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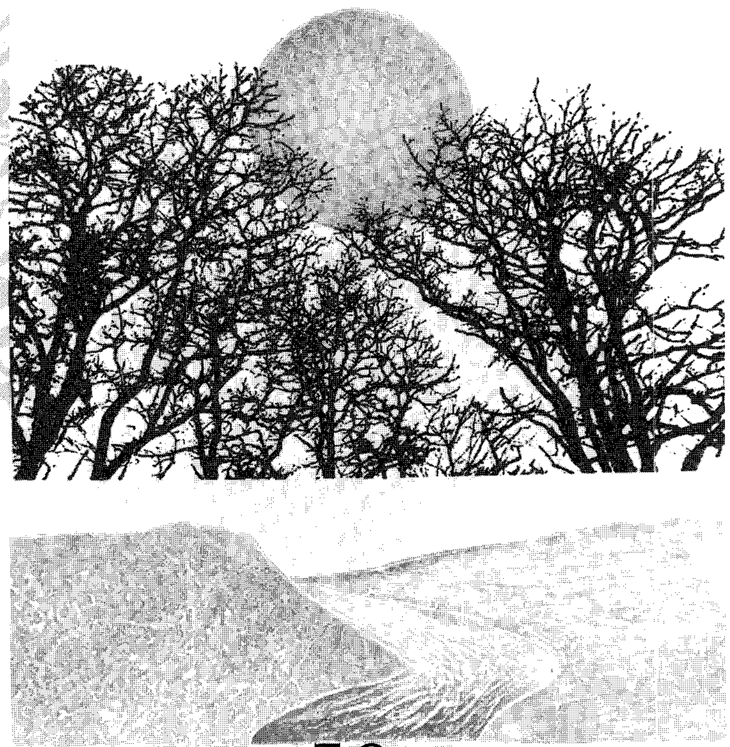
SENSITIVITY OF WATER RESOURCES IN THE DELAWARE RIVER BASIN



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

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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

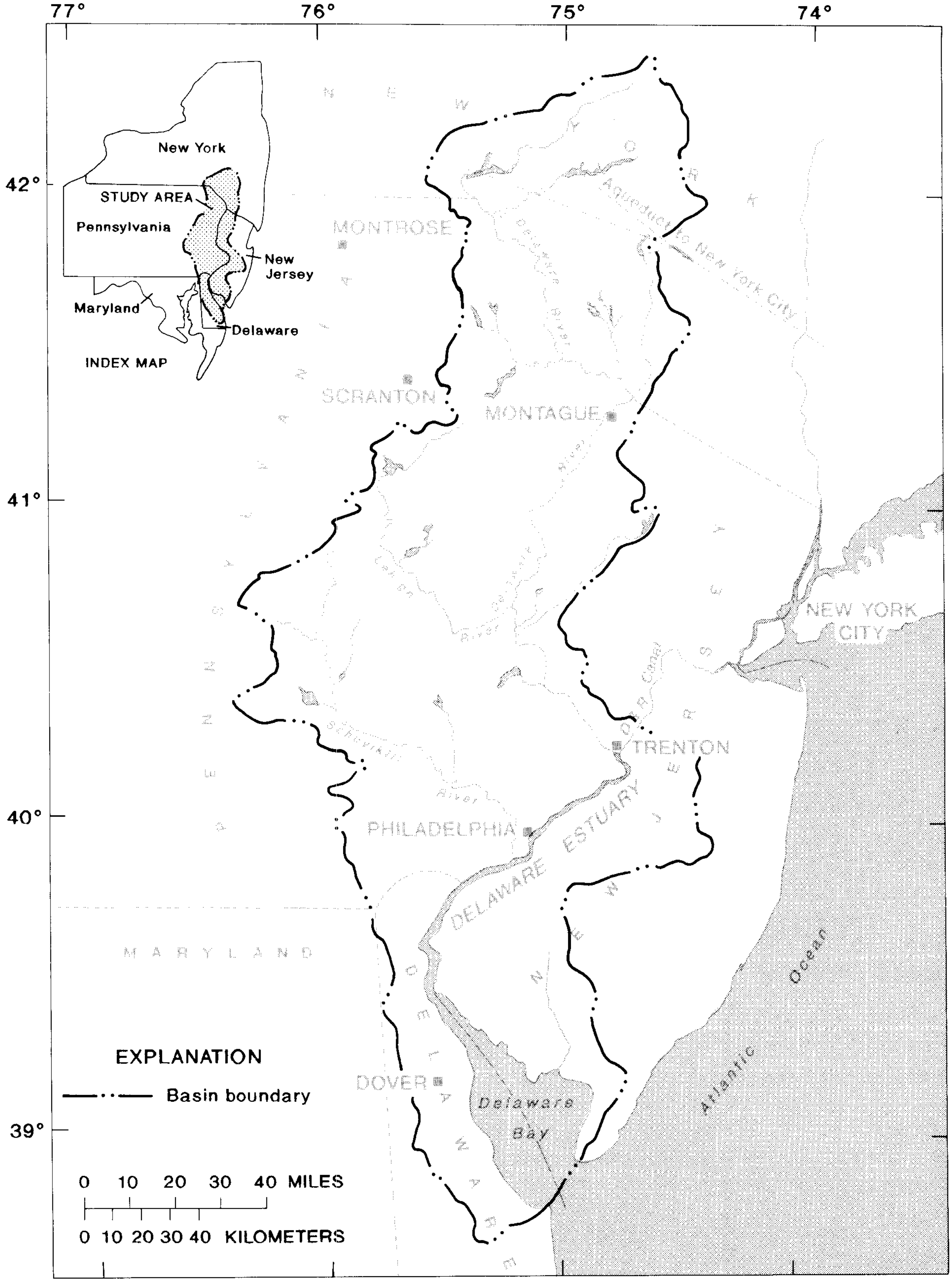
Energy from the sun, in the form of shortwave radiation, readily passes through the Earth's atmosphere. Some of this energy is reradiated from the Earth as longwave (infrared) radiation. Several gases in the atmosphere—principally water vapor (H_2O) and carbon dioxide (CO_2), but also methane (CH_4), 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.

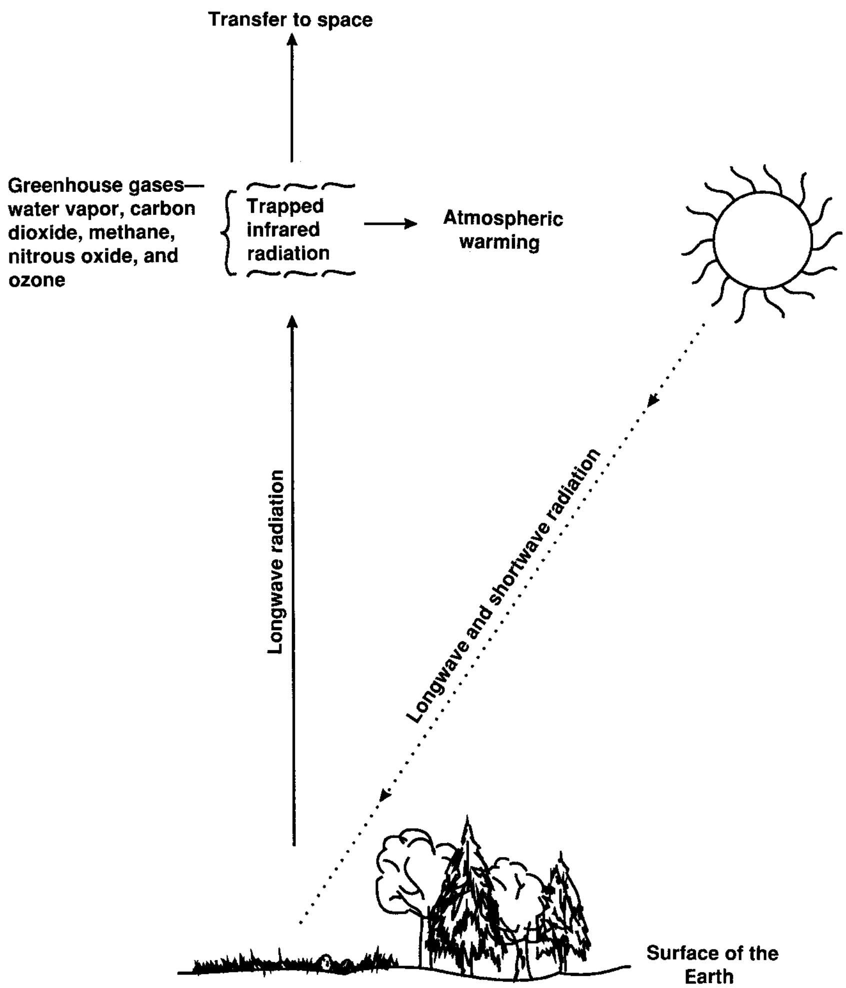


Figure 2. Schematic representation of the greenhouse effect.

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.

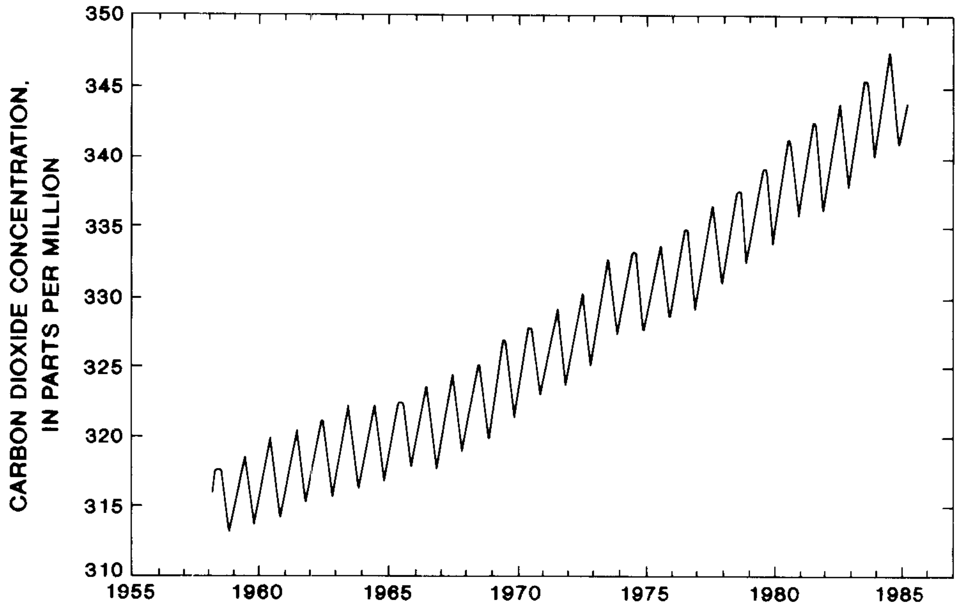


Figure 3. Recent atmospheric carbon dioxide measurements at Mauna Loa Observatory, Hawaii (modified from Gammon and others, 1985; original data from C.D. Keeling, Scripps Institute of Oceanography, La Jolla, Calif.).

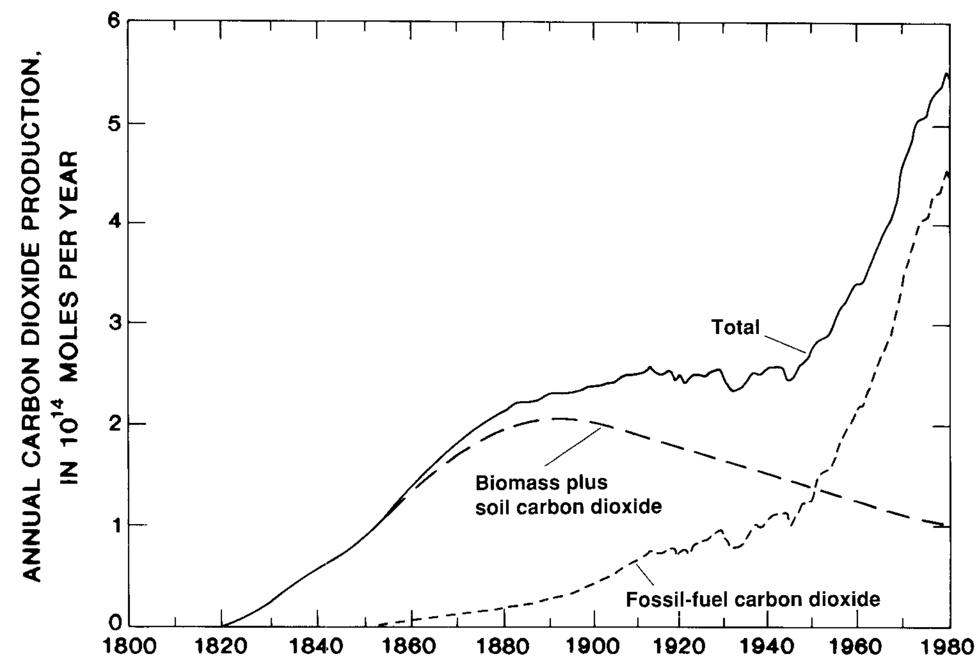


Figure 4. Historical global releases of carbon dioxide from terrestrial (biomass and soil) and fossil-fuel sources (modified from Peng and others, 1983).

