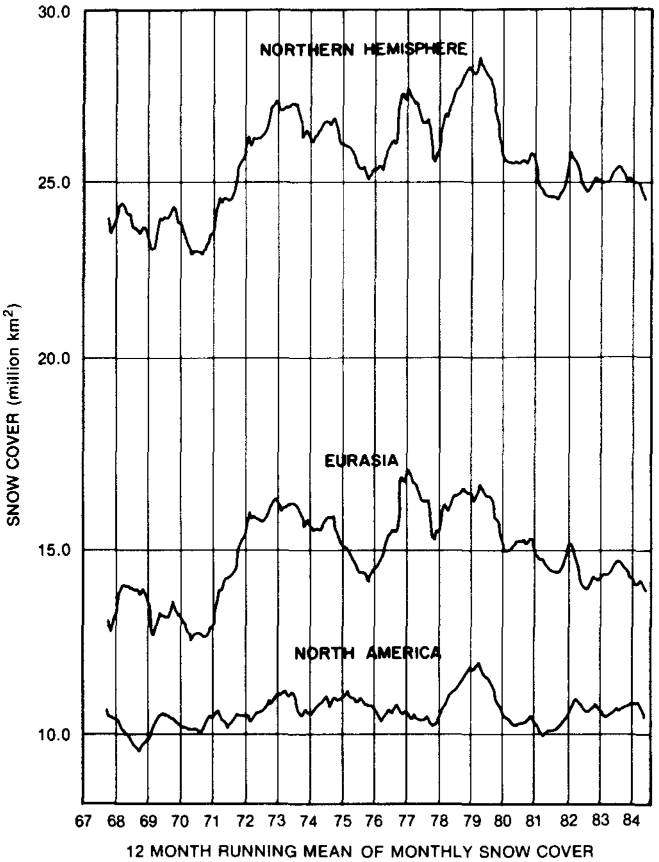

Figure 6.4. Northern Hemisphere snow-covered area, 1967–1984, smoothed by 12-month running means. Source: M. Matson (unpublished).

frequent cyclonic incursions (Brinkmann and Barry 1972), but temperature-snowfall relationships are apparently inconsistent geographically and temporally. At Fairbanks, Alaska, warmer winters tended to be drier after 1930, but were moist prior to that date (Bowling 1984). Nevertheless, temperature and precipitation at Fairbanks for January 1906–1980 showed a weak positive correlation. However, the duration of snow cover is unlikely to be substantially affected by a deeper snowpack. At Barrow, Alaska, for example 30–40 cm of snow disappears in about 10 days from the start of melt (Weller and Holmgren 1974). Warm summers in the Arctic have a greater impact on the net balance of ice caps than 3 or 4 cold and snowy years (Koerner 1980).

Observations show that the frequency of days with snow cover in Britain was at a minimum in the 1920s and 1930s, corresponding to the period of maximum warmth (Manley 1969). Records from four European stations since the 1890s indicate some regional variations in the timing of decades when maximum and minimum snow cover have occurred (Uttinger 1963), but the causes of such variations remain to be investigated. Historical records for Zurich suggest an average of 70 days with snow cover for 1683–1700 compared with 35–42 during the least snowy decades of the 20th century (Pfister 1978).

Based on the empirical data discussed here, CO_2-induced changes in the snow cover should become evident as a reduced extent of spring snow in middle and high latitude land masses. However, the variability in the location of the snow cover boundary in middle latitudes is large over Eurasia in the transition seasons according to Karapet'iants (1978). In April, the standard deviation is just over 3° latitude. Accordingly, a sustained displacement of the April snow boundary poleward by 6° to 67°N at 30°E and 6° to 61°N at 70°E would imply a statistically significant shift.

Extensive snow fields exert a significant effect on the atmosphere, initially through the cooling of the overlying air. Over the northern Great Plains, for example, surface air temperatures are some 10°C lower in the presence of a snow cover than without it (Kukla 1981). There is also a close correlation over Eurasia between the location of the 0°C isotherm and the isoline of 5- or 10-days frequency of snow cover (Afanas'eva et al. 1974). The radiative cooling of the air over snow generates a low-level anticyclonic circulation and an upper-level cold low. Lamb (1955) showed that a snow field $\geq$2500 km from west to east favors the formation of a broad cold trough above it and a warm ridge of high pressure upstream. Such a pattern tends to steer cyclones along the perimeter of the snow cover, augmenting the local snowfall (Williams 1978), and it takes a sustained and vigorous warm air incursion to remove the snow and alter the circulation regime.

A strong statistical relationship between cyclone tracks and the snow margin in the Northern Hemisphere has been identified in October and January and a weaker (nonsignificant) one in April (Carleton 1982). Moreover, during months of more extensive snow cover, Carleton finds an increased frequency of cyclogenesis, implying an augmented surface to atmosphere feedback. Correlations between the Eurasian snow line and an index of synoptic activity (the tropospheric frontal zone) are also apparent in both transition seasons according to Afanas'eva et al. (1979).

The persistence of snow cover also affects circulation and climate. Autumn snow cover affects the temperatures in the ensuing winter, with a much more pronounced effect over Eurasia than North America as a result of the large area involved and the persistence and dominant effect on circulation of the Siberian winter anticyclone (Foster et al. 1983). Conversely, the average winter snow cover is better correlated with average winter temperatures over North America than Eurasia, implying a closer response of winter temperatures to the presence or absence of snow cover. A similar analysis for the United States using data for 1949–1950 to 1980–1981(Walsh et al. 1982) finds that the persistence of snow cover anomalies is greatest in the Far West, with the maximum lag effects on temperature located in the central United States. In the zone of marginal snow cover some 10–20% of the variance of the concurrent monthly temperature, and 5–10% of that of the subsequent month, can be attributed to the snow cover distribution. The effects are largest in the late winter.

There have also been suggestions of an inverse relationship between the extent of Eurasian winter snow cover and the subsequent summer monsoon rainfall over India (Hahn and Shukla 1976; Dey and Bhanu Kumar 1982, 1983). Ropelewski et al. (1984) contend that Dey and Bhanu Kumar's results are biased by their misinterpretation of the NOAA-NESDIS data. Nevertheless, Dickson (1984) reconfirms the Hahn-Shukla finding using data for 1967–1980 which are adjusted for bias in the analyzed snow cover over the Himalayas and excluding the 1969 snow data shown to be in error; he also found a similar correlation of monsoon rainfall with total Eurasian snow cover. Another re-examination of the data for 1967–1980 by Ramage (1983) discounts the proposed relationship on the grounds that the correlations are strongest between winter snow cover and September (rather than early summer) rainfall over peninsular (rather than Northwest) India. Hence, the associations seem to lack a realistic physical linkage.

6.3 SEA ICE

6.3.1 Significance

Sea ice is the second-most spatially extensive component of the cryosphere. Figure 6.5 illustrates the seasonal variations in its extent in the two hemispheres. There is an interesting asymmetry in the seasonal trends of growth and decay; the growth stage is faster than the retreat in the Arctic Ocean, whereas the reverse is true in the Antarctic (Lemke et al. 1980). This relates to the geographical and climatic contrasts between the two polar regions. In the Arctic Ocean, pack ice limits are substantially constrained by the encircling land masses, apart from the North Atlantic sector, whereas in the Southern Ocean the ice margin is unconstrained and there is a circumglobal marginal ice zone over deep ocean waters for much of the year (Figure 6.5). In contrast to this contiguous seasonal sea ice zone around Antarctica, there is seasonal freeze-up in areas well south of the main Arctic ice in the Sea of Okhotsk, Hudson Bay, and the Baltic Sea, for example, as well as ice drift in the Labrador Current southward to Newfoundland (latitude 47°N) and wind drift of Okhotsk Sea ice to northern Hokkaido (latitude 44°N).

Sea ice insulates the atmosphere from the ocean, and its seasonal growth and decay considerably modify the ocean mixed layer. The presence of an ice cover in the Arctic Ocean and adjoining seas is one of the most important factors in the climate system of the Northern Hemisphere, and the question of its long-term stability in the event of a sustained warming trend is a key concern. Its removal, or reduction in extent, would have a significant impact on climatic and oceanic conditions.

In economic terms, Arctic sea ice is particularly significant as an obstacle and hazard to marine transportation, necessitating aerial reconnaissance, satellite monitoring, and the deployment of icebreakers in support of shipping operations. Conversely, stable, level ice affords a platform for wintertime offshore drilling operations for oil and gas exploration and for traditional hunting activities of native peoples. The ice margin is also the location of great biological productivity.

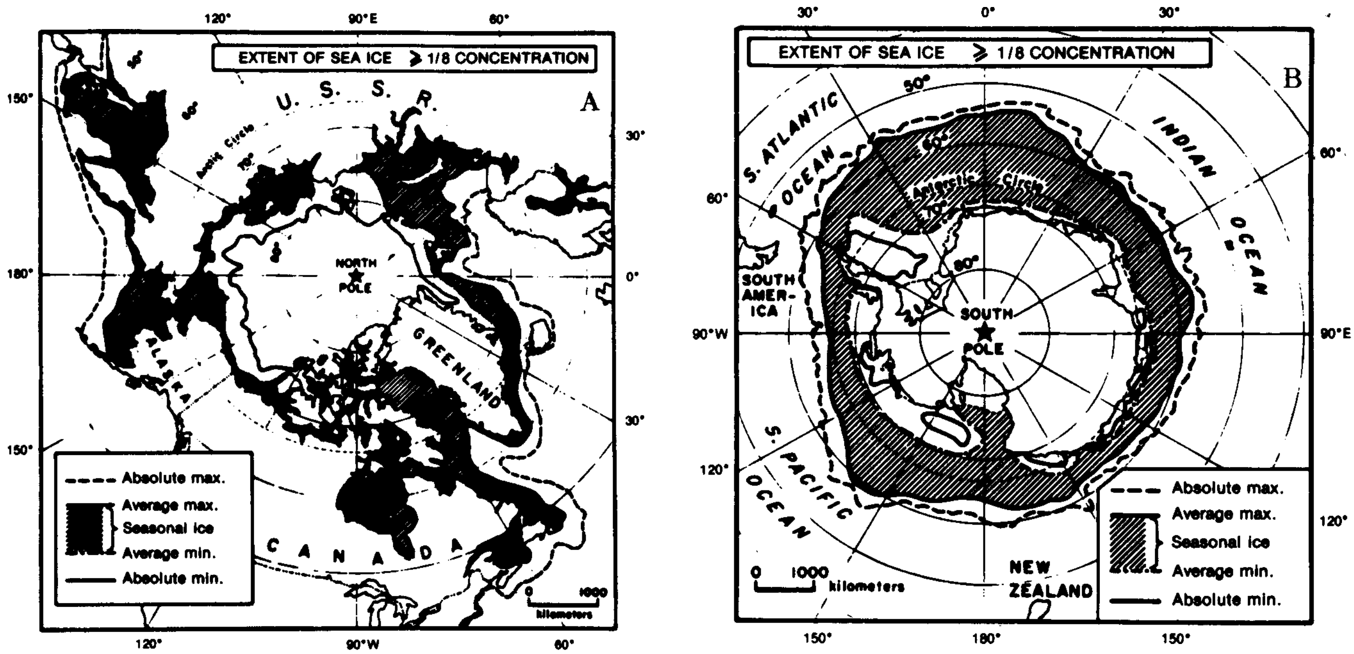

Figure 6.5. Distribution of (A) Northern and (B) Southern Hemisphere sea ice and its average and extreme seasonal variability. Source: Barry (1980), modified from Central Intelligence Agency (1978).

6.3.2 Sea Ice Observations

Sea ice limits have long been observed by ships, and harbors have reported the dates of the appearance and disappearance of coastal ice. Historical sea ice records have been used as climatic indicators by Lamb (1977) and others, although the early observations present many problems of interpretation. An overview of such data sources has been given elsewhere (Barry, in press). More recently, ice conditions have been reported regularly in the marine synoptic observations, by special aerial reconnaissance flights (as coded observer reports, photography, and remote sensing data), and by coastal radar. Since the early 1970s, satellite remote sensing, particularly NOAA Very High Resolution Radiometer (VHRR) and Nimbus microwave data, have been routinely incorporated into the U.S. Navy (now Navy-NOAA Joint Ice Center) weekly ice charts for the Northern and Southern Hemispheres (Figure 6.6).

Maps of sea ice generally depict the boundaries for various ice concentration classes (0–10/10), and they also may distinguish different age categories and the degree of ridging intensity. The complete nomenclature adopted by the World Meteorological Organization (1970) provides for the description of stages of ice development, and ice forms; processes of motion, deformation, and melting; and for types of opening. A new symbology for sea ice charts, known as the "Egg" code, allows for plotting of ice concentration, age, and floe size (Danish Meteorological Institute 1982).

6.3.3 Space-Time Coverage of Data

The longest historical records are those of drift ice duration on the coasts of Iceland (Koch 1945). Lamb (1977, p. 583) tabulates these data, with some additions, annually from A.D. 1600 to 1975; the series is believed to be complete from 1780. A new analysis of the Icelandic data has recently been completed, and a decadal ice index has been presented for 1601 to 1780 (Ogilvie 1984). A related record of the Storis drift (ice from the East Greenland Current) northward along the west coast of Greenland exists from 1820 to 1930 (Speerschneider 1931; see Lamb 1977). The annual winter maximum extent of Baltic Sea ice has been tabulated from 1720 to 1956 (Betin and Preobazhensky 1959; summarized by Lamb 1977, p. 586–589), and a brief analysis of these records for 1830–1951 has been made by Jurva (1952).

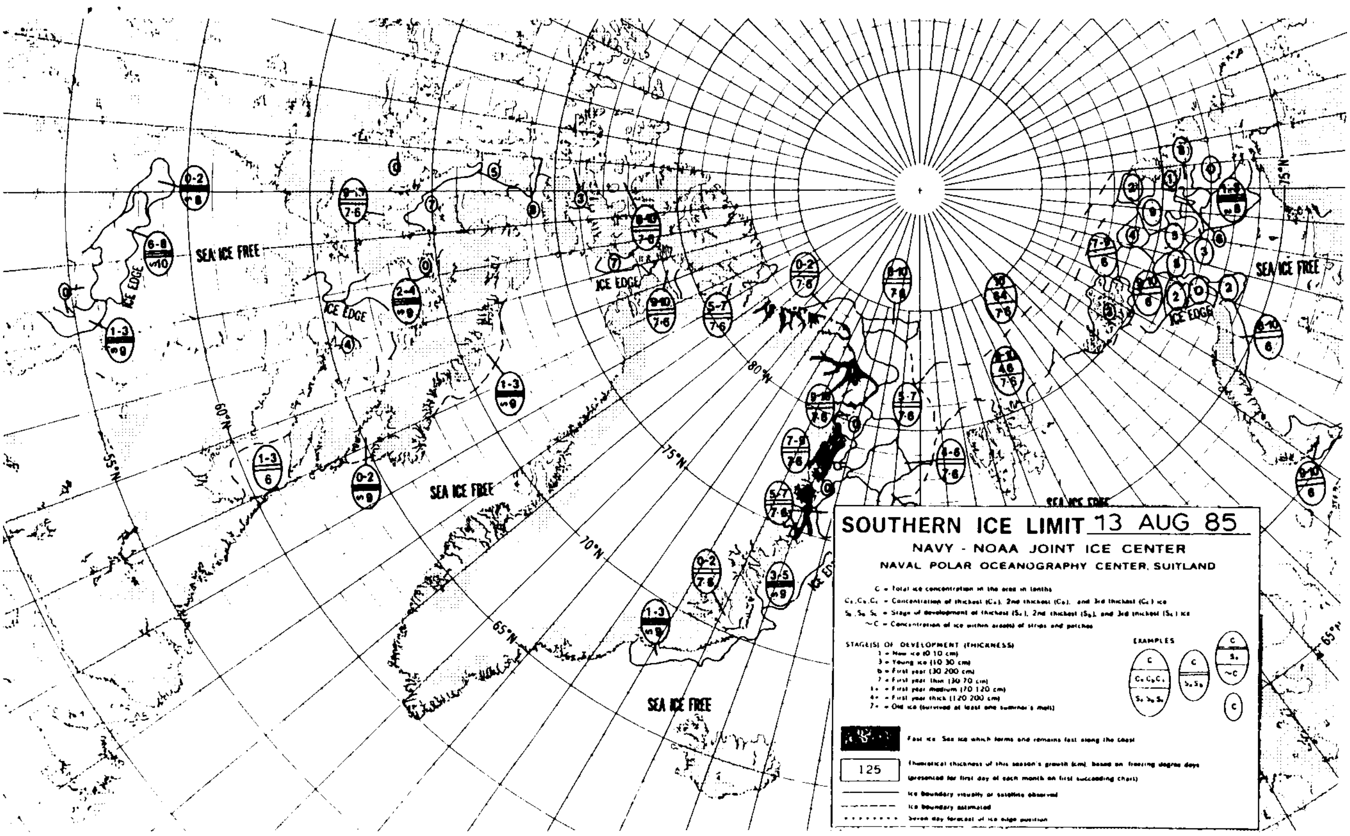

Figure 6.6. Example of a Northern Hemisphere weekly sea ice chart. Source: Navy-NOAA Joint Ice Center.

Numerous chart series now provide both monthly and approximately weekly maps of ice conditions in virtually all seasonally ice-covered sea areas (Barry, in press). The longest series was that published for the Arctic Ocean by the Danish Meteorological Institute (1901–1956) for 1901–1939 and 1946–1950, with more limited coverage of the North Atlantic sector from 1877 (Ryder 1896). The monthly ice limits for the North Atlantic from 1901 have been digitized by Kelly (1979). Problems with the early 20th century ice information are illustrated by the data on Arctic Ocean ice extent presented by Zakharov (1981). The 57-year record of ice area for late August (smoothed by a 5-year binomially weighted moving average) has been correlated (Figure 6.7) with similarly smoothed June-August temperature data for 65–85°N (Kelly and Jones 1981). There is a strong inverse correlation (−0.81) for 1951–1980, but the correlation drops to −0.13 for the whole period 1924–1980. One must conclude either, that one or both sets of data are unreliable because of the limited spatial coverage of the observations in the Arctic prior to about 1950 or that the strong statistical ice-climate relationship for 1951–1980 is fortuitous and the variables are essentially unrelated. The temperature data set is discussed by Wigley et al. (Chapter 4 of this volume). The Soviet ice data are apparently derived from the Danish charts for the Arctic, probably supplemented by other records for the Siberian Northern Sea Route.

The most consistent data on ice extent are probably those available in the Navy-NOAA weekly charts for the Arctic (from 1972) and the Antarctic (from 1973), but information on ice concentration on these maps is of variable quality. Figures 6.8 and 6.9 show time-series of the ice area data. Short-term trends are apparent and there may be a negative relationship between ice extent in the two regions, although its significance is in doubt due to the secular trends (Ropelewski 1985). The Navy-NOAA maps for the Antarctic have been digitized to provide data sets of ice extent versus longitude and ice area by several different workers, including Jacka (1983) and Ropelewski (1983). Despite the

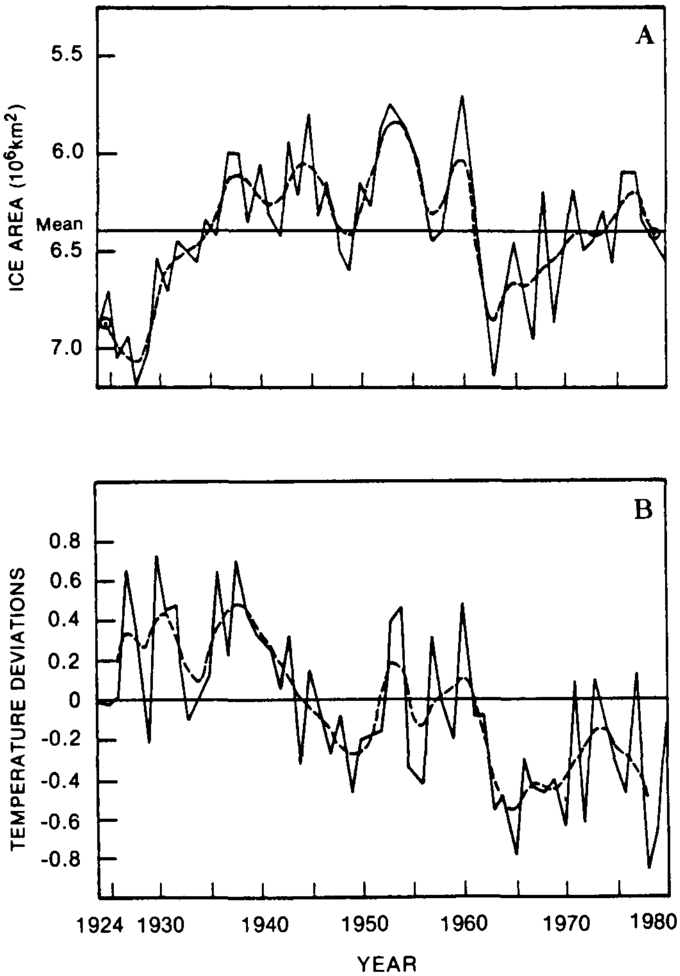

Figure 6.7. (A) Arctic Ocean ice extent in late August 1924–1980 *(from Zakharov 1981) and (B) summer temperature departures* from the 1946–1960 mean for latitudes 65°–85°N (Kelly and Jones 1981). Dashed line denotes 5-year binomially weighted running mean (Barry 1983).

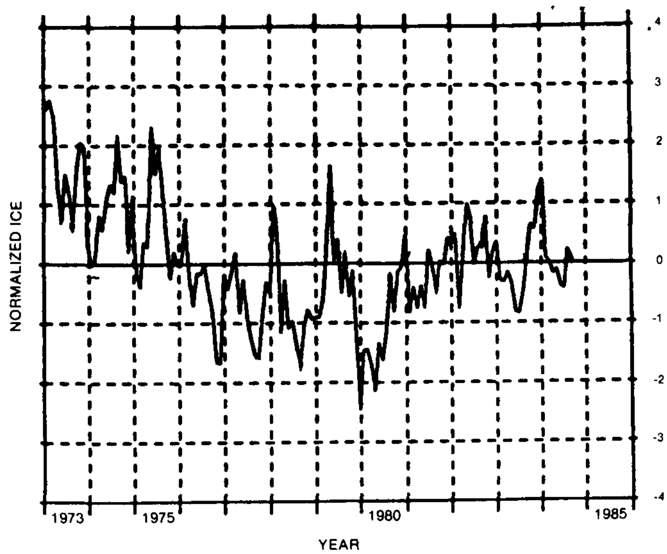

Figure 6.8. Time series of monthly standardized anomalies of sea ice area (standard deviations from the mean) for the entire Arctic region. Source: Ropelewski (1985). (Reprinted by permission of Pergamon Press, Ltd.)

common source of these data sets, there are distinct differences between them in the estimates of mean annual and monthly ice areas, especially in the transition seasons (Sturman and Anderson 1985). The availability from 1979 of the Scanning Multifrequency (dual polarization) Microwave Radiometer (SMMR) data from Nimbus 7 with 10 channels of radiometric data, is now providing new information on total ice concentration, the multiyear ice fraction, and surface temperature (Cavalieri and Zwally 1985). This type of data will continue to be available in the future via a similar system on satellites of the Defense Meteorological Satellite Program.

