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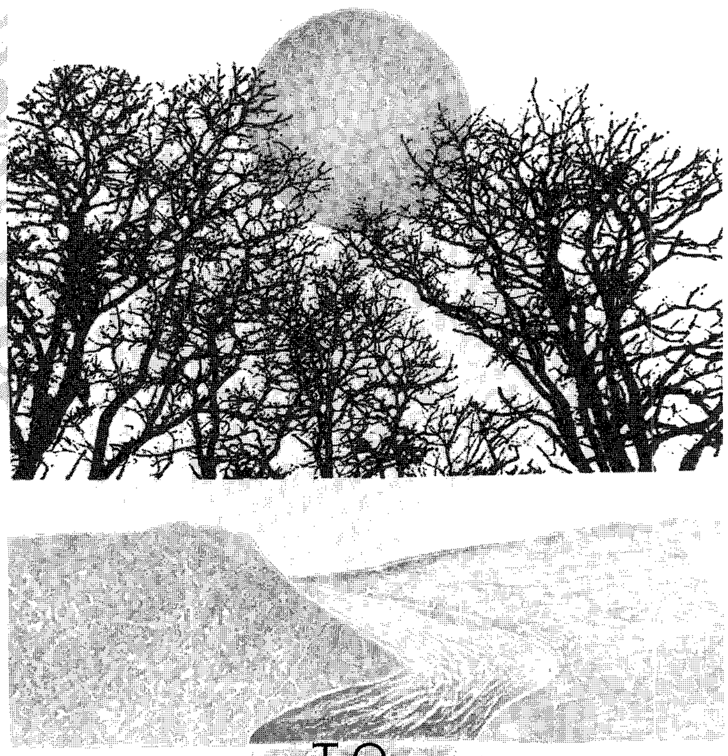
# SENSITIVITY OF WATER RESOURCES IN THE DELAWARE RIVER BASIN



# TO CLIMATE VARIABILITY AND CHANGE

U.S. Geological Survey
Water-Supply Paper 2422

# Sensitivity of Water Resources in the Delaware River Basin to Climate Variability and Change

By MARK A. AYERS, DAVID M. WOLOCK, GREGORY J. McCABE, LAUREN E. HAY, and GARY D. TASKER

U.S. GEOLOGICAL SURVEY WATER-SUPPLY PAPER 2422

U.S. DEPARTMENT OF THE INTERIOR
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# CONTENTS



## FIGURES

## TABLES



## CONVERSION FACTORS AND VERTICAL DATUM

| Multiply | By | To obtain |
|---|---|---|
| millimeters (mm) | 0.03937 | inches |
| centimeters (cm) | 0.3937 | inches |
| meters (m) | 3.281 | feet |
| kilometers (km) | 0.6214 | miles |
| square kilometers ($km^2$) | 0.3861 | square miles |
| gallons (gal) | 0.003785 | cubic meters |
| million gallons per day (Mgal/d) | 0.0438 | cubic meters per second ($m^3$/s) |

*Sea level*: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929—a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

# Sensitivity of Water Resources in the Delaware River Basin to Climate Variability and Change

*By* Mark A. Ayers, David M. Wolock, Gregory J. McCabe, Lauren E. Hay, *and* Gary D. Tasker

## Abstract

Because of the greenhouse effect, projected increases in atmospheric carbon dioxide levels might cause global warming, which in turn could result in changes in precipitation patterns and evapotranspiration and in increases in sea level. This report describes the greenhouse effect; discusses the problems and uncertainties associated with the detection, prediction, and effects of climate change; and presents the results of sensitivity analyses of how climate change might affect water resources in the Delaware River basin.

Sensitivity analyses suggest that potentially serious shortfalls of certain water resources in the basin could result if some scenarios for climate change come true. The results of model simulations of the basin streamflow demonstrate the difficulty in distinguishing the effects that climate change versus natural climate variability have on streamflow and water supply. The future direction of basin changes in most water resources, furthermore, cannot be precisely determined because of uncertainty in current projections of regional temperature and precipitation. This large uncertainty indicates that, for resource planning, information defining the sensitivities of water resources to a range of climate change is most relevant. The sensitivity analyses could be useful in developing contingency plans for evaluating and responding to changes, should they occur.

## INTRODUCTION

Over the past several decades, scientists have gained increasing insight into how the Earth and global processes have been changing through time. These changes result from many interrelated causes and effects, such as changes in solar activity, in the Earth's orbit, in volcanic activity, in landmass distribution, in mountain formation, in ocean circulation patterns, in the climate system, and in the type and distribution of biological species. Natural variations are inherent in these processes, and current (1991) limited knowledge makes it difficult to predict the magnitude, rate, and timing of these changes.

Human population and technological activities have reached a level that can be considered global in their effects. The Earth's climate system and some biological systems are being affected by the expanding industrial and agricultural activities needed to support an increasing world population. The effects include deforestation, desertification, reduction in biodiversity, depletion of stratospheric ozone, increases in greenhouse gases, and changes in sea level.

Many studies indicate that increases in atmospheric concentrations of greenhouse gases, particularly carbon dioxide, could cause global atmospheric warming (Manabe and Stouffer, 1980; Schlesinger and Gates, 1980; Hansen and others, 1981, 1984; Manabe and others, 1981; Wetherald and Manabe, 1981; Smagorinsky, 1982; Washington and Meehl, 1984; Schlesinger and Mitchell, 1985; Dickinson, 1986; Peng and others, 1987; Rind, 1988; Mitchell, 1989). Because the processes involved in global warming and their interactions are complex and difficult to quantify, they present a major scientific challenge to improvement of our understanding during the next few decades. The uncertainty associated with the effects of climate change on water-resource systems is also large. Water-resource planners and managers would benefit from research that increases our understanding of the interactions of water resources with climate and from a thorough evaluation of the sensitivities of water-resource systems to a full range of potential climate changes.

The first part of this report describes the greenhouse effect and reviews the problems and uncertainties associated with the detection, modeling, and prediction of climate change due to increasing greenhouse gases. The report further discusses the hydrologic implications of climate change and describes the Delaware River basin. The report then presents the results of a research project in the Delaware River basin (fig. 1; Ayers and Leavesley, 1988; Moss and Lins, 1988) to illustrate the sensitivity of water resources to potential changes in climate. Examples are given to illustrate how variability and potential changes in temperature, precipitation, and the transpiration rate of vegetation affect the soil moisture, streamflow, drought, and water supply of this river basin. The potential effects of sea-level rise on saltwater intrusion, aquifer recharge, tidal wetlands, and coastal flooding and erosion are also examined.

## THE GREENHOUSE EFFECT AND CLIMATE CHANGE

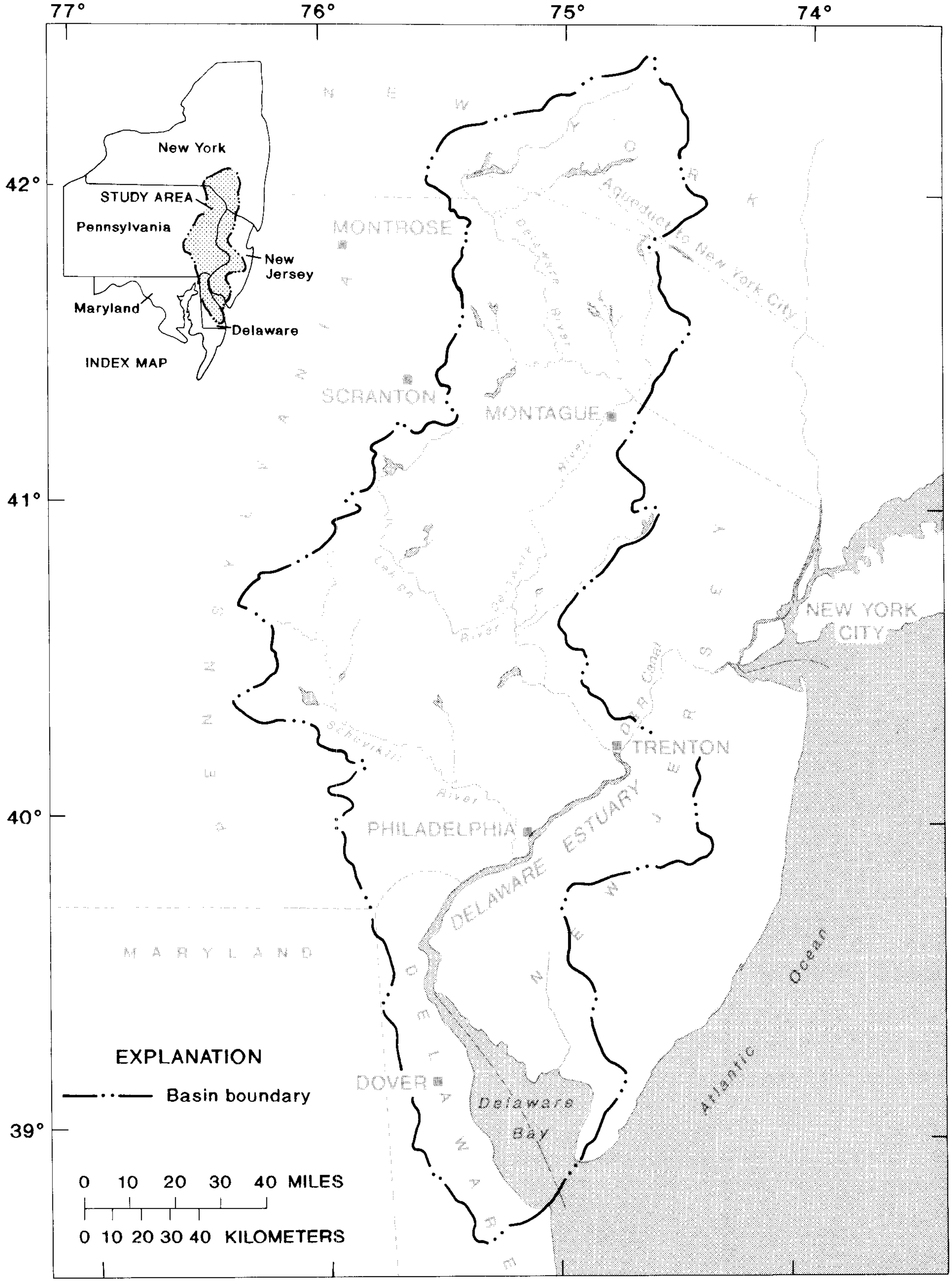
Energy from the sun, in the form of shortwave radiation, readily passes through the Earth's atmosphere. Some of this energy is reradiated from the Earth as longwave (infrared) radiation. Several gases in the atmosphere—principally water vapor ($H_2O$) and carbon dioxide ($CO_2$), but also methane ($CH_4$), nitrous oxide ($N_2O$), and ozone ($O_3$)—absorb or trap the reradiated energy rather than allowing it to pass through the atmosphere back into space (fig. 2). The result is an increase in ambient air temperature, similar to that which occurs when glass in a greenhouse traps reradiated energy; hence the term "greenhouse effect." This process is vital to life systems on the Earth, for without the greenhouse effect of the atmosphere, the average temperature of the Earth would be about 33°C less than its current average of about 15°C (Hansen and others, 1984).

The atmospheric concentration of each greenhouse gas is a result of the interaction of the biological systems on the Earth with hydrologic and other biogeochemical cycles that involve various sinks and sources of the gases. Changes in atmospheric concentrations of greenhouse gases throughout the Earth's history, whether natural or recently human induced, have coincided with changes in the Earth's climate. The most popular paradigm for the correlation between greenhouse-gas concentrations and climate variations is that as concentrations of greenhouse gases increase, the atmosphere becomes an increasingly efficient trap for reradiated energy, and global atmospheric warming could occur (Mitchell, 1989).