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

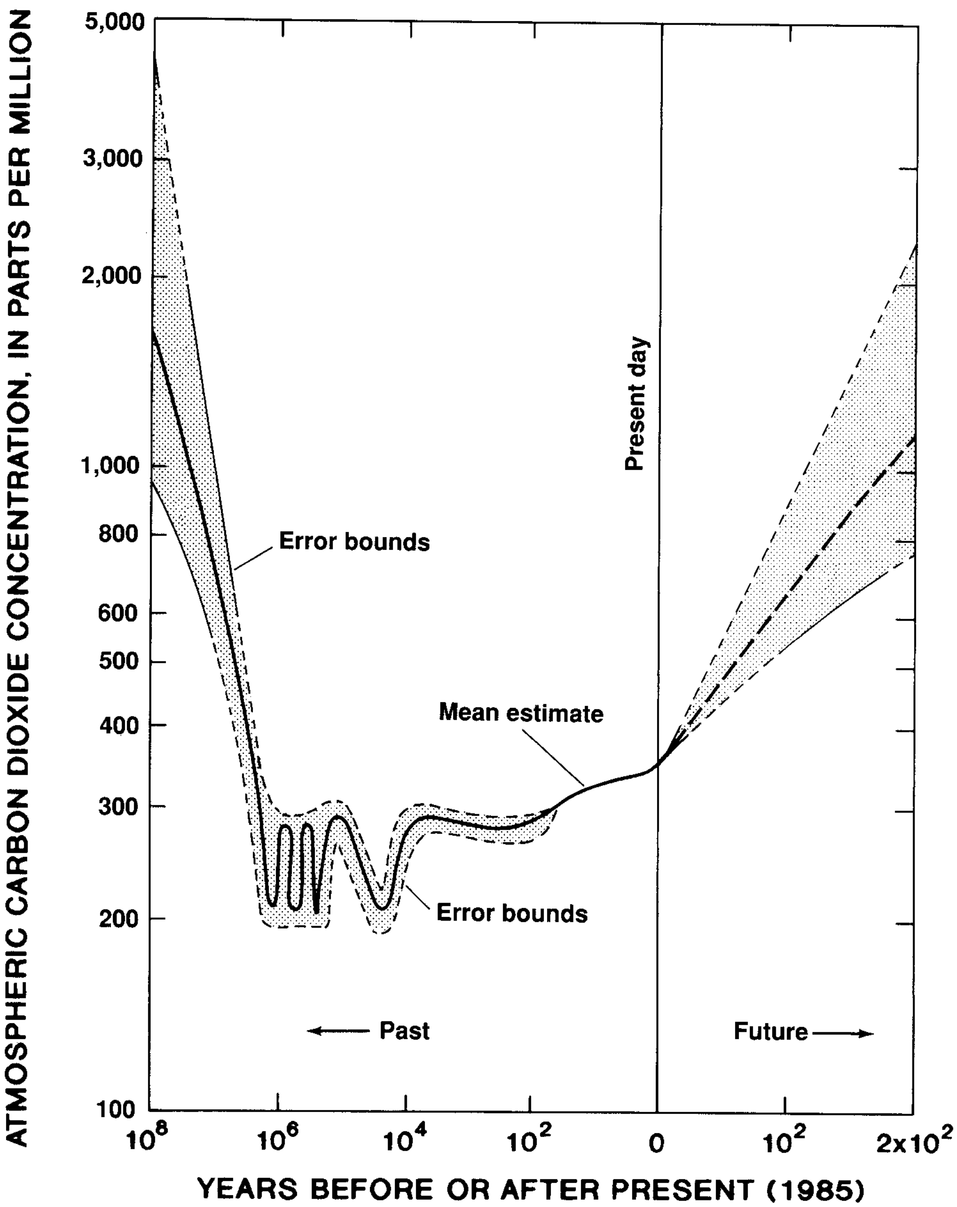


Figure 5. Range in estimated historical fluctuations and future projections of atmospheric carbon dioxide (modified from Gammon and others, 1985).

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.

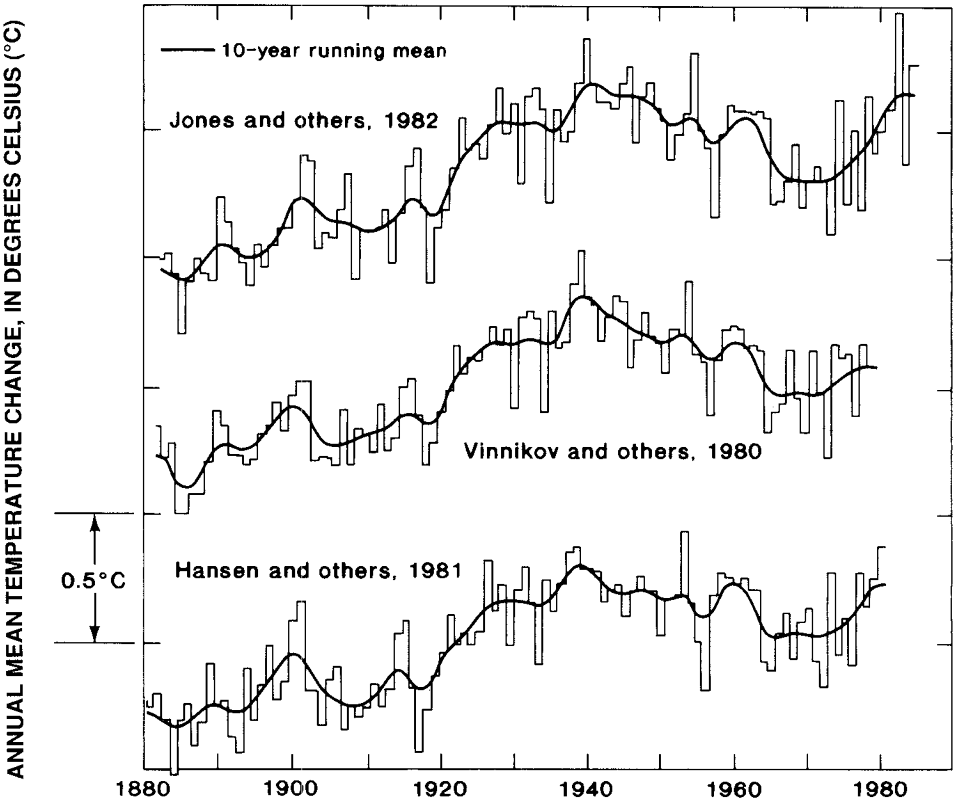


Figure 6. Three estimates of Northern Hemisphere temperature changes from climate records (modified from Wigley and others, 1985).

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

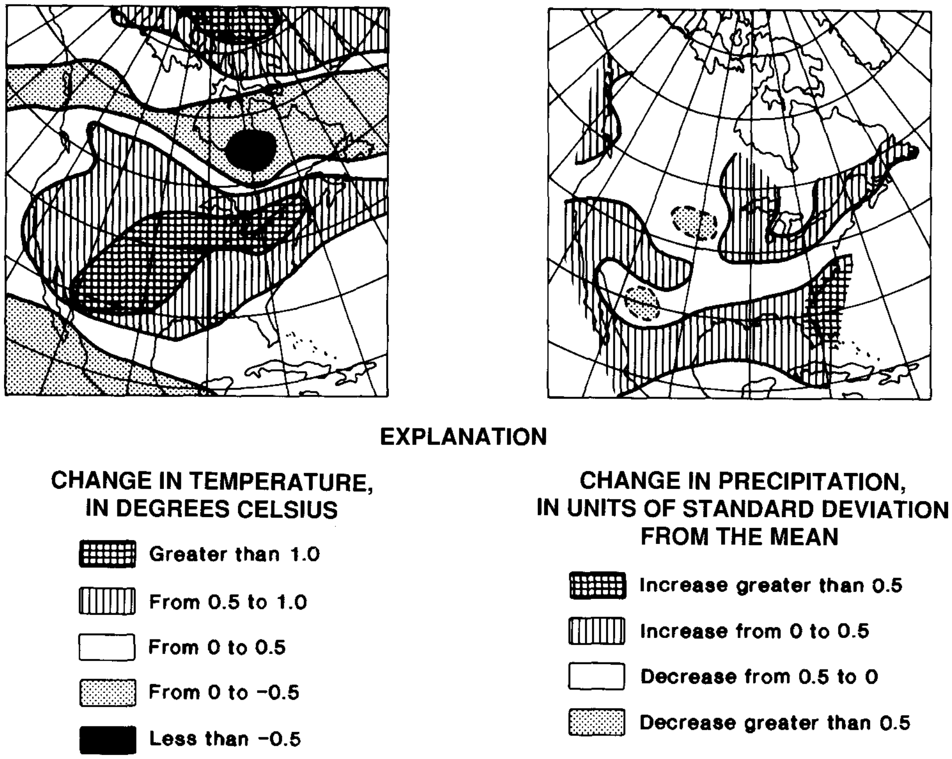


Figure 7. Estimated changes in summer temperature and precipitation for a warmer climate, calculated from differences between mean values of 1901–20 and 1934–53 climate records for North America (modified from Webb and Wigley, 1985).

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

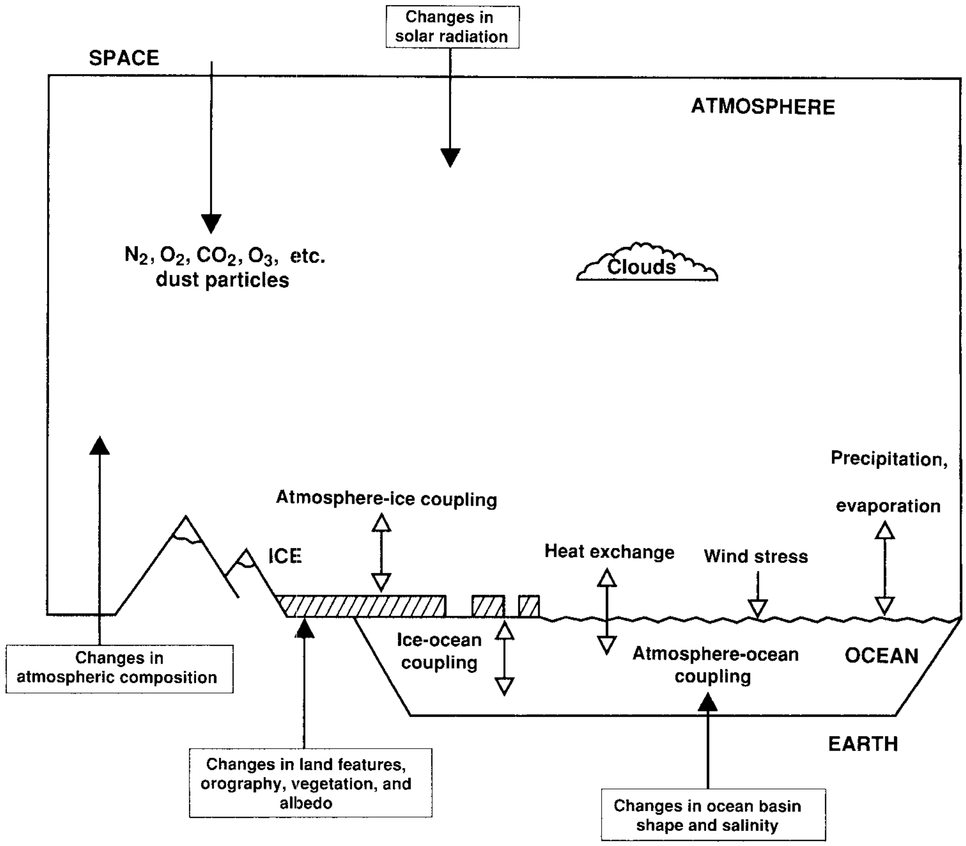


Figure 8. Schematic representation of the Earth's climate system and interactions among its principal components (modified from National Research Council, 1975).

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

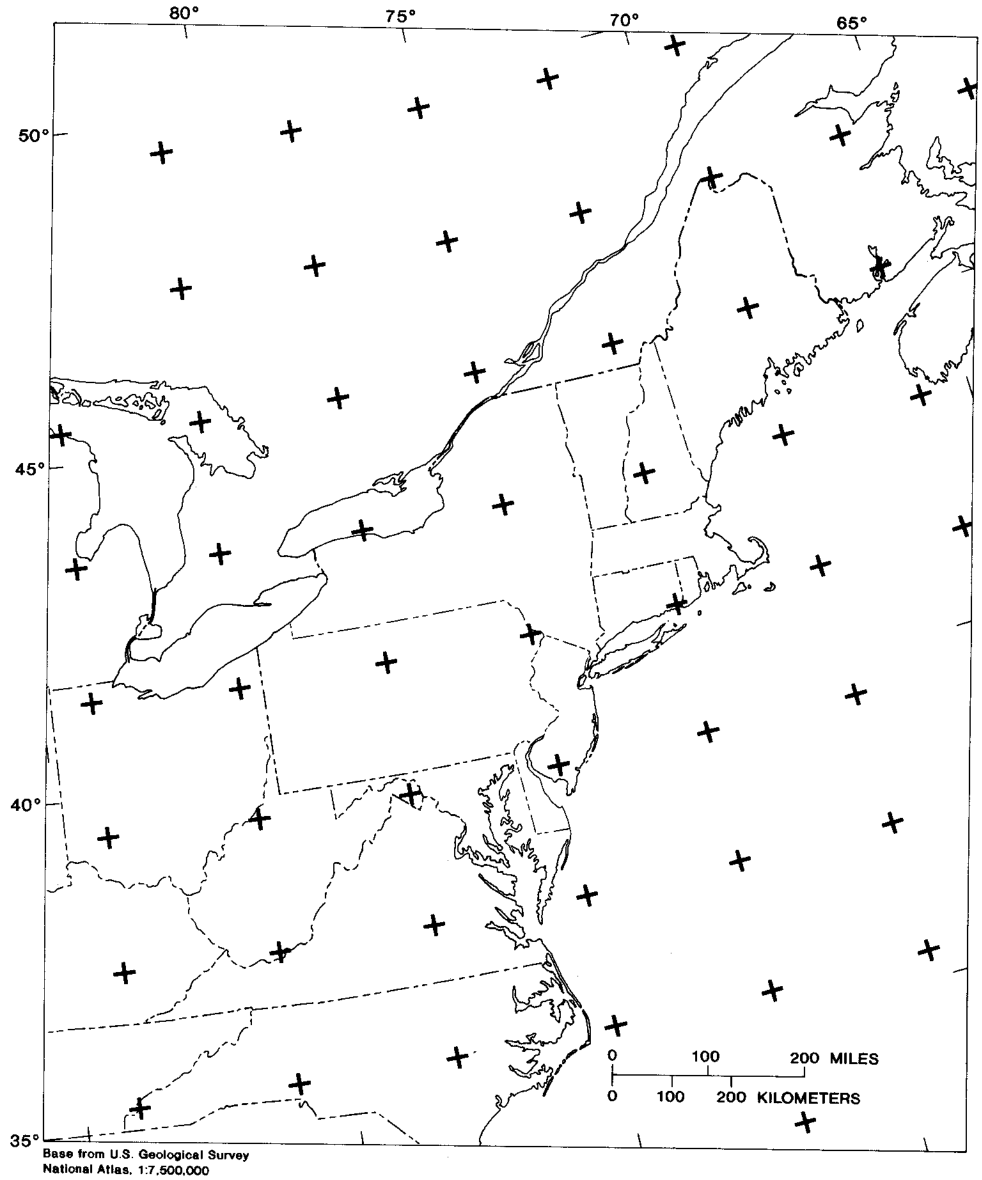


Figure 9. Example of computational grid for a general circulation model with approximate spacing of 2° latitude by 3° longitude.