6.3.4 Sea Ice and Climate Interactions

Sea ice forms when the water temperature, for typical salinities (~35‰), falls to about −1.8°C. During the course of a winter, first year ice in the Arctic thickens to between 1.6 and 2.6 m if the growth is undisturbed, although areas of young ice are highly susceptible to deformation and pressure ridging. Undeformed multiyear ice has a steady-state thickness of between 2.5 and 5 m (Maykut and Untersteiner 1971) and, theoretically, can reach 12 m thickness in about 65 years (Walker and Wadhams 1979). However, field sampling suggests that only 25% of the ice is undeformed, and accordingly, the Arctic Ocean ice is a variable mixture of first-year and multiyear floes (Weeks and Ackley, in press). The age distribution and thickness are still not well known, and hence the mass balance of Arctic sea ice and its temporal variability are uncertain. In the marginal Arctic seas and in the Antarctic most of the ice is first-year ice; observations at Mawson (67°S) indicate annual thicknesses up to 1.3 to 1.6 m (Allison 1981). In the Weddell Sea, however, about 45% of the near-edge ice surveyed by Ackley (1979) had a thickness of 1.6–3.2 m, indicating multiyear ice, and a further 17% was thicker ridged ice.

A significant factor in ice-covered seas is the fraction of open water and young ice in winter. Even a small amount of open water contributes greatly to

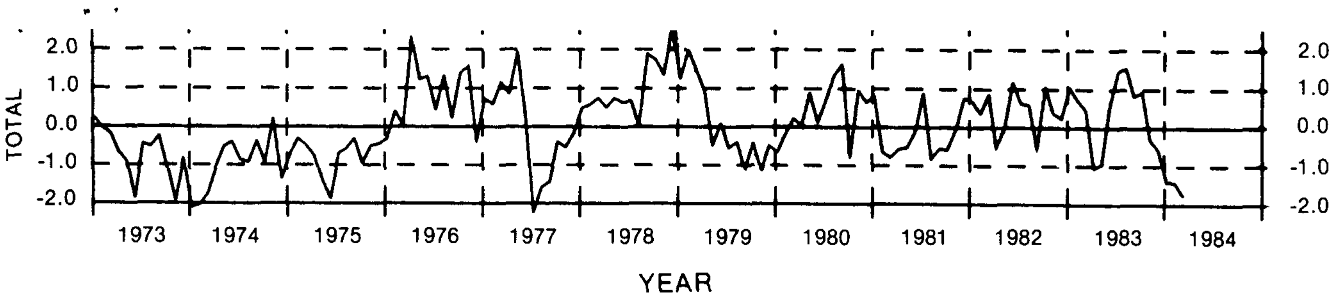

Figure 6.9. Time series of standarized sea ice areas for the Antarctic. The sea ice areas are standarized by the 1973–1984 base period mean and standard deviations. Asterisks designate September of each year. Source: Ropelewski (1985). (Reprinted by permission of Pergamon Press, Ltd.)

the ocean-atmosphere turbulent heat flux. For the Arctic, Maykut (1982) estimates that half of the regional turbulent heat flux occurs through 20–80 cm thick ice, which occupies only 6% of the ocean area in winter. Cavalieri and Zwally (1985) identify significant areas of ice divergence in the central Arctic in winter from SMMR data and show substantial interannual variations in the multiyear ice edge. In the Antarctic, a major area of low ice concentration and open water (polynya) has been observed in the eastern Weddell Sea sector during September-November (spring) in several, but not all, years with satellite microwave data (Carsey 1980). Its occurrence seems to be related to appropriate atmospheric wind regimes and their effect on ice drift and divergence (Parkinson 1983).

The problem of interactions between atmosphere and ice has been approached by descriptive synoptic case studies, by simulation experiments with circulation models, and by statistical analyses. In the last category, Lemke et al. (1980) indicate from an 11-year data set for the Arctic, and a 6-year Antarctic set, that the space-time structure of monthly sea ice anomalies can be accounted for by a combination of white-noise atmospheric forcing (from weather disturbances), seasonal forcing (representing oceanic effects in both polar regions, plus solar radiation in the Arctic), and lateral diffusion and advection of ice, particularly in the Antarctic. The location of the ice edge does respond to synoptic-scale atmospheric forcing, but the major temporal component of the variability in ice area is dominated by the seasonal cycle which is most pronounced in the Weddell Sea and Ross Sea in the Antarctic and in the Baffin Bay and Barents Sea sectors of the Arctic (Lemke et al. 1980). These sections also show the greatest interannual variability.

The dramatic seasonal expansion and contraction of the Antarctic sea ice has been explained by Gordon and Taylor (1975) and Gordon (1981) in terms of (1) the Ekman divergence due to the curl of the wind stress, and (2) the surface heat balance. The first factor is associated with divergence-generating open water areas within the ice. In early winter, open water will tend to freeze over, and the ice therefore expands northward. In summer, open water will provide a heat source and encourage lateral ice decay, although Gordon's (1981) calculations suggest that only half of the required heat flux can be provided from the atmosphere whereas the remainder must come from deep water below the relatively weak pycnocline. Ackley (1981) states that the evidence relating to the ice advance is supportive of the mechanism proposed by Gordon and Taylor (1975), but a modeling simulation by Hibler and Ackley (1983) indicates greater complexity. Their results suggest that the large seasonal fluctuation in ice extent in the Weddell Sea is associated with the ice dynamics. However, it is the rapid decay that depends critically on the role of leads and lateral ice advection, whereas the ice advance is primarily associated with freezing *in situ*, except in the western sector of the Weddell Sea.

Hibler and Walsh (1982) are currently examining the response of an Arctic Ocean ice model to atmospheric forcing using observed daily data from the National Meteorological Center sea level pressure grids and modified temperature grids. A 20-year simulation (Walsh et al. 1984) reproduces ice anomalies that broadly resemble the observational record of Zakharov (1981). However, the model gives an excess of simulated ice in the North Atlantic in winter (as a result of absence of oceanic heat transport in the model) and excess melt in the central Arctic in summer (due to the lack of an explicit formulation of snow cover). Subsequently, Hibler and Bryan (1984) have shown that more accurate simulation of the ice edge location is obtainable given the coupling of the thermodynamic-dynamic ice model with a full ocean circulation model.

The causes of anomalies in ice extent in the North Atlantic sector have been examined on synoptic and monthly time scales by many investigators (e.g., Strübing 1967; Vinje 1977). Aagaard (1972) has shown that in years with high pressure over Greenland in late-winter and spring, there is enhanced southward current and ice transport off east Greenland because of the anticyclonic curl of the wind stress; according to Vowinckel (1963), two-thirds of the total ice export is due to the current and only one-third is due to wind-induced export. He also shows from correlation analysis of ice extent between April and August, and August with the following April, that freeze and melt processes tend to dampen variations caused by ice transport.

In the Sea of Okhotsk, where the effect of ocean currents is weak, the early or late appearance of ice on the north coast of Hokkaido is dependent mainly on the strength of the northerly wind components (Akagawa 1973). More zonal circulation is associated with late ice appearance. Pack ice disappearance is apparently more related to wind direction, with west-southwesterly winds leading to early ice removal.

The effects of sea ice anomalies on the atmosphere have been examined empirically by Wiese (1924) and Johnson (1980) and in a GCM experiment by Herman and Johnson (1978). Wiese showed that, in the North Atlantic, cyclone tracks are further south in years with heavy ice than in years with light ice. This result has also been found by Herman and Johnson using the Goddard Laboratory for Atmospheres (GLA) GCM. They compare a mean of two mid-winter simulations for maximum ice in the northern oceans with the mean of six simulations given minimum mid-winter ice extent as a control experiment. They demonstrate both local and hemispheric effects on the atmospheric circulation. With maximum sea ice there is an increase of sea level pressure over the ice in the Barents Sea, Sea of Okhotsk, and Davis Strait areas and a southward shift of cyclones in the North Atlantic, giving increased precipitation over northern Europe. On the hemispheric scale, deeper lows around Iceland and the Gulf of Alaska at 700 mb are coupled with stronger subtropical anticyclones and enhanced poleward energy flux in midlatitudes. Reduced ice would imply changes in the opposite sense. Herman and Johnson note, however, that the model omits feedback processes, including changes in sea surface temperatures. Moreover, observed ice anomalies do not tend to occur simultaneously in all sectors (Walsh and Johnson 1979a; Lemke et al. 1980; Smirnov 1980; Walsh and Sater 1981). In another data study, Johnson (1980) compared 5 years each of heavy and light Arctic ice with the concurrent winter circulation and found similarity with the GLA GCM study for the Atlantic, but less agreement for the Pacific. She notes the importance of regional links between the atmosphere and ice extent, with northerly airflows causing ice advection, in Davis Strait and the Bering Sea, but indeterminate relationships for the sector east of Greenland, as implied by the empirical studies described earlier. Oceanic anomalies in the northwestern North Atlantic are of importance for climatic anomalies affecting Europe, and a major program—the Marginal Ice Zone Experiment (MIZEX)—is being conducted in this region (Wadhams et al. 1981).

A detailed analysis of Northern Hemisphere synoptic-scale ice atmosphere interactions has recently been completed by Carleton (1985). Based on satellite imagery analysis of cloud vortex systems for 2 years of mid-season months he shows that, for the hemisphere as a whole, cyclones either are displaced equatorward or are more numerous in years when sea ice is more extensive in January and April. A reverse pattern is observed with the sea ice margin in July and October, however, but the distribution of cyclonic activity in October shows that surface-atmosphere interactions in the autumn occur in relation to the continental snow cover margin. These results confirm the earlier findings of Wiese (1924), although analysis of more years is needed to establish the variability of such interactions.

Results of recent studies of satellite data indicate a reduction in Antarctic sea ice compared with ship observations in the 1930s. Kukla and Gavin (1981) have suggested that the apparent decrease in summertime sea ice extent in the 1970s in the sector

35°W–115°E compared with the 1930s may be associated with a warming trend in Antarctic temperatures. Antarctic stations show a 0.6°C rise for 1958 to 1978, but the changes are still within one standard deviation of the long term mean (Budd 1980). Although these changes are consistent with the postulated CO_2-induced warming effect, they are still within the range of natural variability. Continued monitoring of sea ice extent over the next 10–15 years should establish whether or not these trends are significant. Zwally et al. (1983a) have shown that there is no long-term decrease in Antarctic sea ice extent over the 1973–1981 satellite record. Regional interannual variability is found to be large, and changes in atmospheric circulation or other climatic factors seem more likely to be responsible for the changes in ice conditions (Cavalieri and Parkinson 1981; Zwally et al. 1983b). As a result of this large interannual variability, the Committee on Glaciology (National Research Council [NRC] 1984a, p. 36) suggests that "the signal-to-noise ratio of the southern sea ice extent may not be large enough to provide a useful tool to detect the early development of CO_2 effects."

Using empirical data for Laurie Island (Scotia Bay, 60°S), Budd (1975) calculated that a ±1°C change in annual mean temperature corresponds to a ±70-day range in the duration of sea ice at its maximum extent. Assuming this to be generally representative for the Southern Ocean, Budd indicated that a 1°C increase in mean annual temperature corresponds approximately to a 2.5° latitude retreat at latitude 67°S. A 1° latitude change in the ice limit represents a change of about 15% in hemispheric sea ice cover according to Streten (1973). However, Raper et al. (1984) have shown that ice conditions at Laurie Island are poorly correlated with hemispheric ice anomalies. Ice extent is correlated significantly with temperatures in spring when the ice is near its maximum extent but the correlation is *positive*, not inverse as proposed by Budd. Ice duration at Laurie Island seems to depend particularly on the development of the pressure trough in the lee of the Andes, with longer duration associated with a deep trough and with weaker sea level westerly winds (Rogers 1983). Accordingly, the relationships between climate and hemispheric ice extent must be complex.

In the Northern Hemisphere, where the overall pattern of ice extent is strongly influenced by the interactive oceanic and atmospheric circulations, most investigations have been regional in character, although Zakharov (1981) provides an overview of Russian work on long-term ice-climate trends in the Arctic. For the Arctic Ocean ice in summer, the data of Zhakarov suggest an average decrease of about 10^6 km^2 in annual minimum ice area for each 1° rise in summer temperature in the zone 65–85°N. Soviet workers have found that interannual Arctic ice anomalies correlate more highly with temperature anomalies in summer than in the cold season (Zakharov and Strokina 1978). Walsh and Johnson (1979b) have also shown the strongest tendency for ice anomalies to respond to atmospheric forcing between late February and late July with lags of up to several months.

The north polar region experienced significant warming in the 1920–1930s with a three-fold amplification above the Northern Hemisphere annual average (Kelly et al. 1982). Sea ice observations in the Soviet Union indicate a northward shift of the summer pack ice limits in the Barents-Kara-Laptev seas (Zakharov and Strokina 1978) but, as discussed above (cf. Figure 6.7), data limitations prevent the establishment of long-term climate-ice relationships over a longer than 30-year time scale.

Simple estimates of changes in sea ice thickness that are due to warming can be based on empirical relationships with freezing degree days (Bilello 1961). Under present winter conditions at Barrow, Alaska, Bilello's equation gives an ice thickness of 247 cm. Assuming an ice growth period of 270 days, a 5°C rise would lead to reduction of 15%, and a 10°C rise a reduction of 34%, in ice thickness. Analogous empirical relationships have been developed for late summer retreat of the pack ice margin off Barrow. There is a correlation of 0.81 between distance to the limit of 4/8 ice concentration north of Point Barrow on 15 September and the thawing degree day total (see Barry 1977). The mean distance for 1953–1975 was 155 km. A rise in mean summer temperature of 5°C would increase this to 600-km, if the relationship is linear, and a 600 km retreat around the Arctic Basin would leave a core of pack ice only about 1500 km in diameter. A similar calculation was made earlier by Zubenok (1963) for

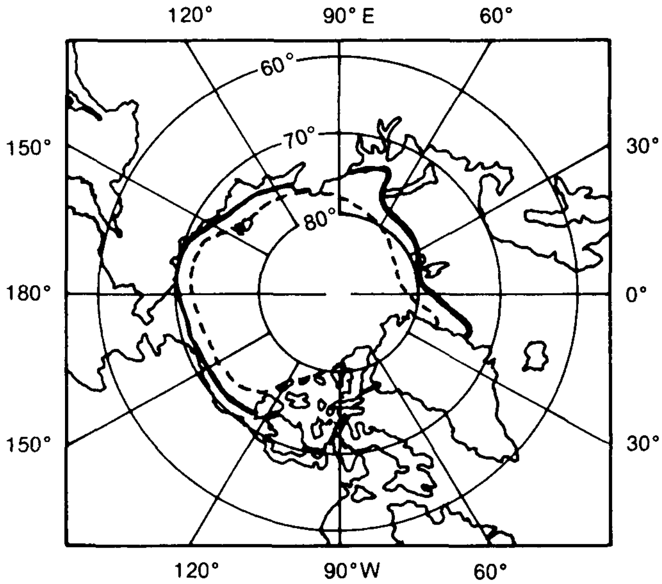

Figure 6.10. Present limit of summer Arctic ice (—) and estimated limit for a 2°C summer warming (- - -). Source: Zubenok (1963).

a 2°C summer warming; Figure 6.10 illustrates his results.

The long-term stability of Arctic and Antarctic pack ice to climatic change has been considered from a modeling standpoint by Parkinson and Kellogg (1979) and Parkinson and Bindschadler (1984). By using a large-scale thermodynamic-dynamic seasonal sea ice model with mean monthly atmospheric forcing (temperature, wind stress), a reasonable simulation of the annual ice cycle has been achieved (Parkinson and Washington 1979). When a 5°C warming is incorporated, the Arctic pack ice is reduced (after 5 model years) to a seasonal ice cover which is absent in summer but reforms in the winter. However, in this model, which has limited treatment of ice dynamics and dynamical interactions, the ice limit is unrealistically forced by the monthly mean air temperature. The Antarctic experiments (Parkinson and Bindschadler 1984) use a modified version of the model. The sea ice area decreases by about half for a 5°C warming, essentially disappearing in summer; retreat rates are nonlinear with change of temperature, with the sensitivity decreasing at higher temperatures (partly as a result of the smaller perimeter of the remaining winter ice). These results are suggestive of the qualitative effects to be expected for a CO_2 doubling, but coupled ice-ocean-atmosphere models are required to establish the quantitative effects more precisely.

Experiments with a GFDL model version, including seasonal solar radiation variations, realistic geography, and a simple ocean mixed-layer, for a quadrupling of CO_2, show strong seasonal differences in warming effects (Manabe and Stouffer 1980). With this model, the Arctic Basin experiences the greatest temperature effect in early winter, thereby reducing the annual temperature variation, with a secondary warming in April around 65°N. The sea ice cover is greatly reduced in extent and thickness with no ice present in the Arctic from August through October (Figure 6.11). However, the experiment is still not fully realistic, particularly with regard to prescribed clouds (see Shine and Crane, 1984), surface albedos, and sea ice thermodynamics.

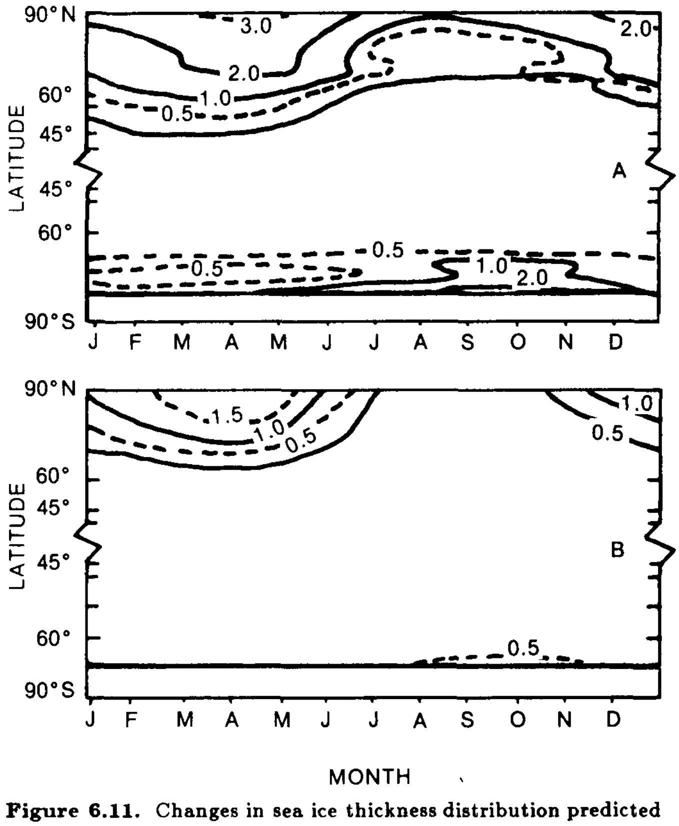

Figure 6.11. Changes in sea ice thickness distribution predicted by the GFDL general circulation model for (A) one times and (B) four times the present CO_2 concentration. Source: Manabe and Stouffer (1980).

The GFDL and most other GCMs essentially treat none of the sea ice thermodynamical processes. Semtner (1984) demonstrates that the use of his 1976 zero-layer model, which neglects heat storage in the ice but adjusts the ice conductivity and albedo values to obtain a correct mean annual ice thickness, gives significant errors in the annual amplitude of ice thickness and its phase. Semtner describes a detailed three-layer ice model that includes the effects of latent heat storage (in brine pockets in the ice) and of the cold content (sensible heat) of the snow and ice in spring. For a CO_2 quadrupling, simulations with this model did not lead to total melting of the ice at latitude 85°N in summer, in contrast to the Manabe and Stouffer (1980) results shown in Figure 6.11. Semtner's three-layer ice model also shows a more realistic simulation of the Arctic ice regime (locally) for the present CO_2 level.

In the current GCMs that treat sea ice, the prediction of sea ice extent is generally successful in the Arctic, but is much less realistic in the Southern Ocean. The errors here are considerably larger if the ocean heat transport is omitted as in the work of Manabe and Stouffer (1980). A major problem in modeling Southern Ocean ice is the poor simulation of Antarctic climatology in general by current atmospheric GCMs (Schlesinger 1984). Sea ice extent and its seasonal variation are well portrayed in the Goddard Institute for Space Studies (GISS) GCM of Hansen et al. (1983), and the modeled surface energy budgets in the Arctic compare well with the observational data set used by Maykut and Untersteiner (1971), as discussed by Barry et al. (1984).

6.4 FRESHWATER ICE

6.4.1 Significance

Ice forms on rivers and lakes in response to seasonal cooling. The sizes of the ice bodies involved are too small to exert other than localized climatic effects. However, the freeze-up and break-up processes respond to large-scale and local weather factors, such that considerable interannual variability exists in the dates of appearance and disappearance of the ice. This is of economic significance in connection with hydroelectric facilities, river traffic, sports fishing, and winter recreation. From an assessment point of view, long series of ice observations can serve as a proxy climatic record, and the monitoring of trends in freeze-up and break-up may provide a convenient integrated and seasonally specific index of CO_2-induced climatic perturbations. Information on river ice conditions is less useful as a climatic proxy because ice formation is strongly dependent on the river flow regime which is affected by precipitation, snow melt, and watershed runoff as well as being subject to human interference that directly modifies the channel flow or that indirectly affects the runoff via land use practices.

6.4.2 Observations

Data tabulations usually refer to the establishment and disappearance of a stable ice cover, and some also give the first appearance and initial break-up of ice. Problems exist in the somewhat subjective nature of these observations[3] and, for large lakes, in the uncertain range of view from the observing point. In older records, the precise location of observations may be unknown and the site may have been moved. The establishment of ice cover on a lake is influenced not only by climatic factors but also by topographic location, surface area (wind fetch), depth, and water inflow and outflow. This information—especially lake depth—is commonly not available, although, according to Mellor (1983), some indication of the depth of shallow lakes in the Arctic can be obtained from airborne radar imagery during late winter. The timing of break-up is modified by snow depth on the ice as well as by ice thickness. Altitude effects on mean freeze-up and break-up dates have been examined for lakes in the eastern Alps by Eckel (1955). On average, freeze-up is 4 days earlier per 100 m increase in altitude and break-up is 7 days later.

The appearance and disappearance of lake ice is readily observed by satellite. Even small lakes (a few km^2) can be monitored because the freeze-up usually occurs after the development of a snow cover and the breakup after the disappearance of the snow. Up to now, however, this technique has not been systematically applied, although the use of aerial reconnaissance for lake ice conditions in subarctic Canada was developed in the 1960s at the University of Wisconsin (McFadden 1965).

[3] B. Singh, 1973. "Break-up and Freeze-up Dates: A Case Study of Inconsistencies in Climatological Data." Unpublished M.A. thesis, University of Manitoba, Winnipeg.