### Recent Measurements and Causes of Increasing Greenhouse Gases

Present concern about global atmospheric warming stems from measurements of increasing concentrations of $CO_2$, $CH_4$, and other greenhouse gases released by human activities (MacCracken and Luther, 1985; Bolin and others, 1986; Lins and others, 1988). Researchers have measured atmospheric concentrations of the gases and have gathered proxy data from ice cores and other sources that indicate that $CO_2$ and other gases have increased substantially since the Industrial Revolution of the mid-1800's.

Actual measurements of atmospheric $CO_2$ at Mauna Loa Observatory, Hawaii (Keeling and others, 1982), show an annual cycle related to biological activity and a substantial increase since the monitoring station was installed in 1958 (fig. 3). The release of $CO_2$ from activities such as (1) combustion of fossil fuels (oil, coal, gas) for power generation, heating, transportation, industry, and other growing activities of modern technology and (2) deforestation, biomass burning, and decomposition of terrestrial and soil organic matter have led to an increase in atmospheric $CO_2$ of about 20 percent since the mid-1800's (Neftel and others, 1985). Although deforestation and other losses of terrestrial biomass and soil organic matter have contributed to increases in atmospheric $CO_2$ (fig. 4), the principal source of $CO_2$ since the early 1900's has been the combustion of fossil fuels. Fossil-fuel combustion is likely to increase in the future to satisfy the expanding technological needs of a growing world population (Peng and others, 1983).




Base from Delaware River Basin Commission, 1986

**Figure 1.** Location of Delaware River basin.

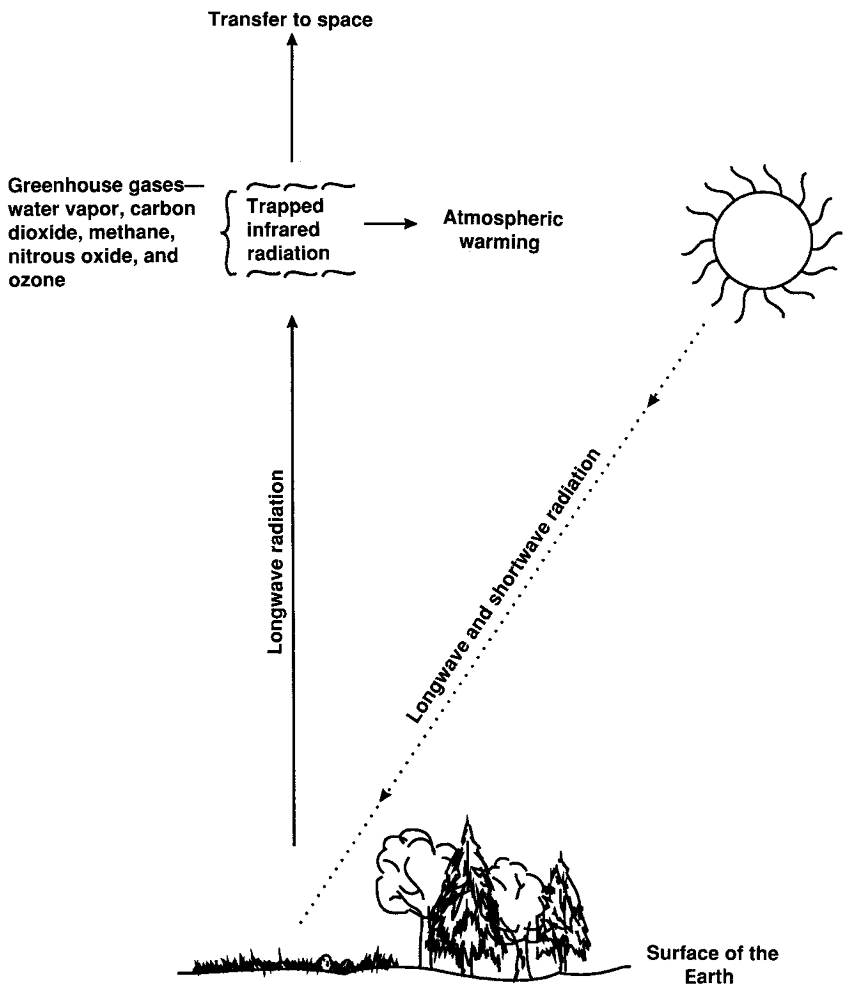



**Figure 2.** Schematic representation of the greenhouse effect.

Atmospheric concentrations of $CH_4$ have increased by 40 to 100 percent in the last century (Lacis and others, 1981; Lins and others, 1988). The increases in $CH_4$ (Pearman and Fraser, 1988) likely result from such activities as (1) releases during fossil-fuel production and storage; (2) deforestation and biomass burning to support increased agriculture; (3) increased anaerobic decomposition of organic matter associated with increased rice production, ruminant animal production, landfills, sewage treatment, and termite populations (in deforested areas); and (4) other agricultural and industrial activities. Rice paddies are a significant and growing source of $CH_4$ (Pearman and Fraser, 1988). So too are the cutting and burning of tropical forests for cattle production, which in turn results in increased termite activity—all of which generate $CH_4$. Large amounts of $CH_4$ are tied up as hydrates of $CH_4$ in latticelike geologic structures of clathrates found in frozen tundra and in ocean sediments on the continental shelves. Clathrates release $CH_4$ when warmed, but the importance of clathrates as a source or sink of $CH_4$ and the effect of global warming on the rate of $CH_4$ release by clathrates are unknown.

Although $CH_4$ is roughly 25 times more efficient in absorbing infrared radiation than $CO_2$ is, $CH_4$ is about 1,000 times less abundant than $CO_2$. Therefore, $CO_2$ remains the single most important greenhouse gas of concern (excluding water vapor). Increases in concentrations of chlorofluorocarbons and other greenhouse gases, such as $N_2O$, in the past century are estimated to have contributed to only 11 percent of the greenhouse effect in comparison with $CO_2$ and $CH_4$ (Mitchell, 1989). The relative effects of chlorofluorocarbons likely will become more pronounced in the future.

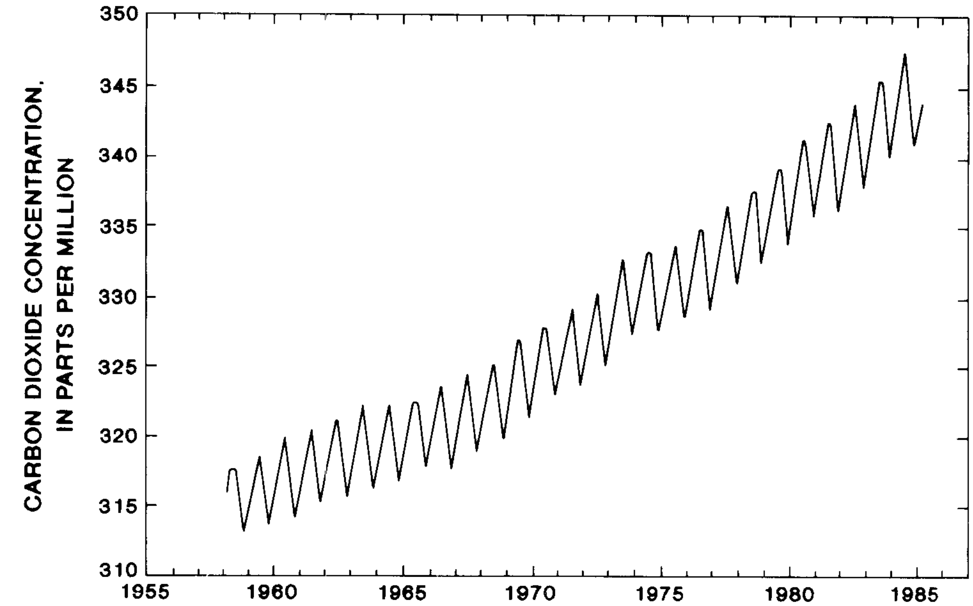



**Figure 3.** Recent atmospheric carbon dioxide measurements at Mauna Loa Observatory, Hawaii (modified from Gammon and others, 1985; original data from C.D. Keeling, Scripps Institute of Oceanography, La Jolla, Calif.).

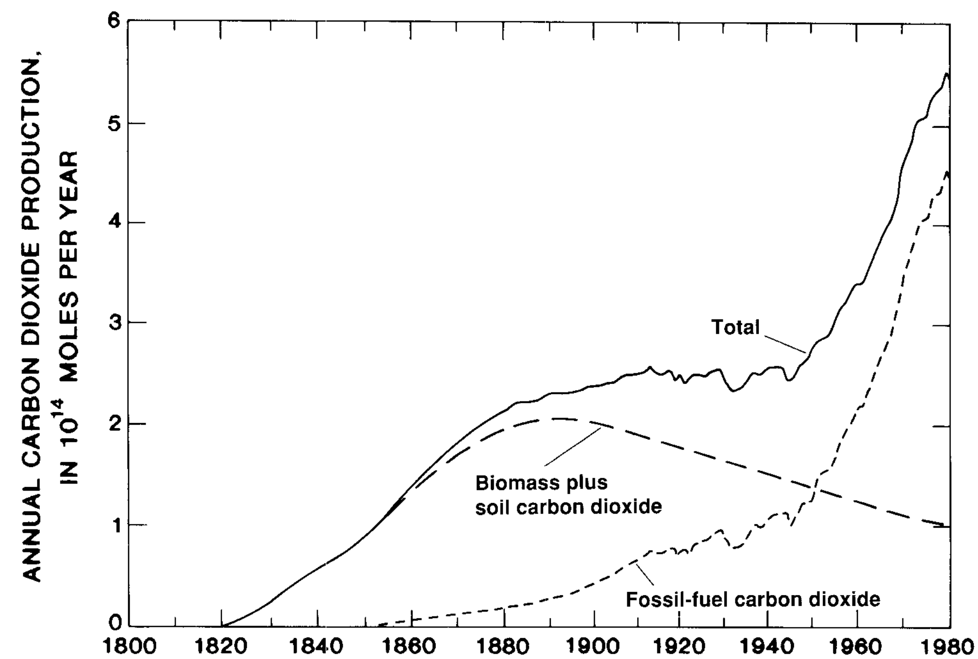



**Figure 4.** Historical global releases of carbon dioxide from terrestrial (biomass and soil) and fossil-fuel sources (modified from Peng and others, 1983).