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

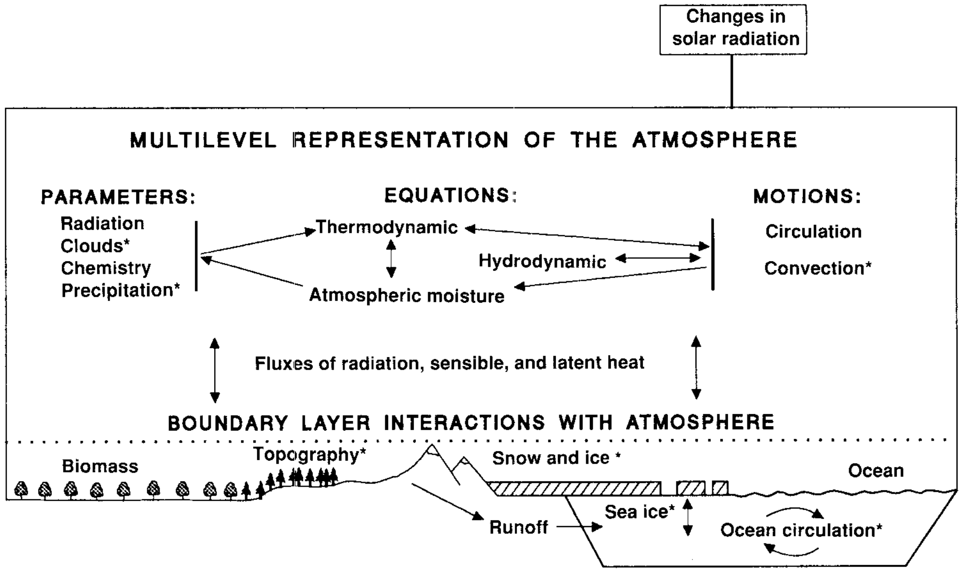


Figure 10. Schematic representation of climate-system elements simulated in general circulation models. The most problematic processes are marked with asterisks.

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-

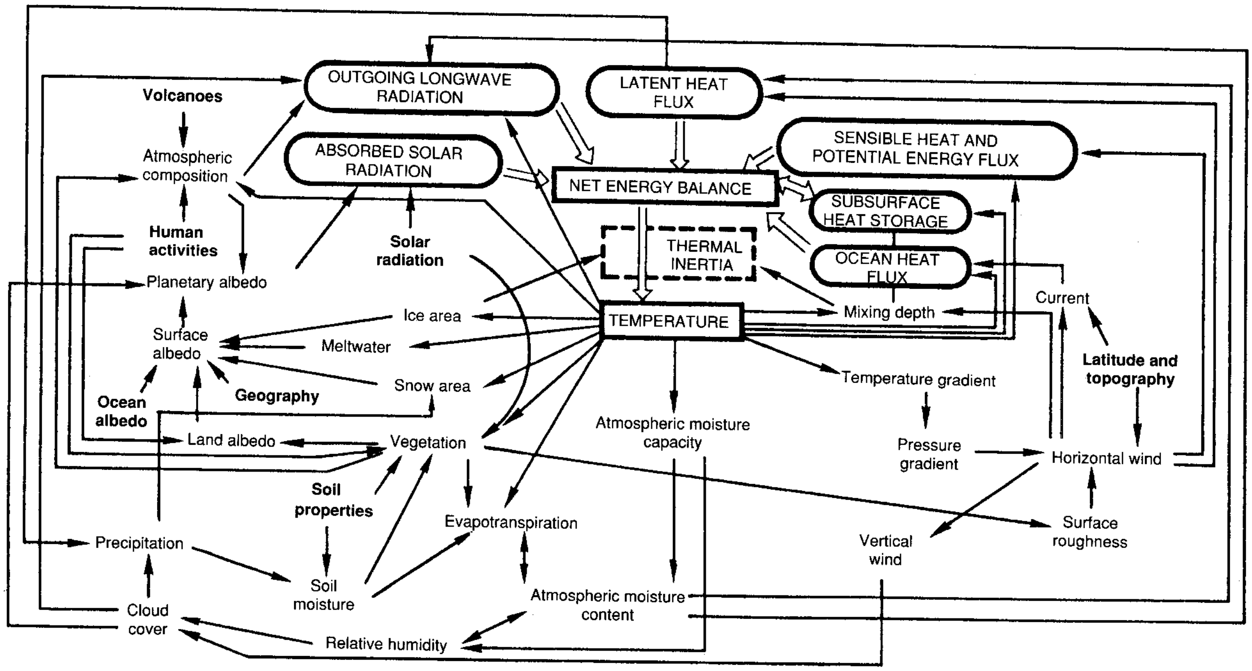


Figure 11. Schematic representation of climatic cause-and-effect (feedback) linkages and variables often included in numerical models of the climate system. Many of these linkages are oversimplified and otherwise treated inadequately (Robock, 1985).

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-

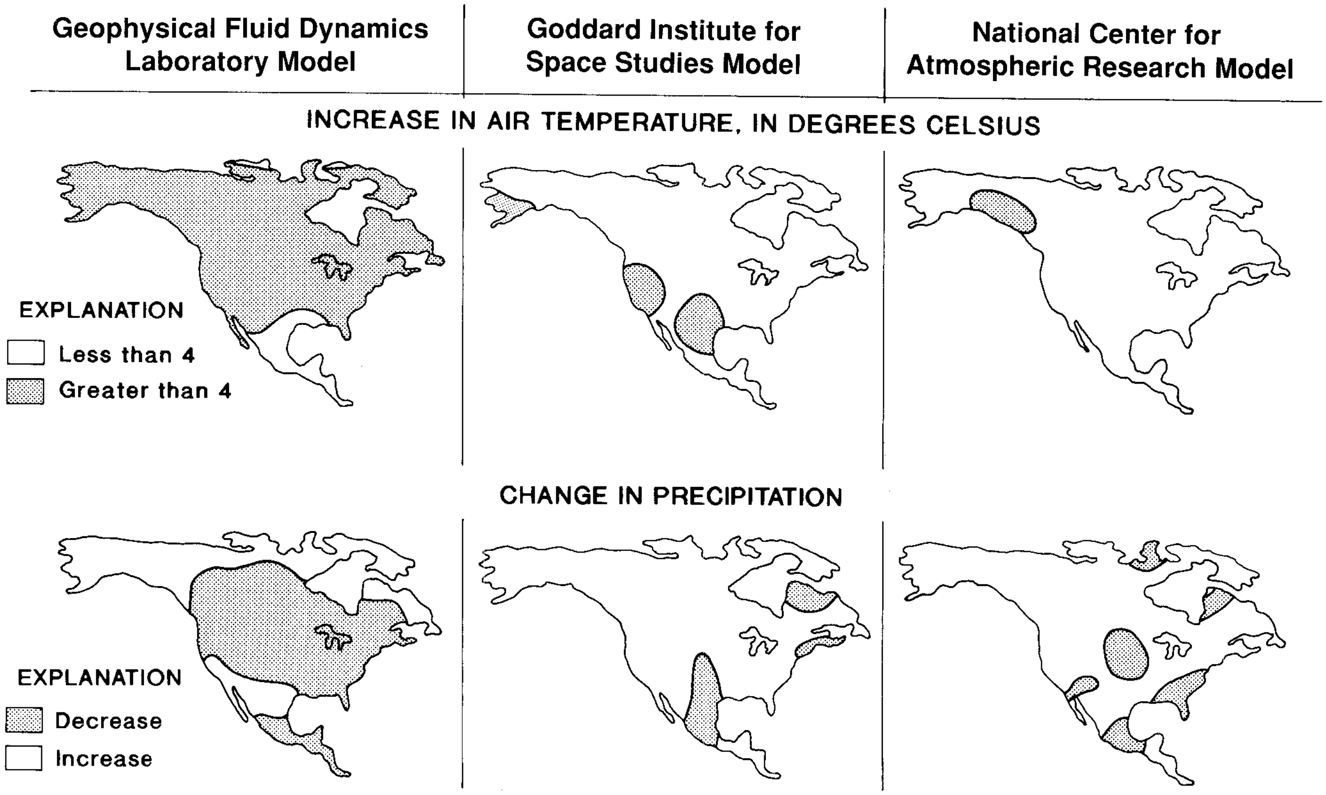


Figure 12. Estimated summer temperature and precipitation changes over North America for doubled-carbon-dioxide conditions in three general circulation models (modified from Schlesinger and Mitchell, 1985).

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

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

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

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.

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.

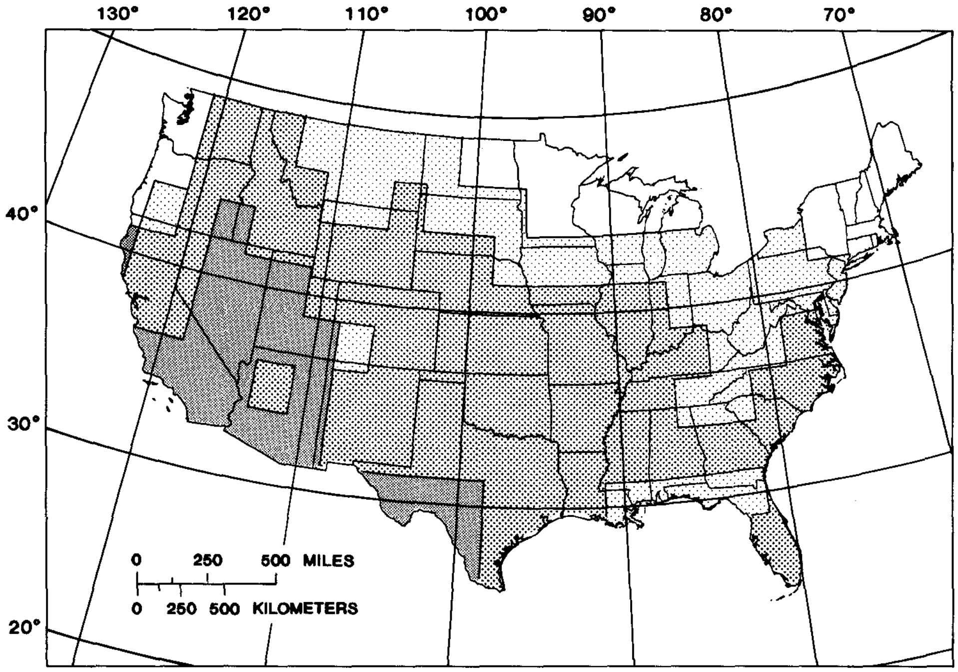
A second analysis was performed to examine the effects of climate changes on the mean annual Thornthwaite moisture index for the conterminous United States (McCabe and others, 1990). To represent current climate conditions, the investigators developed a $2.5° \times 2.5°$ grid of mean annual temperature 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 ≤10
- >-30 to ≤-10
- >-50 to ≤-30
- ≤-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

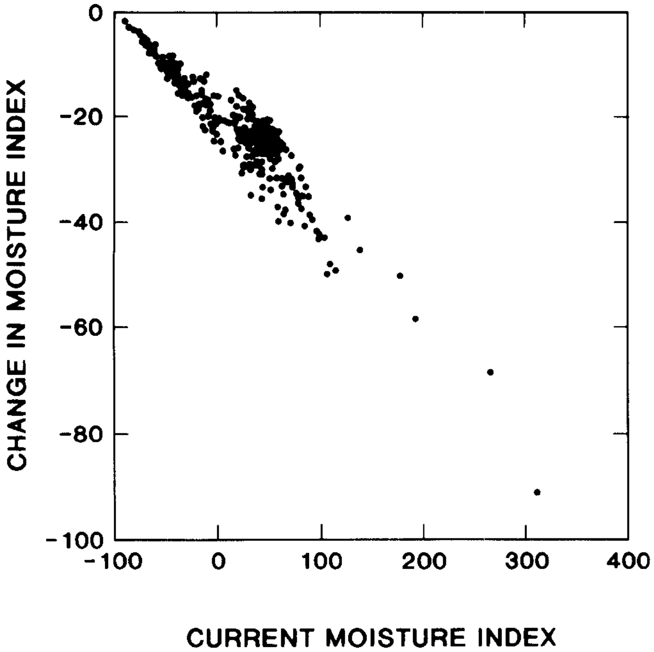


Figure 14. Relation between current Thornthwaite moisture index for each National Weather Service climate division and change in moisture index caused by a 4-degree-Celsius increase in mean annual temperature.

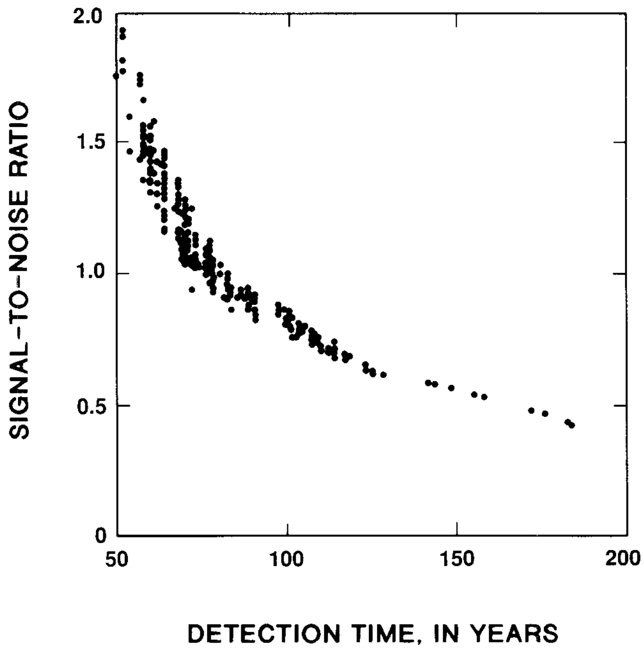


Figure 15. Relation between detection time and signal-to-noise ratio for all National Weather Service climate divisions. Detection time is the time required for a significant (at significance level 0.05) trend in 50 percent of the simulations. Signal-to-noise ratio is the difference between the current 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-