6.4.3 Space-Time Coverage of Data

Apart from a few isolated long records of lake and river freeze-up and break-up in Europe and Japan, most observations span about 30–100 years. There are extensive lake ice records from Finland and more recently from Canada (Allen 1977). The longest lake records that have been analyzed are those of freeze-up of Lake Suwa, Japan, from the mid-15th century (Tanaka and Yoshino 1982) and of Lake Kallavesi, Finland (Simojoki 1961). In North America, there are records for lakes in Wisconsin (Wing 1943), although these may lack homogeneity, and also in Maine (Fobes 1948). Records of river freeze-up and break-up exist for the Neva at St. Petersburg (Leningrad) for 1718–1834 (Jackson 1835) and the Memel, 1811–1914 (Sperling 1956), but these have not yielded climatic assessments.

Lake records in North America and Eurasia are potentially of considerable value as they provide data over a wide latitudinal and longitudinal range where the meteorological station network is sparse.

6.4.4 Lake Ice-Climate Interactions

Lake freeze-up depends on the heat storage in the lake and therefore its depth, the rate and temperature of any inflow, and water-air energy fluxes (Michel 1971). Useful empirical relationships have been developed between temperature indices and the date of ice formation (Simojoki 1940; Williams 1971). Simojoki reported a detailed study of data from Finnish lakes for 1892–1931. He determined the correlation of freeze-up and break-up with pentad temperatures, individually and cumulatively, for up to 50 days before these events. He also examined the role of other meteorological and non-meteorological factors. In general, the mean daily temperature is found to correlate well with freeze-up and break-up, although Williams (1971) used freezing degree days. Thawing degree days are found to be less satisfactory for break-up, especially where river inflow and wind are important factors or where there is considerable year-to-year variability in the depth of snow pack on the ice.

For 27 lakes widely distributed across Canada, Tramoni et al. (in press) have carried out regression analyses of freeze-up and break-up dates in relation to several different temperature indices for nearby meteorological stations. They found that freezing degree day totals gave the highest correlations (0.42 to 0.81) for 13 lakes, whereas the number of days below freezing gave the highest correlations (0.35 to 0.95) for 11 lakes located mostly in the prairies and boreal forest zones. In general, the date of break-up correlates less well with the temperature indices, but it is less likely to show extremes and, consequently, may provide a more coherent climatic index. Preliminary analyses of Canadian and Finnish data indicate considerable decadal variability in freeze-up and break-up.

There is clearly a useful potential for lake ice cover data to serve as an integrative index of seasonal temperature trends, particularly for the transition seasons. More work is needed, however, to convert the results described above into predictive indices in the reverse sense. Based on the study of Tanaka and Yoshino (1982) for Lake Suwa, Japan, a range of about 60 days in freeze-up date corresponds to a 4°C range in December-January mean temperature. Simojoki (1959) showed about a 10-day advance in the break-up of Lake Kallavesi, Finland, in the 1940–1950s compared with the 1860s; this was associated with a warming of about 2°C. For lakes in Canada (45–53°N) and Finland (61–65°N) a 1°C temperature change for the 80 days before the normal freeze-up and break-up would delay this event by about 9 days (Palecki et al. 1985).

6.5 GROUND ICE AND PERMAFROST

6.5.1 Significance

In high latitude and high mountain areas where the mean annual air temperature is below about −1°C, a layer in the ground will remain frozen throughout the year, overlain by a seasonally frozen active layer that thaws in summer (French 1976). The global extent of perennially frozen ground (permafrost) is still not completely known, but it underlies approximately 20% of the northern land area (Gavrilova 1981; Péwé 1983). Figure 6.12 shows the general distribution. Thicknesses exceed 600 m along the Arctic coast of northeastern Siberia and Alaska, but toward the margins, permafrost becomes thinner and horizontally discontinuous. The marginal zones will be more immediately subject to any melting

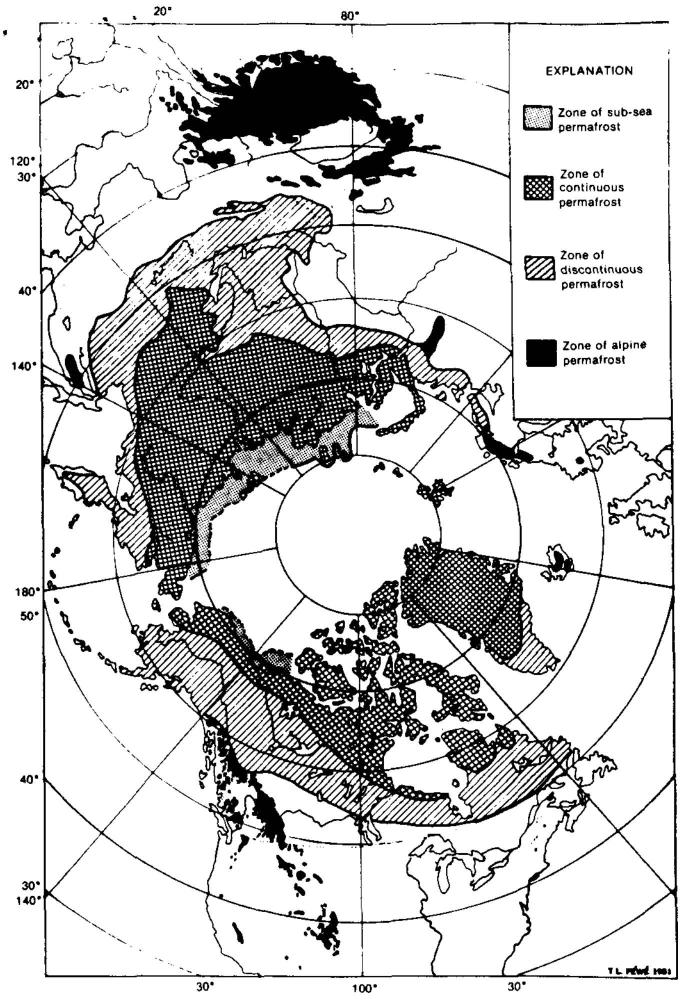

Figure 6.12. Distribution of permafrost in the Northern Hemisphere (Péwé 1983). In the Southern Hemisphere, small areas of alpine permafrost exist along the Andean Cordillera and all of Antarctica is believed to be underlain by permafrost.

caused by a warming trend (Brown and Andrews 1982).

Only a fraction of the permafrost zone consists of actual ground ice. The remainder (dry permafrost) is simply soil or rock at subfreezing temperatures. The ice volume is generally greatest in the uppermost permafrost layers (Pollard and French 1980) and mainly comprises pore and segregated ice in earth material. Estimates of ice volume by Shumskiy and Vtyurin (1963) were in the range 0.2–0.5 × 10^6 km^3. Newer data have been used to determine the equivalent global sea level contribution (Table 6.1). Estimates of permafrost area and ice volume for northern Alaska (Brown 1968), the Mackenzie delta (Pollard and French 1980), and western Siberia (Vtyurin 1978) indicate conversion factors for volume-area of 0.0035, 0.0043, and 0.0027, respectively. Use of a value of 0.0027 for continuous permafrost, and weighting the thinner discontinuous permafrost additionally by 0.25, gives a global ground ice volume of 0.033 × 10^6 km^3 and a sea level equivalent of 8 cm. The higher value given in Table 6.1 is based on the maximum volume-area ratio cited by Vtyurin (1978) for western Siberia. The time scale for release of this contribution as a result of a warming trend would be a few hundred years because most of the ice volume is near the surface.

Across much of Alaska, Canada, northern and western China, and Siberia, including the Arctic offshore areas of subsea permafrost, permafrost is a major environmental factor. Seasonal frost heaving of the active layer and the induced melting of the permafrost by buildings, highways, and other structures, as well as by removal of vegetation cover (Grave 1983), present serious problems for all kinds of construction unless special techniques are used. In addition, mining, oil and gas drilling, and agricultural activities are hampered, and water supply and sewage disposal problems augment the costs of northern development projects.

6.5.2 Observations

Information on permafrost occurrence has been assembled by geographers and geologists from field observations, with extensive recent data collected through mining exploration, engineering construction activities, and geophysical surveys in the Arctic and sub-Arctic. However, the boundaries of discontinuous and patchy permafrost are still subject to revision, especially in mountain areas and offshore. Reliable maps of permafrost extent date from about 1950, and therefore large-scale temporal changes are scarcely detectable. Permafrost thickness is determined from bore holes drilled in connection with mining or other engineering activities (Osterkamp and Payne 1981), althouth much of this information remains proprietary. Thickness also can be estimated from knowledge of the geothermal temperature gradient and the mean annual ground (or air) temperature.

Measurements of bore hole temperatures in permafrost can be used as an indicator of net changes in temperature regime. Gold and Lachenbruch (1973) infer a 2–4°C warming over 75–100 years at Cape

Thompson, Alaska, where the upper 25% of the 400-m thick permafrost is unstable with respect to an equilibrium profile of temperature with depth (for the present mean annual surface temperature of −5°C). Maritime influences may have biased this estimate, however. At Prudhoe Bay similar data imply a 1.8°C warming over the last 100 years (Lachenbruch et al. 1982).

6.5.3 Space-Time Coverage of Data

As a result of the long history of Russian and later Soviet exploration, settlement, and scientific study in Siberia, there is considerable information on Eurasian permafrost extent and thickness. Conditions in Tibet and western China, however, are less well known. Permafrost studies only began in the 1940s in North America. Information for mountain areas is sketchy (Harris 1979; Péwé 1983), and offshore subsea permafrost in the Arctic is still being mapped (Vigdorchik 1980a,1980b; Neave and Sellman 1983). The mapping of permafrost distribution and thickness still must rely heavily on climatic predictive methods. Conditions beneath the Greenland and Antarctic ice sheets are known primarily by inference and ice modeling calculations.

Geothermal data for Canada are routinely published by the Earth Physics Branch of the Department of Energy, Mines and Resources, and there are no other comparable readily accessible archives.

6.5.4 Permafrost-Climate Interactions

Permafrost may occur where mean annual air temperatures (MAAT) are less than −1 or −2°C and is generally continuous where MAAT ≤ -7°C. In addition, its extent and thickness are affected by moisture content in the ground, vegetation cover, winter snow depth, and aspect. Much of the permafrost that presently exists has formed during previous colder conditions and is therefore relic, but it is also forming under present-day polar climates where glaciers retreat or land emergence exposes unfrozen ground. Washburn (1973) concluded that most continuous permafrost is in balance with the present climate at its upper surface, but changes at the base depend on the present climate and geothermal heat flow; in contrast, most discontinuous permafrost is probably unstable or "in such delicate equilibrium that the slightest climatic or surface change will have drastic disequilibrium effects" (Washburn 1973, p. 48). However, in northern areas, deepening of the summer active layer may also have significant impacts (Goodwin et al. 1984). Thawing and retreat of permafrost has been reported in the upper Mackenzie Valley (Mackay 1975) and along the southern margin of its occurrence in Manitoba (Thie 1974), but such observations are not readily quantified and generalized. Based on average latitudinal gradients of air temperature, an average northward displacement of the southern permafrost boundary by 150 ± 50 km could be expected for a 1°C warming. Harris (in press) considers that by using available ground temperature data and a simple classification of stability, it is feasible to map zones of permafrost that are stable (temperatures below −5°C), metastable (−2° to −5°C), and unstable (above −2°). Figure 6.13 shows that much of Canada is underlain by unstable permafrost, including not only most of the discontinuous zone, but also some continuous permafrost west of the Hudson Bay.

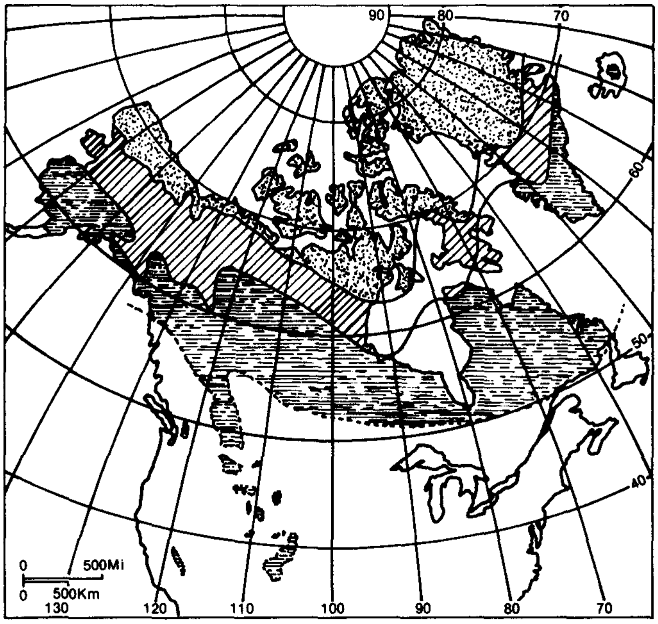

Figure 6.13. The distribution of zones of stable and unstable permafrost in North America. Source: Harris (in press).

The sensitivity of permafrost thawing to the seasonal pattern of climate change has been addressed by Smith and Riseborough (1983) using a microclimate model. Degradation is shown to be most rapid for a uniform warming in all months, whereas winter warming has less effect due to the insulating effect of snow cover. Using a one-dimensional surface-equilibrium temperature model for a representative soil profile at sites devoid of trees at Barrow and Fairbanks, Goodwin et al. (1984) simulated the effects of seasonal changes in temperature, winter snow depth, and summer dryness on thaw depth and ground temperatures. At Barrow, a 6°C summer-only warming is shown to be more effective in increasing thaw depth than a 3°C year-round warming, whereas at Fairbanks, where there is a deeper active layer and long snow-free season, the two cases are equally effective. In general, the sensitivity of thaw depth to a climatic warming increases northward. Changes in snow depth, between a 50% reduction and a doubling, give only small changes in thaw depth, but drier summers resulting in higher soil temperatures would greatly degrade permafrost in interior Alaska.

The potential rates of permafrost thawing have been established by Osterkamp (1984) to be of the order of 2 centuries or less for 25-m-thick permafrost in the discontinuous zone of interior Alaska, assuming warming from −0.4 to 0°C in 3–4 years, followed by a further 2.6°C rise.

6.6 GLACIERS AND ICE SHEETS

6.6.1 Significance

Ice sheets are the greatest potential source of global freshwater, holding approximately 88% of the global total. This corresponds to 80 m of world sea level equivalent, with Antarctica accounting for 90% of this, Greenland almost 10% and other ice bodies and glaciers less than 0.5% (Table 6.1). Because of their size in relation to annual rates of snow accumulation and melt, the residence time of water in ice sheets is of the order of 10^5–10^6 years. Consequently, any climatic perturbations produce slow responses on the time scale of glacial-interglacial periods. Valley glaciers respond rapidly to climatic fluctuations, and significant changes in length and volume occur on the time scale of a few decades. However, glaciers have no feedback effects on global climate, although their recession may have contributed one third to one half of the observed 20th century rise in sea level (Meier 1984). Rapid glacier recession may lead to significant long-term hydrological effects where the runoff from glacierized basins is used for irrigation or hydropower.

A major concern, first addressed by Mercer (1978), involves the possibility of a collapse of the West Antarctic ice sheet, which is grounded on bedrock below sea level (Figure 6.14), leading to a potential rise in world sea level of 6–7 m in a few hundred years with flooding of coastal lowlands (Schneider and Chen 1980).

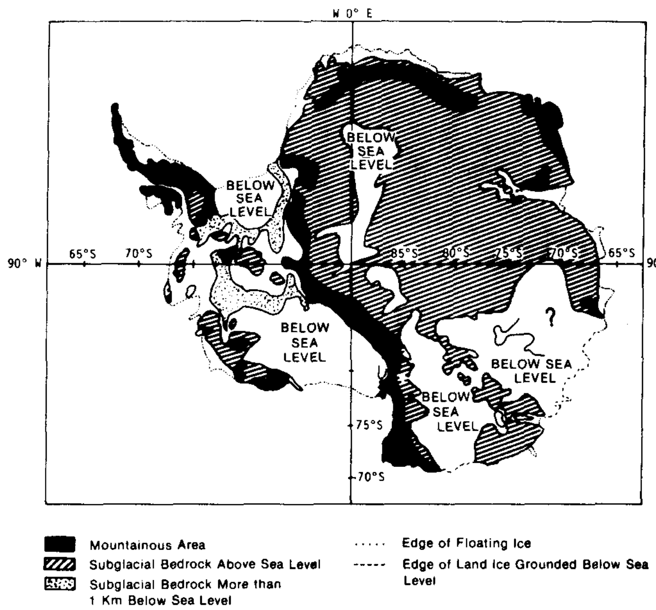

Figure 6.14. Generalized map of subglacial topography in Antarctic showing extensive areas below sea level. Source: Bentley (1983).

6.6.2 Observations

Direct observations of land ice extent have been made primarily during this century, and in only a few cases are changes in ice masses routinely monitored. The European Alps, where the changes of a few glaciers have been documented over several centuries, is an exception. Measurements of ice volume and changes in mass balance are considerably scarcer, and prior to the beginning of the International Hydrological Decade (1965) there are records for only 37 glaciers worldwide (Meier 1984).

Ice volume can be determined on the basis of airborne radio-echo sounding transects of bedrock elevation (Robin 1975); these now cover much of Greenland but only about 25% of Antarctica. The standard techniques cannot be used on temperate glaciers because of meltwater effects. Conventional mass balance studies have been performed by measuring winter accumulation and summer ablation (UNESCO/IASH 1970d), but the field techniques are slow, approximate, and inapplicable to large ice sheets, in which iceberg calving is a major contributor to mass loss. For this reason, we do not know whether the Greenland and Antarctic ice sheets are growing or shrinking. The Greenland ice sheet seems close to a balanced state, whereas computations for the Antarctic ice sheet indicate that it may be growing (Meier 1983; Bentley 1984). Satellite techniques to determine the elevation of an ice sheet with high vertical and horizontal resolution via radar altimetry have been developed, but up to now the satellite coverage has provided only for the mapping of southern Greenland and East Antarctica north of 72°S (Zwally et al. 1983c). Repeated surveys, at intervals of about 10 years, would make a major contribution to resolving much of the present uncertainty concerning ice volume changes.

6.6.3 Space-Time Coverage of Data

Data on glacier balance and volume change spanning more than 50 years are available for about 25 glaciers (Müller 1977; Meier 1983, 1984), but these are mostly in midlatitudes of the Northern Hemisphere. Direct measurements of mass balance began only in 1945–1946. There are more numerous and longer records of glacier advance and retreat (Müller 1977), but these are unreliable indicators of volume change. Composite statistics of such advances and retreat for the Alps (Figure 6.15), however, provide a general measure of the response of glaciers to climatic fluctuations (Hoinkes 1968; Gamper and Suter 1978). On the geological time scale, the advances of glaciers during colder epochs are quite well documented, but the maximum retreat of ice during the Holocene is poorly determined because of subsequent ice advance and the covering or obliteration of the geomorphological evidence (Hollin and Barry 1979; Gibbons et al. 1984).

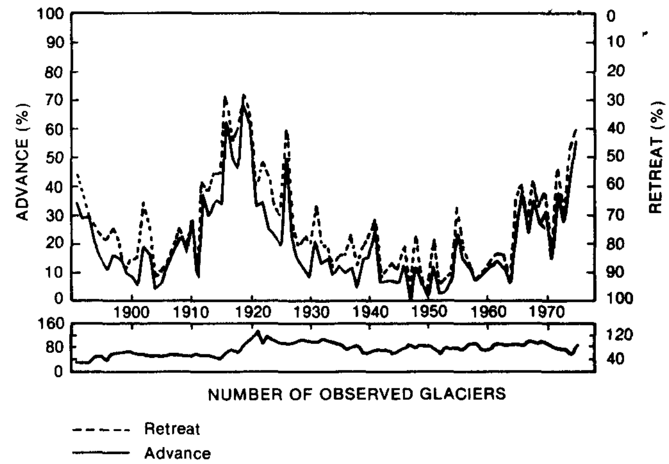

Figure 6.15. Percentage of Swiss glaciers showing terminal advance or retreat, 1890s–1970s, based on data provided by the Permanent Service for the Fluctuations of Glaciers, Zurich. Source: Barry (1981).

6.6.4 Climate and Ice Sheet/Glacier Interactions

Relationships between global climate and ice sheet growth or decay are complex. For a land-based ice sheet, the mass balance is determined by cold season accumulation of snow and warm season ablation, which primarily is due to net radiation and turbulent heat fluxes to the ice during warm air advection. Results of model studies have shown that over long time scales (10^4–10^5 years), ice growth and decay probably resulted from changes in the Earth's orbital parameters, which affected the incoming solar radiation in higher midlatitudes, plus vertical movement of the Earth's crust because of the ice load, acting together with ice-albedo feedback, and changes in precipitation produced by the ice sheet itself (Budd and Smith 1981). Nevertheless, questions relating to internal ice oscillations still arise. Van der Veen and Oerlemans (1984) have shown that two alternative flow regimes can be generated in a two-dimensional polar ice sheet model by temperature-ice flow feedbacks. These may result from temperature-dependent changes in ice strain rate or through the acceleration of the flow and the augmention of the mass discharge by basal meltwater. Cyclical behavior could not be investigated although the time scales involved in such feedbacks will be long.

Where ice masses terminate in the ocean, iceberg calving is a major contributor to the mass loss. In this situation, the ice margin may extend out into deep water as a floating ice shelf, such as that in the Ross Sea. Further complexity exists in West Antarctica where much of the bedrock is well below sea level (Figure 6.14). The West Antarctic ice sheet is stable so long as the Ross Ice Shelf is constrained by drag along its lateral boundaries and pinned by local grounding. It has been argued (Hughes 1973; Mercer 1978) that a retreat of the grounding line or unpinning either by a rise in sea level or by a thinning of the ice shelf could lead to rapid and irreversible retreat of the West Antarctic ice sheet and a 6–7 m rise in sea level. Mercer (1978) postulates that this situation pertained during the last interglacial (centered about 125,000 yr B.P.) as a result of slightly greater global warmth, whereas Hughes (1983) considers from geological evidence that the West Antarctic ice sheet extended to the edge of the continental shelf during the last glaciation, when sea level was lower, and that two-thirds of it disintegrated between 17,000 and 7,000 years B.P. as the sea level rose due to the melt of the northern ice sheets. A time-dependent model simulation of this retreat (Fastook 1984) predicts that the present configuration, with ice shelf buttressing of the marine-based west Antarctic ice, is stable except in the Amundsen Sea sector.

Most of the discharge of the West Antarctic ice sheet is via the five major ice streams (faster flowing ice) entering the Ross Ice Shelf, the Rutford Ice Stream entering Ronne-Filchner shelf of the Weddell Sea, and the Pine Island glacier entering the Amundsen Ice Shelf. There is a wide diversity of opinion as to the present mass balance of these systems (Bentley 1983, 1985; NRC 1984a) principally because of the limited data. Calculations by Thomas et al. (1979, 1980) indicate that the major ice shelves will buffer against ice sheet collapse, by lengthening the time scale to something of the order of 3–4 centuries. Moreover, Bentley (1985) demonstrates the physical constraints on iceberg removal. A time-dependent numerical model study of ice stream E, which enters the Ross Ice Shelf, indicates that, following an extensive retreat during the Holocene, the ice stream is now close to dynamic equilibrium (Lingle 1984). This conclusion is based on the good agreement between the observed back pressure of the ice shelf and the model calculations. The model ice stream is shown to be more sensitive to changes in this back pressure than to changes in accumulation rate.