## Projections of Atmospheric Carbon Dioxide Concentrations

Except for the last century, atmospheric concentrations of $CO_2$ have remained between 200 and 300 parts per million (ppm; fig. 5) for the past million years (Gammon and others, 1985). For reasons discussed above, atmospheric $CO_2$ concentrations have increased to about 350 ppm and are expected to double before the end of the next century (Smagorinsky, 1982); such concentrations probably have not existed on the Earth in more than a million

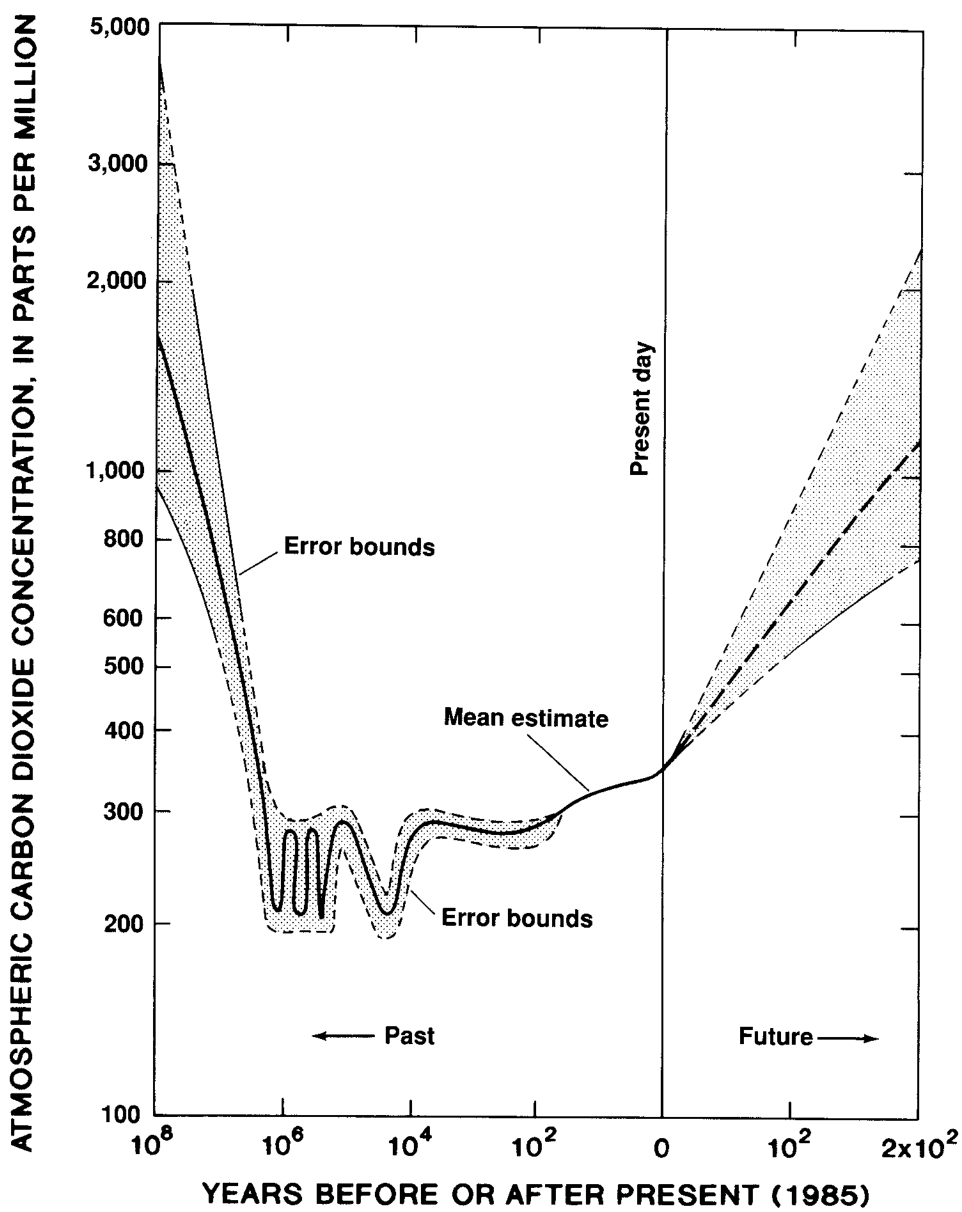



**Figure 5.** Range in estimated historical fluctuations and future projections of atmospheric carbon dioxide (modified from Gammon and others, 1985).

years. Scientists are concerned that these increased $CO_2$ concentrations could induce global warming through the greenhouse effect and create profound climate changes over many regions of the Earth. Researchers have turned to the climate records to look for evidence of recent climate change.

## HISTORICAL EVIDENCE OF CLIMATE CHANGE

Climate is the long-term average of the variations in daily weather for an area. Historical temperature and precipitation records are crucial to quantifying climate variability and to determining how much change might have occurred during the last century. The records are barely sufficient to define present-day variability, and unfortunately, reliable long-term records for trend detection are uncommon (Karl, 1989).

### Detecting Change

Detection of $CO_2$-induced warming from climate records has been a high-priority issue for research. A primary focus has been on selection of accurate and detailed records to look for trends.

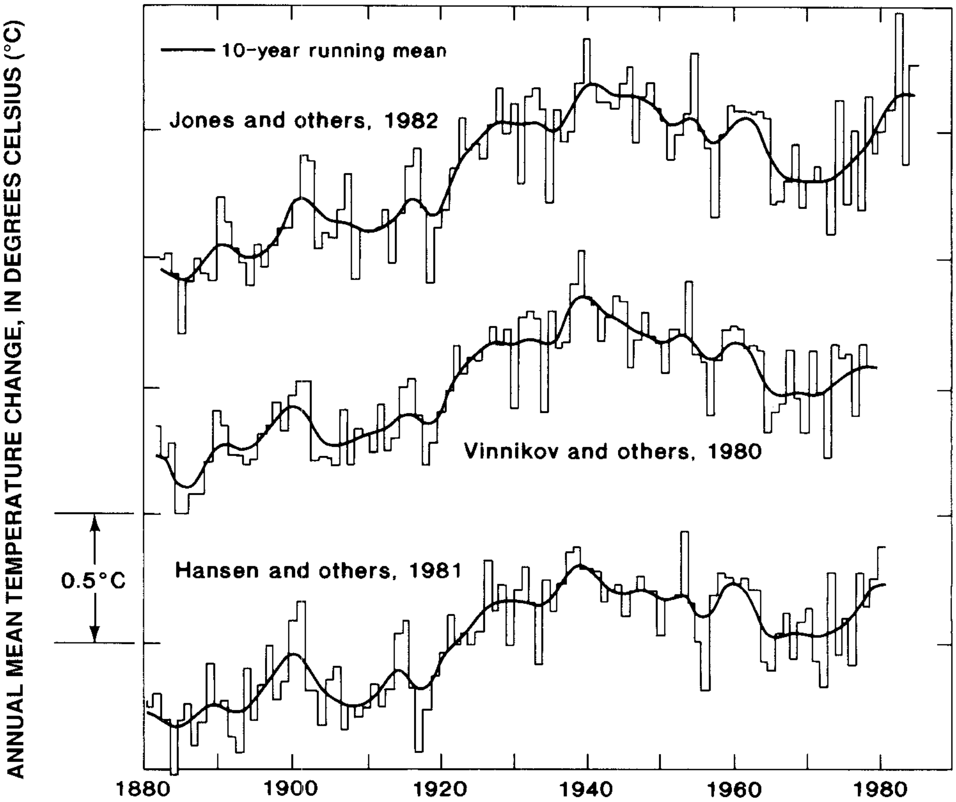



**Figure 6.** Three estimates of Northern Hemisphere temperature changes from climate records (modified from Wigley and others, 1985).

Three compilations of data (Vinnikov and others, 1980; Hansen and others, 1981; Jones and others, 1982) indicate a warming of about 0.5°C from 1880 through 1980 (fig. 6). More recent work by Hansen and Lebedeff (1988) and by Jones and Wigley (1990) seems to substantiate this increase. Other investigators, however, caution that the historical records might be insufficient to draw definite conclusions because accuracy of the data varies with changes in measurement techniques, measurement sites, observers, and local land use (Karl, 1988, 1989).

One of the location factors that most confounds interpretation of long-term temperature records is the heat-island effect, namely, the increase of average temperature in and near urban areas, largely due to absorption of thermal energy by pavement and buildings. For example, in a comparison of temperature records from 31 urban and 31 rural areas in California, Goodridge (1985) found that the heat-island effect was present in many records collected in and near populated areas. Communities of only a few thousand people can induce the heat-island effect in temperature records (Balling and Idso, 1989). Balling and Idso (1989) contend that approximately 74 percent of the 0.5°C warming estimated for the last 100 years is due to the heat-island effect rather than actual global warming. Jones and Wigley (1990) have expended considerable effort to eliminate this and other types of bias from both the land-based and marine air-temperature records, and they conclude that the 0.5°C warming since the late 1800's is real. Jones and Wigley also point out that many questions remain, especially about the cause of the warming trend, whether the trend will persist, and whether the trend is related to the greenhouse effect. These questions can be answered only by research and by decades of additional data collection (MacCracken and Luther, 1985).

## Projecting Change From Climate Records

To gain a perspective on regional climate changes resulting from global warming, researchers have used records of past climates as indicators of a

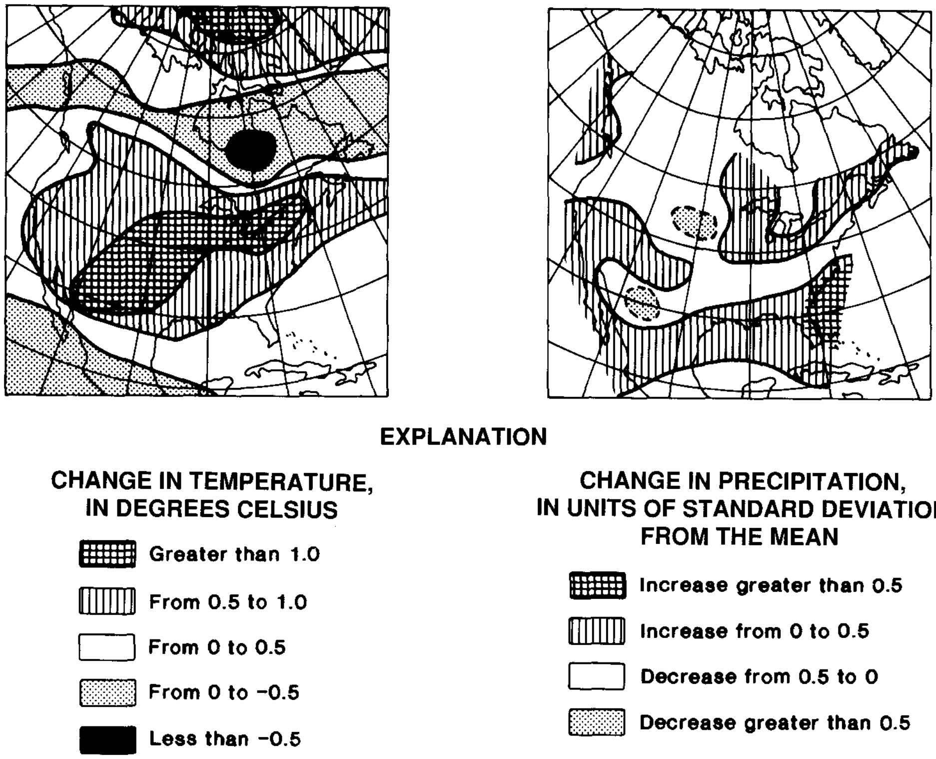



**Figure 7.** Estimated changes in summer temperature and precipitation for a warmer climate, calculated from differences between mean values of 1901–20 and 1934–53 climate records for North America (modified from Webb and Wigley, 1985).

warmer world (Palutikof and others, 1984; MacCracken, 1985; Webb and Wigley, 1985; Wigley and others, 1985; Lins and others, 1990). Palutikof and others (1984) used differences between the mean values of the 1901–20 and the 1934–53 climate records for North America to represent regional changes that might be expected for an additional 0.5°C warming of the Northern Hemisphere (fig. 7). Patterns on the resulting maps indicate that temperatures in most of the United States would be warmer by 0.5°C to 1.0°C. A band of cooling would extend across Canada between latitudes 50°N. and 60°N. Changes in precipitation associated with the warming indicate gains in some regions and losses in other regions (fig. 7).