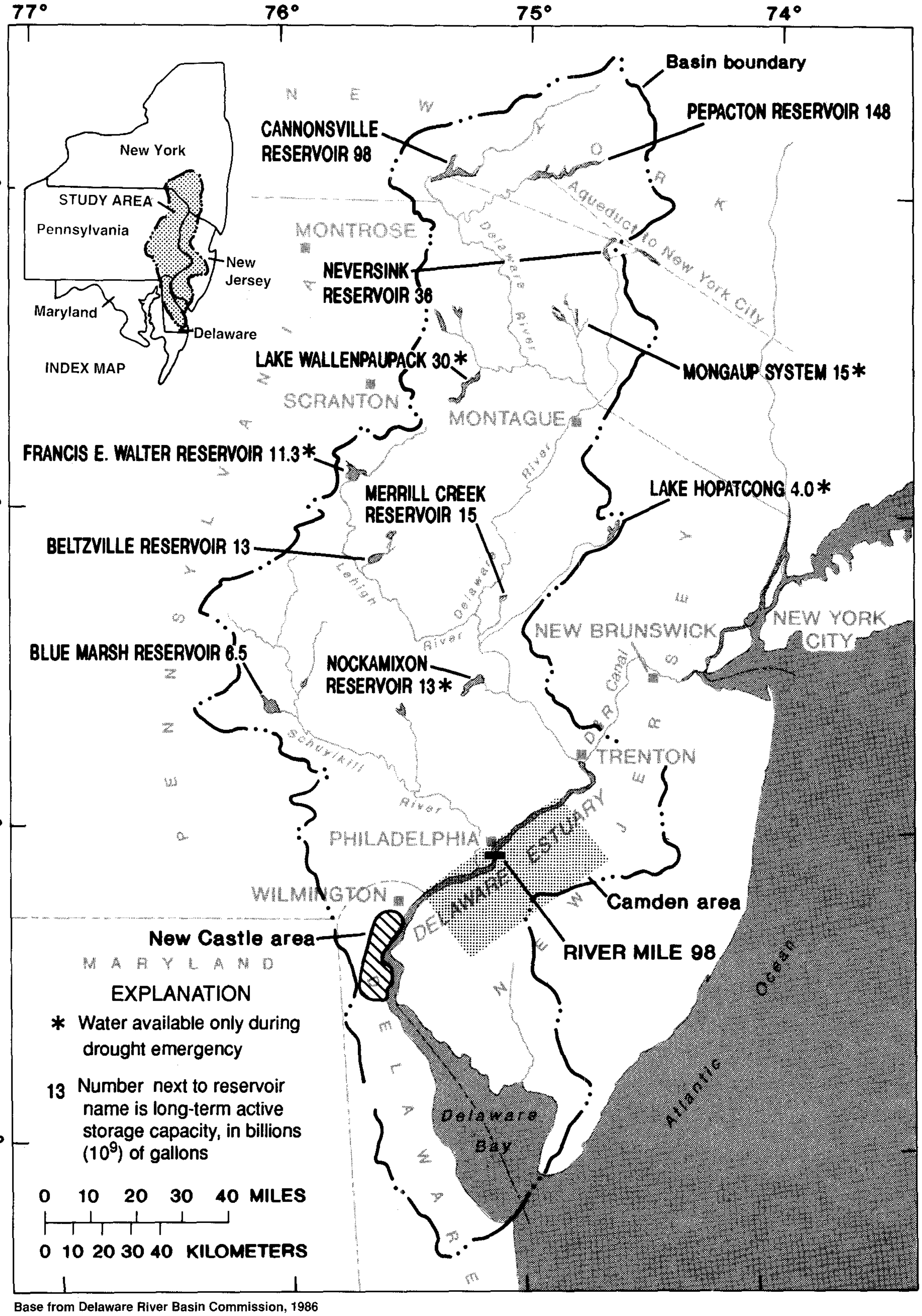


Figure 16. Major water-supply features and long-term active storage capacity of major reservoirs in the Delaware River basin.

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

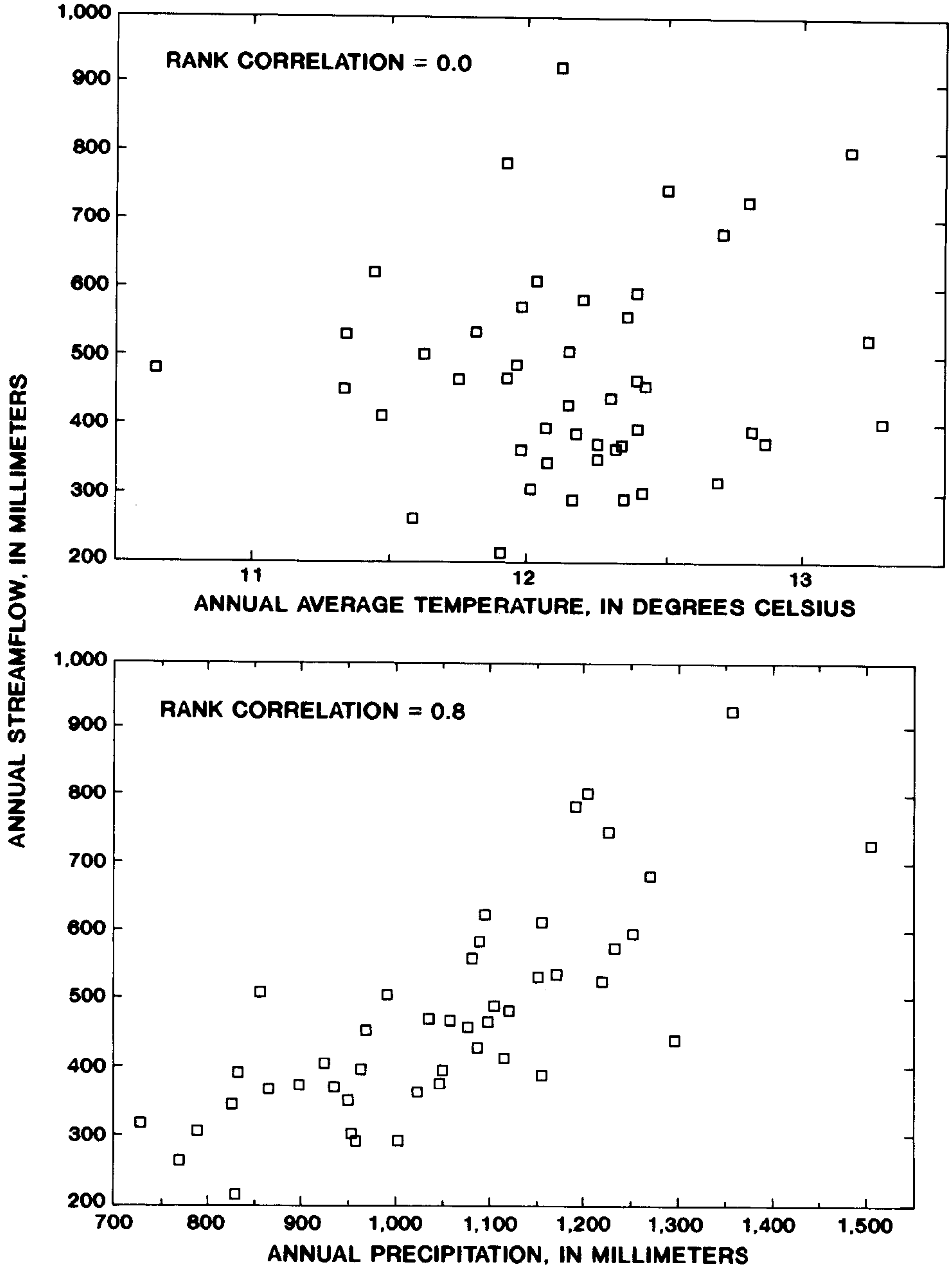


Figure 17. Annual flow of a Pennsylvania stream, Neshaminy Creek, in relation to mean annual air temperature and total annual precipitation.

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 saltwater intrusion, aquifer recharge, tidal wetlands, and coastal flooding and erosion are also described.

Potential Changes in Soil Moisture and Irrigation Demand

Soil moisture becomes an increasingly important component of the water budget during summer (June, July, August), when the climatic and biological demand (evapotranspiration) for water is highest (McCabe and Ayers, 1989). Low soil moisture usually leads to plant stress, reduced streamflow, and decreased ground-water recharge. Irrigation is necessary to maintain enough soil moisture to produce crops in areas of sandy, well-drained soils in the southern part of the basin.

Estimates of soil-moisture conditions during summer months were calculated by means of a monthly water-balance model (McCabe and Ayers, 1989) for present conditions, for various hypothetical (prescribed) changes in temperature and precipitation, and for three GCM projections (GFDL, GISS, and OSU models). The model used a Thornthwaite water-balance approach (Thornthwaite and Mather, 1955; Mather, 1978) based on average monthly temperature and precipitation for three climate stations in the basin. Results of the model simulations indicate that atmospheric warming will cause an increase in potential evapotranspiration and, unless precipitation increases, a decrease in soil moisture during the summer. Even with a 20-percent increase in precipitation, simulated mean summer soil moisture for the basin would likely decrease for any temperature increases greater than 2°C (McCabe and Ayers, 1989). Simulated increases in stomatal resistance of plants to transpiration also could counteract the effects of a simulated temperature increase, if the net effect of increased CO_2 is reduced plant transpiration.

Effects of reduced soil moisture in summer on simulated streamflow and ground-water recharge would persist into the autumn. More autumn precipitation would be needed to replenish soil-moisture deficits, and thus less water would be available for autumn and early winter streamflow and ground-water recharge.

About 60,000 acres of cropland and 250 golf courses are irrigated each year in the Delaware River basin (R. Limbeck, Delaware River Basin Commission, oral commun., August 1990). Irrigation water use (57 Mgal/d) is not a large part of the total water use in the basin (0.7 percent of 7,700 Mgal/d; J. Featherstone, Delaware River Basin Commission, written commun., Feb. 1988), but because about 90 percent of irrigation water applied is lost to evapotranspiration, irrigation constitutes a substantial part of consumptive water use in the basin (19 percent of 302 Mgal/d). This percentage does not include consumptive water use for lawn watering, which is a substantial but largely unknown part of commercial and municipal consumptive water use. A warming trend likely would result in a need for more frequent irrigation and, in turn, in an increase in consumptive water use in the basin, unless the irrigated area or application rates decrease accordingly.

In a simulation study by McCabe and Wolock (1992), an irrigation-demand model was developed to define the potential effects and relative sensitivity of long-term changes in climate on mean annual irrigation demand in southern New Jersey. The simulation model was based on a daily Thornthwaite water-balance approach that was driven by a stochastic model of daily temperature and precipitation. The climate-change scenarios consisted of combinations of gradual increases in mean annual temperature of 2°C and 4°C, gradual changes in mean annual precipitation ranging from −20 to +20 percent, and a gradual increase of 20 percent in stomatal resistance of plants to transpiration. Changes were linear and were applied uniformly throughout the year. Other factors that affect soil moisture, such as vegetation type, were not considered in this approach. The approach also did not account for additional possible and largely unknown effects of climate change, such as the effects of increasing CO_2 on the variability of temperature and precipitation and on other transpiration and growth changes of plants.

The daily Thornthwaite water-balance model was used to calculate soil moisture for two soil layers and to determine when irrigation was required (Thornthwaite and Mather, 1955; Mather, 1978). The stochastic model generated daily potential evapotranspiration (from temperature) and daily precipitation. The total storage capacity of soil moisture (field capacity) of both layers, 150 mm, was based on the capacity of sandy-loam soil and an average crop rooting depth of 610 mm. The capacity of the top layer was 25 mm; water was withdrawn from the top layer at a rate equal to potential evapotranspiration. The capacity of the bottom layer was 125

Table 3. Simulated average annual volume of irrigation needed in southern New Jersey for simulated changes in temperature, precipitation, and stomatal resistance

[Changes from current (1948–88) conditions. —, not available; °C, degrees Celsius]

Change in temperature	Change in stomatal resistance	Irrigation volume in millimeters, for indicated change in precipitation		
		+20 percent	No change	−20 percent
None	None	—	182	—
+2°C	None	204	242	279
+4°C	None	269	311	362
+2°C	+20%	99	134	176
+4°C	+20%	156	194	231

mm; water was withdrawn at a decreasing rate proportional to the ratio of soil moisture to capacity of the bottom layer.

On the basis of irrigation guidelines for New Jersey, irrigation was applied when daily soil moisture storage dropped to 50 percent of field capacity or below. The amount of water applied during each simulated irrigation event was 75 mm, the amount needed to bring the soil moisture storage to field capacity.

Hypothetical changes in mean annual temperature and precipitation without any change in stomatal resistance of plants to transpiration resulted in increases in mean annual irrigation demand, even with a 20-percent increase in mean annual precipitation (table 3). The increases ranged from 22 mm/year (a 12-percent increase) to 180 mm/year (a 99-percent increase). The effect on mean annual irrigation demand was greater from changes in mean annual temperature than from changes in mean annual precipitation. The greater influence of temperature changes in this case is important because climate models are more in agreement in regard to regional changes in mean annual temperature than to regional changes in precipitation.

A simulated increase in stomatal resistance of plants to transpiration, however, counteracted the effects of increased temperature and decreased precipitation on irrigation demand (table 3). When a 20-percent increase in stomatal resistance was assumed, changes in mean annual irrigation demand ranged from −83 mm per year (a 46-percent decrease) to +49 mm per year (a 27-percent increase). A regression analysis of the simulated results indicates that changes in stomatal resistance had more effect on changes in annual irrigation demand than did the specified changes in mean annual temperature and precipitation.

Regression models of the results of the irrigation-demand simulations were developed. A comparison of regression coefficients derived from the regression analyses indicates that a 5.1-percent increase in stomatal resistance can counteract the effects of a 1°C increase in mean annual temperature on irrigation demand and that a 2.8-percent increase in stomatal resistance can counteract a 100-mm (9.3-percent) decrease in mean annual precipitation. The regression also indicates that, when no change occurs in stomatal resistance, a 16.9-percent (182-mm) increase in mean annual precipitation can counteract the effects of a 1°C increase in mean annual temperature on irrigation demand.

Potential Changes in Streamflow

We have demonstrated that 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 and that changes in regional precipitation patterns could have a larger effect on streamflow than changes in temperature. The following discussion will focus on the sensitivities of streamflow in the basin to changes in climate.

Relation of Changes in Temperature, Precipitation, and Stomatal Resistance to Streamflow Variability

A set of 200 simulations (each 60 years long) was performed with the daily stochastic climate model of the basin as input to the physically based daily streamflow model of the basin (TOPMODEL; Wolock and others, 1989) to evaluate how uncertainty in changes in temperature, precipitation, and plant stomatal resistance to transpiration will affect daily streamflow (in this example, maximum daily streamflow). Each simulation represented a ran-

Table 4. Percentage of variance in Kendall's tau for maximum daily flow in the basin due to components of uncertainty

Source of variability	Percentage of the variance explained
Climate-change components	
Average daily precipitation	40
Storm-period duration	4
Interstorm-period duration	11
Mean annual temperature	6
Stomatal resistance	1
Natural variability	28
Nonlinearities	10

domly selected combination of climate-change variables in the following hypothetical ranges of model parameters: average daily precipitation, storm-period duration, and interstorm-period duration, −20 to +20 percent of current basin values; mean annual temperature, 0°C to +6°C of current basin values; and plant stomatal resistance to transpiration, 0 to +20 percent of current basin values. These ranges reflect current estimates of the potential for climate change.

For each simulation, Kendall's tau (a measure of trend in the data) was calculated for the predicted annual time series of maximum daily streamflow. The variability in Kendall's tau for the 200 simulations was partitioned into three categories: variability in climate-change variables, natural variability, and variability inherent in the methodology (table 4).

The largest explained variance (40 percent) is associated with average daily precipitation, indicating that the largest uncertainty in forecasting the effects of climate change is due to the current lack of confidence in projecting how average daily precipitation will change. The combined effects of average daily precipitation, storm-period duration, and interstorm-period duration account for 55 percent of the explained variance or uncertainty. The second-largest explained variance (28 percent) is due to natural variability in streamflow response to climate. Temperature changes accounted for only 6 percent of the uncertainty, and stomatal resistance only 1 percent. Simulation results for mean daily streamflow and for 7-day low flow (not shown here) were similar to the results for maximum daily streamflow.