The sensitivity of the West Antarctic ice sheet to a warming is quite uncertain, because the effects involve air and ocean temperatures, sea ice extent, and changes in accumulation rate. The accumulation rate may increase in the coastal areas of Antarctica (see Oerlemans 1982a; Limbert 1984), which would favor a more positive balance. A rise in air temperature of up to 5°C would probably have little direct effect in view of the very low mean annual and summer air temperatures over all of Antarctica, including the Ross Ice Shelf, although estimates of surface heating are required to quantify this effect (Thomas 1984). Paterson (1984) notes that intervals of positive air temperatures in West Antarctica at present are associated with specific synoptic events and that any melt refreezes. Basal heating of the ice shelves by the ocean is also uncertain, but MacAyeal (1984) considers that the Ross Ice Shelf is well buffered against changes in bottom melt. (See also the separate DOE-sponsored report prepared by the National Research Council [1985]).

Much less attention has been given to the Greenland Ice Sheet. Ambach (1980) considers that a temperature rise of 1.5°C could give a substantial change in its mass balance, although the time scale of response would be long. However, recent ablation estimates for Greenland are much lower than those assumed by Ambach (Radok et al. 1982). A major problem in estimating total net ablation in Greenland is caused by the refreezing of meltwater and its reappearance as superimposed ice below the equilibrium line.

The relationships between glacier mass balance and meteorological parameters are complex (Krenke 1974; Kuhn 1981). Glaciers in midlatitudes are controlled largely by summer ablation, although small ice caps in the Canadian Arctic are mostly accumulation controlled (Koerner 1979). Summer ablation, which is determined by solar radiation, warm air advection, and condensation, as well as accumulation, will be affected by direct and secondary effects of a CO_2-induced warming. If we utilize simple empirical correlations between temperature and glacier behavior (Hoinkes and Steinacker 1975; Posamentier 1977), quantitative estimates of the effect of any particular degree of warming can be made. Grosval'd and Kotlyakov (1980) concluded that, on alpine glaciers in the Caucasus and Soviet Central Asia, the shorter accumulation period will be counter-balanced by increased winter precipitation, at least up to about the year 2020. However, they consider that small glaciers and ice caps on the Arctic islands may disappear because of increased summer melting. Using temperature predictions of Budyko et al. (1979), for a 12°–15°C rise in high-latitude temperatures by 2020, and an ablation formula of Krenke and Khodakov (1966):

$$\text{ablation (mm w.e.)} = (\overline{T} + 9.5)^3 \tag{6.1}$$

where $\overline{T}$ is the mean July temperature (°C); ablation rates of 4–6 m y^{-1} of water equivalent (w.e.) are estimated for Franz Josef Land and Novaya Zemlya. Such rates, if sustained, could lead to the disappearance of most small Arctic ice bodies (150–250 m thick) within about 50 years. However, the temperature increase postulated by Budyko is considerably higher than most consensus estimates of CO_2-induced warming.

Observational evidence for the Alps shows that glaciers did respond strongly to the general warming between 1920 and 1950 (Hoinkes 1968). Results of Gamper and Suter (1978) and Orombelli and Porter (1982) indicate that 1°C changes in mean temperature give rise to fluctuations of the order of 1 km in Alpine glacier termini locations. Similar evidence exists in several other glacierized mountain ranges. Nevertheless, the retreat and advance of individual glaciers may be asynchronous to the same climatic forcing because of differences in glacier length, elevation, slope, and speed of motion (Paterson 1982), although surge behavior in glaciers does not appear to be climatically induced. Glacier flow regimes have been analyzed using a time-dependent model by Nye (1960), Smith and Budd (1981), and others. Using changes in observed parameters for four European glaciers of differing characteristics and responses, Smith and Budd estimated that average summer temperatures increased between 0.5° and 1.0°C since the Little Ice Age about 300 years ago. Such models can be used in a reverse sense to predict possible future changes in individual glaciers, but regional patterns of ice response are probably best assessed via simpler, more general regression relationships of the types described above. For example, in southern Alaska, the Wolverine Glacier thickened substantially between 1976 and 1981 during a period of warmer and more snowy winters (Mayo and Trabant 1984). They concluded that global warming need not result in a simple worldwide glacier recession, but in a complex mix of growth and thinning with asynchronous advances and retreats.

The question of the ice melt contribution to global sea level change cannot be answered definitively at present. Reasons for this include the inadequacy of data on ice volumes in the world's glaciers and two major ice sheets and on the time change of their mass balances; the limited observations on, and incomplete theoretical understanding of, the floating ice shelves (in the Ross Sea and other parts of Antarctica) and their role in the stability of the West Antarctic ice sheet; the limited understanding of possible warming effects on Antarctic climate, Southern Ocean temperatures and circulation, and the mass balance of the Antarctic ice sheet and ice shelves. A special DOE-sponsored report prepared by the Polar Research Board (NRC 1985) addresses the role of the various components of global sea level change and their uncertainties.

Considering first the computations of mass balance for the whole Antarctic ice sheet made since the International Geophysical Year (1957–1958), five determinations range from +240 to +1320 km^3 of water per year with an average value of +900 km^3 (Meier 1983). This corresponds to an annual sea level drop of 2.25 mm (22.5 cm/century), with possible extremes of 0.6 and 3.3 mm. A new calculation by Budd and Smith (1985) indicates a net influx of 2000 km^3 per year nearly balanced by outflow with a discrepancy of 0 to ±20%. The positive mass balance implies a sea level *decrease* of about 0 to 1.2 mm y^{-1}. Bentley (1985) presents additional evidence for the current buildup of the West Antarctic inland ice. However, the first results from an iceberg-monitoring program indicate an iceberg calving rate of 2.3×10^{15} kg y^{-1}, which is 3–4 times higher than previous estimates (Kristensen 1983) and probably in excess of net accumulation (Orheim 1985). This illustrates the need for additional, careful determination of all mass balance components. The information is similarly limited for Greenland, but the evidence suggests a near zero change (Radok et al. 1982; Meier 1983).

Mountain glaciers and ice caps that account for only 3.5% of the land ice areas and 0.4% of the ice volume may be losing mass at a rate of about 666 km^3 of water per year according to Meier

(1984). This corresponds to an annual sea level rise of 0.46 ($\pm$0.26) mm and could, therefore, account for the discrepancy between the observed rise of 1.2 ± 0.3 mm y^{-1} and the calculated thermal expansion effect of 0.6 ± 0.3 (Meier 1983, 1984; also see Chapter 5 of this volume). However, it must be emphasized that these calculations are based on a marginally adequate data set of long-term glacier balance and volume change (see Section 6.6.3) that has been scaled to derive global estimates. Mountain glaciers in coastal Alaska, central Asia, and the southern Andes appear to contribute 60% of the total.

As a basis for future assessments, Meier (1984) notes that a 1.5–4.5°C warming could lead to a sea level rise of 1.7 to 5.2 mm y^{-1} due to the melting of mountain glaciers (based on the observed range of melt rate coefficients per 1°C rise in summer air temperatures). If this warming occurred in 100 years, the sea level rise would be 9 to 26 cm, assuming a linear increase in annual meltwater.

6.7 SUMMARY AND RESEARCH RECOMMENDATIONS

Global snow and ice phenomena can be subdivided into categories that respond either rapidly or slowly to changes in climate. Snow cover and floating ice are in the first category; ground ice, glaciers, and ice sheets are in the second. Hence, the monitoring requirements to detect changes are substantially different. Table 6.3, which is based on several assessments of requirements for large-scale climate research, illustrates the major variables that need monitoring and the needed accuracies and resolutions for the data (National Aeronautics and Space Administration 1979; Jenne 1982). The specified accuracies and resolutions should be taken as general guides rather than as firm standards. For example, higher resolution sea ice data, routinely available from passive microwave sensors, are necessary for climate monitoring according to a recent report (World Meteorological Organization 1984). Detailed records of global snow and ice are generally of much shorter duration than those for most other climate system parameters, and consequently, there are many unknowns in terms of their variability and uncertainties about the representativeness of short-term empirical studies and modeling results of climate-cryosphere interactions. These questions are summarized in Table 6.4. In most areas there is a need both for the improvement of existing data bases and their availability and for the continuation of the collection of consistent data in the future.

The scientific priorities that can be identified from the foregoing discussion are listed below in unranked order. It should be noted that recent reports (NRC 1983a; 1983b, 1984b) address the needs for snow, ice, and permafrost research in a wider context.

First, there are the factors controlling longer term fluctuations in sea ice extent and thickness and, particularly, the stability of the Arctic pack ice to changes in climatic, oceanic, or hydrologic regimes. Sea ice extent plays a dominant role in snow and ice-climate feedback effects and, therefore, this variable is of major significance for global climate. There is a need for more remote sensing and field data on ice thickness, summer snow melt on the ice, and the age distribution of ice (see also NRC 1983a). Coordinated measurements of oceanographic, atmospheric, and ice variables are needed in high latitudes of both hemispheres on a more extensive and continuing basis than the present Arctic MIZEX program provides. Untersteiner (1984), for example, has documented a scientific plan for air-sea-ice research in the 1980s. Coupled ice-ocean-atmosphere models will need to take into account many of the detailed physical processes known to be important in determining the annual cycle of ice growth and decay. At present, there are no adequate coupled models for predicting the changes in sea ice area and thickness that can be expected for a CO_2 doubling. Given the complex factors controling sea ice cover, any CO_2-signal is unlikely to be detected from this variable over at least the next several decades.

Second, is the monitoring of the annual freeze-up and break-up of inland lakes, particularly in North America and Eurasia. Following further checking of statistical relationships between lake freeze-up and break-up and climatic variables, it should be feasible to monitor trends in freeze-up and break-up dates from satellite measurements. These can provide an integrated index of thermal effects resulting from the CO_2-signal in the transition seasons over large areas of the northern continents.

Table 6.3
Cryospheric Data Requirements for Climate Research

Variable	Purpose	Current Climatic Conditions	Month-Decade Changes	Longer-Term Changes	Human Impacts	Accuracy Desired/Basic	Horizontal Resolution (km)	Time Resolution
Snow Cover Extent	V,M	X	X	X	X	3/5%	250	3 d
Snow Water Content	V	X	X	X	X	0.5/1 cm w.e.[b]	250	1 wk
Surface Albedo	V		X	X	X	0.02/0.04	250	3 d
Sea Ice Boundary	M		(X)	(X)	(X)	20/50 km		1 wk
Sea Ice Concentration	V,M	X	X	X	X	3/5%	250	1 wk
Drift of Sea Ice	P		(X)			2/5 km d^{-1}	500	3 d
Sea Ice Thickness	V,M			(X)	(X)	10–20 cm	250	1 mo
Sea Ice Melting	P	(X)			(X)	Not defined	250	1 wk
Multiyear Fraction[a]	M,P		(X)	(X)		0.1/0.1	250	1 wk
Ice Sheet Elevation	M			X	X	0.1/1 m	2–200	1–10 y
Ice Sheet Boundary	M	X			X	1/5 km		1–10 y
Ice Movement	P,M			X	X	1–10/100	Point m y^{-1}	1 y targets
Ice Sheet Thickness[a]	M			X	X	10/100 m	250	1–10 y
Permafrost Boundary[a]	M	(X)		(X)	(X)	10/50 km		1–10 y
Lake Ice Duration[a]	M	(X)	(X)	(X)	(X)	Not defined	Selected lakes (250 km)	3 d

Note: Modified after Jenne (1982). M—monitoring requirement; P—process study requirement; V—verification of models requirement ( )—requirement not listed in Jenne (1982).
[a] Variable not listed in Jenne (1982).
[b] w.e. = water equivalent.

Table 6.4
Summary of Knowns, Unknowns and Actions Needed

	Snow Cover	Sea Ice	Freshwater Ice
1) Known	Approximate shift in hemispheric snowline for 2 × CO_2 (in absence of cloudiness change).	Forcing mechanisms of seasonal cycles and synoptic scale anomalies of ice extent	Natural variability of ice cover duration for 25-50+ years for some lakes
2) Unknown	Changes in regional, and seasonal snow cover for 2 × CO_2.	Changes in area and thickness for 2 × CO_2; Arctic ice stability.	Regional changes in ice cover duration for 2 × CO_2.
3) Uncertain	Long-term natural variability. Feedback effect on clouds clouds and circulation.	Causes of variations in extent, thickness, age structure, mass balance (GCMs highly parameterized). Arctic snow melt and cloud interactions.	Generality of regression equations (freeze/break-up and temperature) for individual sites.
4) Data (tools) Available	Hemispheric data since 1966, longer station records. GCMs with surface hydrology and predicted cloud cover.	Ice limits since 1972/3; longer records for some sea areas. Sea ice models—dynamic and thermodynamic.	Data status uncertain outside North America and Scandinavia.
5) Data (tools) Required	Extended digital data sets Surface albedo data products. Better snow/cloud discrimination algorithms.	Long-term area and mass balance changes. Fully coupled atmosphere-ocean-ice models.	Identification of representative lakes for monitoring
6) Recommended Research	Data set compilation. Snow (and cloud) discrimination algorithms. Albedo algorithm development. EBM and GCM experiment.	Continuation of passive microwave remote sensing; radar and sonar surveys; upgraded polar meteorological data and archives.	Compilation/analysis of lake ice and climate data.
7) Impacts on 1,2,3	Better definition of (3). Improved modeling assessment of (2). Higher quality data for (3).	Better understanding of (3). Improved modeling of (2). Better data for (1) and (3).	Improved prediction of (2) to make monitoring future changes feasible

Table 6.4
Continued.

	Ground Ice/Permafrost	Glaciers	Ice Sheets
1) Known	Approximate shift of boundaries for 2 × CO_2.	Response of glacier length to climatic forcing	Large-scale flow behavior of idealized ice sheets.
2) Unknown	Rate of change of thickness, especially in marginal areas.	Regional responses of ice volume to 2 × CO_2.	Stability of ice shelves and W. Antarctic ice sheet to 2 × CO_2; contributions to global sea level change.
3) Uncertain	Age of ice/permafrost and its stability for present climate.	Rate of volume change and sea level impact.	Volume changes of entire Greenland/Antarctic ice sheets; ice sheet histories.
4) Data (tools) Available	Approximate extent of continuous, discontinuous permafrost.	Global inventory; a few mass balance records.	Numerical ice sheet models. Limited accumulation/ablation and flow data.
5) Data (tools) Required	Better data on thickness, thermal profiles.	Global network of mass balance in relation to climate.	More data on temperature and velocity profiles, ice history, especially of shelves.
6) Recommended Research	Extraction, compilation of data, modeling of climatic effects for various ground conditions.	Extension of case-specific modeling of glacier response to climate. Augmented glacier monitoring.	Complete altimetry survey at 5–10 year intervals; additional surveys of motion, etc. in boreholes.
7) Impact on 1,2,3	Improved definition of (3) and prediction of (2).	Better assessment of sea level trends.	Better assessment of sea level trends and model assessments of ice shelf/sheet stability.

Their potential in other land areas remains to be explored.

Third, the assessment of the regional and seasonal responses of snow cover extent, depth, and duration to CO_2 doubling should be studied. Better understanding of the natural variability over most of this century could be obtained by extraction and compilation of the relevant data from meteorological archives. The necessity for global snow cover data in relation to the CO_2 problem is recommended strongly by the Committee on Glaciology (NRC 1983a). Reductions in hemispheric snow cover extent and duration could be indicative of a CO_2-signal within the next several decades.

Fourth, modeling studies, supported by additional field measurements should be conducted, to ascertain the stability of the West Antarctic ice sheet and adjacent ice shelves to a CO_2 doubling. This question is critical to the longer time scale (~100 years) of global sea level change. This topic has been the subject of recent special workshops (NRC 1984a).

Fifth, better observational data are necessary to determine mass balance and volume changes on the two major ice sheets and a representative coverage of the world's glaciers to assess their actual and potential contributions to sea level rise as global temperatures increase. A recent Polar Research Board workshop specifically addressed this topic (NRC 1985).

Finally, improved mapping and observation of permafrost are critical to be able to monitor long-term changes in ground temperature (NRC 1983b).

In summary, from the detection point-of-view, the clearest indication of CO_2-induced climate changes in the cryosphere will be provided by trends in annual lake freeze-up and break-up dates. These show a strong relationship to transition season temperatures. Snow cover and sea ice are important components of the global climate system, but these cryospheric variables are each affected by many climatic factors and show large interannual and regional variability. Trends in the Northern Hemisphere snow-covered area would, however, be the second most likely source of cryospheric evidence for a CO_2 signal by the early part of the 21st century. Information on changes in land ice volumes, ground ice and permafrost conditions is needed for assessing long-term climatic trends over periods of 50 to 100 years, especially relating to global sea level changes. However, these parameters cannot yet be monitored on a routine global basis.

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7. LONG-TERM CHANGES IN PRECIPITATION PATTERNS

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CONTENTS

7.1 INTRODUCTION

A stable precipitation climate seems a prerequisite for stable economic structures and agricultural systems. Less obvious perhaps, is the use of the rain water for replenishment of ground water supplies, drinking water, hydroelectric power, and so forth. Furthermore, a change in the precipitation climate in one area can have significant effects in distant regions through the transport mechanism provided by rivers. Perhaps most subtle is the fact that evaporation/condensation processes have fundamental effects on the energy budgets of both the oceans and the atmosphere. In the case of the latter, condensation provides heat energy to help drive the system. Substantial changes in position and amount of this heat release could alter the "typical" atmospheric climate as we know it today.

Model projections of atmospheric conditions with increased levels of carbon dioxide (CO_2) all seem to suggest a small increase in the intensity of the hydrological cycle (cf. National Research Council [NRC] 1983). This presumably means slightly more precipitation on a global average basis. However, it seems likely that regional changes will dominate this average signal, for example, a tendency toward summer dryness in midlatitude areas (Manabe et al. 1981; Hayashi 1982). The implications of these precipitation changes are not as obvious as one might guess. For instance, Revelle and Waggoner (1983) have suggested that, in the absence of compensating effects, a modest warming (2°C) could *reduce* runoff by 15–30+% in drainage basins in the western United States. They have also estimated that the reduction in runoff due to a warming can be doubled by a small decrease (10%) in precipitation. Alternatively, the reduction in runoff could presumably be cancelled by an increase in rainfall by the same amount (cf. NRC 1983, Tables 1.7 and 1.8,) or even *increased* by 40–60% by reduced evapotranspiration (Idso and Brazel 1984). Such scenarios are highly speculative, at best, and at this stage do not seem to find general acceptance in the scientific community. More importantly, the model projections of such small precipitation changes which give rise to the scenarios are attended by large uncertainties.

To determine whether increasing amounts of CO_2 are changing the distribution of precipitation, it is necessary to know first the past history of the rainfall regime. On a global average basis this appears impossible because the majority of the ocean regions (72% of the Earth's surface) have inadequate observations of rainfall. The situation is apparently better over some portions of the land masses, so it may be feasible to detect regional changes in precipitation that are due to CO_2 effects. However, as will become evident, sampling and measurement problems and the discontinuity of the precipitation process itself, plus relatively large, natural decade-to-decade variability, make such analyses highly uncertain.

The purpose of this chapter is to summarize briefly what is known about the large-scale patterns of precipitation. No attempt will be made to detect a "CO_2 signal"; indeed, the text should convince the reader that this may be impossible for the case of precipitation. With these facts in mind, the following sections of this chapter briefly describe available precipitation data and some of the associated measurement problems (see also Chapter 3 of this volume). Selected studies of regional precipitation changes since the turn of the century will then be reviewed. These summaries are augmented, whenever necessary, by a set of new regional studies carried out specifically for inclusion in this chapter. A subsequent section investigates the possibility of a coherent precipitation signal in the set of regional averages themselves. The chapter closes with a set of conclusions and recommendations.

It is important here to note two significant shortcomings in the following text. A decision was made to discuss annual averaged precipitation data. This should provide a general view of the space and time structure of precipitation. However, as pointed out by E.B. Kraus (1954), the precipitation process is highly seasonal, so that the annual average may embody a variety of different physical situations, for example, both monsoon and extra-tropical frontal activity over Australia. Suppose CO_2 affected only one of these situations (e.g., the monsoon). The use of the annual average would tend to obscure this effect. A second problem has to do with the initial selection of regional areas for analysis. The goal of the paper was to look at large-scale change, and for that purpose, regional selection was used. A better approach, at the next level of detail, would be to select the regions based on the type of physical situation which gives rise to rainfall over them. Both the seasonal and regional stratification process will be important considerations in a formal attempt to detect a CO_2 signal in the precipitation regime. However, that is beyond the scope of this chapter.

7.2 DATA

7.2.1 Abundance

There are a surprising number of long time series of rainfall measurements. For instance, a search of one source of such data (World Weather Records) revealed 186 stations around the globe that had 80 concurrent years of data with less than 5% missing values. Corona (1978) found over 1300 usable station records, albeit with gaps, for his study of Northern Hemisphere precipitation changes. The regional distribution of data can be even more impressive when multiple archives are used; for example, Tabony (1981) obtained 182 stations in Europe alone covering the period 1861–1970. The geographical distribution of the long precipitation records is relatively good, with northern Canada and the higher latitudes of Asia representing the main areas void of data. In summary, there are a lot of precipitation data over most of the world's continents. However, no single, integrated precipitation data archive seems to exist at this time, although there is work in that direction (e.g., Bradley et al. 1984).

The major gap in the precipitation data sets occurs over the oceans. Ships report synoptic weather and include precipitation information in a qualitative manner (e.g., clear sky, showers, heavy continuous rain). Attempts have been made to convert these observations to quantitative precipitation estimates (e.g., Tucker 1961; Reed and Elliott 1977; Reed 1979) with some success. For example, Dorman and Bourke (1979) have used the ship data to compute a precipitation climatology for the Pacific Ocean from 30°S to 60°N, whereas Elliott and Reed (1984) have attempted to construct a similar estimate for all of the world's oceans. However, the conversion procedure used to obtain these climatologies is fraught with problems, and most authors agree that the procedure does not work well in areas with sparse data, for instance, in most of the Southern Hemisphere oceans. Finally, the estimation of a stable climatology over the oceans does not mean that we can detect variations relative to the long-term mean. In short, our knowledge of precipitation changes over the oceans is minimal.

The problems discussed above, plus the probable poor confidence limits on the precipitation climatology, mean that it is unlikely that the relatively small precipitation changes expected from a CO_2 increase could ever be detected on a global basis. Detection of regional changes would seem more promising and will be discussed in Section 7.3. For now, however, one can only hope that promising satellite systems may one day allow truly global estimates of the planet's precipitation regime. But these systems must be in operation for at least several decades before they will produce enough data to begin to study variations about a climatological mean that they themselves must define.

7.2.2 Problems

A large number of problems attend the measurements of precipitation over the land masses, perhaps more so than most surface climatological variables. Some of these difficulties are discussed below. Rodda (1969), Rainbird (1967), and others list and discuss more completely many of the problems associated with the rainfall data.