Climate-record scenarios are one approach to estimating future regional changes in climate variables; however, their use has been criticized as potentially inaccurate because past climate records probably do not reflect the same processes (and weather conditions) that may result from increased $CO_2$ concentrations in the future. Critics argue that only climate-system models that account for the effects of elevated $CO_2$ concentrations and associated feedbacks will provide realistic estimates of climate change.

## SIMULATION OF THE EARTH'S CLIMATE SYSTEM

The Earth's climate results from the interaction, over time, of many physical processes (fig. 8). Among the factors interacting to determine climate are the patterns of atmospheric and oceanic circulation; variability in solar radiation; the atmospheric content of gases, aerosols, and dust; and changes in planetary albedo (reflectivity) resulting from natural variations in the area covered by clouds, vegetation, ice, and snow.

Atmospheric circulation determines the horizontal and vertical fluxes of heat and moisture as well as the exchange of heat and moisture with the land and ocean surfaces. Oceanic circulation strongly affects the temperatures of the sea surface and lower atmosphere and the rates of exchange between the atmosphere and ocean. Solar radiation varies over time and affects the energy received at

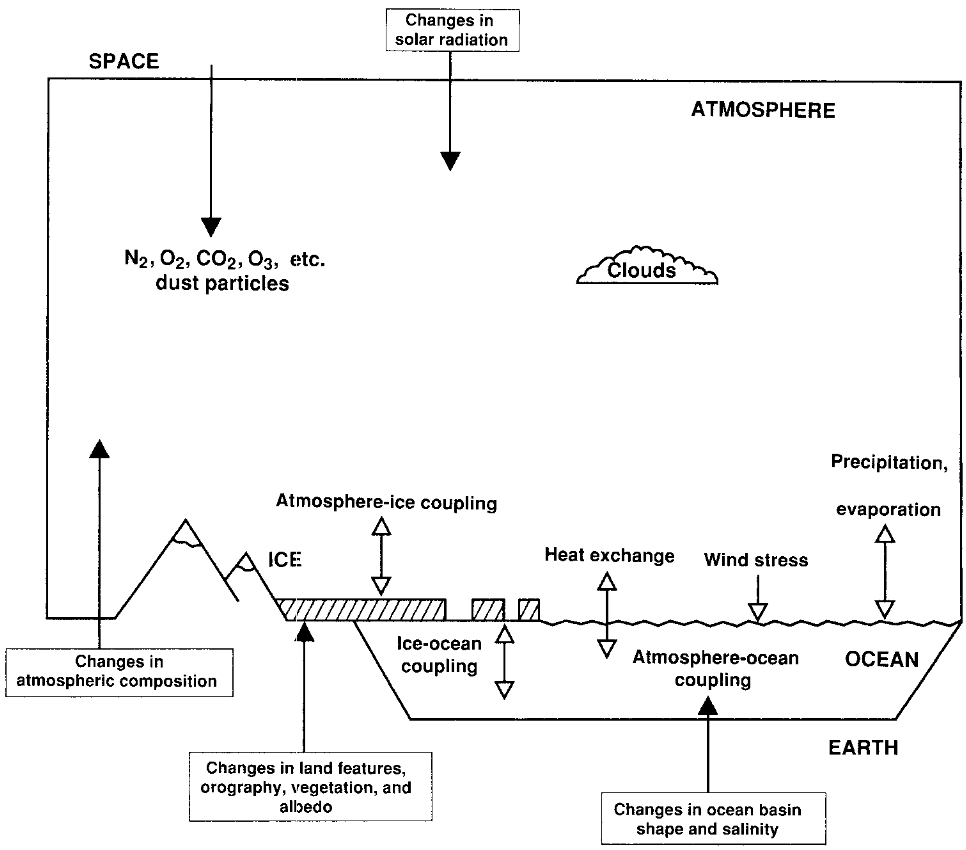



**Figure 8.** Schematic representation of the Earth's climate system and interactions among its principal components (modified from National Research Council, 1975).

the upper limits of the atmosphere. Atmospheric concentrations of gases, aerosols, and dust, in turn, affect the amounts and types of radiation transmitted or absorbed by the atmosphere. Changes in natural processes (volcanic and biological) and in human activities affect atmospheric concentrations of gases, aerosols, and dust. Changes in land-surface and cloud-cover characteristics affect the distribution and exchange of energy and moisture, and the albedo of the Earth's surface. These and other factors affect atmospheric and oceanic circulation; hence, climate processes are replete with combinations of interactions, feedbacks, and changes. Fluctuations in these factors and processes lead to climate variability (Jones and Wigley, 1990).

Most year-to-year variations in climate stem from processes affecting atmospheric circulation. Variations with a period of 2 to 8 years stem from changes in vertical circulation of the oceans and, hence, in sea-surface temperatures; the El Niño/Southern Oscillation phenomenon (Enfield, 1989) is one example. Variations on the order of decades result from the large thermal inertia of the oceans interacting with the more rapid fluctuations in processes, such as the solar cycles, that affect atmospheric and oceanic circulation. Solar radiation varies by about 0.1 percent over the 11-year sunspot cycle, and it may vary by larger amounts (0.6 percent) over longer periods. Changes in the Earth's tilt and orbit around the sun (Cooperative Holocene Mapping Project Members, 1988) also affect the amount of radiation received by each hemisphere on a time scale of centuries. Short-term solar variations, however, are not likely to be a significant determinant of climate change when compared with the effects of a doubling of atmospheric $CO_2$ (Jones and Wigley, 1990).

During the last few decades, the complexity of atmospheric, land, and ocean interactions and the potential for climate change caused by human activities have prompted the development of complex numerical models that simulate the major components of the climate system. Experimentation with climate models continues to provide new insight

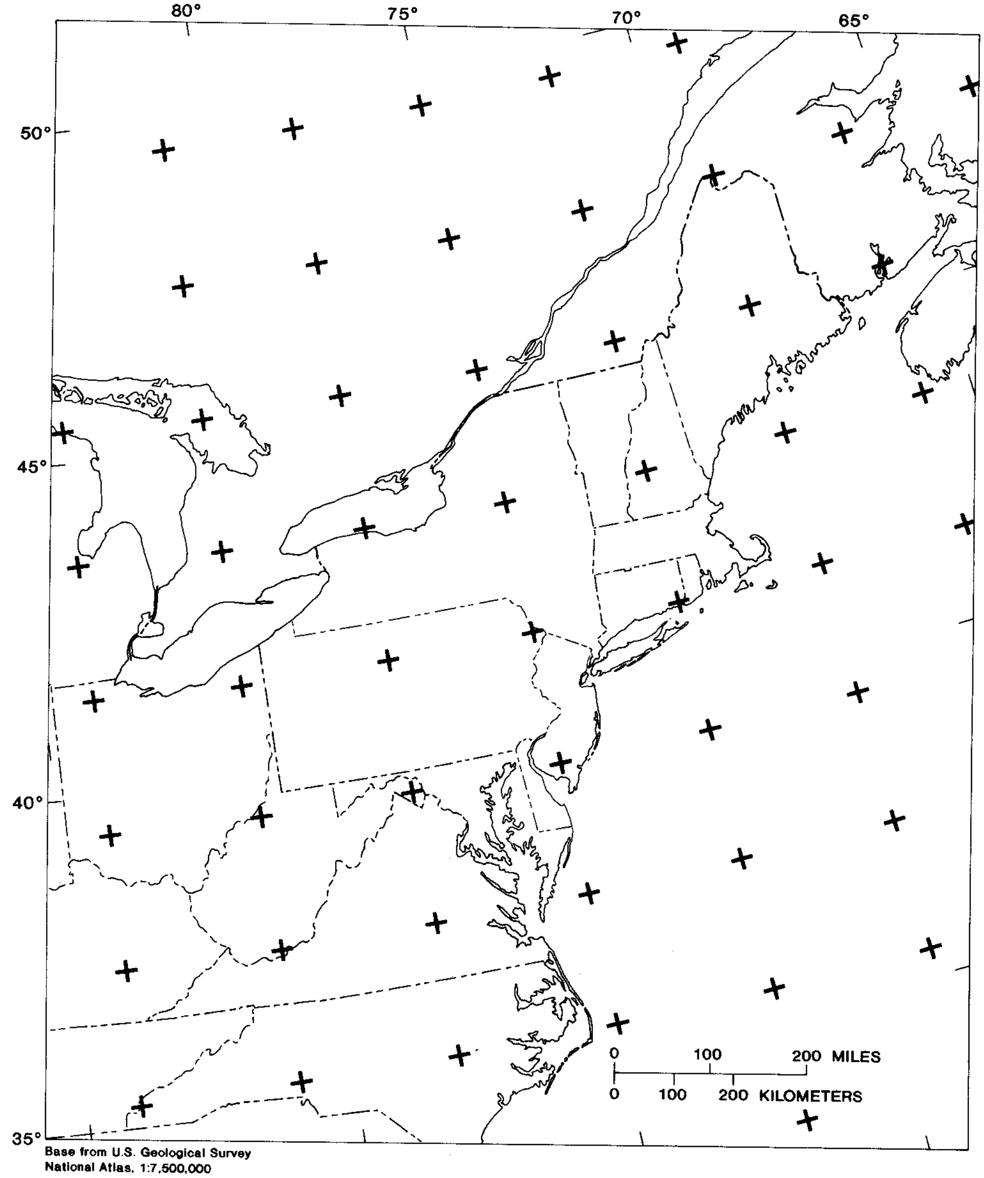



**Figure 9.** Example of computational grid for a general circulation model with approximate spacing of 2° latitude by 3° longitude.

into the importance of various components and the many factors and interactions, such as increase in concentrations of greenhouse gases, that can influence climate.

## General Circulation Models

Numerical simulation models of the global climate system, referred to as general circulation models (GCM's; Dickinson, 1986; Lamb, 1987; Mitchell, 1989), are the most advanced approach used to evaluate the effects of increasing atmospheric $CO_2$ and other greenhouse gases on climate. GCM's are computationally intensive computer programs, yet they are only coarse representations of energy and water cycles at and above the Earth's surface. Spacing of the computational grids varies from as large as about 8° latitude by 10° longitude to as small as about 2° latitude by 3° longitude (fig. 9). Even the smallest grids are greatly simplified representations of surface processes and topography.

