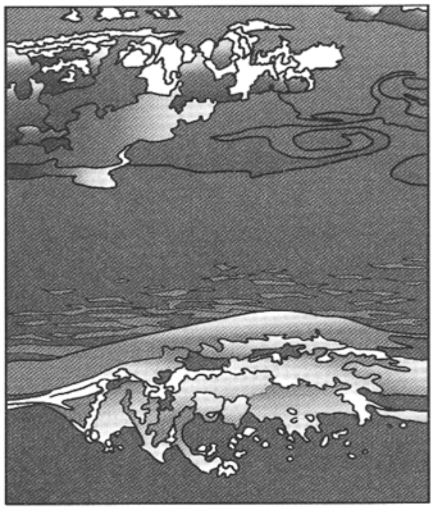
CHAPTER 6



Modeling the Atmosphere–Ocean System

Key Questions

- What are Global Climate Models and why do we need them?
- What types of variables do Global Climate Models take into account?
- What can Global Climate Models tell us about Earth's climate?
- What affects the accuracy of Global Climate Models?
- Can we rely on what Global Climate Models tell us about the greenhouse effect?

Chapter Overview

In this chapter we continue the discussion of numerical models that we began in Chapter 3. Three-dimensional Global Climate Models (GCMs) are the primary tool used for projecting potential future climate changes that may occur from increasing concentrations of atmospheric greenhouse gases. We begin by presenting a brief history of the development of GCMs. We then describe, in very general terms, how these models are constructed and what information these models include. In particular we focus on the vast array of variables these models must take into account to reflect the complexities of the earth system. Once we know what variables they account for, we can begin to understand how the models produce reasonable simulations of the present-day climate. We then illustrate the way these models have been used for climate change experiments. Finally, we discuss some of the limitations of the models and assess what they can and cannot tell us about our possible future.

Why Numerical Models?

We have talked in earlier chapters about systems and the way we can study them. We have described Earth's energy budget, the latitudinal distribution of net radiation, and the resulting circulations of air and water over Earth's surface. We have seen how some very basic physics allows us to make deductions of how the climate system should operate, and we have seen how these deductions are borne out by climate observations. From this it is possible to account for the very broad-scale distributions of temperature and precipitation across the planet and explain how they vary geographically and seasonally. While there is still undoubtedly much more to learn, we have a reasonably good knowledge of the physical and chemical processes at work in our atmosphere and oceans. None of this knowledge, however, allows us to predict how the system will change in the future.

To be able to predict the future, we need a model of the system. In our simple systems model (Figure 6-1) we need to be able to change the input, or some part of the process, and see how this plays out to affect the climate or the model output. We can construct a variety of different types of models. The systems diagrams you have seen,

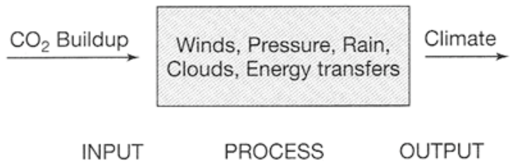


FIGURE 6-1

Simple system model for CO_2 build-up and climate.

and will see, throughout the book are *conceptual models.* A conceptual model is a qualitative description of a system that provides some understanding of how the system works but does not allow one to describe its detailed behavior. An example of such a model is the Daisyworld system described in Chapter 2. The feedback diagrams shown there allow us to find the stable and unstable equilibrium points in the system, but they do not allow us to calculate the actual planetary surface temperature or how the system would evolve in time as solar luminosity changes. This kind of detail can be added to the model, the results of which are shown in Figure 2-14. Doing so, however, requires the use of mathematics.

Consider also the ice–albedo feedback described in Chapter 3. This is an important feedback in the Earth system so you will see this diagram again in later chapters. It is simple to see that if we increase temperatures we reduce the ice cover, and we reduce the albedo, which further increases temperatures—a positive feedback. But by how much? Unless we can quantify those linkages in some way, we can't really tell what will happen. Will just a slight change in temperature cause a change in ice cover? Is there some threshold that has to be overcome first? If we change the temperature from −40°C to −38°C over the central Arctic, presumably this will not have the same effect as changing the temperature from 0°C to +2°C at the ice margin. We can quantify these linkages statistically. That is, we can derive an empirical (observable) relationship between the system components—an *empirical* or a *statistical model.* A statistical model involves mathematics but in a limited way that is based on observations rather than fundamental physical principles. Based on observations we could discover, for example, that the sea ice edge advances or retreats 100 km for every 1°C temperature change (remember from Chapter 4 that sea ice is the thin layer of ice that forms over the polar oceans).