Uncertainty in the predictions of climate change will decrease as our understanding of climate-change processes and feedbacks increases. Predictions of the effects of climatic change on streamflow will become more certain, but uncertainty due to natural variability will never be eliminated.

Sensitivity of Streamflow to Changes in Climate and Stomatal Resistance

The Thornthwaite monthly water-balance model and combinations of hypothetical temperature and precipitation changes were used to estimate potential changes in streamflow in the Delaware River basin (McCabe and Ayers, 1989). The simulated effects of CO_2-induced warming on annual streamflow indicated that a 3-percent increase in precipitation would be necessary to counteract the effects of each 1.0°C warming. A warming of 4°C could increase evapotranspiration and reduce annual basin streamflow by as much as 25 percent if current precipitation patterns remain unchanged. The magnitude of change in streamflow in each simulated climate-change scenario was highly dependent on the direction and magnitude of the precipitation changes in combination with the temperature increases.

Results from the daily (Wolock and others, 1989) and monthly (McCabe and Ayers, 1989) streamflow simulation of the basin indicate that atmospheric warming would increase the proportion of winter precipitation that falls as rain, which would increase winter streamflow, reduce snow accumulation, and reduce spring streamflow in the northern part of the basin. These findings agree with findings from other studies (Schaake and Kaczmarek, 1979; Gleick, 1987, 1989; Lettenmaier and others, 1988; Schaake, 1990), in which only changes in temperature and precipitation were considered. The effect of CO_2-induced warming on increased winter streamflows in the northern part of the basin for the GFDL model projection of climate change at Montrose, Pa., is shown in figure 18. Although the GFDL-projected annual streamflow decreases only 7 percent from the current average, there is a substantial shift in the timing of the projected streamflow.

In the southern two-thirds of the basin, snow accumulation generally is negligible, as illustrated for Trenton (fig. 19). The GFDL model projection of CO_2-induced warming resulted in streamflow changes that reflect increased summer evapotranspiration, reduced fall precipitation and soil moisture, and increased winter precipitation.

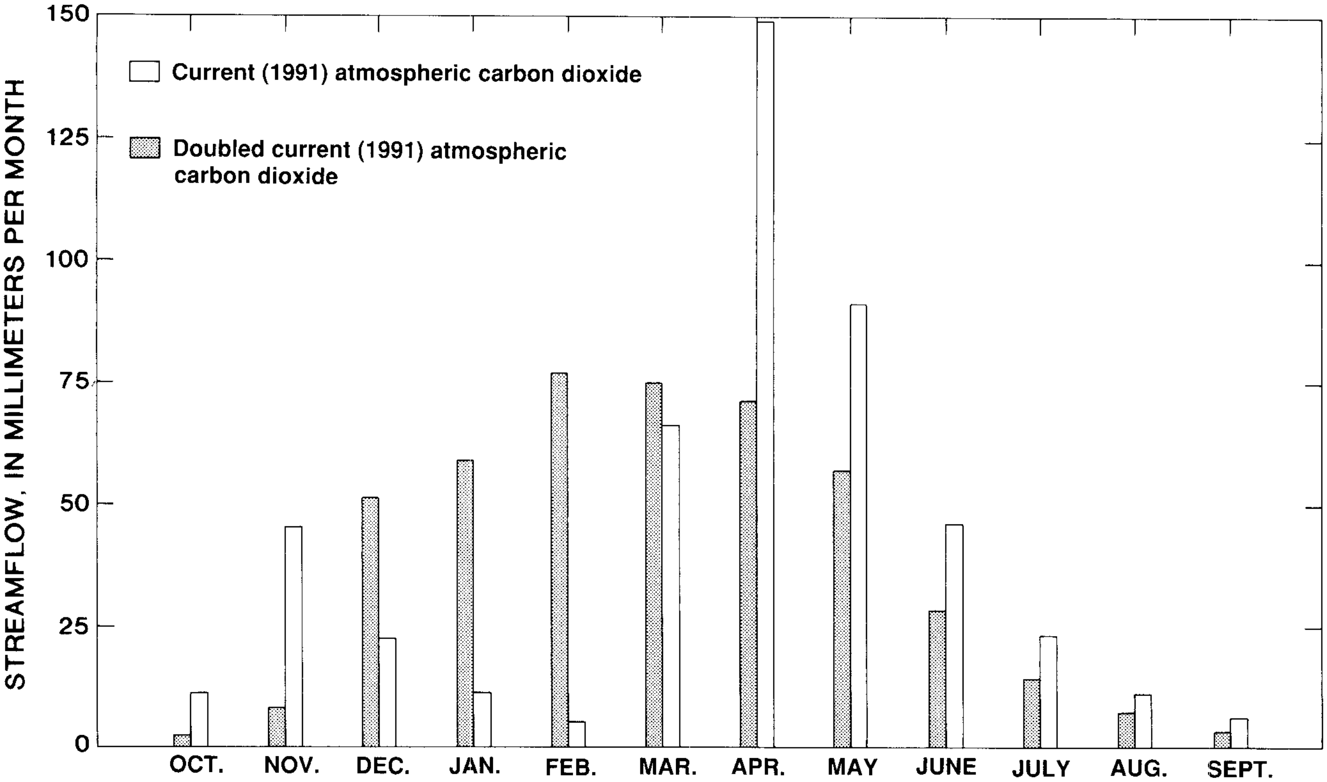


Figure 18. Simulated monthly average streamflow at Montrose, Pa., for present climate conditions and a scenario for a Geophysical Fluid Dynamics Laboratory general circulation model of doubled atmospheric carbon dioxide.

Although the two areas illustrated above appear to respond differently to the GFDL model projection of climate change, a reduction in annual streamflow of about 7 percent was simulated for both areas. Projections from two other GCM's produced simulated changes in annual streamflow for the basin ranging from −39 percent for the GISS model to +9 percent for the OSU model. The wide disparity in the three GCM projections is largely due to the difference in projected changes in precipitation.

Changes in the 7-day low flow that would result from hypothetical changes in climate and plant stomatal resistance to transpiration were simulated with the stochastic daily climate model as input again to the daily streamflow model of the basin (TOPMODEL; Wolock and others, 1989). Fifty 60-year simulations were performed for several hypothetical scenarios (table 5). The percentage change in 7-day low flow was calculated for each run of each hypothetical climate-change scenario.

The 20th, 50th (median), and 80th percentile values of the percentage change in the 7-day low flow for current conditions (no change) are −11, +3, and +12, respectively (table 5). These results can be interpreted as a forecast of the potential-change after 60 years in the 7-day low flow due to natural variability alone. The results indicate that the 60-percent confidence interval for the forecast is between −11 and +12 percent, a range of 23 percentage points.

The effect of a 3°C warming on 7-day low flow is indicated by the changes in the 20th, 50th, and 80th percentile values to −27, −13, and −2 percent, respectively (table 5). The negative shift in this distribution of forecasted change in 7-day low flow indicates that the low flow is more likely to decrease with the 3°C warming.

The range in percentage change in 7-day low flow is strongly affected by changes in precipitation and stomatal resistance to transpiration that may accompany a global or regional warming (table 5). If precipitation were to increase 20 percent with the warming, then the range defined by the 20th and 80th percentile values for percentage change in the 7-day low flow would be 0 to 24. If precipitation were to decrease 20 percent with the warming, then the range would be −61 to −34 percent. Finally, if warming were accompanied by a 20-percent increase

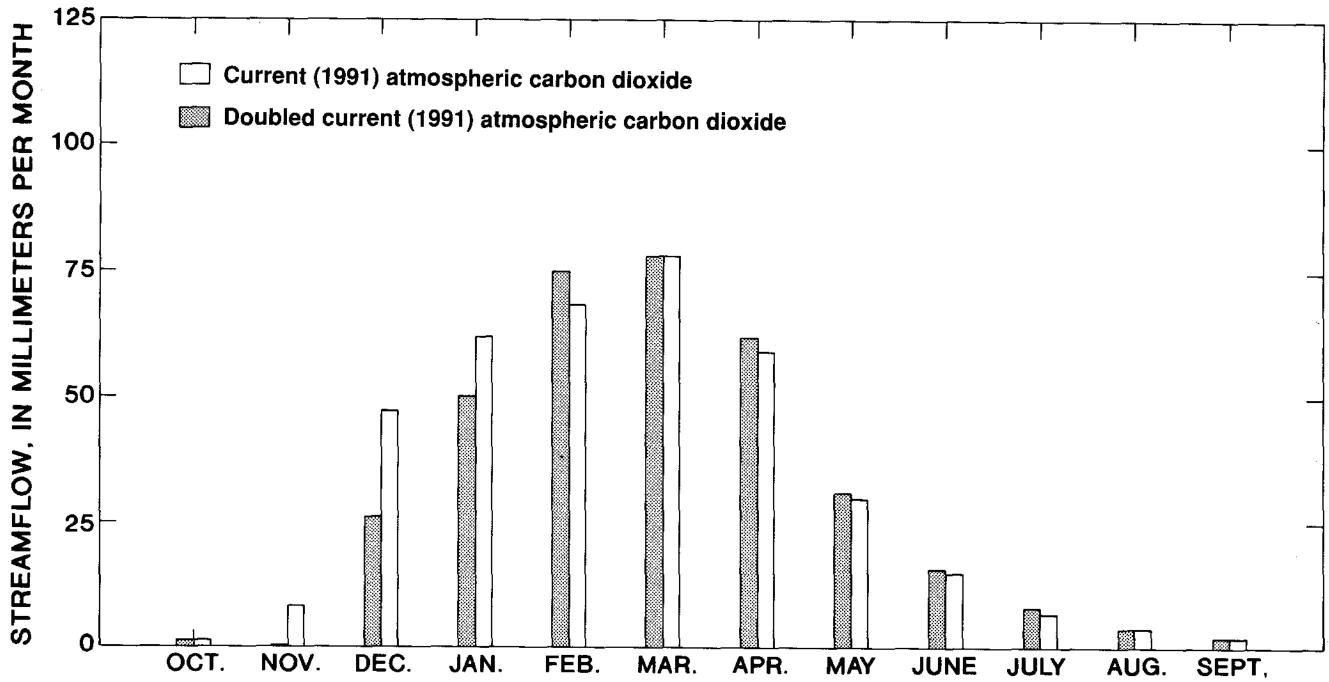


Figure 19. Simulated monthly average streamflow at Trenton, N.J., for present climate conditions and a scenario for a Geophysical Fluid Dynamics Laboratory general circulation model of doubled atmospheric carbon dioxide.

Table 5. Range in percentage change in the 7-day low flow among the fifty 60-year simulations for each hypothetical scenario

[Median 7-day low flow for the first year of the 50 simulations was 1.62 mm. °C, degrees Celsius]

Climate-change scenario			Percentage change in indicated percentiles		
Temperature	**Precipitation**	**Stomatal resistance**	**20th**	**50th**	**80th**
No change	No change	No change	−11	3	12
+3°C	No change	No change	−27	−13	−2
+3°C	+20 percent	No change	0	12	24
+3°C	−20 percent	No change	−61	−43	−34
+3°C	No change	+20 percent	−5	9	21

in stomatal resistance, the range would be −5 to +21 percent.

Three categories of uncertainty—the direct CO_2 effects on change in vegetation transpiration, the indirect CO_2 effects on change in precipitation, and the natural variability in climate—have a demonstrated effect on streamflow predictions. Incorporation of a 20-percent increase in stomatal resistance raised the median 7-day low flow change by 22 percentage points. Including a 20-percent increase in precipitation raised the median 7-day low flow change by 25 percentage points. For each climate-change scenario, the range between the 20th and 80th percentile values (natural variability) was about 25 percentage points. Thus, uncertainty in the predictions of streamflow due to the three categories is of the same magnitude, given the ranges used in the simulations.

These results are consistent with other findings (Gleick, 1989). The directions and magnitudes of streamflow changes were highly dependent on the changes in precipitation and stomatal resistance that were assumed to accompany the hypothetical warming. A component of uncertainty that previous studies have ignored, however, is the effect of natural variability. In forecasts of the effects of climate change on streamflow, for the Delaware River basin, the amount of uncertainty due to natural variability is of the same order of magnitude as that due to uncertainty in how the precipitation characteristics or the effects of CO_2 on vegetative transpiration will change. As workers improve their projections of precipitation that accompanies increasing concentrations of atmospheric CO_2, the uncertainty in forecasts of effects on streamflow will be reduced accordingly. In addition, as research progresses on how increasing atmospheric CO_2 affects vegetation and transpiration, the uncertainty associated with stomatal resistance will decrease. The uncertainty in forecasts due to natural variability, however, is probably irreducible.

Potential Difficulty in Detecting Changes in Streamflow

Changes in annual streamflow that would result from changes in climate were simulated with the stochastic daily model of climate as input to the daily streamflow model of the basin (TOPMODEL; Wolock and others, 1989). The stochastic climate model was forced to generate daily precipitation statistics equivalent to the statistics of the 1954–88 period of record, and in the simulations, a warming of 0.5°C per decade was assumed to apply to the temperature statistics for that same period. Differences in streamflow between the first and last years of each of the fifty 60-year simulations indicated that most model runs resulted in streamflow decreases (fig. 20). Natural variability in precipitation, however, was large enough to cause some of the model runs to yield streamflow increases and thus partially mask the detection of streamflow decreases resulting from the 3.0°C warming.

Three GCM projections (GISS, GFDL, and OSU models) of climate change were used in another series of TOPMODEL simulations to determine the extent to which a significant decreasing trend in streamflow would be masked by the variability in precipitation. All three GCM's projected a warming greater than 3.0°C. GISS and GFDL projected decreased and unchanged precipitation, respectively, along with the warming, whereas OSU projected slightly increased precipitation. The resulting changes in streamflow (fig. 21) would be detectable much sooner with the first two GCM projections than with the third.

In another TOPMODEL analysis, fifty 60-year simulations were performed for an assumed warming of 3°C and no change in precipitation characteristics. Each of the 50 simulations represents a different future realization of the same climate-change projection (warming of 3°C). Annual and monthly maximum daily streamflow and Kendall's tau trend statistics were calculated for each of the 60-year streamflow time series to detect trends in streamflow.