The major processes and components used in the GCM's are shown in figure 10. Those processes that are most difficult to quantify (and therefore the

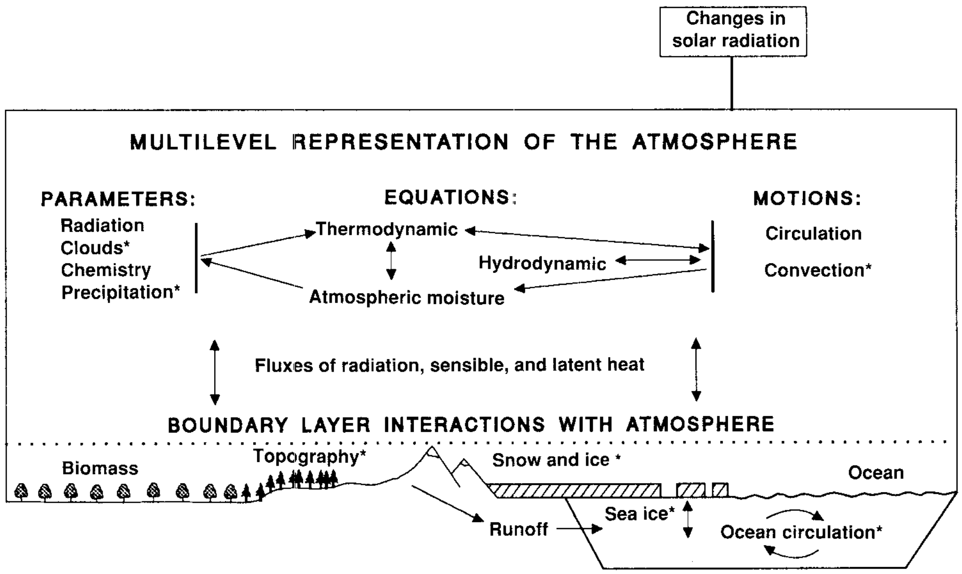



**Figure 10.** Schematic representation of climate-system elements simulated in general circulation models. The most problematic processes are marked with asterisks.

most problematic) are marked with asterisks. At present, cloud processes are probably the most difficult, and yet the most important, atmospheric components to represent in the GCM's (Cess and others, 1989; Fouquart and others, 1990). Ocean processes (including sea-ice interactions) probably are next in complexity and importance, followed by land-surface interactions (Avissar and Verstraete, 1990) involving vegetation and snowfields.

The representations of climate feedbacks of energy and moisture from land, vegetation, ice, and ocean surfaces to the atmosphere (fig. 11) are critical to the overall performance of GCM's (MacCracken, 1985; Verstraete and Dickinson, 1986; Peng and others, 1987). A comparison of 14 GCM's (Cess and others, 1989) shows that a major source of uncertainty is the treatment of cloud-radiation interactions. Warming induced by $CO_2$ could create feedback effects that accelerate the warming, such as increased atmospheric water vapor, increased high-altitude cloud cover, or decreased reflection of solar energy because of retreating ice and snowfields (Mitchell, 1989). Other factors, such as increased snow cover or increased low-altitude cloud cover, could retard the warming (Hansen and others, 1984; Monastersky, 1989). Knowledge of these complex interactions and, hence, model representations are inadequate as yet to determine with much confidence what effects the various feedbacks will produce.

## Projecting Climate Change With General Circulation Models

Early GCM results helped to form the consensus that increased atmospheric $CO_2$ concentrations might result in global atmospheric warming (Smagorinsky, 1982). A doubling of current atmospheric $CO_2$ concentrations in the GCM's is projected to result in a global average temperature increase between 1.5°C and 4.5°C. GCM's are greatly simplified approximations of the climate system, however, and many uncertainties are inherent in their use. Compared with actual climate conditions, GCM's are able to reproduce large-scale circulation patterns (Lamb, 1987) and temperature distributions (Rind and others, 1990) better than other modeled variables. Confidence in modeled precipitation, runoff, and soil moisture, especially at regional scales, is low.

The uncertainty of GCM projections is reflected, in part, by the relatively large differences among GCM estimates of climate change. For example, projected estimates in summer temperature and precipitation for doubled-$CO_2$ conditions are illustrated in figure 12 for three GCM's: the Geophysical Fluid Dynamics Laboratory (GFDL) model, the Goddard Institute for Space Studies (GISS) model, and the National Center for Atmospheric Research (NCAR) model. The estimates of temperature change all show increases across North Amer-

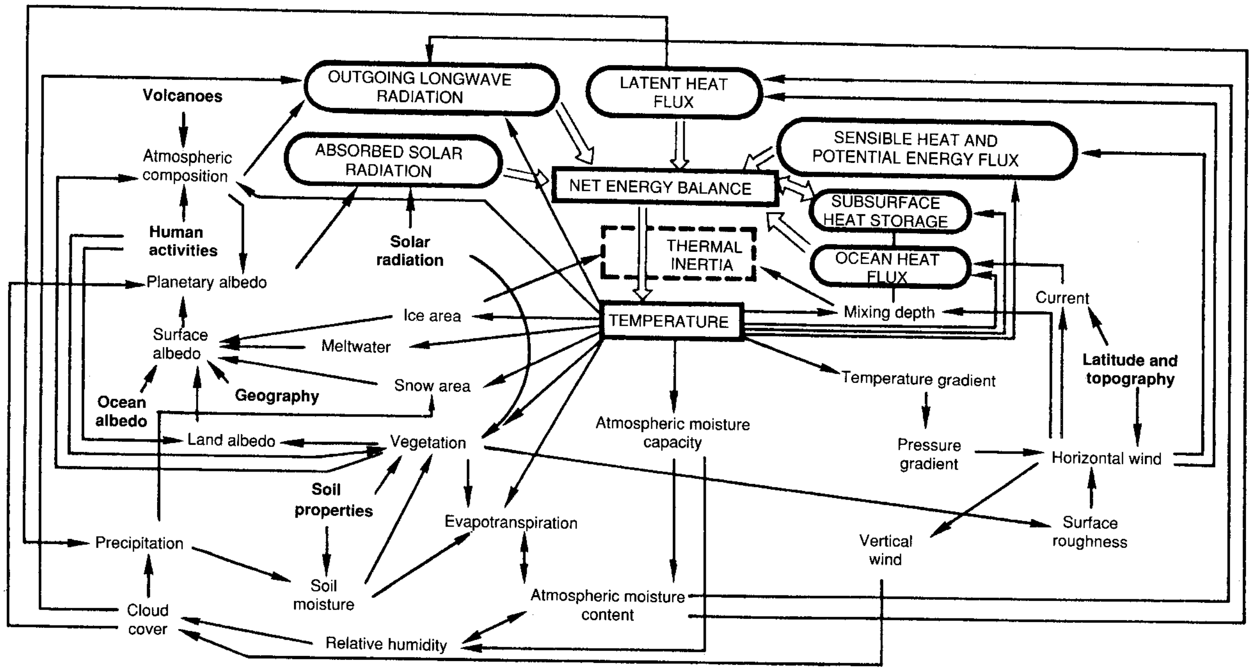



**Figure 11.** Schematic representation of climatic cause-and-effect (feedback) linkages and variables often included in numerical models of the climate system. Many of these linkages are oversimplified and otherwise treated inadequately (Robock, 1985).

ica, but the values range from about +2.0°C to +6.0°C. The estimates of precipitation change are less consistent and range from a decrease of 1.0 mm/d to an increase of 1.0 mm/d. This difference in estimates of precipitation change creates large uncertainty in quantifying the effects of the $CO_2$-induced warming on water resources.

With improvements in GCM's, research presumably will refine estimates of regional-scale changes in temperature and precipitation. Until then, use of regional temperature and precipitation projections from current GCM's to obtain local estimates of future water availability could be risky (Rind, 1988).

## HYDROLOGIC IMPLICATIONS OF CLIMATE CHANGE

Atmospheric warming induced by $CO_2$ could alter climate conditions and affect hydrologic processes and cycles in many regions of the Earth (National Academy of Sciences, 1977). If changes in temperature and precipitation are large, they could result in changes in the availability of freshwater in many regions. Regions where the water resources are used heavily could be harmed by even small changes in climate. The water-resource research community is seeking to answer several technical questions concerning the potential effects of climate change on water resources:

1. Which regions will be most affected by climate changes?
2. What will be the direction and magnitude of regional changes in temperature, precipitation, and other factors?
3. When will the changes occur?
4. What will be the seasonal distribution of the changes?
5. What data and approaches will be needed to distinguish the effects of climate change on regional water resources from the effects of natural climate variability?

Unfortunately, current GCM projections are too uncertain to answer the first four questions with much confidence. Related to the fifth question, however, is the current need for research on the interactions of watershed and water-resource systems and their sensitivity to a range of potential climate changes. The GCM outputs are valuable in setting ranges of potential changes to test hydrologic sensitivities (Gleick, 1989).

The effects of increasing atmospheric $CO_2$ on changes in watershed systems will be complex (Gle-

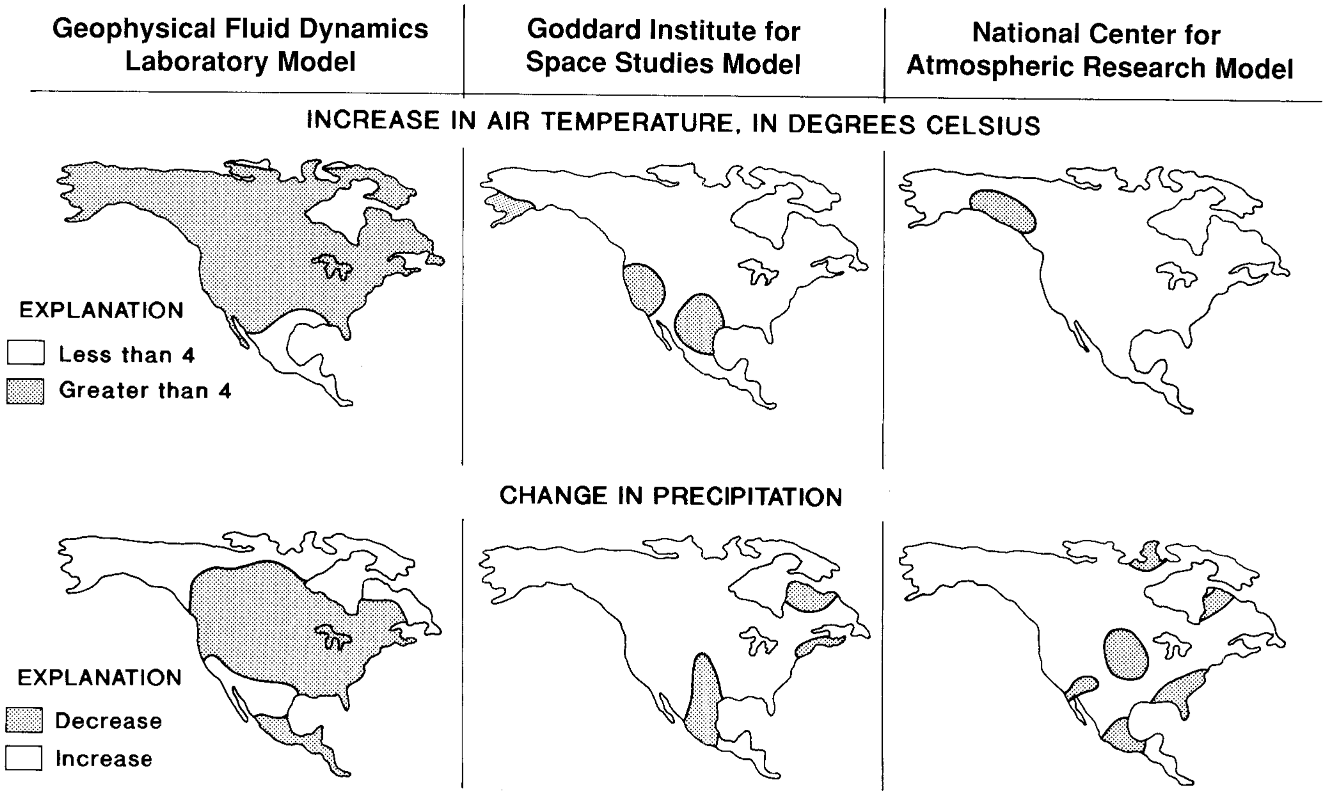



**Figure 12.** Estimated summer temperature and precipitation changes over North America for doubled-carbon-dioxide conditions in three general circulation models (modified from Schlesinger and Mitchell, 1985).

ick, 1989; Lins and others, 1990; Waggoner, 1990). Increasing atmospheric $CO_2$ concentrations could directly affect transpiration processes of plants and contribute to several competing changes (Strain and Cure, 1985). Increased efficiency of water use by plants (Rogers and others, 1983; Eamus and Jarvis, 1989) could result from the possible increase in stomatal resistance to transpiration directly due to higher concentrations of $CO_2$ (that is, plant stomata do not open as wide under higher ambient concentrations of $CO_2$) and to a decrease in stomatal density (that is, to fewer stomata per unit leaf area) (Kimball and Idso, 1983; Eamus and Jarvis, 1989; Martin and others, 1989). Other factors, however, could negate these potential gains in plant water-use efficiency: a potential increase in leaf temperatures due to reduced transpiration rates (Idso and others, 1985), a potential increase in leaf area due to $CO_2$-fertilization effects (Patterson and Flint, 1980; LaMarche and others, 1984), and perhaps species changes in vegetation communities (Emanuel and others, 1985). The net effect of increasing $CO_2$ on evapotranspiration—and hence on soil moisture, streamflow, and ground-water recharge—is difficult to project from the current understanding of these interacting processes.