1. Representativeness—Precipitation is a discontinuous process. Thus, given two closely located measurement stations, one may record a significant amount of rain while the other records none. It is thought that a representative value of rainfall over a region can be obtained by averaging together the measurements from a large number of stations. Few places in the world have the required station density to make one comfortable with this process, particularly in convective rain regimes. Interesting reviews of different methods used to obtain regional averages from individual station data have been provided by Singh and Birsoy (1975) and Corona (1978). It is clear from these works that the method of obtaining the areal averages can significantly affect the eventual data analysis, yet there appears to be no generally agreed upon averaging method.

2. Station location—Local orographic conditions can strongly affect the rates and amounts of precipitation. In recent times, the location of the measurement site relative to large urban centers also has been important because the pollutants and heat from such centers may alter rainfall in the region downwind of the center (e.g. Changnon 1969).
3. Instrumentation—Precipitation is measured by collecting rain in a graduated "bucket." The location of the bucket relative to the ground and to surrounding structures can be critical to the observations. Hence, changes in instrument (or station) location can affect the data. These and other problems are particularly important in remote areas and early in the records of many stations. Even in modern times different nations use different rain gauges. These different instruments do not give the same estimates of precipitation (cf. Larson 1971; Sanderson 1975). The measurement errors can be in the range 1–20+%.

In summary, the numerical estimates of precipitation available to us from the historical archives likely contain numerous, potentially serious problems. In view of this likelihood, it seems essential to remain highly skeptical of virtually any analysis of these types of data (including that reported here).

7.3 REGIONAL VARIATIONS IN PRECIPITATION

The intent of this section is to investigate changes in precipitation on the largest regional scales possible, because one can speculate that CO_2 induced changes are most likely to occur on these scales. Results of recent studies on precipitation variations in regions where the data density is relatively high are examined first. New results from a uniform study on the long-term changes in coherent precipitation patterns in these areas are reported where appropriate.

The procedures used in this latter study are based upon a method of determining the existence of a spatially coherent signal of uniform sign in a set of time series from different locations, in this case, yearly averaged data on rainfall amounts at different stations for the period 1891–1970. The study regions are shown in Figure 7.1. The results of the analysis are then used to form a single time series representative of changes in strength of the coherent precipitation signal. In most cases, this signal constitutes a relatively small fraction of the variance in the regional precipitation data. The implicit assumption in the approach is that any CO_2-induced signal will appear in this measure of coherent precipitation. Additional details on the methods are given by Barnett (1983).

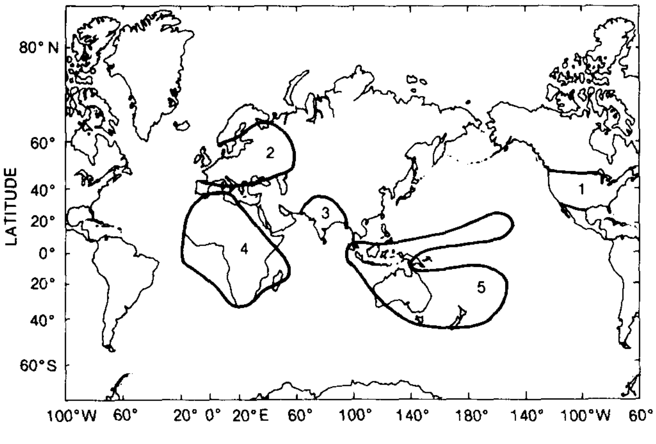

Figure 7.1. Definition of study regions.

The annual average precipitation data were computed from monthly averages listed in the World Weather Records summaries (available in digital form from the National Center for Atmospheric Research). Missing monthly data resulted in missing annual average data. Missing annual values were replaced by the long-term station mean; a procedure required for less than 5% of the annual averaged data. Each of the original time series was visually inspected for abnormalities.

Let the annual rainfall data be denoted by R_{ij}, where i is spatial location and j is $1, 2\ldots, 80$ a year counter. The fractional departure in rainfall is then,

$$\delta R_{ij} = \frac{R_{ij} - \langle R_{ij}\rangle_j}{\langle R_{ij}\rangle_j} \quad , \tag{7.1}$$

which normalizes the data to account for the wide range of annual precipitation amounts.

Next, compute the correlation matrix

$$C_{ij} = \frac{\langle \delta R_{il} \delta R_{jl}\rangle_l}{\langle \delta R_{il}^2\rangle_l^{1/2} \langle \delta R_{jl}^2\rangle_l^{1/2}}, \tag{7.2}$$

and then its eigenvectors (B_{ni}) and eigenvalues (λ_n). The variance associated with the nth mode is just $\lambda_n/\Sigma\lambda_m$.

It can be shown that if there exists a signal that is coherent and has the same sign over the entire domain of i, then all components of B_{1i} will have the same sign (Backus and Preisendorfer 1978). Thus, the elements of B_1 are inspected for like sign and a statement made as to the existence (or lack thereof) of a signal coherent and of the same sign among the set of precipitation stations under analysis.

The time series of the strength of the large-scale signal was computed according to

$$T_j = \frac{\sum_i \mid B_{1i} \mid \delta R'_{ij}}{\sum_i \mid B_{1i} \mid}, \tag{7.3}$$

where ($\delta R', T_j$) are in units of standard deviations. Note that T_j is very closely allied with the first principal component and, when there are *no* missing data, differs only by a normalization factor. The T_j were smoothed with a 5-year running mean filter prior to presentation in the main body of the text.

This method of analyzing the data is applied to a number of different regions in the following subsections.

7.3.1 Continental United States: Region 1

The data coverage over the United States is generally excellent so it is little wonder that numerous studies of precipitation variability over this area have been made. For purposes of this report, it is appropriate to concentrate on the results associated with large-scale, long-term changes. Results of previous studies with this emphasis suggest that there exists a coherent pattern of variation in annual average precipitation over the entire continental United States (Walsh and Mostek 1980; Diaz and Fulbright 1981; Karl and Koscielny 1982; Vines 1984). The strength of this pattern in terms of the variance it accounts for in the total precipitation field is 10–30%. Given the nature of the precipitation process and the data problems noted above, the existence of such a pattern is remarkable. The central part of the country contributes more variance to the coherent pattern than do other regions, whereas the eastern seaboard is only marginally represented (cf. Figure 7.2; Walsh et al. 1981; Karl and Koscielny 1982).

The time variation of the pattern described above is of key importance if one aims to detect a CO_2 effect. Diaz and Quayle (1980) have shown that, over most of the country, changes in the mean annual precipitation from 1921 to 1954 and from 1955 to 1977 are generally not significant. This is in agreement with the results of Corona (1978, 1979). One of the reasons for this result may well be the high variance of yearly means within each of their base periods. In contrast, Mitchell et al. (1979), using a 300-year proxy record of precipitation, suggest that a significant fraction of rainfall variance in the central and western United States may be due to solar effects. The results of Corona (1978) and Vines (1984), based on the far shorter instrumental record, are in broad agreement with this conclusion. However, Williams (1978), working with a 64-year record, finds the power spectra of these station data to be indistinguishable from a white noise process. The same result was found by Karl and Koscielny (1982) using 86 years of data.

In an attempt to resolve the above discrepancies, empirical orthogonal function (EOF) analysis was applied to the averaged, normalized annual precipitation anomalies for nine continental United States climatic divisions. The results were as follows:

1. The first eigenvector had components of like sign in all nine subdivisions, supporting the existence of a coherent pattern of spatial variability of the same sign in annual average precipitation anomalies over the United States, in agreement with the work of Walsh and Mostek (1980), Diaz and Fulbright (1981), and Karl and Koscielny (1982). The pattern accounted, on the average, for 31% of the normalized variance and was statistically distinct from the other eigenfunctions. As noted above, the eastern seaboard had a meager participation in this pattern, whose center of action was the middle third of the country (Figure 7.2).
2. An area average precipitation index for the continental United States was computed using the mode 1 EOF components as weights and then smoothed with a 5-year running mean (Figure 7.3). Two features of the curve are apparent: decadal variations of the order of one

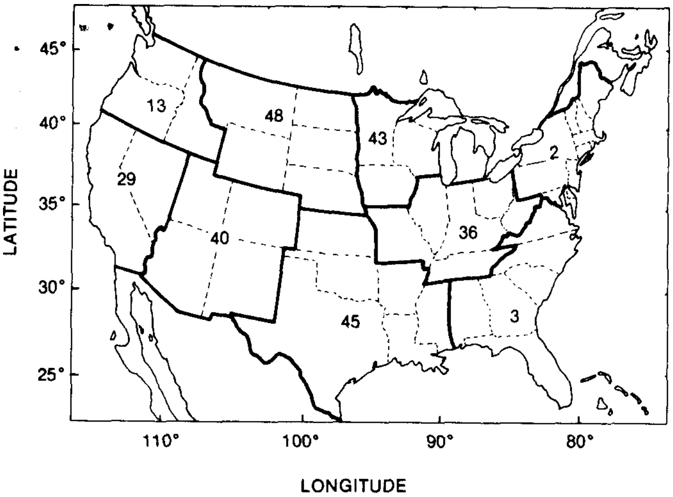

Figure 7.2. The heavy lines show the nine climatic divisions used to study precipitation in the continental United States. The numbers in each division represent the mode 1 eigenvector components (× 100).

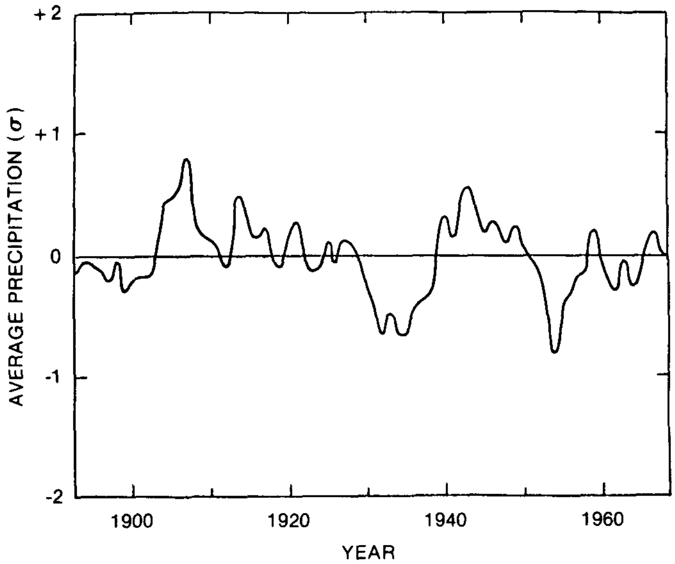

Figure 7.3. Time series of annual average precipitation anomaly for the continental United States. The average is a linear combination of the division data constructed according to Equation 7.3 and expressed in units of standard deviation (σ).

standard deviation are superimposed on a decreasing trend in precipitation (-0.2 standard deviations per century). The decade-to-decade changes make this trend value statistically insignificant, which is in agreement with previous results. Note the strong drought signature in the 1930s and 1950s, which is in agreement with the results of Mitchell et al. (1979). However, no such signal exists in the 1910s, thus making the existence of a 20+ year cycle in this U.S.-wide index open to question. Unfortunately, the existence of such a periodicity is very difficult to determine from a relatively short instrumental record containing only three to four realizations of the period of interest.

In summary, there appears to be a spatially coherent signal in the normalized precipitation anomaly field over the continental United States. The time dependence of this signal is characterized by large decade-to-decade fluctuations of the order of one standard deviation. These variations represent a very large natural noise, against which detection of a relatively small CO_2-induced signal will be exceedingly difficult.

7.3.2 Europe and Western Asia: Region 2

The coherent properties of the precipitation field in Europe have been demonstrated in the study of Tabony (1981). He performed an EOF analysis on

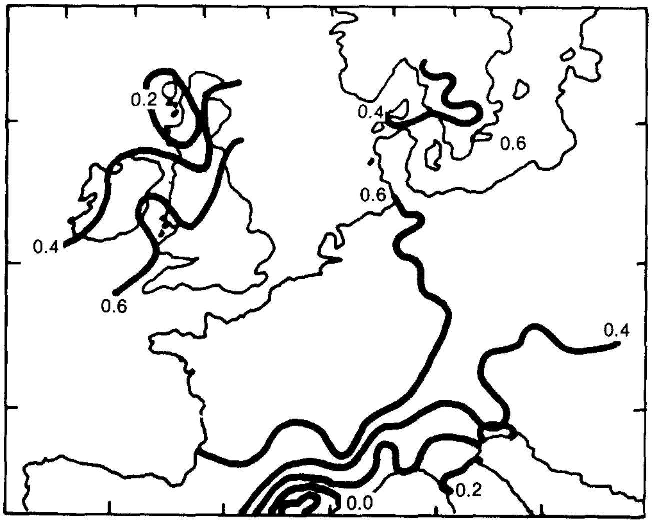

Figure 7.4. First eigenvector of annual rainfall from 1861 to 1970. This mode captured 28.1% of the variance in a 182-station data set. Source: Tabony (1981).

data from 182 stations spanning the period 1861–1970. The first EOF pattern is shown in Figure 7.4 and accounts for 28.1% of the variance in the data set. The fact that all components of the first eigenvector had the same sign showed the existence of a pattern of uniform rainfall variation over most of Europe. Gradients in the pattern only appear in the northwest portion of the British Isles and in the Mediterranean area. Wigley et al. (1984), among others, have shown that the coherency of the pattern does indeed include all of the British Isles.

The time variation of the coherent European pattern was examined by Tabony (1981) using several different tools. Of interest here is the long-term behavior of the coherent precipitation pattern. This is shown in Figure 7.5 by decadal means of the EOF principal component amplitudes associated with the first eigenvector. These data suggest a weak, positive, but statistically insignificant trend of increasing precipitation (cf. Tabony 1981, Table II). More prominent in the record are fluctuations on interannual time scales, which show significant spectral peaks around periods of 5 years and 2.1–2.4 years, which are time scales comparable with those of the Southern Oscillation and Quasi-Biennial Oscillation, respectively.

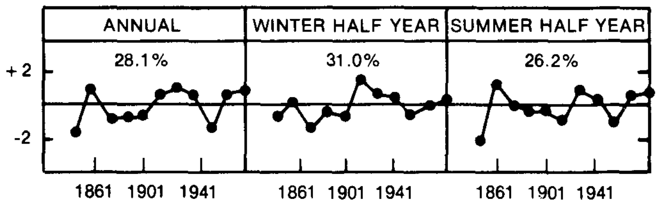

Figure 7.5. Decadal means of first principal component of patterns of European rainfall, in relative units. The percentage of variance explained is given for annual and semi-annual time periods. Source: Tabony (1981).

Gruza and Apasova (1981) also examined long-term (1891–1979) trends in Eurasian precipitation and found positive values that appear to be marginally significant. The increasing trends were dominated by winter (January) precipitation; this result is in moderate agreement with those of Tabony. However, examination of the spatial distribution of "raw" trend values given by Gruza and Apasova did not show the large spatial coherence found in the EOF analysis. This disagreement is not particularly surprising given the noisy nature of the precipitation field, the strength of the coherent signal, and the difference in areas and periods of data coverage.

In summary, the large multi-year and decadal fluctuations in the coherent European precipitation pattern will make detection of a small CO_2-induced signal very difficult in this region.

7.3.3 India: Region 3

The precipitation patterns over India have been intensively studied over the years. These studies have often centered on prediction of the monsoon rainfall, see, Blanford (1884), Walker (1924), and many others in recent times. The strong dependence of the economy of this area on the monsoon precipitation, plus excellent data coverage, have both been responsible for this interest.

Studies regarding long-term variation in Indian precipitation are in reasonable accord. Kidson (1975) and Stoeckenius (1981) present results that suggest coherent precipitation variations exist over the subcontinent. Weare (1979) and Rasmusson and Carpenter (1983) performed EOF analyses of station and divisional data, respectively, and found coherent precipitation patterns over all of India west of 90°E. These results appear to be stable as the periods of record analyzed were 26, 105, and 80 years, respectively.

The reasons for the variations in year-to-year precipitation are largely related to waxing and waning of the summer monsoon (e.g., Bhalme and Mooley, 1980). Both variations are intimately associated with global scale Southern Oscillation events (e.g. Walker 1924; Pant and Parthasarathy 1981; Angell 1981; Rasmusson and Carpenter 1983). However, more local processes also seem important (e.g., Bhalme and Mooley 1980; Weare 1979).

The time dependence of the coherent rainfall pattern discussed above has been examined by Mooley et al. (1981) using 306 rain gauges and is reproduced in Figure 7.6. The authors state that this series is both homogeneous and random. Its serial, or lag one, correlation is 0.013. This value is so low as to suggest no significant trend in the Indian precipitation field. It also prompts their conclusion that the year-to-year variations in precipitation are indeed random. Furthermore, the large noise levels seen in Figure 7.6 (e.g. the 90% confidence levels on the mean) again suggest the difficulty of detecting the rather small changes that may be due to CO_2 effects.

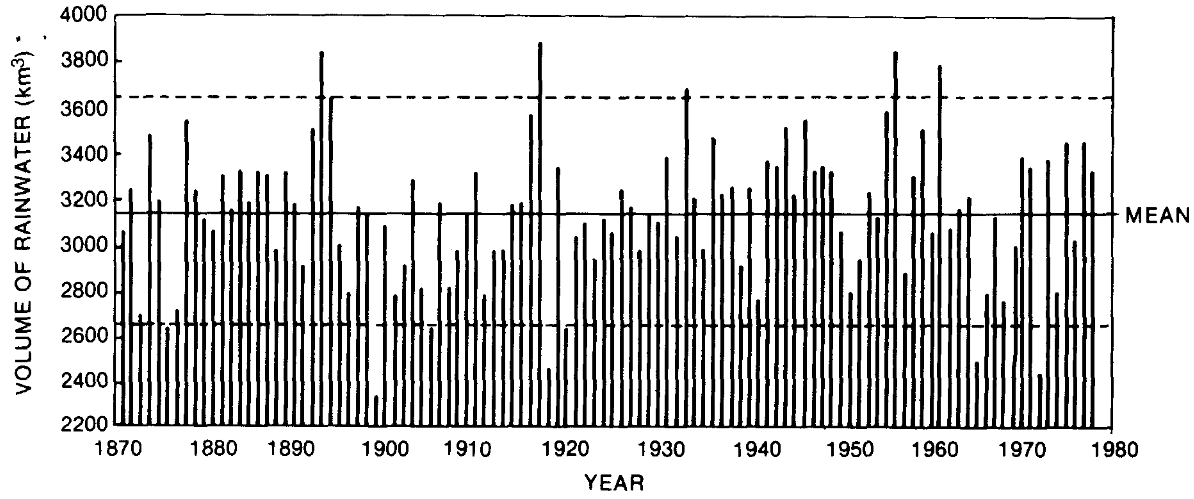

Figure 7.6. Time series of annual average precipitation anomaly for India from 1871 to 1978. The mean value is 3143.5 km^3, and the 90% confidence limits on the mean are shown by the dashed lines. Source: Mooley et al. (1981).

7.3.4 Africa: Region 4

The semiarid areas of Africa are regions with highly variable rainfall and histories of recurrent and intense droughts. The most notable characteristics of the droughts are a random distribution in time, spatial coherence over large areas, and, in some regions, persistence for many years. In some years the spatial scale of drought seems to encompass the entire continent (WMO 1983).

The quantitative details supporting the statements presented above are both numerous and impressively robust. Strong spatial coherence seems characteristic of the precipitation field over Africa. This has been amply demonstrated for the subtropical region by Nicholson (1978, 1979, 1980, 1983), Stoeckenius (1981), Kraus (1977), and others. A larger scale coherence of continental dimension has been well documented by Nicholson (1980, 1981). She also shows that the coherency has been interrupted at the northern fringes of the continent, as well as toward its eastern limits.

The recent severe drought in the Sahel caused numerous investigators to study the past variation of rainfall in that region and elsewhere in Africa. Nicholson (1980, 1983) has shown that the region has experienced prolonged drought during the periods 1910–1915, the 1940s, and since 1960. By contrast, the periods around 1900, the 1920s, and 1950s were times of relatively plentiful precipitation. These results are in general agreement with those of Bunting et al. (1976), Kraus (1977), Lamb (1985), and Dyer (1982).

The causes of the periodic droughts and wet spells seem partially associated with massive modulations of the Hadley cell over the continent (cf. Nicholson 1981; Kanamitsu and Krishnamurti 1978; Kidson 1977). In a series of papers, Lamb (1978a, 1978b, 1983) showed that the anomalies associated with these wet and dry spells involve key atmospheric and oceanic features of the entire tropical Atlantic. The long time scales of these oscillations are not explained. More recently, Charney (1975) and others have suggested the existence of a biophysical feedback to explain the persistent drought of the last two decades, i.e., overgrazing enhances the (already) negative radiation balance, thereby increasing subsidence, reducing rainfall, and making the region less suitable for vegetation.

The regional calculations for Africa produced results similar to those noted above.

1. A coherent signal was found to exist over the continent. However, the current analysis, because of initial demands on the length of record required for analysis, tended to have a majority of stations in the southern half of the continent. The regions generally studied by authors mentioned above were not particularly well represented. The study should be redone using the more voluminous data described by Nicholson.[1]

[1] S.E. Nicholson, personal communication, 1981.

2. The temporal variations of the coherent pattern (Figure 7.7) show a marked similarity to those described by the authors noted above. In addition, a strong trend (-0.86 ± 0.18 in units of standard deviations (σ) per century) is evident. The data limitations noted above suggest this strong trend should be considered critically. It is only moderately apparent in the regional time series (e.g., Nicholson 1981; Dyer 1982; Lamb 1985) but quite clear in measures of the Nile discharge (e.g. , Kraus 1956).

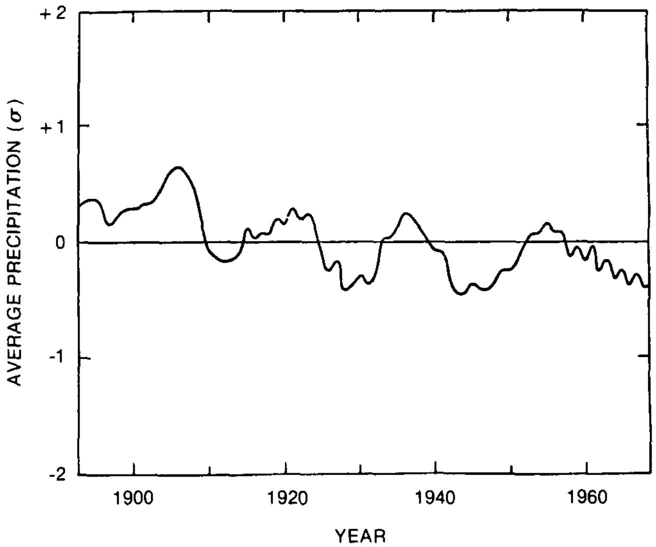

Figure 7.7. Time series of annual average precipitation anomaly for Africa. The average is a linear combination of the station data constructed according to Equation 7.3 and expressed in units of standard deviations (σ).