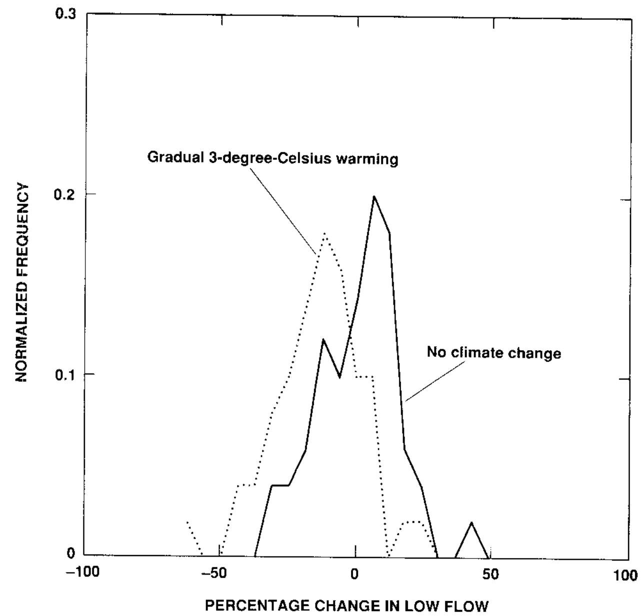


Figure 20. Normalized frequency distribution of the percentage change in 7-day low flow in the Delaware River basin for simulations with no climate change and with gradual 3-degree-Celsius warming, both with present precipitation amounts and variability.

These simulations (table 6) illustrated two important characteristics of the sensitivity of basin streamflow to climate change. First, there are seasonal differences in the expected effects of the hypothetical warming on streamflow. Maximum daily streamflow increases with time (has more positive than negative trends in Kendall's tau) in midwinter, decreases in spring and summer, and changes little in autumn and early winter. These seasonal differences in trend primarily reflect changes in snowfall accumulation and snowmelt. With a warming, more winter precipitation falls as rain than as snow, and snowmelt occurs earlier. The warming effect is strongest in the northern part of the basin (above lat 40.8°N.), where snow accumulation currently is significant. Second, the natural variability in precipitation masks the effects of increasing temperature. The percentage of simulations that showed no significant increase or decrease

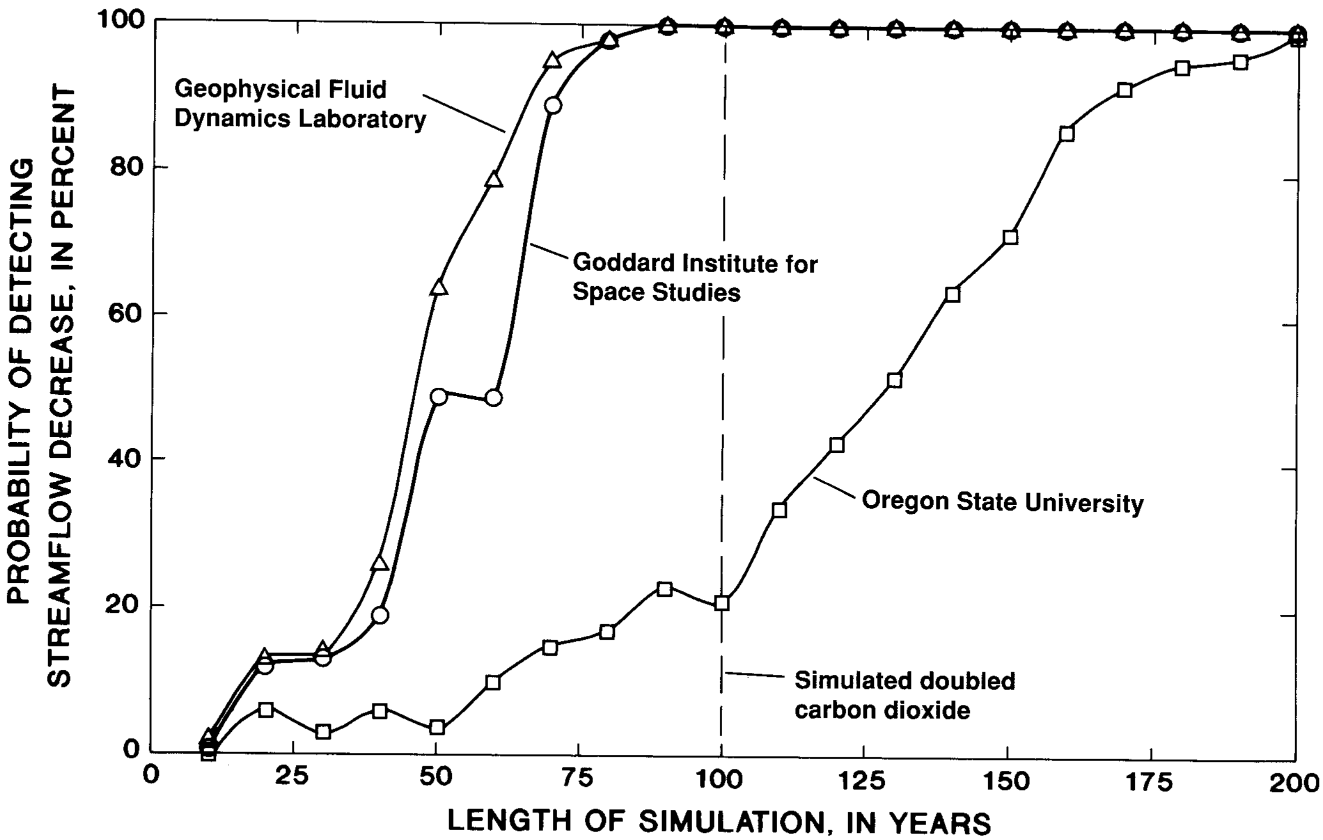


Figure 21. Probability of detecting streamflow change in the Delaware River basin over the next 200 years according to three scenarios of general circulation models.

Table 6. Numbers and types of trends in maximum daily streamflow for 50 simulations and six streamflow records in Delaware River basin

[Positive (+) or negative (−) trend at significance level equals 0.05; 0, no trend; °N., degrees north, columns 1–6 refer to different gages]

Month	Number of simulated trends			Trends in observed streamflow records					
				Latitude greater than 40.8°N.				Latitude less than 40.8°N.	
	+	−	0	1	2	3	4	5	6
January	12	2	86	0	0	0	0	0	0
February	34	8	58	+	+	+	+	0	0
March	4	22	74	0	0	0	0	0	0
April	0	42	58	0	+	0	0	0	0
May	0	20	80	0	+	0	+	0	0
June	0	14	86	0	0	0	+	0	+
July	0	14	86	−	0	0	0	0	0
August	0	14	86	−	0	0	0	0	0
September	0	8	92	−	0	0	0	0	0
October	0	8	92	0	0	0	0	0	0
November	0	8	92	0	0	0	0	0	0
December	4	8	88	0	0	0	0	+	+
Annual	0	16	84						

in Kendall's tau was greater than 58 in any month and averaged 84.

Six U.S. Geological Survey streamflow gaging stations in the basin were identified as having at least 75 years of streamflow record and few effects from diversions, reservoirs, or urbanization. Data from these stations reveal a positive trend for February maximum daily streamflow during the past 75 years for the more northern watersheds. This trend is similar to the simulated trend in table 6. These six stations represent only a few streams in the basin, and no other general statement can be made about historical trends in streamflow.

The effects of global or regional warming on basin streamflow cannot be determined precisely as yet, principally because of the unreliability of precipitation projections. Furthermore, streamflow changes due to climate change will be extremely difficult to separate from the large natural variability in precipitation.

Potential Changes in Drought

A monthly water-balance model of the Delaware River basin was developed to analyze the effects of climate change on drought in the basin (Tasker, 1990). In these analyses, drought was defined in terms of either the contents of the New York City water-supply reservoirs or of specified (target) streamflows of the Delaware River at Trenton. The model was developed for five major subbasins of the basin, and algorithms were used to simulate the reservoir operations (reservoir contents and releases) of each subbasin, diversions out of the basin, consumptive water use in each subbasin, and ground-water pumpage from the confined aquifers in the basin (Ayers, Tasker, and others, 1990). A stochastic model of monthly temperature and precipitation also was developed for the five major subbasins as input to each subbasin component of the monthly flow model of the basin (Tasker, 1990). The stochastic climate model was based on monthly statistics of temperature and precipitation for 1895 to 1988 from the U.S. Climate Division data base of the National Weather Service.

Sensitivity of Drought to Changes in Climate

Simulations with the monthly flow model of the basin indicate that the risk of drought is highly dependent on the changes in precipitation associated with a warming trend (Ayers, Tasker, and others, 1990; Tasker, 1990). In this analysis, drought was defined as the percentage of time (for example, 9.5 percent is 1,140 out of the 12,000 simulated months) that the New York City reservoir contents are below volumes specified by a seasonal rule curve (Delaware River Basin Commission, 1985). Current (1986) water-use and climate variability were assumed in fifty 20-year simulations. Results indicate that, under current climate conditions, the basin is in drought about 9.5 percent of the time (table 7).

Table 7. Simulated percentage of time Delaware River basin is in drought, based on New York City reservoir contents, for various changes in climate and consumptive water use

[Drought is defined as a period during which New York City reservoir contents are below levels specified by a seasonal rule curve (Delaware River Basin Commission, 1985). °C, degrees Celsius; —, no data]

Temperature change/ from current conditions	Percentage of time in drought for given precipitation change		
	No change	−10 percent	+10 percent
1986 water use			
None	9.5	—	—
+2°C	17.7	45.8	5.6
+4°C	27.2	61.1	10.3
2040 water use			
None	9.6	—	—
+2°C	17.8	46.0	5.7
+4°C	29.3	61.3	10.4

An increase of 2°C or 4°C in mean temperature increases the time in drought by 1.9 or 2.9 times, respectively. When the warming is accompanied by a 10-percent decrease in precipitation, the time in drought increases 4.8 or 6.4 times, respectively, whereas a 10-percent increase in precipitation significantly decreases the effect of the temperature increases on drought.

To place these simulated climate changes in perspective with the natural variability of the basin, a precipitation change of ±10 percent is about half the standard deviation of annual precipitation at Philadelphia, whereas a temperature change of +4°C is about 6 times the standard deviation of mean annual temperature at Philadelphia. The data in table 7 indicate that a greater sensitivity of basin drought to changes in precipitation than to temperature. Given the large uncertainty associated with climate-change projections, the scenarios in table 7 represent a hypothetical but reasonable range. Consequently, the large range in change (from a 40-percent decrease to a 540-percent increase) in time the basin is in drought exemplifies the large uncertainty that water-supply planners face regarding climate change.

Sensitivity of Drought to Changes in Consumptive Water Use

Changes in drought simulated by the monthly flow model (Tasker, 1990) were far less sensitive to the estimated (year 2040) growth in consumptive water use in the basin than to changes in temperature and precipitation (table 7). In that analysis, drought was defined by the contents of the New York City reservoirs. The effects of projected growth in consumptive water use in the upper part of the basin were small (less than a 2-percent increase in time in drought) relative to the effects of climate changes. In other simulations, in which drought was defined by a failure to meet the target flows of the Delaware River at Trenton (not shown in table 7), the effects of projected growth in consumptive water use were somewhat greater but still small (less than a 15-percent increase in time in drought) relative to the effects of climate changes.

Sensitivity of Drought to Changes in Reservoir Capacity and Use of Ground Water

Reservoirs for water supply, flood control, and low-flow augmentation have been built on several tributary streams of the Delaware River (fig. 16). About 390 billion gallons of total long-term active storage capacity is available in the basin—a volume equal to about 50 days of average flow of the Delaware River at Trenton. The three reservoirs of the New York City system (Cannonsville, Pepacton, and Neversink) in the northern part of the basin account for about 72 percent of this capacity.

Table 8. Simulated percentage of time Delaware River basin is in drought for various scenarios of reservoir use, diversions, and ground-water use under current (1948–88) climate conditions

[Drought is defined by failure to meet the target discharges of the Delaware River at Trenton, N.J., based on 1986 water-use conditions]

Scenario	Percentage of time in drought
Without reservoirs or out-of-basin diversions	5.00
With upper basin reservoirs and out-of-basin diversions, but without lower basin reservoirs	3.38
With existing reservoirs and diversions, and with:	
No increases in storage or ground-water usage	.69
Increases in ground-water usage	.63
Increases in storage in F.E. Walter reservoir	.38
Increases in storage and ground-water usage	.29

Results of other simulations with the monthly flow model (Tasker, 1990), in which drought was also defined by a failure to meet the target flows of the Delaware River at Trenton, indicate that the current system of reservoirs has improved basin water supply substantially during low-flow periods (table 8). Without the current reservoirs and diversions out of the basin, the simulated time the basin would be in drought was 5.00 percent. The presence of reservoirs in the upper basin (above Montague, N.J.), principally the New York City system, reduced the simulated time in drought to 3.38 percent. The entire system of reservoirs and diversions, as existed in 1988, reduced the simulated time in drought to 0.69 percent, a 7-fold decrease in drought compared with unregulated conditions (table 8).

The plans of the Delaware River Basin Commission to increase storage in the Francis E. Walter Reservoir (fig. 16) were simulated with the monthly flow model, and results indicate an improvement in low-flow augmentation. The simulated reservoir decreased the time in drought to 0.38 percent, a 13-fold decrease compared with unregulated conditions (table 8).

Other simulations indicate that increased ground-water pumpage from the confined Coastal Plain aquifers during drought could be an additional source to offset drought-induced water shortages, but not as effective an alternative as the increased storage in the Francis E. Walter Reservoir (Ayers, Tasker, and others, 1990). Together, however, the increased ground-water pumpage and reservoir storage decreased the time in drought to 0.29 percent, a 17-fold decrease in drought compared with unregulated conditions (table 8). The simulations were based on an increase in ground-water pumpage from the confined Coastal Plain aquifers instead of a restriction (as is currently mandated for periods of drought). The model allowed for a constant pumpage of 55 Mgal/d (30 percent less than current ground-water withdrawals) and a 4-fold increase in pumpage when the Delaware River at Trenton was below the target flow. Because ground-water recharge induced from the river lags considerably behind the onset of pumping, the simulation delayed most of the induced recharge to periods of higher runoff and hence simulated a gain in streamflow during drought conditions.

Table 9. Simulated percentage of time salt front is above river kilometer 160 or New York City reservoir storage is zero for various percentage reductions in target flows of Delaware River at Montague, N.J., for 4-degree-Celsius warming and 10-percent decrease in precipitation

	Percentage reduction in specified flows[1]						
	0	10	20	30	40	50	60
Percentage of time salt front above river kilometer 160	0.0	0.0	0.0	0.0	0.5	2.0	5.0
Percentage of time reservoir storage zero	4.0	3.0	2.0	1.5	1.0	.05	.0

[1]From current (1988) conditions.

Sensitivity of Drought to Changes in Specified Flows

Releases from major reservoirs in the basin are managed, according to a rule curve, to meet specified flows in the Delaware River at Montague and Trenton and thus prevent upstream movement of the salt front (the point in the estuary where chloride concentrations are 250 mg/L) into freshwater reaches where surface- and ground-water supplies now derive their water (Delaware River Basin Commission, 1985). New York City reservoirs are the most highly managed of the basin's reservoirs.