Some researchers have attempted to evaluate the sensitivity of streamflow to the direct effect of increased $CO_2$ on increasing stomatal resistance to transpiration by variably adjusting potential evapotranspiration in their models (Aston, 1984; Idso and Brazel, 1984; Wigley and Jones, 1985; Gleick, 1989; Wolock and Hornberger, 1991). They generally concluded that the direct effect of increasing $CO_2$ on plants could counteract the reductions in streamflow due to $CO_2$-induced warming. These researchers, however, admittedly assume with some unknown confidence that the net effect of increased $CO_2$ concentrations is reduced transpiration. As discussed above, there is no conclusive evidence on what the net effects of increased $CO_2$ will be on transpiration.

The indirect effects of increasing $CO_2$ concentrations on watershed systems will result from global warming and other changes in regional climate (Gleick, 1989). Changes in watershed systems will stem largely from changes in temperature and precipitation, although changes in other weather conditions, such as cloud cover, wind, and humidity, will have some effect (Gleick, 1989; Rosenberg and others, 1989). Changes in temperature could affect the rate of evapotranspiration in warm seasons and the amount of snow accumulation and the timing of snowmelt in cold seasons and, in turn, would affect soil moisture, streamflow, and ground-water recharge (Gleick, 1987; Lettenmaier and others, 1988; McCabe and Ayers, 1989; Vacarro, 1991). Changes in precipitation could affect soil moisture, streamflow, and ground-water recharge (Wigley and Jones, 1985) and, in turn, the frequency and duration of floods and droughts (Gleick, 1989). For example, increased amounts of precipitation could offset increased evapotranspiration demands of a warming trend, whereas decreased or unchanged amounts of precipitation during a warming trend most likely would result in decreased streamflow and ground-water recharge.

## THE THORNTHWAITE MOISTURE INDEX

The Thornthwaite moisture index (Thornthwaite and Mather, 1955) is an indicator of the supply of water (precipitation) in an area relative to the climatic demand for water (potential evapotranspiration). It is one of the simplest water-resource indicators that can be derived from available climate data (temperature and precipitation). Use of the moisture index minimizes the assumptions inherent in computer simulations of the potential effects of climate change on complex hydrologic indicators such as streamflow. The index does not require any watershed data.

The Thornthwaite moisture index (Im) used in this study is given by

$$Im = 100[(P/PE)-1],$$

where P is annual precipitation and PE is annual potential evapotranspiration (Mather, 1978) for a climate station. Positive values indicate a humid climate with a water surplus; negative values, an arid climate with a water deficit. A moisture index of zero indicates that annual precipitation is just enough to satisfy the demand for water under prevailing climate conditions.

**Table 1.** Comparisons of current (1950–83) mean annual Thornthwaite moisture indices at three study sites with indices derived from projections under doubled atmospheric carbon dioxide from three general circulation models

[GISS, Goddard Institute for Space Studies; GFDL, Geophysical Fluid Dynamics Laboratory; OSU, Oregon State University]

| Site | Current carbon dioxide | Doubled carbon dioxide | | |
|---|---|---|---|---|
| | | GISS | GFDL | OSU |
| Salisbury, Md. | 39 | 13 | 12 | 40 |
| Philadelphia, Pa. | 40 | 11 | 10 | 38 |
| Scranton, Pa. | 45 | 10 | 8 | 38 |

### Potential Changes in the Thornthwaite Moisture Index Due to Climate Change

McCabe and Wolock (1991b) examined how potential changes in temperature and precipitation might affect the annual Thornthwaite moisture index in the Delaware River basin. They gradually adjusted mean annual temperature and precipitation at three climate stations in the basin with changes derived from three GCM's: the GISS, GFDL, and Oregon State University (OSU) models.

Mean annual moisture indices for current conditions were compared with indices derived from the steady-state GCM projections (table 1). The GISS and GFDL models indicate decreases in the mean annual moisture index, which are attributable principally to the projected increases in temperature and the absence of offsetting increases in precipitation. The OSU model did not produce significant changes in the mean annual moisture index, because of a projected increase in precipitation that offsets the temperature increase.

Some research suggests a relation between the moisture index and the vegetation typical of an area (Mather, 1978; Mather and Feddema, 1986). Judging from the mean annual moisture index values derived from doubled-$CO_2$ conditions for the GISS and GFDL models, hypothetical moisture conditions in the basin could be more typical of a tallgrass prairie than of the current hardwood forest vegetation. Even if the moisture conditions were to change, however, the changes in vegetation likely would be slow because the response time of forests to climate change is on the order of centuries (Shugart and others, 1986).

Differences among GCM's also are evident in simulated results of gradually induced changes in temperature and precipitation (McCabe and Wolock, 1991b). The GCM model projections were used to induce a gradual change in a stochastically generated time series of the climate-station records for an assumed doubling of $CO_2$ in 100 years. The stochas-

**Table 2.** Number of years until likelihood of detecting significant trends (significance level equals 0.05) in annual Thornthwaite moisture index is 50 and 100 percent

[GISS, Goddard Institute for Space Studies; GFDL, Geophysical Fluid Dynamics Laboratory; OSU, Oregon State University; >, greater than]

| Model | Years until likelihood equals 50 percent (or 100 percent) | | | | | |
|---|---|---|---|---|---|---|
| | Salisbury, Pa. | | Philadelphia, Pa. | | Scranton, Pa. | |
| GISS | 75 | (125) | 55 | (105) | 55 | (95) |
| GFDL | 65 | (125) | 55 | (105) | 55 | (95) |
| OSU | >200 | (>200) | 185 | (>200) | 145 | (>200) |

tic model used the mean and standard deviation of annual temperature and precipitation for the three climate stations. The number of years required to reach 50- and 100-percent probability levels of detecting a statistically significant trend in the moisture index was calculated (table 2).

With the GISS or GFDL model projections, a 50-percent probability of detecting a trend in the moisture index is possible in 55 to 75 years. With the OSU model projection, the 50-percent probability is not attained for at least 145 years because of a smaller temperature increase than projected by the other two GCM's, coupled with an increase in precipitation projected by the OSU model. The site-to-site differences in trend detectability are caused principally by differences in the annual precipitation variability from site to site.

These study results indicate that temperature and precipitation under doubled-$CO_2$ conditions will result in decreased Thornthwaite moisture indices, implying drier conditions in the basin. The amount of decrease depends on the GCM used.

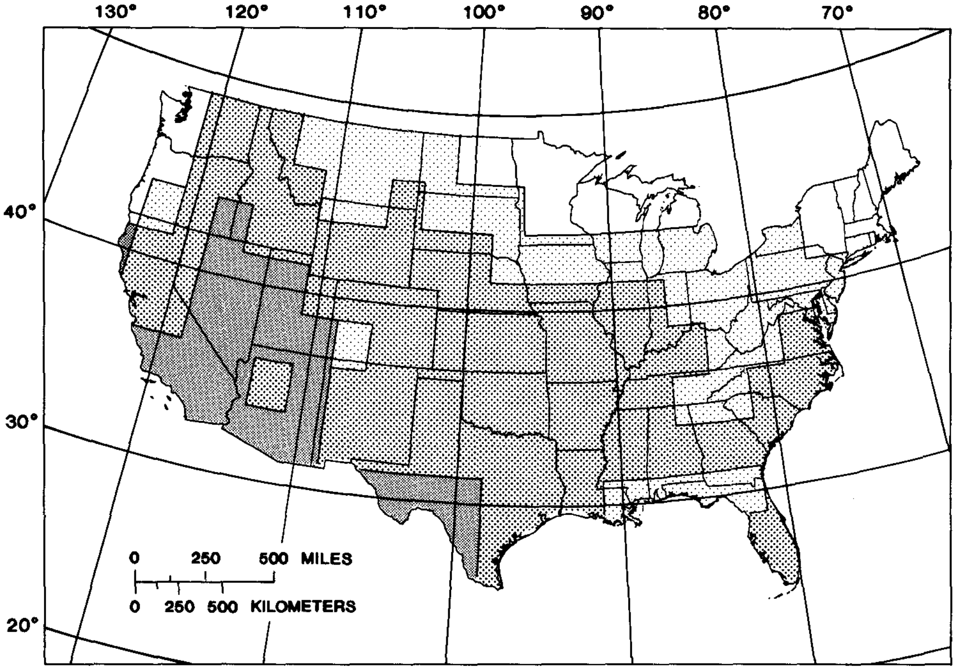
A second analysis was performed to examine the effects of climate changes on the mean annual Thornthwaite moisture index for the conterminous United States (McCabe and others, 1990). To represent current climate conditions, the investigators developed a 2.5°×2.5° grid of mean annual temperature and mean annual precipitation for the United States from maps of the 1951–80 normals of the National Weather Service climate divisions. They then applied the changes in the gridded mean annual temperature and precipitation projected from three GCM's—the GFDL, GISS, and OSU models—for doubled-$CO_2$ conditions and compared the changes in computed mean annual values of the Thornthwaite moisture index across the country.

The study results for doubled-$CO_2$ conditions indicate that the mean annual moisture index generally will decrease, implying drier climate conditions on average for most of the United States. The pattern of expected decrease is consistent among the three GCM's, although the amount of decrease varies with the GCM used. Only the changes projected by the GFDL model are illustrated (fig. 13). The greatest decrease in the moisture index was simulated for the Pacific Northwest, Great Lakes, and New England States—areas that currently are moist and have high ratios of precipitation to potential evapotranspiration (P/PE). The annual moisture index in the Southwestern States would change little, mainly because this region currently has a small P/PE ratio and is associated with annual moisture-index values that already approach the lowest possible value. In addition, simulations indicate that the boundary between negative and positive moisture-index values (not shown) shifted eastward from about long 100°W. to about long 95°W.

The study results indicate that changes in the moisture index are related mainly to changes in mean annual potential evapotranspiration caused by changes in mean annual temperature, rather than to changes in mean annual precipitation. Similar findings were reported by Rind and others (1990) using two drought indices with GISS model projections.