In summary, the precipitation field over Africa is characterized by very large-scale, spatially coherent structures. These patterns of drought and wetness persist for years and thus constitute a large background noise against which to attempt to detect a CO_2-induced effect. The strong trend toward decreasing rainfall over the continent, *if real*, represents a natural low-frequency signal that requires explanation: Is it the result of natural variation or could it be due to increasing CO_2 concentration?

7.3.5 Australia/Indo Pacific: Region 5

Studies of large-scale precipitation patterns in Region 5 are fairly common in the literature because much of this area encompasses the domain of the Southern Oscillation. The results of Kidson (1975) suggest that a large-scale, coherent pattern of rainfall anomaly exists over the area. The sense of this pattern is exactly opposite that which is occurring in the central equatorial Pacific. This pattern was also noticed partially by Walker (1924) and numerous others since then (see Rasmusson and Carpenter [1983] for a review). In general, annual precipitation variations are coherent over all of Australia, although the signal is stronger in the eastern part of that country (cf. Gentilli 1971; Pittock 1975).

The temporal variations of precipitation in the Australia and Indo Pacific area have a wide ranging character. Along the equator the rainfall spectra are "red"; that is, the rainfall variations have a persistent character, whereas in the subtropics the spectra are "white," indicating random variations (Doberitz 1968; Fleer 1981). The temporal variations over the Australian continent exhibit two major maxima (1920s and 1960s) and one large minimum (1940s) in 60 years of record (e.g., Pittock 1975). Results of all of these previous studies suggest that energetic, long period fluctuations exist in the regional precipitation regime and are closely associated with the Southern Oscillation.

Twenty stations in Region 5 were analyzed by the methods described above to consolidate the results from the different parts of the region. The main conclusions from this analysis are as follows:

1. A coherent spatial signal was found, as expected, in the Indo Pacific with maximum strength in the eastern Australia and Samoa region. However, the pattern was generally weak, accounting for only 14% of the variance in the data set.
2. The time variation of the pattern strength (Figure 7.8) shows the temporal character noted above. The trend was virtually nil, a factor of 10 smaller than its standard error of estimate.

In summary, the Australian and Indo Pacific regions might appear to be regions in which to look for CO_2-induced effects in precipitation patterns. There is virtually no trend in the rainfall field. However, the temporal changes are dominated by the Southern Oscillation whose 3–8 year time scale constitutes a natural, low frequency signal against which a CO_2 signal would have to be detected. It has been suggested that this signal could be filtered from the data, thereby greatly reducing the noise

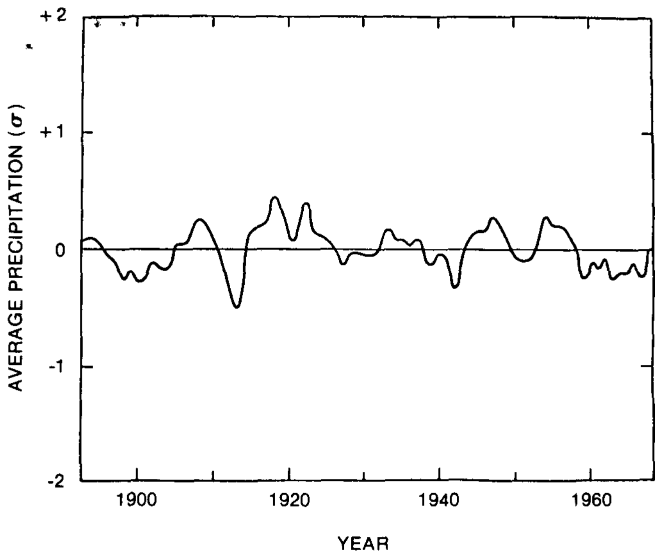

Figure 7.8. Time series of annual average precipitation anomaly for Australia and the Indo Pacific. The average is a linear combination of the division data constructed according to Equation 7.3 and expressed in units of standard deviations (σ).

level and enhancing the chances for detection (cf. Wigley et al., Chapter 4 of this volume.)

7.3.6 Other Regions

Data density was adequate to define the existence of large-scale rainfall regimes in two other parts of the world: (1) Japan and eastern China and (2) central and southern South America. No significant trends were obvious in any of these areas. Furthermore, the natural variability in the areas was as high, or higher, than that shown for any of the areas discussed above. In short, neither of the areas would be a good place to search for a CO_2-induced signal.

7.4 NEAR-GLOBAL VARIATIONS

It is natural to wonder whether the rainfall variations discussed in Section 7.3 have a degree of commonality; that is, is there a coherent signal within the regional averages themselves? This possibility was studied by Gruza and Apasova (1981). They constructed gridded precipitation fields for the Northern Hemisphere (0–85°N) for the months of January and July from 1891 to 1979. In general, no data were available over ocean areas. The linear trend coefficients at each grid point were computed. The resultant distribution of plus and minus trends generally showed the regional coherence in precipitation patterns noted above. However, the regions themselves did not appear to be positively correlated; the areas of decreasing and increasing precipitation were approximately equal. Although the trends for January and July differed, the authors concluded that there is no substantial linear trend in the precipitation field for the 85-year period.

Corona (1979) developed a monthly time history of precipitation over the Northern Hemisphere land masses from 1935 to 1975. His results (cf. his Figure 27) suggest a depression in precipitation in the early 1940s but otherwise no significant trend in the hemispheric average. Ambe (1967) studied secular variations in the arid portions of the globe and came to the same conclusion. However, on a somewhat smaller spatial scale, Kraus (1955) presented data to suggest rather large, long-term trends in the precipitation patterns of many tropical regions of the world (e.g., Africa, Australia), particularly during the period 1900–1940.

A quantitative estimate of the degree of coherence among the various regions discussed in Section 7.3 was obtained by applying the methods described in Section 7.3 to the regional averages themselves. The results are shown in Table 7.1. It is clear that there is *not* a coherent, near-global signal of the same sign in the regional precipitation data because all eigenvector components do not have the same sign. However, the results do show that the precipitation regimes of the Americas tend to fluctuate in unison (both have large positive eigencomponents). The same can be said for the India/Australia/Indo Pacific variations (large negative eigencomponents). The difference in sign between these two super regions shows them to be largely out of phase with each other; that is, more than normal precipitation in the Americas generally coincides with less than normal precipitation in the Indo Pacific region and vice versa. Note that Africa is allied with the Americas, whereas Eurasia and the Far East are not, for all intents and purposes, related with either super region. The results presented above are in broad agreement with those of other workers (e.g., Fleer 1981; Kidson 1975; Stoeckenius 1981).

The use of regional averages, even of highly correlated station data, can be criticized on physical grounds because the different stations making up

Table 7.1
Regional Eigenstructure

Region Name	Mode 1 Component (× 100)
United States	36
Eurasia	−6
India	−58
Japan and eastern China	−6
Africa	22
South American	50
Australia and Indo Pacific	−48

the average come from very different climatic regions and thus may have different physical processes that affect rainfall. To investigate this situation, the individual stations were averaged over specific climatic zones, as defined in Landsberg et al. (1963). A total of 22 such climatically consistent regions were defined, and the analysis described above was repeated. Two main results were found: (1) The general relation among the super regions discussed above was reaffirmed. (2) Some of the large regions discussed in Sections 7.3.1 to 7.3.6 had large variability within them so that the idea of *uniform* variation in precipitation over them is dependent, to some extent, on the distribution of data stations, areal averaging techniques (cf. Section 7.2), and analysis method (particularly Africa). These cautions have been expressed earlier in this chapter.

7.5 CONCLUSIONS AND RESEARCH RECOMMENDATIONS

Past work and current results indicate the existence of large-scale regional coherence in the precipitation field of the planet. One might guess in advance that any CO_2-induced climate changes should be manifest in one or more of these regional signals.

Unfortunately, these coherent precipitation variations are characterized by large natural variability associated with decadal time scales or trends. Even the way in which the rainfall data are averaged and subsequently analyzed causes unexpectedly large perturbations in attempts to form quantitative estimates of regional changes. Both of these types of noise will make it difficult to detect the presence of a small CO_2-induced signal.

Performing massive averaging of the rainfall data in search of a global precipitation signal seems senseless at this stage. The historical data coverage is *not* global. Furthermore, the data from the land masses do not support the concept of a globally coherent precipitation signal.

There are several future areas of activity that deserve special attention if any attempt is made to search for a CO_2-induced signal in precipitation. These are as follows:

1. The precipitation data base must be expanded and homogenized, particularly over the oceans. Detailed studies of this data base should be undertaken so that meaningful estimates of regional averages can be made. Development of such averages, based on high data densities and geophysically oriented mapping techniques, may reduce, somewhat, the noise of individual station data through which a CO_2 signal must be sought.
2. It seems appropriate that greater effort be made to look at the magnification of minor precipitation changes by different parts of the biogeochemical system. This may not be an easy task. For instance, the results of preliminary work of Revelle and Waggonner (1983) have shown that small changes in rainfall, coupled with air temperature changes due to increased levels of CO_2, could produce a relatively large change in runoff and subsequent river flow in the western United States. However, Idso and Brazel (1984) included biological considerations and arrived at the opposite conclusion. Both studies should be considered to be speculative. Nevertheless, changes in lake levels and chemistry, vegetation types, and so forth, may be far more sensitive indicators of changing precipitation regimes than relatively crude measurements from a group of widely spread, imperfectly operated rain gauges.
3. Precipitation integrators or proxies [which need not magnify small precipitation changes as in (2) above] seem well worth substantial study, as they may relate to the CO_2 signal detection problems. River discharge is a likely candidate. Examples of what might be done in this area appear in studies of the Nile discharge, tree rings, and so forth, which offer less direct but still potentially useful measures of precipitation climate. Serious consideration should be given to determining whether such proxies can be used in a signal detection study.

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8. DETECTING THE CLIMATIC EFFECTS OF CARBON DIOXIDE: VOLUME SUMMARY

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CONTENTS

8.1 INTRODUCTION

The atmospheric carbon dioxide (CO_2) concentration has been increasing measurably as a result of human activities since the mid-19th century. As a result of these agricultural, industrial, and energy-related emissions, the CO_2 concentration in 1985 is estimated to be about 25% greater than the pre-anthropogenic baseline value, which is presumed to have been relatively stable for at least a few centuries preceding 1850 (Trabalka 1985).

Theoretical studies and empirical evidence indicate that increasing the atmospheric concentration of gases such as CO_2 which absorb infrared radiation will lead to a warming of the surface and to changes in precipitation patterns, snow cover, sea level, and many other climate variables. The best numerical models available indicate that global average surface temperatures would rise a few degrees Celsius if the CO_2 concentration were to increase from 300 to 600 parts per million by volume (ppm) and a new equilibrium climatic state were to become established (Schlesinger and Mitchell 1985).

The warming and other climate changes that might be expected as a result of the time-dependent increase in CO_2 concentration that has already occurred must currently be estimated using simple models that make many significant approximations. These simple models indicate that the global warming to be expected from the CO_2 increase over the last 150 years is probably in the range from about 0.3 to 1.1°C (see Chapter 4 of this volume and Hoffert and Flannery [1985]). The large range results from uncertainties in the initial CO_2 concentration (260–280 ppm), differences in the treatment of the ocean, and the differing model estimates of the climatic sensitivity to doubled CO_2 (1.5–4.5°C). As important, and equally uncertain, the results from the climate models suggest that, even without further growth in the CO_2 concentration, the 25% increase that has already occurred has committed the Earth to a further warming of about 0.1–1.0°C. The climatic effects of increasing concentrations of other infrared-absorbing trace gases would be in addition to the changes that are due to the increasing CO_2 concentration (Wang et al. 1985).

A further warming of several degrees Celsius, as is projected by models to occur over the next few centuries, could cause serious ecological and social perturbations. Mean surface temperature in historical times has been relatively stable, but even the minor fluctuations of temperature and precipitation experienced in the past have had significant societal impacts. A useful and thorough analysis of these potential impacts requires that the ranges, the rates, and the spatial resolution of projected climate changes be adequately specified (White 1985).

An important aspect of developing more accurate estimates is to compare available model projections with the climate changes that should already have been induced by the increased CO_2 concentration. In the companion volume to this report, MacCracken and Luther (1985) describe steps recommended to improve projections of future climate changes. This volume describes the efforts to identify such changes and to relate them to the model results as a test of model accuracy.

Over the past 10 years, there has been increasing recognition of the difficulties inherent in detecting and then attributing any climate change to the rising CO_2 concentration, especially in the presence of potential changes induced simultaneously by natural variations in solar irradiance or in the stratospheric aerosol loading resulting from volcanic eruptions. The spatial and temporal resolution of the data bases needed to undertake the search for the CO_2 signal is still insufficient, although significant progress is being made to improve the situation. New approaches have been developed for analyzing the data and for assessing the role of the oceans in delaying the expected equilibrium warming. Climatic variables other than temperature are beginning to be examined. This chapter summarizes what we have learned as a result of the studies described in detail in Chapters 1 through 7, in which complete references appear.

8.2 DETECTION STRATEGIES (CHAPTER 1)

Analyses of geological and other proxy indicators of climatic conditions over the Earth's history demonstrate that substantial variations have occurred over all time and space scales. Over historical times there have been variations in the mean temperature amounting to as much as perhaps a degree Celsius on the global scale and a few degrees over continental size regions. The causes of these variations have not been established, although atmosphere-ocean interactions, changes in stratospheric aerosol loading, and variations in solar irradiance may have contributed to the generation of this background level of variability (or "noise") from which the CO_2-induced perturbation (or "signal") must be distinguished.

The CO_2-induced perturbation has two stages. The primary effect consists of alteration of the local radiation environment. The secondary consequence consists of large-scale shifts in the atmospheric and oceanic circulations resulting from the integrated effect of the spatially nonuniform changes in radiative fluxes. Local radiation perturbations may, in places, be masked or even reversed by the changes in the general circulation. Developing a workable strategy for detecting this complex CO_2 signal from the climatic background variations as early as possible is important if the validity of climate models is to be evaluated and the necessary understanding is to become available so that impact studies can be conducted and policy options developed.

There are three elements of a CO_2-detection strategy, all intended to contribute to better projections of the future climate. First, monitor climatic conditions, assemble representative and accurate data bases of relevant variables, and analyze these records to determine the changes that have occurred, especially for those variables that are expected to change by statistically significant amounts. Second, identify and quantify the various factors—both natural and anthropogenic—that may be contributing to climate changes, fluctuations, and oscillations. Third, isolate that part of the climate change attributable to an increasing CO_2 concentration (together with concentrations of other trace gases) from changes and variations arising from other factors (both of high and low frequency), and then compare the results with the theoretically projected CO_2 effects.

This strategy requires knowledge of, to varying degrees, the time history of the changing CO_2 concentration and projections of how the climate should be changing in response to the changing CO_2 level. Although uncertainties exist, the CO_2 concentration was probably between 260 and 280 ppm at the beginning of the preindustrial era (Trabalka 1985). Climate models indicate that the increasing CO_2 concentration will induce a global warming—occurring at differing rates at different locations—but likely to be larger than elsewhere along zones of melting snow and sea ice. Seasonal variations in the warming are related, in part, to the varying patterns of the poleward retreat of snow and ice fields. While the troposphere is expected to warm, the stratosphere should cool. Average precipitation should increase on a global basis, but the seasonal and regional patterns of the changes are quite uncertain. The upper ocean should warm and sea ice should retreat; sea level is expected to rise as a result of thermal expansion and glacial melting.

A number of approaches have been proposed to search for this multifaceted CO_2 signal. In order to prove the cause-effect relationship, these theoretical approaches all require detailed quantitative projections of the expected climate changes as a function of the increasing CO_2 and trace gas concentrations. The limitations and uncertainties currently present in these projections (see Chapters 4 to 6 in MacCracken and Luther [1985]) make the search for the CO_2 signal particularly difficult. One of the problems is that long-term temperature records are available from only limited regions and model-projected perturbations are much less certain on a regional and seasonal basis than on annual average and global scales.

Attempts to identify the presumed CO_2 signal have become increasingly complex. The initial trials consisted of a traditional signal-to-noise analysis of the Northern Hemisphere land surface temperature series. This is especially difficult because the CO_2 concentration started increasing at about the same time as the early observational network was being established. As a result, no adequate, consistent record of baseline unperturbed climate exists to which the recent climate can be compared. In addition, averaging the early observations and the recent records over intervals of several decades, which is necessary because of the variability (noise) in the data, reduces the projected CO_2 signal.

A second approach focusing on noise reduction has also been pursued. This approach attempts to develop a less variable climatic record by subtracting variations that may be induced by factors other than CO_2 and then to apply signal-to-noise analysis. This approach requires that the climatic effects of these other perturbations be defined, either by signal-to-noise analysis (or other statistical analyses) or by model studies. Both of these approaches suffer limitations similar to those involved in detecting the CO_2 signal and, as a result, noise reduction analyses can produce uncertain identification of the CO_2 signal.

To bring some stability to the single variable dependence of the signal-to-noise and noise-reduction approaches, a multi-component, or "fingerprint," approach has been proposed. With this method, the records of several different variables, including the spatial patterns of the changes, are analyzed to determine if, as a coupled set, the projected changes due to increasing CO_2 are occurring. Although this approach has some mathematical advantages, it requires that adequate sets of several variables be available, and it places more dependence on (and is a more stringent test of) model projections of the expected CO_2 signal.

Each of the above approaches has advantages and disadvantages; all can make contributions, but none now provides a unique, unequivocal identification of projected CO_2 signal. The following sections review the application, or potential application, of these approaches to the different data bases described in the preceding chapters of this volume.

8.3 THE DIRECT EFFECTS OF CARBON DIOXIDE ON RADIATION (CHAPTER 2)

The direct effect of increasing CO_2 and trace gas concentrations is the alteration of the Earth's radiation balance (see Chapter 2 of this volume and Luther and Ellingson 1985), which in turn alters the climate and further alters the longwave radiances. Changes in the concentrations of these gases will have a relatively small influence on the absorption and scattering of solar radiation by the atmosphere. The local variability of solar radiation at the surface as a result of variations in cloudiness and other variables is so great that a surface network to detect the projected flux change would not be feasible; numerical representations of this effect can be much better tested by comparison of their predictions to laboratory experiments.

Larger changes are expected to occur in the spectral radiance of longwave (infrared) radiation in the atmosphere. Carbon dioxide and various trace gases (primarily methane, nitrous oxide, and various chlorocarbons) as well as water vapor and ozone, which may change as a result of climatic and chemical interactions, can absorb and emit longwave radiation at particular wavelengths, depending on their particular molecular structure and the atmospheric conditions. Thus, changes in the concentrations and distributions of these substances alter the upward and downward fluxes of infrared radiation, including, along with changes in temperature, alteration of the spectral radiance (i.e., the distribution of energy over wavelength).

Changes in fluxes occur throughout the atmosphere. These changes may be observable on a routine basis either at the surface or from space, whe[illegible] satellite platforms may facilitate observation. M[illegible]surement of changes in both the broadband [illegible] spectral components of the longwave fluxes at [illegible] the surface and in space have been consider[illegible] a basis for detecting the radiative and clim[illegible]fects of increasing CO_2 and trace gas co[illegible]tions. However, such measurements do [illegible]vide a practical approach because of the [illegible]ues of the flux perturbations. Given the [illegible] slow changes in atmospheric CO_2 concen[illegible] the tendency of the Earth-atmospher[illegible] adjust the outgoing longwave flux to balance the net incoming solar flux, changes in the outgoing broadband flux would be expected to occur mainly to compensate for planetary albedo variations—not primarily for CO_2 changes. Any broadband CO_2 signal would likely be several orders of magnitude smaller than both the net outgoing radiation and its seasonal variations, and hence be undetectable with any plausible observational and analysis system. Although the downward infrared flux at the surface for a doubling of the CO_2 concentration (and associated other atmospheric perturbations) could increase by several percent, the variations from place to place and day to day are so large that advanced analysis techniques would have to be developed and applied for several decades in order to separate the CO_2 signal from the noise.

Calculations with detailed radiation models indicate that at some wavelengths, however, the fractional change in flux radiance measured with high-resolution instruments could be quite large, and changes of the radiance as a function of wavelength could provide important information about the specific causes of the changes (e.g., changed gas concentration, increased temperature, etc.). Just as is the case for attempting to detect the broadband flux change, important limitations arise in attempting to detect these changes at the surface using spectral measurements. The complexities and uncertainties related to water vapor continuum absorption would make interpretation of the causes of flux changes difficult. In addition, spatial and temporal variability of the flux changes (and perhaps of the CO_2 *concentration itself*) would necessitate an extensive network of intercalibrated stations and relatively long records; such an effort would be expensive and does not appear a promising method for evaluating model projections.

Satellite measurement of shifts in the spectral distribution of outgoing infrared radiation at the top of the atmosphere on a several-decade time scale may provide an indication that CO_2 is increasing (an observation made much more easily and accurately from the surface), that the upward flux is being emitted from higher levels in the atmosphere, and that the atmosphere is warming on a nearly global basis. Proper interpretation of such results, which might well be difficult, would also require that the atmospheric distribution of temperature, clouds, water vapor, and other constituents be monitored, however. Thus, a very large data base would be required for both surface and satellite approaches.

Limitations to what could be learned arise for several reasons. An important source of uncertainty is the lack of adequate understanding about the radiative properties of atmospheric gases, particularly for the water vapor absorption continuum and for overlaps between water vapor and other gases. For measurements made from space, it is also extremely difficult to maintain calibration over long periods of time. In addition, measurements from space would not provide significant information about radiative and climatic changes near the surface, which is where climate models are in the greatest need of verification.

Thus, while a detection effort making use of satellite observations of spectral radiance may be feasible, it is not likely to provide the timely information needed to verify the performance of climate models.

8.4 DATA BASES AVAILABLE FOR ISOLATING THE CLIMATIC EFFECTS OF CARBON DIOXIDE (CHAPTER 3)

Successful detection of potential CO_2-induced perturbations is dependent upon data bases that describe the behavior of the climate. These data bases must include records of observations over extended periods and large areas of the globe so that the relatively small and slowly emerging CO_2 signal can be distinguished from the other local, regional, and temporal variations in the climatic variables. This statistically based requirement severely limits the set of climatic variables that can be considered. In this volume, the data bases for temperature, precipitation, and some oceanic and cryospheric parameters have been examined.

Surface air temperature has been the most widely and consistently observed measure of the climate. Over the past 10 years, intensive efforts have been made to expand and improve the quality of the hemispheric and global air temperature record. Over land, this has involved studies of historical documents and determination of possible influences of changing station environments, changing measurement procedures, and other biases and errors in the data bases. Monthly and annual average temperature anomalies at individual stations have been interpolated to latitude-longitude grids. This short-term averaging in time and space has helped reduce the variations due to local effects.

Within the last few years, considerable progress has been made in reviewing and assembling temperatures of the ocean surface layer and of the air temperature over the ocean. Although instrument-related and other potential biases in these data require attention, the observations have greatly expanded the data coverage and representativeness of the temperature record.