Sensitivity analyses were performed for changes in contents of the New York City reservoirs and position of the salt front caused by reductions in specified flows of the Delaware River at Montague. The monthly flow model (Tasker, 1990) was used to simulate the system changes that would result from a 4°C warming and a 10-percent reduction in precipitation. Simulation results indicate that the percentage of time that New York City reservoir storage would reach zero declined from 4.0 percent of the time for zero reduction in specified flows (current requirements) to 1.5 percent of the time for a 30-percent reduction in specified flows (table 9). A reduction in specified flows reduces the amount of reservoir releases needed to satisfy the specified flow and hence conserves reservoir storage. At the same time, the simulated percentage of the time the salt front intruded to river kilometer 160 or above was zero for a 30-percent reduction in specified flows and increased only with a 40-percent or greater reduction in specified flows. The position of the salt front as simulated in this monthly model is not as sensitive to reductions in specified flows as the reservoir contents are. These results indicate that the time the basin would be in drought might be substantially reduced by lowering current or future requirements in meeting target flows of the Delaware River.

Potential Effects of Sea-Level Rise

Sea-level rise is likely to accompany a global warming (National Research Council, 1984; Robin, 1986; Titus, 1986). Concerns associated with potential sea-level rise are increases in estuary salinity, increases in saltwater intrusion into coastal aquifer systems, inundation of coastal wetlands and marshes, increases in tidal and storm flooding in coastal areas, and increases in coastal erosion.

Historical and Potential Changes in Sea Level

Changes in global mean sea level computed since 1900 correlate closely with global mean surface-air temperatures (fig. 22); the correlation coefficient between temperature and sea level is 0.8 when the sea-level record is lagged 18 years behind the temperature record (Robin, 1986). The major factors contributing to sea-level rise are estimated to be thermal expansion of the ocean and melting of small glaciers and ice caps (Gornitz and others, 1982; Meier, 1984; Robin, 1986), both due to global warming. The rise in sea level of 10 to 15 centimeters (cm) over the last 100 years may indicate that global warming is occurring, but does not prove that increased atmospheric CO_2 is the cause of the warming.

Estimates of future sea-level rise, derived from the relation between temperature and sea level in figure 22, range from 15 to 30 cm of rise per 1.0°C of warming (Robin, 1986). A warming of 1.5°C to 4.5°C for doubled CO_2 concentrations could result in an estimated sea-level rise of 23 to 137 cm. Thermal expansion of the ocean and melting of small glaciers and ice caps would be the expected major contributors to the rise.

Titus (1986) qualitatively summarized recent estimates of sea-level rise with four scenarios (fig. 23). These estimates do not include local land uplift or subsidence. For example, the rate of subsidence in the mid-Atlantic coast is 10 to 15 cm per century,

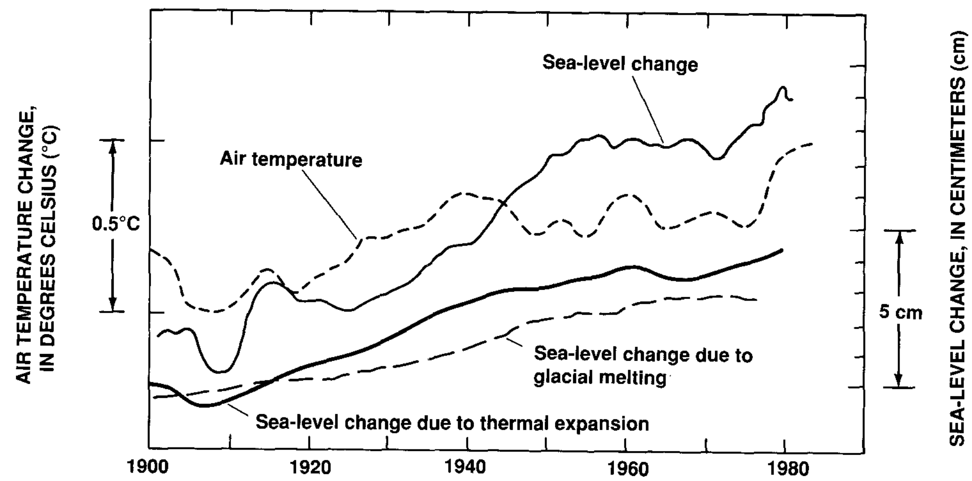


Figure 22. Changes in air temperature (Jones and others, 1982) and sea level (Gornitz and others, 1982) since 1900 and contribution to sea-level rise from thermal expansion of oceans (Gornitz and others, 1982) and melting of glaciers (Meier, 1984).

and along the gulf coast the rate is higher. The latest "consensus" estimate of rise in sea level, however, is 34±40 cm in response to a warming due to doubled atmospheric CO_2 (Meier, 1990). The consensus mean corresponds to the low to middle estimates in figure 23, but the estimates range from a rise of 74 cm to a fall of 6 cm. The large uncertainty is an expression of scientists' lack of understanding of all the processes involved. Nevertheless, the sensitivities of several hydrologic systems to the potential effects of sea-level rise are discussed below.

Effects on Estuary Salinity

By R.A. Walters and M.A. Ayers

Sea-level rise would affect conditions at the mouth of the Delaware Bay (fig. 24) and would effectively push saltwater farther upstream for a given amount of freshwater flow (fig. 25). For example, a rise of 74 cm would cause the salt front (the point in the estuary where chloride concentration is 250 mg/L) to move about 13 km farther upstream than it normally extends (Hull and others, 1986). Although this scenario is based on the highest current estimates of sea-level rise, such a shift in the salt front would have serious implications for the freshwater withdrawals by the city of Philadelphia and by many industrial users in these reaches.

The estimates of Hull and others (1986) are based on a one-dimensional model analysis of the Delaware Bay and Estuary. Walters (1992a, b) developed a three-dimensional model that indicates that a 100-cm rise in sea level would substantially amplify the tidal oscillation in the upper half of the estuary (above river kilometer 97). The model indicates that the current tidal cycle of about ±100 cm would increase to ±130 cm. This amplification causes an increase in longitudinal dispersion of saltwater in the model and a simulated increase in saltwater intrusion in the Delaware system beyond what the one-dimensional analysis indicates. At a minimum, the three-dimensional analysis indicates a significant sensitivity of the Delaware system to what would be small changes in channel geometry brought on by a 100-cm rise in sea level.

These scenarios do not consider any change in freshwater flows to the estuary. A decrease in flow would increase upstream salinity movement, whereas an increase in flow would help counteract

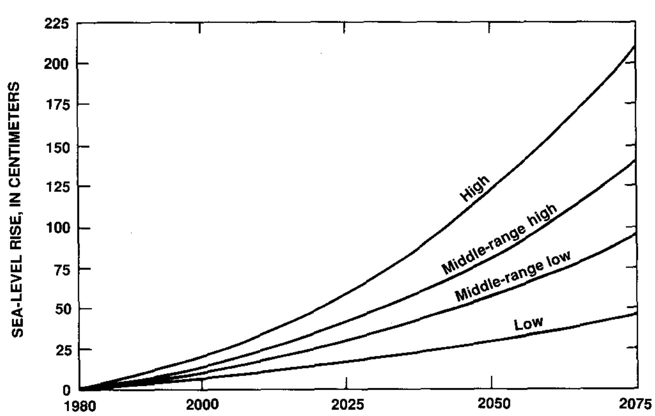


Figure 23. Projections of sea-level rise, 1980–2075 (modified from Titus, 1986).

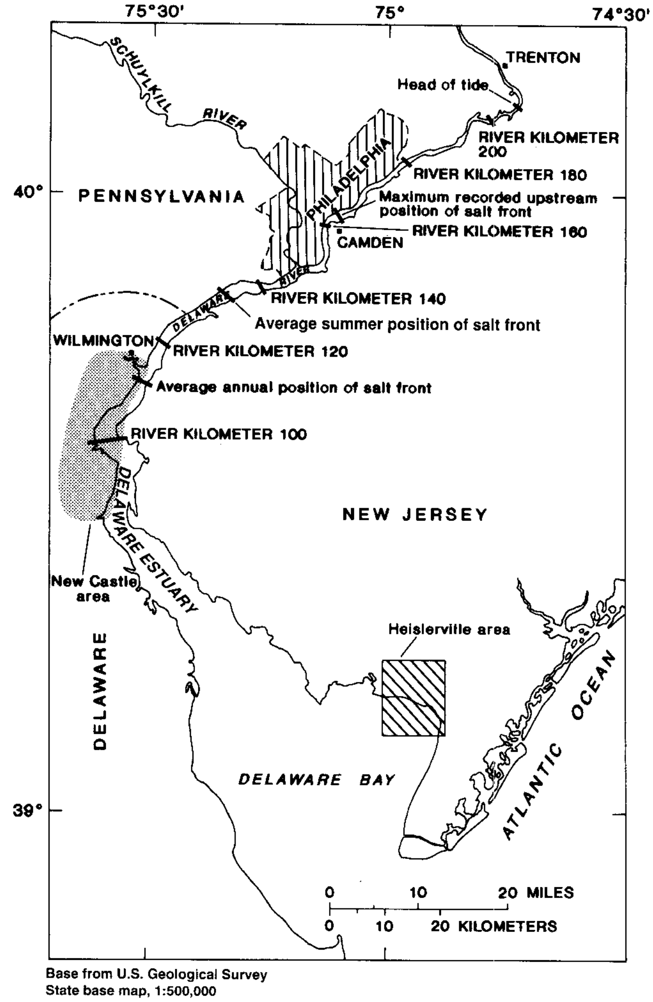


Figure 24. Location of study areas and average annual, average summer, and maximum recorded upstream positions of salt front in the Delaware Estuary (modified from Ayers and Leavesley, 1988).

upstream salinity movement due to sea-level rise. In essence, additional runoff may be needed to counteract the upstream movement of the salt front during drought conditions, and this need could amplify the shortage of water supplies in the basin during periods of drought.

Effects on Coastal Aquifers

By W.H. Werkheiser and M.A. Ayers

Any upstream movement of the salt front in the Delaware Estuary caused by sea-level rise could increase saltwater intrusion into adjacent aquifers. For example, about 70 percent of the recharge to a New Jersey aquifer system in the Camden, N.J., area currently is induced from the freshwater part of the Delaware Estuary by ground-water pumpage (figs. 16 and 24; Ayers and Leavesley, 1988). Upstream movement of the salt front would result in saltwater intrusion into previously fresh parts of the aquifer system and would seriously affect major water supplies. Aquifers already affected by saltwater intrusion could undergo accelerated rates of intrusion with a rise in sea level.

One area of concern is near New Castle, Del. (figs. 16 and 24), where a semiconfined aquifer system currently receives about 11 percent of its recharge (Phillips, 1987) from the estuary (usually saline in this reach). Saltwater intrusion moves inland along tributary streams in this area because saline estuary water flows up the tributaries during high tide. Chloride concentrations in the aquifer extend farther inland in some tributaries than in others because the silty confining unit that underlies the estuary does not extend upstream; thus, these areas receive more saline recharge than those where the underlying silty confining unit is present. Local geology is a major factor affecting saltwater intrusion now and would also determine the effect of sea-level rise on future intrusion into coastal aquifers.

A quasi-three-dimensional ground-water flow model of this aquifer system (Phillips, 1987) was used to test the sensitivity of the flow system to two scenarios of sea-level rise (W. Werkheiser, U.S. Geological Survey, written commun., May 1989). In the first scenario, a 150-cm rise in estuary (sea) level was assumed to be accompanied by no change in estuary shoreline boundary. The flow model of the aquifer system estimated that 12.5 percent of the ground-water recharge would be from the estuary—only 1.5 percent more than at present. In the second scenario, however, a 150-cm rise in estuary level was assumed to be accompanied by a new

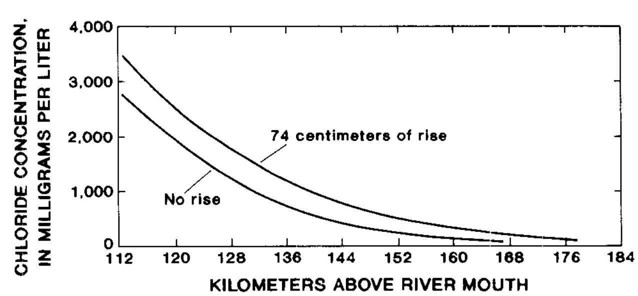


Figure 25. Effect of hypothetical 74-centimeter rise of sea level on 30-day chlorinity in the Delaware Estuary for 1960's drought-equivalent flow in the Delaware River (modified from Hull and others, 1986).

shoreline boundary resulting from the subsequent estimated inundation. The flow model of the aquifer system estimated that more than 30 percent of the aquifer recharge would be from the estuary—almost triple the recharge amount under present conditions or for the first scenario.

The modeling results indicate that this aquifer system is more sensitive to inundation than to sea-level rise alone because the absence of a silty confining unit in areas inundated by a rise in estuary levels allows greater recharge of saline water. Werkheiser (oral commun., 1990), however, tried to verify the extent of current inundation with field reconnaissance and existing maps of the study area. He concluded that the lack of detailed topographic maps (with a 30- to 50-cm contour-interval range) and detailed description of the silty confining unit limited confidence in estimates of current inundation or of potential inundation due to the 150-cm rise. Hence, accurate estimates of either the current amount of recharge or the change in recharge that might result from sea-level rise are practically impossible because of a lack of detailed topographic and geologic data in the study area. No doubt such data are lacking in many other coastal areas.

Effects on Coastal Wetlands

An area along the Delaware Bay near Heislerville, N.J. (fig. 24), illustrates the effects that sea-level rise could have on coastal wetlands. For illustration purposes, a sea-level rise of 150 cm and a new shoreline near the 150-cm contour interval are assumed. The mapped difference between the present and new shorelines (fig. 26) indicates that large areas of tidal wetlands and marshes could be inundated and presumably lost unless they can migrate inland with the rise. The State of New Jersey estimates that 30 to 80 percent of these important ecosystems in the estuary could be lost should a rise of 150 cm occur. Because these systems serve as important filters of sediment and provide habitat and food for a wide range of organisms, a decrease in the extent of coastal wetlands would harm estuarine and coastal water quality, fisheries, and ecology.

Effects on Coastal Flooding and Erosion

Sea-level rise also could increase coastal flooding. Tidal or storm-induced flood surges at an increased sea level would cause water to reach higher elevations and inundate larger areas than at present. Near Heislerville, for example, an assumed sea-level rise of 150 cm shows that the estimated new 100-year flood elevations would result in more extensive inundation of developed areas (fig. 27). In this and other more developed coastal areas, sea-level rise will result in more frequent inundation of areas that were developed without considering the potential for sea-level rise.