## Detecting Changes in the Thornthwaite Moisture Index

In another study (McCabe and Wolock, 1992), the Thornthwaite moisture index was used to examine the effects of a hypothetical 4°C increase in mean annual temperature on moisture conditions in the conterminous United States. The goals of this study were (1) to illustrate how the natural year-to-year variability in temperature and precipitation can confound the detection of climate-change effects on a hydrologic index and (2) to identify characteristics of areas where significant changes in the moisture index are likely to be detected first. The effects of a




Base from U.S. Geological Survey digital data, 1:2,000,000, 1972
Albers Equal-Area Conic projection
Standard parallels 29°30' and 45°30', central meridian-96°00'

**EXPLANATION**

**CHANGE IN ANNUAL THORNTHWAITE MOISTURE INDEX—**
**>, greater than; <, less than; ≤, less than or equal to**

- $>-10$ to $\leq 10$
- $>-30$ to $\leq -10$
- $>-50$ to $\leq -30$
- $\leq -50$

**Figure 13.** Change in annual Thornthwaite moisture index from current to doubled-carbon-dioxide conditions, as predicted by use of Geophysical Fluid Dynamics Laboratory general circulation model for conterminous United States (McCabe and others, 1990).

gradual increase in air temperature, with no accompanying change in precipitation, on the Thornthwaite moisture index were simulated. A simulated increase in temperature at a rate of 4°C per 100 years was used to induce a gradual change in a stochastically generated time series of the climate-division records. The stochastic model incorporated the mean and standard deviation of annual temperature and precipitation for each climate division.

As in the GCM projections, the 4°C temperature increase resulted in simulated increases in potential evapotranspiration and decreases in the moisture index across the United States. Similarly, decreases in the moisture index were greatest in cool and wet regions (Pacific Northwest, Great Lakes, and New England States) and least in hot and dry regions (Southwestern States).

The time required to detect significant trends in the moisture index was a function of both the magnitude of change in the moisture index (fig. 14) and the natural year-to-year variability of the moisture index. In general, the time required to detect significant trends was short when the ratio of the magnitude of change in the moisture index to the

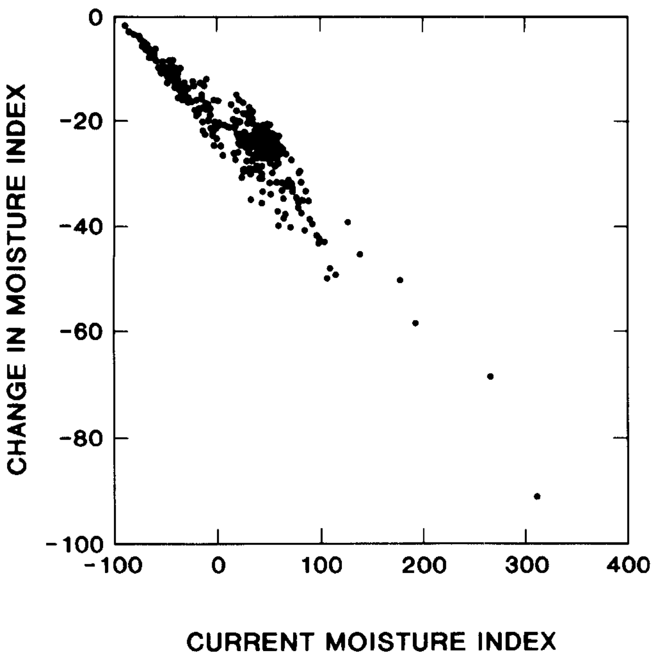



**Figure 14.** Relation between current Thornthwaite moisture index for each National Weather Service climate division and change in moisture index caused by a 4-degree-Celsius increase in mean annual temperature.

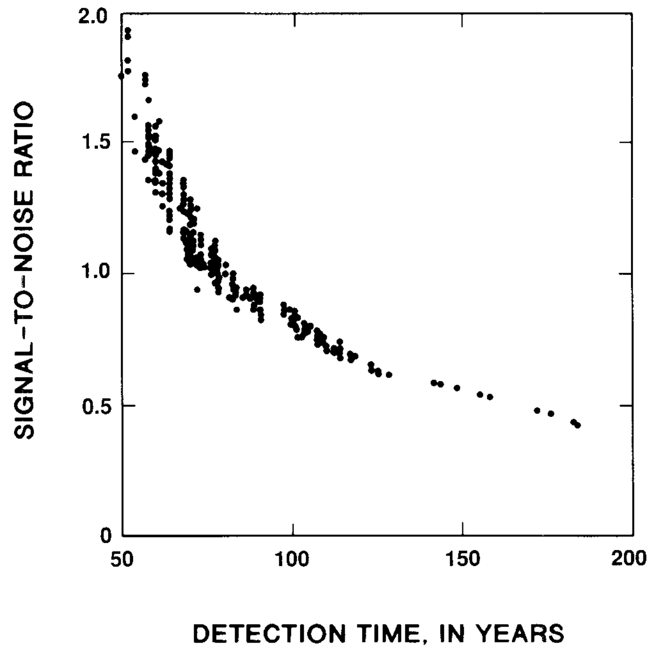



**Figure 15.** Relation between detection time and signal-to-noise ratio for all National Weather Service climate divisions. Detection time is the time required for a significant (at significance level 0.05) trend in 50 percent of the simulations. Signal-to-noise ratio is the difference between the current moisture index and the moisture index that is 4-degrees Celsius warmer, divided by the standard deviation of the current index.

magnitude of variability (signal-to-noise ratio) was large (fig. 15). Results of the research show that natural variability strongly influences the detectability of the effects of the temperature increase on the Thornthwaite moisture index.

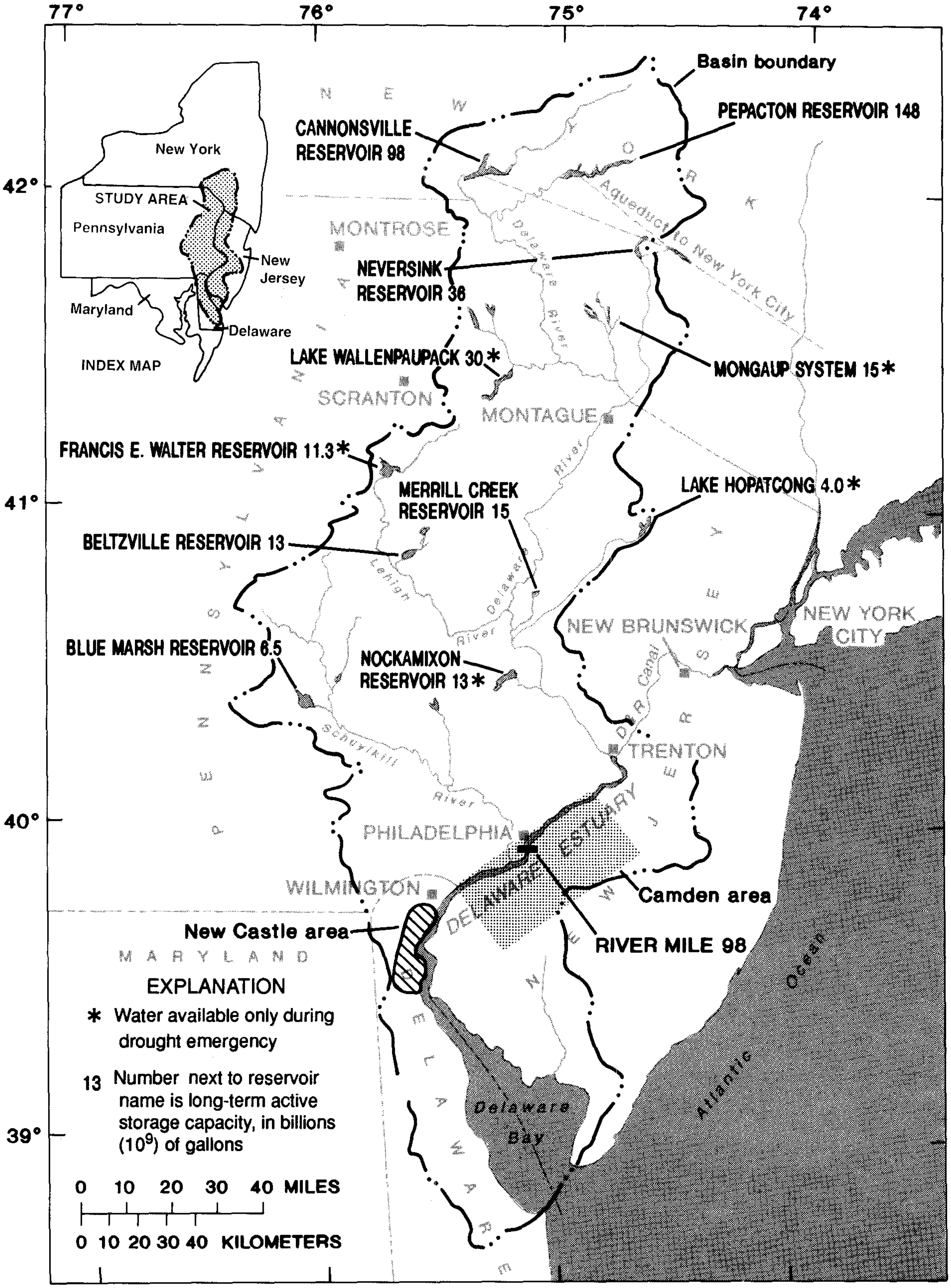
## DELAWARE RIVER BASIN CHARACTERISTICS

The Delaware River (fig. 16) flows freely for more than 300 kilometers (km) from southern New York State to Trenton, N.J. Downstream from Trenton, the river is a tidal estuary for 190 km before entering the Atlantic Ocean at the mouth of Delaware Bay. The watershed area above Trenton is 17,560 $km^2$ and above the mouth is 33,061 $km^2$.

The basin is in a humid-temperate climate with a mean annual temperature of approximately 12°C and mean annual precipitation of about 1,200 mm (Jenner and Lins, 1991). Mean annual temperature varies mostly with latitude from about 13°C in the southern part of the basin to about 7°C in the northern part. The number of days with precipitation generally increases from the southern to the northern part of the basin, whereas the average intensity of precipitation for days when precipitation falls generally decreases from the southern to the northern part. If mean annual precipitation across the basin is compared, the net effect of the intensity-duration relations tends to mask day-to-day and site-to-site variability across the basin. Mean annual precipitation in the basin ranges from 1,000 to 1,300 mm. Much of that variation, however, is due to elevation differences in the basin rather than to latitude or longitude differences.

Soils, vegetation, and topography differ considerably in the basin (Parker and others, 1964). The basin lies in five physiographic provinces (Parker and others, 1964) and three different ecoregions (Omernik, 1986). In the southern part of the basin, the Coastal Plain physiographic province is characterized by nearly flat topography and thick, sandy-loam soils. Coastal Plain streams respond slowly to rainfall; in fact, nearly all streamflow is derived from ground-water discharge. About one-third of the Coastal Plain is forested with hardwoods and softwoods.

In the middle part of the basin, the topography varies from the rolling hills of the Piedmont physio-




Base from Delaware River Basin Commission, 1986

**Figure 16.** Major water-supply features and long-term active storage capacity of major reservoirs in the Delaware River basin.

graphic province to a series of parallel ridges, oriented northeast-southwest, in the New England physiographic province. Both provinces are characterized by relatively thin, clayey-loam soils, and streams respond quickly to rainfall. About one-third of the middle part of the basin is forested, primarily with hardwoods.

In the northern part of the basin, the Appalachian Plateaus and Valley and Ridge physiographic provinces are characterized by mountainous topography. Hillslopes are steep and covered by well-drained soils. Streamflow response to rainfall is between that of the middle and southern parts of the basin. The northern part is the only part of the basin where snow accumulation is substantial in most years and the only part that once was glaciated. The northern part has numerous lakes and is mostly forested in hardwoods.