Although extending back in time only into the 1950s, the data base of tropospheric and stratospheric temperatures offers the potential for detection studies because of significantly reduced diurnal and spatial variability, that is, the background climatic noise. In addition, the projected stratospheric temperature change due to the increasing CO_2 concentration is larger and of opposite sign to projected tropospheric changes, so that it may be an important element in a multi-component analysis scheme. The main disadvantage, however, is that stratospheric temperatures respond mainly to radiative perturbations (induced by changes in CO_2, ozone, aerosol loading, etc.), and therefore do not provide a strong indication of how accurately climate models may be treating tropospheric and surface processes and feedback mechanisms. In addition, the reliability and accuracy of radiosonde data decrease in the stratosphere.

The precipitation record is also being substantially improved, but there remain many potential biases arising because the spatial scale of natural variations and perturbations in precipitation is smaller than the resolution provided by the observation network. The potential for detecting changes in precipitation by averaging over very large regions is discussed more fully in Chapter 7 of this volume and in Section 8.8.

Data bases needed to search for climate change in oceanic and cryospheric variables are discussed in Sections 8.6 and 8.7 in the context of the search for changes in these fields.

In addition to data bases of climatic variables, the noise-reduction approach, in particular, requires that data bases describing changes in other potentially influential factors be available. Aerosols injected by volcanoes into the stratosphere and changes in solar irradiance are the two phenomena believed to have noticeable climatic effects, although changes in tropospheric aerosol loading or in surface characteristics may also be having some influence. Correlations of various measures of aerosol loading (e.g., acidity in ice cores) and major volcanic eruptions suggest that most, but perhaps not all, major volcanic eruptions of the last century have been identified, but the global distribution of the resulting aerosol burden (much less its effect on planetary albedo) is generally poorly known. Estimated effects of the aerosol on solar radiation often come from limited observations, sometimes indirectly. The temporal resolution of the volcanic aerosol record does not match the monthly resolution of the temperature record. This may be partly due to estimation methods that provide only seasonal or annual resolution of the deposition of the aerosol, which may occur months or even years after the eruption.

Important progress is now being made in estimating changes of solar irradiance. Previous techniques have had to rely on surface measurements, which are of limited accuracy because of absorption and scattering of solar radiation by the atmosphere. Observations of solar diameter and satellite measurements of solar irradiance over the last 5 years are beginning to allow development of possible relationships between irradiance and proxy indicators such as sunspots so that a useful reconstructed record may become available for detection studies.

Much less can be and is being done to develop records of changes in other variables affecting the Earth's heat budget; records are needed that describe tropospheric aerosol loading (including source and composition) and surface characteristics such as albedo, land use, and moisture availability (i.e., extent of irrigation). Monitoring of the elements of the Earth's energy balance, including particularly the planetary albedo, are essential to help address these difficulties.

Although limitations remain, a review of recent activities indicates that important progress is being made in assembling and improving the data bases needed to undertake the search for the CO_2 signal.

8.5 SEARCHING THE TEMPERATURE RECORD (CHAPTER 4)

Most efforts to identify a potential CO_2-induced signal have examined the surface air temperature data set, for which the records are longest and most complete. Because of the high temporal variability that weather introduces into single station data, analysis of such records, even in the few cases when they are long and not biased by local effects, makes determination of the gradual temperature change due to rising CO_2 very difficult. To reduce the variability, spatial averaging of the local anomalies is usually performed in order to generate large-scale (e.g., polar, midlatitude, and tropical), hemispheric, and global averages. Such averaging presumes that the available observations, frequently with incomplete coverage, reasonably and consistently represent changes over the entire domain. This may not always be valid.

Analyses of the changes in Northern Hemisphere mean annual temperature computed from land station data over the last 100 years indicate that temperatures changed little from 1850 to 1880, decreased in the late 1800s, increased until the middle 1940s, decreased again until about 1970, and have generally increased thereafter. This pattern is not in accord with the monotonic temperature rise expected to result from the steady increase of the CO_2 concentration; influences other than CO_2 must also have contributed to the hemispheric temperature changes.

Marine air temperatures generally indicate a similar, perhaps steadier, cooling in the late 1800s, and warming during the first half of the 20th century. There is a high variability in the records on land and over the oceans prior to this century that may result from the limited spatial coverage of the observations. There may, however, be factors other than CO_2-induced climatic effects that are also causing the warming.

The combined record of land and ocean anomalies in surface air temperature since about 1900 indicates an increasing, but oscillatory, trend of about 0.4 to 0.6°C per century in the Northern Hemisphere (the 1940s to 1970s cooling, primarily of land areas, is less evident in the combined record). In those parts of the Southern Hemisphere for which data are available, there has been an overall warming since at least 1900 at a rate of between 0.3 and 0.7°C per century. These ranges in the trend reflect uncertainties in the data bases arising from such factors as uneven and changing station location (especially prior to about 1920) and from merging of data having different origin, length, and geographic coverage (especially in the Southern Hemisphere). Possible bias also arises from changes in station environments (e.g., warming due to urbanization).

The warming trend that has been observed is generally consistent with the warming expected from an increasing CO_2 concentration, but application of the various detection strategies has not yet resulted in unequivocal and quantitative identification of the CO_2 signal. This failure to convincingly detect the CO_2 effect occurs in part because it is difficult to account for what appear to be coincidental long-term climate fluctuations spanning several decades to a few centuries that are evident, for example, in long-term records at the few individual sites for which such records are available. The fluctuations are also suggested by ice core, tree ring, and some other proxy data. These records show that some of the warming and cooling over the last 100 years is probably a result of longer term fluctations unrelated to CO_2 that may either be obscuring or amplifying the CO_2 effect. Another diffuculty arises because of the limitations in the data set, including greatly varying spatial coverage over the last century. As a result, the hypothesis that most of the fluctuations may be due to factors other than CO_2 cannot now be rejected.

Signal-to-noise studies have divided the record into multidecadal segments. Comparison of, for example, the average hemispheric surface air temperature from 1955 to 1985 with earlier intervals of similar length shows no statistically significant temperature increase.

A wide range of noise-reduction studies have been applied to the hemispheric mean temperature record, and several have claimed identification of the CO_2 signal. Because of known physical links with climate, stratospheric aerosol loading, overall atmospheric turbidity, and solar irradiance have been studied as factors that may be causing much of the high-frequency variance obscuring the low-frequency CO_2 effect. The indirect nature and poor quality of the records of these factors and the limitations in our understanding of their climatic effects (both in magnitude and phase) together introduce substantial uncertainties into such analyses. As a result, although several studies have suggested the presence of a CO_2-induced effect, in most cases falling into the lower range predicted by models, these analyses are strongly dependent—and in varying ways—on largely arbitrary estimates of the climatic effects of changes in aerosol loading and solar irradiance which, to a large extent, are supposedly masking the presence of the expected CO_2-induced warming. The conflicts in the detailed interpretations, if not in the general conclusions, are discomforting and the results of the different diagnostic analyses can hardly be taken as mutually reenforcing. Even though one of these diagnostic analyses may be essentially correct, there is no way to tell which one until data bases are improved and the results of individual analyses reconciled.

Attempts have also been made to search for the CO_2 signal in the record of free air temperatures.

However, because of the short record, unknown natural variability, and instrumental problems, the free atmosphere temperatures have also not yet demonstrated a CO_2-induced effect.

A few multi-component detection analyses have been attempted. Their results have been limited by uncertainties in both the model results and by the limited availability of various data sets. For example, examination of the seasonal and latitudinal pattern of the temperature change has been attempted, but its dependence on uncertain model results—in addition to the other problems raised by analysis of the hemispherically averaged data set itself—results in severe constraints in demonstrating detection. A second approach that involves contrasting of tropospheric warming and stratospheric cooling has also not yet been successful in identifying the CO_2 signal.

Thus, although the data appear to indicate an oscillatory warming since the end of the last century, a unique and quantitative cause-effect attribution to increasing CO_2 and trace gas concentrations has not yet been convincingly demonstrated. In addition, even if the warming has been due to CO_2, its magnitude is almost a factor of 2 less than expected from those model results in which the time lag induced by the oceans is only a few decades; for model calculations in which the ocean lag time is estimated to be more than a century, there is much nearer, but still not excellent, agreement with observations.

8.6 SEARCHING THE OCEANIC RECORD (CHAPTER 5)

The ocean is an essential component of the climate system. The increasing CO_2 and trace gas concentrations are, in part, controlled by the oceans; increased evaporation from the warming oceans amplifies the greenhouse feedback, but the oceans' large heat capacity acts to slow the warming. This last interaction is extremely important, but still poorly understood. The atmospheric temperatures are influenced directly by the surface layer of the oceans; for the global air temperature to warm significantly, this surface layer must warm too. The mixed layer temperature, however, is also influenced by interactions with the deep ocean. Because of its large heat capacity and its coupling to the upper ocean layer, the deep ocean influences the rate of atmospheric warming. Additional interactions are less direct. Changes in the hydrologic cycle affect ocean salinity, and changes in atmospheric circulation (wind stress) influence ocean currents; both types of changes can influence how the oceans respond to CO_2-induced warming. Warming of the oceans and changes in ocean water mass as a result of the melting of glaciers and icecaps can increase sea level, whereas snow accumulation on icecaps would tend to decrease the sea level.

Although there is general agreement that the oceans may warm and ocean currents might change, the quantitative response of the oceans to the increasing CO_2 concentration has been much less studied than that of the atmosphere and many important aspects remain uncertain. Projections of the oceanic response are receiving increasing attention, but results to date are limited primarily to estimates of global oceanic warming, sometimes treated as only a function of latitude and depth. Most recent analyses have focused on the search for trends in sea surface temperature.

A serious limitation in oceanic studies is the limited quality of the data bases. The oceans are large, and only some parts are sufficiently well sampled at the surface. Much less information exists on subsurface conditions. Measurement methods have changed over the period of record. For example, the bucket method for measurement of sea surface temperature was gradually replaced by the engine intake method around the middle of this century, introducing a potential warming bias of up to several tenths of a degree Celsius, which is comparable to the expected CO_2-induced signal. Ship tracks have changed to make use of advantageous weather, and ship heights, weights, and structural materials have changed, and with them, the height and the environment surrounding the instruments.

The record of oceanic temperatures has uneven coverage over the Earth and is particularly poor in the middle and high latitudes of the Southern Hemisphere. Intensive efforts are being made to develop representative estimates of changes in the global average sea surface temperature. After correction for instrumental effects, the surface temperature data indicate a cooling of about 0.4°C from the late 1800s to about 1910, a warming of about 0.5°C into the 1950s, followed by a modest cooling. Important questions are whether some of the warming could be a result of the change in measurement technique and whether some of the variation could be unrelated to the changing CO_2 concentration. Thus, it appears premature to indicate that a CO_2-induced warming is present.

The situation is even more uncertain for subsurface ocean temperatures (which should warm), salinity (which should decrease if glacial melt is increasing the mass of ocean water), density (which combines temperature and salinity changes), and circulation (for which predictions are not available). There are very limited and diverse indications of secular changes in subsurface temperature, depending on where the measurements are being made, and apparent trends appear to be well below the noise level. Salinity has also shown mixed changes, with much attention focusing on a recent freshening of deep Atlantic waters. It is not clear whether this is a CO_2-induced perturbation. Density measurements actually suggest that the ocean may be cooling instead of warming, but these changes are still well within the range of natural variability.

Sea level changes provide another potential climatic indicator. Measurements of sea level have been made for many years at a number of stations and the histories of relative sea level can be generated for coastal stations over much, but not all, of the globe. However, the land areas in many locations are still rebounding or subsiding in response to the relatively recent removal of the Pleistocene icecaps, some of which were still present 8,000 years ago. Although partial corrections can be made, these problems introduce important uncertainties into detection analyses. For example, even after these adjustments for coastal deformations, the sea level trends over the last century are not the same over all areas of the globe, indicating that the records at particular stations are still contaminated by factors such as crustal motions and circulation changes that may be as large as, but have little or nothing to do with, any CO_2-induced effect.

On a global basis, recent analyses suggest that relative sea level has been rising at a rate of about 10–25 cm per century over the last 100 years (Chapter 5 and Polar Research Board 1985). Thermal expansion induced by CO_2 warming and ice melt may be in part responsible. The extent of oceanic warming since the middle of the last century, at least at the surface, is becoming better known. Present studies appear to indicate that thermal expansion is contributing to sea level rise, although it may not be the only factor. However, more observations, particularly on the changes of subsurface temperature, are needed to better determine the role of this effect since the rise may, at least partially, also be explained by factors unrelated to CO_2. The angular shifts in the Earth's axis of rotation are generally consistent with the hypothesis that melting of mountain or polar icecaps is a major contributing cause to the sea level rise, but there are also other plausible hypotheses about the causes of the axial shift (Polar Research Board 1985). More data and better projections will be required to enable a CO_2 signal in sea level to be identified.

In summary, although sea level is rising and sea surface temperatures are increasing, both in qualitative accord with expected CO_2 effects, these changes cannot yet be related quantitatively to CO_2-induced perturbations. Detection of the CO_2 signal will require better model predictions and increased efforts to monitor and analyze oceanic conditions.

8.7 SEARCHING THE CRYOSPHERIC RECORD (CHAPTER 6)

Snow and ice cover are indicators of regional temperature and precipitation. Because of the large latent heat of fusion, snow and ice thickness and areal coverage combine to provide a complex and implicit integral of climatic conditions over a period of weeks to months and, in the case of icecaps, millennia. Thus, in some sense, these indicators average out some of the unwanted short-term noise.

Model sensitivity experiments with a doubled CO_2 concentration indicate that, once a new equilibrium climate is established, the thickness, duration, and extent of snow cover and sea ice will be substantially reduced and melting of mountain glaciers and polar icecaps will be accelerated. These large changes result from, but also contribute to, the amplified warming in high latitudes, particularly through snow and ice albedo feedbacks. This amplified climatic response in high latitudes suggests that cryospheric variables may be particularly sensitive indicators of climate change. Because appropriate calculations have not been done, estimation of what the CO_2-induced changes should have been over recent decades has had to be based primarily on interpolation between the widely differing control and doubled CO_2 equilibrium simulations. This may be a poor approximation for several reasons. Snow and land ice accumulation are controlled by precipitation and temperature. In very cold regions such as the Antarctic, small increases in temperatures may lead to more snowfall rather than less. For retreating sea ice and snow cover, the seasonal dependence may lead to a nonlinear relationship. In addition, the models are not yet capable of predicting the large regional variability of snow cover that is now observed. Thus, while we may have a sense of how the cryospheric climate will ultimately change, there are not yet accurate quantitative model projections of what the large-scale CO_2 signal should be, especially on the regional scale.

The available data sets are improving. Snow cover and sea ice have been monitored for almost 20 years from satellites with relatively consistent criteria used to estimate extent. Snow cover can also be reconstructed from surface network data for the last few decades, although considerable gaps in spatial coverage do exist. Prior to World War II, however, knowledge of the snow and sea ice in the Northern and, particularly, the Southern Hemisphere is very incomplete and irregular and, therefore, any reconstruction of past large-scale changes of snow and ice cover would be highly uncertain. In some regions, records of lake, river, and harbor freeze-up and break-up dates may serve as proxy information for air temperature in data poor regions, although precipitation events can also be a contributing factor.

Data on mountain glaciers and the polar icecaps are quite limited. The snow/ice budgets of Greenland and Antarctica, for example, are poorly known and cannot be used to provide definitive estimates of CO_2 effects on sea level (Polar Research Board 1985). Mountain glaciers must be studied on a case by case basis, and their past size estimated via indirect techniques, resulting in poor accuracy of global summaries. There is no evidence that the West Antarctic ice sheet may be collapsing or decaying at rates accelerated by a CO_2 perturbation.

Despite the limitations in model projections of expected effects and in the length of available data bases, analyses have been made to determine whether trends are evident. Substantial variability is evident on hemispheric, regional, and local scales, with anomalies seeming to persist for several years to a decade or more. Thus, clear identification of possible trends is not possible at this time. No trend is apparent in Northern Hemisphere snow cover. The summer extent of Arctic sea ice was at a minimum in the 1950s and has increased since. The average summer sea ice extent around parts of the Antarctic apparently decreased from the 1930s to the 1970s, but much of this decrease may be due to fluctuations induced by factors other than CO_2-induced warming.

Locally, changes may be occurring in some other cryospheric indicators. For example, there are indications of later freeze-up and earlier break-up of ice in undisturbed lakes in Scandinavia and Canada and of higher temperatures affecting the permafrost. These observations have not, however, been confirmed over wider regions. Many mountain glaciers are retreating and thereby contributing to sea level rise, but others are advancing and it is difficult to relate these changes to CO_2-induced perturbations.

In summary, limited data, inadequate understanding of the causes of cryospheric variability, the absence of decreasing trends in snow and ice cover, and a poorly defined CO_2 signal combine to make cryospheric changes an uncertain indicator of the suggested CO_2 signal.

8.8 SEARCHING THE PRECIPITATION RECORD (CHAPTER 7)

The warming of the Earth's surface resulting from increasing CO_2 and trace gas concentrations will increase the intensity of the hydrologic cycle. Climate models project that the doubling of the CO_2 concentration will increase the precipitation and evaporation rates by 5–10% on a global basis. Precipitation increases are likely to occur in low and high latitudes with both increases and decreases occurring in midlatitudes, depending on season and location. Evaporation is projected to increase, particularly during the warm seasons over midlatitude continents. These projections are, however, highly model dependent and should, therefore, be viewed as suggestive, but uncertain, especially in their regional pattern.

Precipitation has been measured at many locations for very long periods. The almost complete lack of measurements over the oceans, however, makes construction of a global average estimate impossible. As a result, comparisons with model projections can be attempted only over land regions. However, a variety of problems can arise with the land-based records. For example, station relocation or changes in the station environment (higher tree height, etc.) can affect the observation. Precipitation exhibits high local variability; as a consequence, a representative record of regional scale precipitation requires a denser network of stations than is readily available. To reduce the effects of the local variability, averaging of observations over wide regions is necessary.

An alternative might be to simply monitor and study the runoff from large watersheds. Unfortunately, such detection studies would be very difficult because virtually all river basins have been altered by societal activities—whether by diking, damming, or removing water for urban or agricultural use. In addition, runoff is dependent on the difference between precipitation and evaporation, which is in turn dependent on climate and land use. Thus, interpretation of runoff data would be difficult, although further investigation may be warranted.

Because of the complicated pattern of the CO_2 signal, precipitation variations must be examined on a regional scale. In virtually all regions examined, the interannual and decadal variability is such that it would be exceedingly difficult to detect a relatively small CO_2 signal. Of particular interest is that some of the variability appears to be correlated to other causal factors, such as the Southern Oscillation. The CO_2 signal would have to rise above these effects unless such variations could be removed via a noise-reduction approach.

Thus, although changes in precipitation amount may be a major climatic effect of an increasing CO_2 concentration, convincing detection would be very difficult.

8.9 DOES THE EVIDENCE CONVERGE?

The findings from each of the lines of inquiry taken individually are, by themselves, insufficient to constitute convincing evidence that the climate models are correctly projecting the effects of the increasing CO_2 concentration on climate. However, to varying degrees, the evidence is generally consistent with, or at least not contradictory to, model projections of such effects. Very little of the observational evidence is in direct conflict with the model calculations, but important aspects of the projected changes are not yet evident.

Temperature records from both the Northern and Southern Hemisphere indicate a warming over the length and area covered by reliable records and the overall trend is within the range expected from climate models, especially if account is taken of the delaying effect of oceanic heat capacity. The coolings of Northern Hemisphere land and ocean areas from the late 1800s to about 1910 and from about 1940 to 1970, however, are unexplained and must either be artifacts of the data sets or result from factors other than CO_2-induced effects, including perhaps internal oscillations of the climate system. The warming in those parts of the Southern Hemisphere where data are available has been somewhat steadier and is more evident than in the Northern Hemisphere, but oscillations also occur there. Given these differences in regional warming, it is difficult to determine to what degree the recent warming is due to CO_2 and to what degree it may be due to other factors.

Sea level rise is in broad accord with the expected CO_2 warming. The apparent contribution from melting of mountain glaciers could be a response to the CO_2-induced warming. The expected effect of thermal expansion is well within the observed range of sea level rise, and so it could be contributing to the observed rise. This conclusion is preliminary, however, because much more careful analysis of the possible role of other causes is needed.

The coupling between temperature and sea level merits much closer investigation, especially with respect to the oceanic role in delaying the climatic warming. If the oceans are active in taking up heat and delaying the climatic warming (which could reconcile the high sensitivity of many recent model simulations with observations), then sea level should be affected strongly by thermal expansion and relatively modestly by glacial melt; if the oceans play only a limited role in delaying the warming (a situation that would suggest that recent model results have too high a sensitivity), then sea level should now be rising mainly due to the glacial melting. More accurate models and better data are needed to permit the use of these variables in a coupled, fingerprint approach that can better resolve the role of the oceans.

In this respect, free air temperature and ocean salinity records may be helpful. Study of changes in the tropospheric lapse rate and in the relative changes of stratospheric and tropospheric temperature may be useful in a multi-component (fingerprint) approach to CO_2 detection. Salinity trends would be helpful in differentiating between the two possible causes of sea level rise—thermal expansion and the melting of ice—although detailed mapping of the mass balance of mountain glaciers and polar icecaps may also provide important data.

Observed precipitation and cryospheric changes are too equivocal to provide support for the presence of the CO_2 signal. Until progress can be made on predicting regional perturbations and explaining the causes of the observed variations, these climatic parameters seem to offer little additional information.

In summary, the observational evidence to date is broadly compatible with many aspects of the model predictions. Increasing CO_2 and trace gas concentrations undoubtedly have a warming effect on climate, but it is not now possible to show definitively to what degree the current climate changes result from the increases in CO_2 and trace gas concentrations and to what extent they result from other factors. Without such a determination, the climate record is not now adequate to narrow the rather large range in temperature sensitivities to CO_2 rise suggested by the current set of climate models.

Some recent analyses have predicted that a convincing case for the detection of CO_2-induced changes should be possible by the end of this century. Several assumptions underlie these claims. Detection will require continued monitoring of climatic variables and factors other than CO_2 that may alter the climate by changing the planetary albedo and atmospheric energy balance. It is presumed that such monitoring will occur. A second prerequisite to CO_2 detection is that the relative quiessence of volcanic activity and solar variability continue for at least the next few decades or until we achieve a much better quantitative understanding of the climatic effects of these factors than is now available. Even so, an increased role by other factors could still continue to hide the CO_2-induced signal. In addition, successful detection will depend on having available reliable indications from climate models of how the increasing CO_2 and trace gas concentrations should have affected the recent climate. The uncertainties and differences among model results present in equilibrium and transient simulations which are now available pose serious limitations to development of convincing quantitative relationships and to statistical detection of CO_2-induced climate change within the next decade.

The following chapter presents recommendations for making progress over the next 10 years in improving understanding and for helping to determine how sensitive the climate is to the changing CO_2 and trace gas concentrations. These studies will be essential in order to determine how long it will take to detect the CO_2-induced signal and to calculate the climatic warming that will ultimately result from the increased concentrations already present.