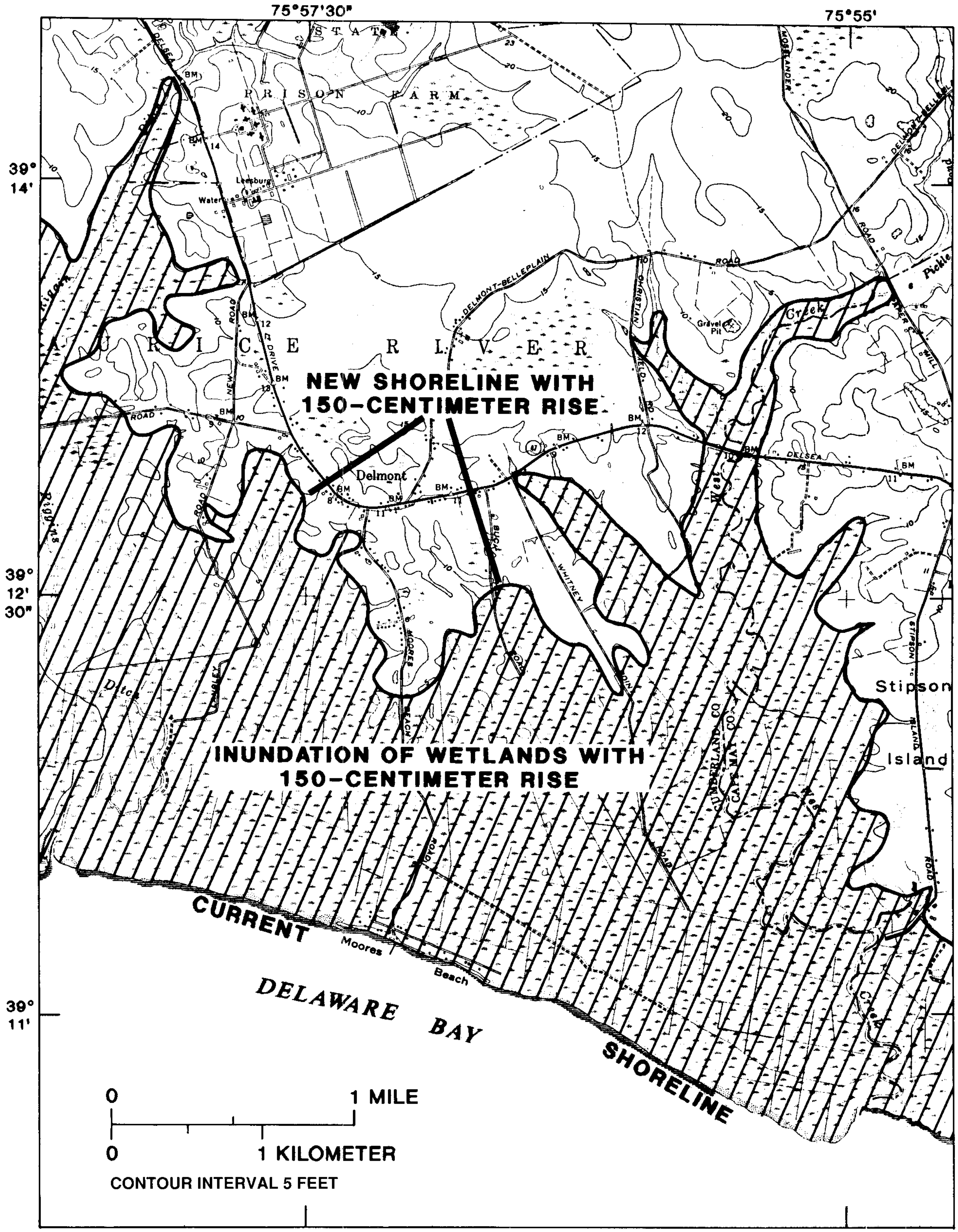
The geomorphic instabilities brought about by a rising sea level, and the resulting changes to shorelines and increased frequency of flooding, likely will result in increased coastal erosion. Coastal erosion already is a problem along much of the Atlantic and gulf coasts.

SUMMARY AND CONCLUSIONS

The greenhouse effect is vital to the support of conditions favorable to Earth's life systems. Atmospheric concentrations of CO_2, a greenhouse gas, are expected to double in the next century and thereby reach levels that probably have not existed on Earth in more than 1 million years. An increase in atmospheric concentrations of CO_2 is expected to cause global warming, increases in sea level, shifts in regional precipitation patterns, and changes in the transpiration efficiencies of plants.

Detecting CO_2-induced warming in climate records has been difficult because the records are suspected to contain errors resulting from changes in observers, site measurements, measurement techniques, and local land use. For example, as much as 74 percent of the 0.5°C warming estimated from the records for the last 100 years might be due to the heat-island effect.

Numerical simulation models of the global climate system, referred to as general circulation models (GCM's), are the most advanced approach used to evaluate the effects of increasing atmospheric CO_2 and other greenhouse gases on climate. Representations of climate feedbacks of energy and moisture are critical to the overall performance of GCM's. 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. GCM's indicate that a doubling of atmospheric CO_2 concentrations will induce a global atmospheric warming of 1.5°C to 4.5°C. Large differences among GCM estimates of regional temperature and precipitation, however,



Base from U.S. Geological Survey
Heislerville, 1:24,000, photo revised 1972

Figure 26. Estimated extent of wetland inundation near Heislerville, N.J., that would result from sea-level rise of 150 centimeters.

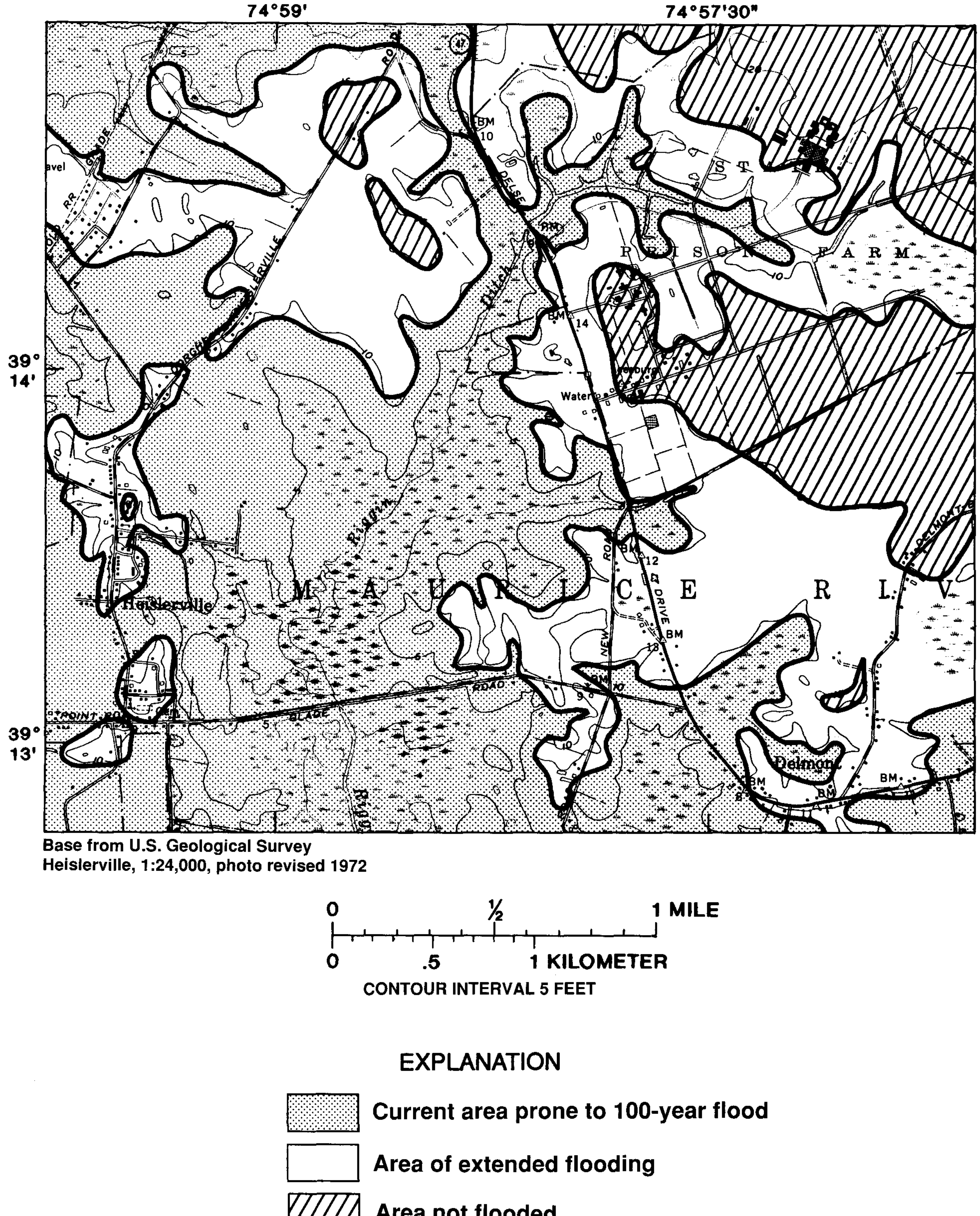


Figure 27. Estimated increase in extent of 100-year flood near Heislerville, N.J., for sea-level rise of 150 centimeters.

exemplify the great uncertainty associated with GCM projections of climate change. The GCM outputs, however, are valuable in setting ranges of potential changes to test hydrologic sensitivities.

GCM results indicate that temperature and precipitation under doubled-CO_2 conditions cause a reduction in Thornthwaite moisture indices, implying drier climate in the Delaware River basin. The amount of decrease depends on the GCM used. Simulation using GCM projections and a hypothetical 4°C increase in temperature both resulted in simulated increases in potential evapotranspiration and decreases in the Thornthwaite moisture indices across the United States. Decreases in the moisture index were greatest in cool, wet regions (Pacific Northwest, Great Lakes, and New England States) and least in hot, dry regions (Southwestern States). Results also show that natural variability strongly influences the detectability of the effects of the temperature increase on the Thornthwaite moisture index.

Simulated changes in mean annual temperature had a greater effect on mean annual irrigation demand for the southern part of the Delaware River basin than did changes in mean annual precipitation. Increased stomatal resistance to transpiration, however, counteracted the effects of increases in temperature and decreases in precipitation.

Daily simulations of basin streamflow indicate that climate has the strongest effect on maximum daily flow, mean daily flow, evapotranspiration, and 7-day low flow. These results indicate that 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.

Three categories of uncertainty—direct CO_2 effects on change in stomatal resistance of vegetation to transpiration, indirect CO_2 effects on change in precipitation, and natural variability in climate—were shown to have about the same magnitude of effect on streamflow predictions. As workers improve their projections of precipitation that accompanies increasing atmospheric CO_2 concentrations, the uncertainty in forecasts of effects on streamflow will be reduced accordingly. Furthermore, as research progresses on how increasing atmospheric CO_2 affects vegetation and transpiration, the uncertainty associated with stomatal resistance will decrease. The uncertainty in forecasts due to natural variability, however, probably is irreducible and will continue to make changes in streamflow due to global warming difficult to distinguish from natural variability.

Simulations of daily and monthly streamflow indicate that a transient warming trend would increase the proportion of winter precipitation that falls as rain, increase winter runoff, reduce snow accumulation, and reduce spring runoff in the northern part of the basin. A warming of 4°C could increase basinwide evapotranspiration and reduce annual basin runoff by as much as 25 percent if current precipitation patterns continue. An increase in precipitation of about 3 percent would be needed to counteract decreases in streamflow that would result from each 1°C of warming.

Simulation results indicate that the Delaware River basin is in drought under current climate conditions about 9.5 percent of the time. An increase of 2°C or 4°C in mean temperature multiplies the time in drought by 1.9 or 2.9, respectively. When the warming is accompanied by a 10-percent decrease in precipitation, the time in drought increases 4.8 and 6.4 times, respectively, whereas a 10-percent increase in precipitation significantly offsets the effect of the temperature increases on time in drought. Basin drought was most sensitive to changes in precipitation and less sensitive to increases in temperature, reservoir capacity, groundwater pumpage during drought, and consumptive water use, in that order. Sensitivity analysis results further indicate that the time the basin would be in drought might be substantially reduced by lowering current or future requirements in meeting target flows of the Delaware River (based on the position of the salt front as simulated in a monthly model).

Global warming will likely cause a rise in sea level through ocean thermal expansion and melting of small glaciers and ice caps. Sea level has risen 10 to 15 cm over the last 100 years. According to current consensus, an estimated future sea-level rise of 34 ± 40 cm would occur in response to a warming caused by a doubling of atmospheric CO_2, excluding effects of land-surface subsidence.

A sea-level rise would cause the salt front in the Delaware Estuary to move upstream. For example, a sea-level rise of 74 cm would cause the normal position of the salt front to move upstream about 13 km, which would affect surface-water supplies and could increase saltwater intrusion into adjacent aquifer systems. Analysis of the effects of sea-level rise with a three-dimensional estuary model indicates that the position of the salt front is very sensitive to small changes in channel geometry.

Analysis of the effects of sea-level rise with a quasi-three-dimensional ground-water flow model of an aquifer system in New Castle, Del., indicates that changes in recharge of saltwater to the aquifer were sensitive not so much to the rise itself as to the area inundated by the rise. Clearly limiting the analysis was a lack of detailed (30- to 50-cm contour-interval) maps of coastal areas, which would be needed to define the areas of inundation accurately.

A sea-level rise could result in loss of large areas of tidal wetlands and marshes, unless these features migrate landward with the rise. In many coastal areas, a sea-level rise would increase the frequency of flooding in areas developed without considering rising sea level and would increase coastal erosion, which already is a problem in many areas.

Distinguishing CO_2-induced changes in climate from natural variability and measurement error is difficult. Furthermore, GCM projections of regional changes are inconsistent and, therefore, considerably lacking in certainty. Results of sensitivity analyses of climate-change effects on water resources described herein, however, have serious implications for future availability of water resources for some climate-change scenarios. Unless more reliable estimates of future precipitation amounts and variability are obtainable along with global warming estimates, the future direction of changes in basin water resources cannot be determined precisely.

Current resource planning and protection policies could either preserve or endanger the future availability of water-related resources. Mitigative measures for one resource may not be beneficial for other resources. For example, in response to sea-level rise, building a dike around a well field or other cultural feature might inadvertently prevent the migration of a wetland system.

Resource planners and managers face many uncertainties in choosing appropriate estimates of climate change to use in resource planning. The best that planners and managers can do is to develop a better understanding of the interactions of the resources involved and, through use of either conceptual or deterministic system models, to define the systems' sensitivity to a range of potential changes in climate. Contingency plans for each scenario of climate change could define how to account for, evaluate, and respond to each scenario. Then, as confidence is gained about potential changes in temperature, precipitation, and sea-level rise, new change estimates could be applied to system models, and contingency plans could be refined.

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