## Water Supply and Water Use

By D.J. Phelan and M.A. Ayers

The Delaware River basin is a major source of water for more than 15 million people in New York, New Jersey, Pennsylvania, and Delaware. Streams and aquifers in the basin supply water to an estimated 7.3 million people within the basin, and surface-water diversions out of the basin supply about as many more. Complex systems of storage reservoirs and diversion pipes, tunnels, wells, and canals have been installed to improve distribution and availability of this intensively used water supply, especially during extended periods of below-normal precipitation (drought).

In 1986, the estimated basinwide use of water for municipal, domestic, industrial, commercial, agricultural, and power generation purposes amounted to about 7,700 million gallons per day (Mgal/d; J. Featherstone, Delaware River Basin Commission, written commun., Feb. 1988). This value is equivalent to the mean annual streamflow of the Delaware River at Trenton. Most of the water used is returned to basin streams and aquifers, except for 302 Mgal/d (3.9 percent) in consumptive uses within the basin and 692 Mgal/d (9.0 percent) in diversions out of the basin (J. Featherstone, Delaware River Basin Commission, written commun., Feb. 1988). The two major diversions out of the basin are to New York City (645 Mgal/d through the New York City aqueduct) and to northeastern New Jersey (47 Mgal/d through the Delaware and Raritan Canal).

About 53 percent of the annual consumptive water use occurs in June, July, and August, the months of highest evapotranspiration and water use (J. Featherstone, Delaware River Basin Commission, written commun., Feb. 1988). The percentage ranges from 48 in the part of the basin above Trenton to 58 in the Coastal Plain part of the basin. The percentage is higher in the Coastal Plain because a larger proportion of the water is used for agricultural irrigation there.

Most reservoirs in the basin are managed to meet target flows in the Delaware River at Montague and Trenton, N.J. (fig. 16), as part of good-faith agreements among the States and New York City to restrict the upstream movement of the salt front into freshwater reaches where surface- and ground-water supplies now derive their water (Delaware River Basin Commission, 1985). The salt front is defined as the point in the estuary where chloride concentrations are 250 milligrams per liter (mg/L). During the 1961–65 drought (Anderson and others, 1972) and at three other times since 1980, drought warning or emergency water-use restrictions were invoked throughout the basin to minimize the effects of saltwater intrusion.

Future consumptive water use within the Delaware River basin could increase in response to the demands of a growing population. Out-of-basin diversions, however, are not likely to increase in the future above the current allowed maximum of 900 Mgal/d. A rough estimate of the within-basin water use was obtained by assuming that water use will increase linearly with projected population growth in the basin. The consumptive water use for the basin in 2040 is estimated to be about 390 Mgal/d, or 30 percent more than in 1986. Increases in consumptive water use could pose a threat of future water shortages in the basin independent of climate change. A decrease in total basin water supply caused by global warming would increase that threat. This estimate will be discussed in later analyses to place into perspective the sensitivity of water supply to the potential growth in consumptive water use and to potential climatic change.

## Effects of Climate, Topography, and Soils on Hydrologic Characteristics

By D.M. Wolock and C.V. Price

The hydrologic characteristics of drainage basins in humid-temperate regions are primarily

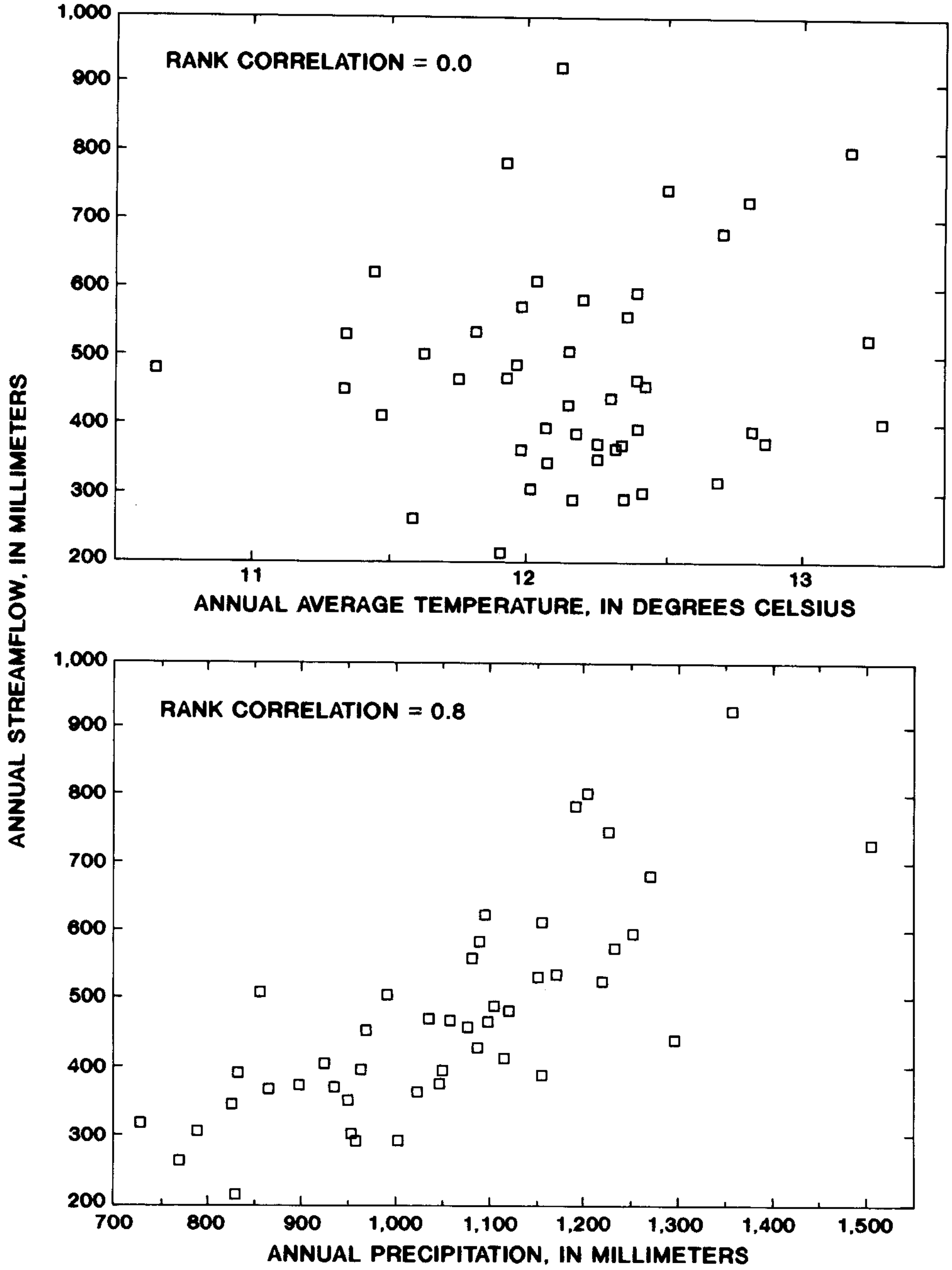



**Figure 17.** Annual flow of a Pennsylvania stream, Neshaminy Creek, in relation to mean annual air temperature and total annual precipitation.

controlled by climate, topography, and soils. Climate determines the input of water to a basin (precipitation) and affects the evaporative loss of water from a basin. Topography determines the effects of gravity on the movement of water in a basin and affects the drainage of water from hillslopes as streamflow or ground water and, in part, the flow path that water follows through a basin. Soil properties determine the water-storage capacity of a basin and, along with topography, affect the quantity and rate of water movement through subsurface flow paths.

A physically based daily streamflow model of the Delaware River basin (Wolock and others, 1989), based on TOPMODEL (Beven and Kirkby, 1979), was used to determine the relative effects of climate, topography, and soils on basin hydrologic characteristics. Actual climatic, topographic, and soil characteristics were determined from more than 100 representative areas of the basin. The parameters for the model were chosen for each run at random from the range of actual characteristics, and a large number of runs then could produce a representative sample of hydrologic characteristics for the basin. The same procedure of sampling a range of basin climatic, topographic, and soil parameter values was followed in all analyses involving TOPMODEL that are referenced later in this report; hence, all sensitivity experiments are comparable and incorporate parameter values representative of all the source areas of basin streamflow.

Data on basin topography, soil properties, and daily precipitation and temperature are required to run TOPMODEL. The topographic parameter, a measure of hillslope steepness and shape, was derived from gridded, digitized elevation data from 1:250,000-scale maps. Soil parameters (saturated hydraulic conductivity, depth to bedrock, and field capacity) were estimated from digitized soil-survey maps at 1:250,000 scale. Daily average temperature and total precipitation were derived from a stochastic climate model of the basin (Wolock and others, 1989) based on data from 28 first-order National Weather Service meteorological stations in and near the basin. The daily stochastic climate model incorporates the spatial differences in intensity and duration of daily precipitation.

One hundred 5-year simulations were run, each simulation starting with a different randomly selected set of climatic, topographic, and soil parameter values. The following hydrologic characteristics were calculated for each 5-year simulation: mean daily depth to water table, mean daily evapotranspiration, mean daily streamflow, maximum daily streamflow, 7-day low flow, and the ratio of surface runoff to total runoff. Rank correlation coefficients between the basin parameters (climatic, topographic, and soil) and the hydrologic characteristics were calculated from the set of 100 simulations. The magnitude of the correlation coefficient indicates the degree of association between the basin parameter and the hydrologic characteristic.

The ranges of correlation coefficients ($r$) derived from the 100 simulations indicate that simulated maximum daily flow ($r$ = 0.61 to 0.83), mean daily flow ($r$ = 0.74 to 0.94), and 7-day low flow ($r$ = 0.65 to 0.77) correlated better with precipitation parameters than with other basin parameters. The ratio of surface runoff to total runoff correlated best with the topographic parameter ($r = 0.72$). Simulated evapotranspiration correlated best with average daily temperature ($r = 0.92$). Simulated moisture deficit for saturated soil correlated best with depth to bedrock ($r = 0.72$).

Grouping the correlation results into the general categories of climate, topography, and soil indicates that climate has the greatest effect on maximum daily flow, mean daily flow, evapotranspiration, and 7-day low flow. Topography has the greatest effect on the ratio of surface runoff to total runoff. Soil has the greatest effect on depth to water table. On the basis of these results, the differences in most of the hydrologic characteristics for small basins in the Delaware River basin are due to differences in climate from one part of the basin to another. Only differences in depth to water table and flow path are not due to differences in basin climate.

### Relation of Streamflow to Temperature and Precipitation

A comparison of mean annual temperature and total annual precipitation with mean annual streamflow in a midbasin watershed (Neshaminy Creek; fig. 17) reveals that year-to-year variability in precipitation largely controls the observed year-to-year variability in streamflow. This relation is true for most hydrologic systems. (The principal exceptions are systems involving extreme snow accumulation, as in mountain glaciers, where temperature determines how much of the precipitation becomes streamflow each year.) Thus, changes in regional precipitation patterns will have a substantial effect on streamflow.

## SENSITIVITY OF WATER RESOURCES TO CLIMATE VARIABILITY AND CHANGE

The remainder of this report presents the results of a research project in the Delaware River basin (Ayers and Leavesley, 1988), which illustrate the sensitivity of water resources to potential changes in the humid-temperate climate of the basin. Many examples illustrate how variability and potential changes in temperature, precipitation, and the transpiration rate of vegetation affect the soil moisture, streamflow, drought, and water supply of the basin. The potential effects of sea-level rise on