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9. RECOMMENDATIONS FOR MONITORING AND ANALYSIS TO DETECT CLIMATE CHANGE INDUCED BY INCREASING CARBON DIOXIDE

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9.1 INTRODUCTION

Diagnostic analyses of observational data are essential to identify, or "detect," the effects of increasing carbon dioxide (CO_2) and trace gas concentrations on the past and present climate and, thereby, to increase confidence in model projections of future climate. Detection of such effects requires that climate changes induced by these greenhouse gases be isolated from natural fluctuations arising as a result of processes within the climate system and from variations induced by forcing factors external to the climate system. The effort to identify the CO_2 and trace gas climate signal must be guided by the results calculated by numerical models (described in a companion State-of-the-Art report *Projecting the Climatic Effects of Increasing Carbon Dioxide* [MacCracken and Luther 1985]). Diagnostic studies of empirical data have the important independent role of determining whether identified changes in climate are in quantitative agreement with model predictions of the effects of increasing CO_2 and trace gas concentrations. This is an essential step in assessing the validity of model projections of the future climate.

To accomplish these objectives, representative data sets of suitable length and quality must be available so that the natural variations of climate can be established (particularly the low-frequency components) and any effects of greenhouse gases can be isolated. In addition, records are required of changes in various external factors that are known or predicted to have influenced the past climate. More theoretical and statistical studies are needed to relate the variations in these factors to the climate record. Such needs have also been identified in earlier studies (e.g., World Meteorological Organization 1982; National Research Council 1983; World Climate Research Programme 1984).

Limitations in the length of available records and possible influences of perturbations caused by unknown variables introduce important constraints on our ability to detect CO_2-induced climate changes. To accomplish both identification and quantitative isolation of the CO_2 and trace gas signal requires the careful application of suitable detection strategies and rigorous statistical techniques.

9.2 DETECTION STRATEGIES

Any detection strategy intended to identify past climatic effects of increasing CO_2 and trace gas concentrations must take into account warming and cooling trends resulting from factors other than CO_2 that may have exerted an influence within the time span of interest. Climatic variations induced by volcanic aerosol injections, solar variability, and atmosphere-ocean interactions have probably been of most importance. The climate of the next century will be a composite of the warming expected because of increasing CO_2 and trace gas concentrations, the effects of high-frequency weather noise, and longer term fluctuations due to internal climatic oscillations and external forcings unrelated to CO_2 and trace gases. Because the possibility cannot be dismissed a priori that the future greenhouse warming could be significantly amplified, or perhaps even temporarily obscured, by other factors, detection strategies must take such a possibility into account.

In studies carried out to date, three main detection strategies have emerged, as explained more fully in Chapter 1.

9.2.1 Signal-to-Noise Ratio Analysis

In this approach, it is expected that, as the CO_2 and trace gas signal becomes sufficiently large, it will emerge out of the "noise," that is, become distinguishable from the background natural variability of climate. Only one variable at a time is considered using this approach. This strategy is currently restricted by the limited record length of most data bases and by the small number of variables for which a model prediction of the signal is well established. Additional research is required to improve the reliability of estimates of the CO_2 and trace gas signal and the quality of the data bases in which the signal is being sought.

9.2.2 Noise-Reduction Analysis

Noise-reduction analysis attempts to reduce the noise level by accounting for and factoring out the climatic fluctuations that are occurring as a result of causes other than increasing concentrations of the greenhouse gases. However, because the causes of natural climatic variability are not yet well established and the climatic impact of individual known or suspected forcing variables (or their proxies) is not quantified, the potential for significant noise removal in a statistically convincing way is not now high. As in the signal-to-noise ratio approach, the noise-reduction method considers only a single variable at a time.

9.2.3 The Fingerprint Strategy

Essentially, the argument for the fingerprint strategy is that, although the evidence from a single climatic variable may be insufficient or equivocal, when a number of variables (or time- or space-dependent facets of variations in a single variable) are considered, confidence that detection has been achieved may be accomplished more readily. From a statistical point of view, coupled changes in a pair or a set of variables may be easier to detect than changes in any one of the variables. In addition, a significant climate change may be identifiable by a multicomponent strategy, even in the absence of any global mean temperature change (as might occur if CO_2 and non-CO_2 effects tended to cancel). Theoretically, it should be possible to select a suitable combination of variables at an appropriate combination of sites and/or levels in the atmosphere to compose a signal that would be uniquely characteristic of increasing CO_2 and trace gas concentrations.

This approach, however, faces two difficulties. First, although this method may not require data sets that are as long as those needed by the first two strategies, it does require data sets including more variables and having better temporal and spatial resolution. Second, this strategy requires a considerable modeling effort to reliably define the multivariate structure of the CO_2 signal. As indicated in MacCracken and Luther (1985), models do not yet exhibit sufficient quantitative agreement to allow definitive application of this strategy, although such application is a long-term goal.

9.3 GOALS AND RECOMMENDATIONS

The record of the past climate can provide important clues to the future climate. Such information, however, is not easily extracted. A multifaceted approach is essential, encompassing development of a detection strategy, assembly of past records, expansion of the monitoring program, empirical analyses, and model simulations. The following sections present recommendations in support of these goals.

9.3.1 Detection Strategies

Each of the three detection strategies currently has specific strengths and weaknesses. Simply waiting several decades, as required by the signal-to-noise strategy, might seem to be an attractive approach. Early detection, however, is essential so that the validity of future climate projections can be evaluated and, if appropriate, policy options considered at an earlier time. This is particularly critical if the oceans are moderating what may be quite large equilibrium climate changes that are inevitable, but will not become apparent until well after emission of the CO_2. Thus, to provide the greatest likelihood for progress in detecting the climatic effects of increasing CO_2 and trace gas concentrations, a comprehensive approach involving the pursuit of all three detection strategies promises to provide the best results. It must encompass a broad spectrum of research, analysis and monitoring efforts, and close collaboration between and a rational balance of modeling and empirical studies.

In pursuing individual detection strategies, the focus should not be solely on behavior of the global mean temperature. This is true largely because the global mean temperature signal may continue to be obscured by low-frequency natural variability in the immediate future, even though the climatic sensitivity to an increasing CO_2 concentration may be high (Chapter 4). To overcome this difficulty, continued efforts to improve data bases of temperature (Chapter 3) and other variables (Chapters 5 through 7) are needed so that a broader range of climatic variables may be considered.

Additional efforts must be devoted to achieving a better understanding of natural climatic variability, which constitutes the noise from which the signal resulting from increasing CO_2 and trace gas concentrations must emerge. Explanation of the decadal and longer time scale components of this variability is critical to the detection issue and demands a better understanding of its causes and spatial patterns. This applies not only to surface temperatures, but to all climatic parameters that might be directly or indirectly related to the effects of increasing greenhouse gas concentrations.

To attribute an identified trend or change to increasing CO_2 and trace gas concentrations, model estimates of projected changes must be improved so that the signal being sought is unambiguous (Chapter 8). Such improved model estimates are especially needed in order to apply the multicomponent (fingerprint) detection strategy. Detection of potential CO_2-induced effects has proven difficult because of unexplained climatic variations and the possibility that factors other than CO_2 may also be influencing the climate. Further progress requires a monitoring and analysis strategy if early detection is to be achieved.

Goal 1: Development of a strategy for detection of CO_2-induced climatic effects.

Recommendation 1A.
Develop a "CO_2 and trace gas climate watch" encompassing a spectrum of research, analysis and monitoring efforts, and including close collaboration between and a rational balance of modeling and empirical studies.

Recommendation 1B.
Pursue signal-to-noise, noise-reduction, and multicomponent (fingerprint) strategies as part of a comprehensive approach to detection of the climatic effects resulting from increasing CO_2 and trace gas concentrations.

Recommendation 1C.
Analyze records of climatic variables in addition to surface air temperature. Ocean temperatures, the intensity of the hydrologic cycle, snow and ice extent, sea level, and free atmosphere variables may be particularly important.

Recommendation 1D.
Document climatic variability on time scales ranging from several years to several centuries. The reliable determination of the baseline climate is an essential prerequisite to effective projection of future perturbed climate states.

9.3.2 Observational Records

The observational component of a comprehensive detection strategy will require improving the available records and encouraging systematic monitoring of the variables of greatest importance. For the near future, surface air temperature will have to remain the backbone of detection studies. Its signal is reasonably well defined and its past record is longer than that for any other variable. However, although considerable progress has been made in recent years, large-scale average temperature sets may still contain potentially serious biases resulting from the uneven areal coverage of stations, the changing sampling procedures over the oceans, and the effects of urbanization and station relocation for stations near or in cities. Initial attempts to remove urbanization effects have altered parts of the hemispheric average record by about 0.1°C (Chapter 3). Further corrections are likely to result in additional small changes, even though they may be important on the regional scale.

Efforts should continue to detect and correct inhomogeneities in data records and to remove any remaining biases. Comparisons between the different large-scale data bases of surface air, free atmosphere, and sea surface temperatures must continue to be made, and the existing discrepancies must be explained and corrected. In particular, the effort to integrate the ocean and land temperature data sets deserves strong encouragement.

To date most efforts to identify the CO_2 signal have concentrated on expanding the areally averaged surface temperature onto a global scale and comparing the results with the model-predicted global warming. These temperature data sets are not long enough for studies of natural climatic fluctuations with periods longer than several decades. Although the regional-scale CO_2 and trace gas signal may not yet be well-defined, thorough analyses of data at individual stations with records of 200 years or longer may be warranted.

Free atmosphere temperatures, either directly measured or calculated from pressure data, are especially important in the fingerprint strategy because they are not affected by urban warming. Analysis and removal of potential instrumental biases in the existing data should be continued.

Assembly and improvement of regional records of parameters other than temperature can also support diagnostic studies of the causes of past climate change and are needed to help implement the fingerprint detection strategy. Of special importance are cryospheric data, which may provide indications of trends in high latitudes where the natural variability of climate is greatest, and precipitation and pressure records, which may provide indications of large-scale shifts in climatic zones.

As a basis for developing the data bases required for extending the analyses described in Chapters 4 to 7, work on data records must be continued.

Goal 2: Improvement, assembly, and integration of climatic data sets.

Recommendation 2A.
Remove biases from available data sets so that nonclimatic changes can be ruled out. Removal of biases from records of free atmosphere thickness, temperature, and humidity data would greatly expand availability of high quality data sets.

Recommendation 2B.
Screen and combine the land and near-surface ocean temperature data sets. For the first time, this could provide a homogeneous, long-term, high-resolution data base covering a large fraction of the area of the Earth.

Recommendation 2C.
Assemble long-term, global-scale data sets of precipitation, surface snow and ice cover, and atmospheric circulation. The quality of these data sets must also be assessed.

9.3.3 Monitoring Requirements

The data needed for detection purposes must be continuously collected, checked for reliability and compatibility with earlier observations, and disseminated in a timely fashion. Frequent intercalibrations and comparisons may be needed to ensure long-term homogeneity of the data series.

Surface air and free atmosphere temperatures, water vapor mixing ratio, atmospheric thickness, sea surface temperature, and sea level are of particular importance for detection analyses. Monitoring of these variables should continue as a part of the World Weather Watch and the World Climate Research Programme (Parker 1985). However, present data coverage is inadequate in the southern middle latitudes and over many ocean areas, and the network should be improved.

Most stations making measurements of the upper troposphere and stratosphere take only two soundings a day and at different solar times. These limitations complicate comparisons. An additional radiosonde launch at local midnight at selected stations would enhance the utility of the upper air data sets for climate studies.

Satellites provide the only truly global coverage; their use in the detection effort should be expanded. Although they have been in operation for only about 20 years and their sensors and flight paths have changed many times, their data are of special importance because they represent the most likely means by which gaps in present data sets can be filled. (A particularly glaring gap in which satellite data could be of use is over the Southern Hemisphere ocean area between 40° and 70°S). In addition to providing necessary radiance data for detection (see Chapter 2), satellites may indirectly be able to provide useful data on variables other than temperature; for example, monitoring outgoing longwave radiation may serve as a useful, large-scale measure of precipitation. Wherever possible, satellite data should be calibrated against the existing observational record and newly acquired data from oceanic buoys, because assurance of homogeneity and compatibility of satellite information is of fundamental importance.

The oceans play a crucial role in controlling decadal and longer time scale climatic fluctuations, and they are likely to delay and modulate the warming expected as CO_2 and trace gas concentrations rise. Improved monitoring of surface and subsurface ocean parameters (particularly the vertical structure of the ocean), rapid dissemination of data, and improvements in ocean modeling are needed. (Such improvements are currently planned as part of the World Ocean Circulation Experiment [WOCE] of the World Climate Research Programme.)

The records of other key internal variables of the climate system, such as snow and ice cover, cloud cover, and surface albedo need to be improved. Changes in cloud cover or surface albedo of only a few percent could significantly alter the Earth's radiation balance and thereby moderate or amplify the projected climatic effects of increasing CO_2 and trace gas concentrations. Sea ice, in addition to its direct effect on the Earth's radiation budget, has a profound influence on high latitude oceanic heat and salinity fluxes and so is of special importance. Existing and planned observational programs (such as WOCE, the International Satellite Cloud Climatology Project [ISCCP], the Marginal Ice Zone Experiment [MIZEX], and the International Satellite Land-Surface Climatology Project [ISLSCP]) need to be carried out to address these monitoring needs.

It is vitally important that parameters most closely related to major causal mechanisms of climate change be monitored continuously and without interruption. Such parameters include, among others, atmospheric CO_2 and trace gas concentrations, stratospheric aerosol loading, and solar irradiance. Under the present satellite program, there is some possibility that broadband solar radiation monitoring will not be carried out in the late 1980s and early 1990s. This lapse would present a serious setback to the detection program and should be avoided.

Both ground-based and satellite monitoring of stratospheric aerosols may help to determine the causes of high-frequency oscillations in the Earth's heat budget and so must be continued. Direct satellite monitoring of the Earth's radiation budget is at present limited to the middle and low latitudes and suffers because of changes of spectral bands and the timing of satellite overpasses.

Monitoring is the basis for the detection studies considered in Chapters 3 through 7. Although extensive, and seemingly unexciting, it is an essential component of global climate studies.

Goal 3: Continued and augmented monitoring of climatic variables.

Recommendation 3A.

Ensure the continuous and uninterrupted monitoring of important climatic variables. Data reliability and compatability with earlier observations must be confirmed if the records are to be useful in climatological studies.

Recommendation 3B.

Continue monitoring of the free atmosphere temperature, thickness, and water vapor mixing ratio. The station network should be expanded over the oceans and especially in the middle and high latitudes of the Southern Hemisphere. Additional special soundings could usefully be made at local midnight at selected stations.

Recommendation 3C.

Improve use of satellite data. Satellite measurements of surface temperature must be calibrated with surface-based observations so that the spatial coverage of the temperature record can be improved, particularly in the Southern Hemisphere. Expanded observation of the Earth's energy budget and planetary albedo are essential. The behavior of ocean temperature and currents should also be monitored. Attention should be paid to normalizing spectral ranges and orbital characteristics of weather satellites with the highest potential utility in studies of climate change.

Recommendation 3D.

Monitor climate system components that can amplify or moderate model projections of expected climate changes resulting from increasing CO_2 and trace gas concentrations. Coverage should be expanded, especially over the oceans; cloudiness and the cryosphere deserve particular attention. Special observation programs can play an important role in building reference data bases.

Recommendation 3E.

Observe factors in addition to the CO_2 concentration that can alter the Earth's radiation balance. Variables to be monitored at the surface and, where appropriate, via satellite, include solar irradiance, surface albedo, aerosol concentrations, and atmospheric concentrations of methane, nitrous oxide, chlorocarbons, ozone, and other trace gases.

9.3.4 Empirical Analyses

Better understanding of the spatial and temporal patterns of observed changes of climate is essential if the climatic effects and ecological and societal impacts of increasing CO_2 and trace gas concentrations are to be understood and their importance evaluated. For all three detection strategies, it is essential that statistical methods be applied with mathematical rigor. In expectation of better scrutinized and more extensive data sets and with the improving reliability of model results, the development and application of improved statistical techniques may be necessary.

Of the various external factors believed to be of major importance, the variations of volcanic and solar activity deserve most attention. Current empirical and modeling studies show some disagreement on the timing and duration of the cooling from injections of volcanic aerosols into the stratosphere. There is also disagreement among data bases that are used as proxy measures of past stratospheric aerosol loading, and these discrepancies require reconciliation as part of an effort to develop a more adequate record of the temporal and spatial distribution of stratospheric aerosol loadings over the last hundred years (Chapter 3).

Proxy measures indicating that solar irradiance may have varied in the past are uncertain. These records can be tested when sufficiently long records of satellite measurements become available. Diagnostic and theoretical analyses will be required to determine whether significant multidecade fluctuations in solar output exist and whether these may be causing climatic variations on these time scales. Empirical analyses of the past climate provide the basis for determining the causes of past climate change and for verifying whether theoretical models used to project future change are sufficiently encompassing and accurate.

Goal 4: Determination of the causes of past climate changes and variations.

Recommendation 4A.

Continue empirical diagnostic studies of the causes and mechanisms of climate change. Development and application of more sophisticated analysis techniques may be needed.

Recommendation 4B.

Test and calibrate proxy records of past variations of stratospheric aerosol loading and of solar activity with satellite data. Empirical analyses of solar output should be intensified to clarify or disprove the existence of decadal or longer time-scale fluctuations.

Recommendation 4C.

Pursue diagnostic studies of the climatic response to changes in solar variability, atmospheric aerosol loading, ocean circulation, and other variables. If these effects can be determined, the CO_2 effect should be more easily identified.

9.3.5 Modeling Requirements

Results from multidimensional atmosphere-ocean climate models suggest that the relatively simple diagnostic analyses that have been conducted to date may have been oversimplified in their representation of the climate's response to external and internal forcing (Hoffert and Flannery 1985). This is especially true with regard to the role that the oceans may play in climatic fluctuations. Improvement of the approximations used in diagnostic analyses may be possible if special model simulations are carried out as part of a coordinated effort.

A fingerprint approach offers the greatest potential to improve on single-variable signal-to-noise and noise-reduction analyses. Practical application of this multicomponent approach requires improvement in modeling of the multicomponent facets of the expected signal from CO_2, trace gases, and other causal factors so that a distinct fingerprint for each can be developed. As discussed in various chapters in MacCracken and Luther (1985), the confidence that can be placed in model predictions of the geographical and seasonal details of the perturbations induced by increasing CO_2 and trace gas concentrations is not yet sufficiently high. Improved estimates are needed to define an unambiguous multicomponent (fingerprint) signal. Exploratory studies of details such as the geographical and seasonal character of temperature variations, day-to-night temperature differences, and the seasonal variation of the snow-and-ice transition zone should be attempted.

Confirming a quantitative relationship between climate changes and CO_2 and trace gas concentrations requires accurate simultion of the expected changes. Present inadequacies severely limit diagnostic analyses seeking to detect the CO_2-induced effects.

Goal 5: Accurate projection of the climatic sensitivity to the changing CO_2 concentration and other causal factors.

Recommendation 5A.

Improve the capability of numerical climate models to project the spatial and temporal evolution of the climate in response to increasing CO_2 and *trace gas concentrations. This is essential in order to provide a reliable estimate of the spatial and seasonal pattern of the CO_2 signal to be detected.*

Recommendation 5B.

Conduct climate model simulations that consider the time-dependent changes in CO_2 and trace gas concentrations. Such results are needed to develop an appropriate, time-dependent, multicomponent climate signal.

Recommendation 5C.

Carry out special climate model simulations considering changes in atmospheric aerosols and solar irradiance. These studies may permit attribution of observed variations to particular causal factors. Sensitivities to variations in internal factors (e.g., clouds, albedo, ocean circulation) also require further investigation.

9.4 SUMMARY

The objective of detection studies is the unambiguous, quantitative confirmation of the model-predicted climatic effects of increasing CO_2 and trace gas concentrations. This has not yet been achieved, although there are strong qualitative indications. That the Northern Hemisphere average surface air temperature has warmed approximately 0.5°C since the end of the last century is almost certain, but attribution of this temperature increase to changing CO_2 or trace gas concentrations remains ambiguous, mainly because of the unexplained temporal pattern of the warming. A substantial increase in the level of confidence in this attribution by single-variable analyses could be left to time, but that might take decades. There is increasing recognition, however, that the increase in CO_2 concentration since the last century may already have committed the world to a substantial further warming. To address this potential concern, an active monitoring and analysis program to detect the climate changes being induced by increasing CO_2 and trace gas concentrations must be maintained.

The principal obstacles in current detection efforts are the inadequate understanding of the causes and extent of natural climatic variability and the limited detail provided by climate modeling studies. We recommend the simultaneous pursuit of a suite of detection strategies and closer cooperation between data analysts and those using models to study the climate of the recent past and future. Such a combined effort should provide the substantial improvement in understanding needed to achieve reliable detection of the CO_2 and trace gas signal.

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GLOSSARY

AAAS	American Association for the Advancement of Science
AR1	First-order autoregressive
AVHRR	Advanced high-resolution radiometer
BT	Bathythermograph
CCM	Community Climate Model
CDIC	Carbon Dioxide Information Center
CFC	Chlorofluorocarbon
CIRES	Cooperative Institute for Research in Environmental Sciences
COADS	Comprehensive ocean-atmosphere data set
DOE	Department of Energy
DVI	Dust veil index
EBM	Energy balance model
EOF	Empirical orthogonal function
ESMR	Electrically scanning microwave radiometer
GCM	General circulation model
GFDL	Geophysical Fluid Dynamics Laboratory
GISS	Goddard Institute for Space Studies
GLA	Goddard Laboratory for Atmospheres
HIRS	High-resolution infrared radiation sounder
IASH	International Association of Scientific Hydrology
IGY	International Geophysical Year
ISCCP	International Satellite Cloud Climatology Project
ISLSCP	International Satellite Land Surface Climatology Project
MAAT	Mean annual air temperatures
MAT	Marine air temperature
MIZEX	Marginal Ice Zone Experiment
NCAR	National Center for Atmospheric Research
NESDIS	National Environmental Satellite Data and Information Service
NMAT	Nighttime marine air temperature
NMC	National Meteorological Center
NOAA	National Oceanographic and Atmospheric Administration
NRC	National Research Council
OWS	Ocean weather ship
RSL	Relative sea level
SAT	Surface air temperature
SCR	Selective chopper radiometer
SL	Sea level
SMMR	Scanning multifrequency microwave radiometer
SOA	State-of-the-Art report
SOF	Statement of findings
SST	Sea surface temperature

UKMO	United Kingdom Meteorological Office
UNESCO	United Nations Educational, Scientific, and Cultural Organization
VAS	Visible infrared spin scan radiometer atmospheric sounder
VEI	Volcanic explosivity index
VHRR	Very high resolution radiometer
WMO	World Meteorological Organization
WOCE	World Ocean Circulation Experiment
WWR	World Weather Records

CITATION INDEX*

* Authors cited by name in the text are listed in this index. Where a coauthor's name is subsumed under an et al. in the text, no entry for that coauthor appears in this index. Complete references appear at the end of each chapter.

SUBJECT INDEX

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