This helps, but it is still not enough. Other feedbacks come into play (Figure 6-2). If the ice retreats 100 km it will expose more ocean surface. The ocean is usually warmer than the air, so there will be more heat transferred from the ocean to the atmosphere. We could consider that there will be more evaporation from the ocean, so atmospheric water vapor (a greenhouse gas) will increase. Maybe this will result in more clouds, which both cool the system because of their albedo and warm the system because they absorb longwave radiation (see Chapter 3). What will be the net effect of all these changes? It would be impossible to develop a simple conceptual model or a statistical model that can answer the question. As we broaden the scope to look at the whole climate system we realize that there are hundreds of different variables interacting in many interconnected systems that may operate differently over different parts of the globe. To build a statistical model you would need observations of every possible combination and change, and if such an incredible database existed, we still couldn't answer the question of what the future may hold. The past climates for which data could exist may not include the climate that may develop in the future.

The only realistic possibility for predicting the climate future is to construct a numerical model based on the physics of the system—a model that takes into account all of the processes and feedback loops that we know about. If we can build such a model, and if it produces a realistic simulation of the present climate, then

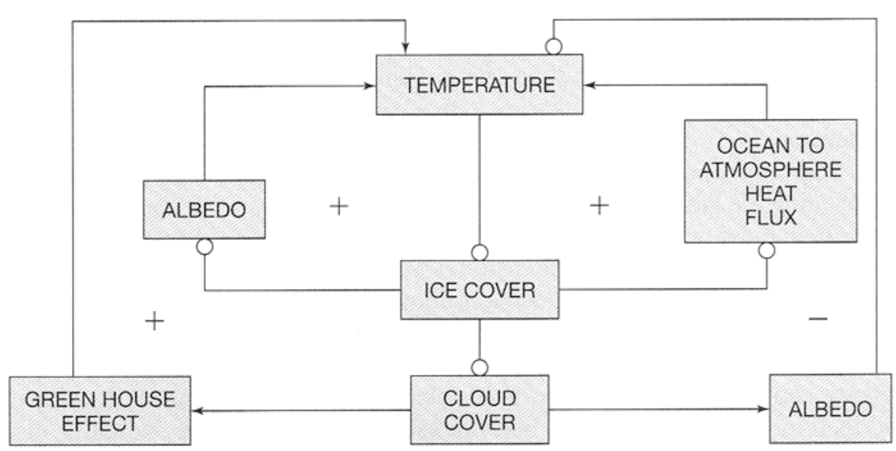


FIGURE 6-2

Feedback diagram for ice-ocean-atmosphere interactions at the ice margin.

we can change the model slightly (e.g., by increasing the greenhouse gas composition of the atmosphere) and expect it to provide a reasonable approximation of how the climate system will respond to the slight change we introduced. In the next section, we introduce models that do so, discuss how they are constructed, and show how we can use them to learn more about the future.

General Circulation Models

What Variables Are Accounted For in the Model?

Chapter 3 described a one-dimensional, radiative-convective climate model (RCM), which took into consideration the vertical energy exchanges in the atmosphere and at the surface. Such models average the incoming and outgoing energy fluxes over the entire globe. They are useful for computing the globally averaged greenhouse effect, but they have nothing to say about temperature variations between different parts of the surface. These models can be made somewhat more sophisticated and realistic by averaging conditions around a set of different latitude bands (e.g., 0–10° N, 10–20° N, 20–30° N, etc.). We could do that for all latitude bands and have a model that not only includes vertical energy exchanges, but also energy transfers between adjacent latitude bands, so that we now have a two-dimensional model (height and latitude). Such models are often described as **energy-balance climate models** (EBMs). They are useful because they allow for additional feedbacks that cannot be calculated in the one-dimensional models (the RCMs). One use of these models is to examine cloud feedback processes by allowing clouds to vary as a function of latitude. They can also be used to explore the ice–albedo feedback referred to above—again by having the albedo change as ice cover extends further equatorward or retreats poleward. You will see the results of some energy balance climate models in the discussion of "Snowball Earth" in Chapter 12. These EBM models cannot be used to make detailed predictions about modern climate change, however, because the atmospheric winds and waves that are responsible for latitudinal energy transfer are inherently three-dimensional, not two-dimensional. If we want to make accurate predictions of future climate, we need to move up to three dimensions.

Three-dimensional climate models take the next step of dividing the latitude bands up into longitudinal blocks as well. In these models you can imagine the surface of the globe divided into a two-dimensional array of longitude/latitude cells, with the atmosphere above each cell divided into discrete layers so that the atmosphere is divided into a three-dimensional grid of boxes. Using these models we can now do the same sorts of calculations as described in Chapter 3, but each box can exchange energy with the boxes above and below, and on all four sides (east–west and north–south). We can include not only the effects of different latitudes, but we can also take geography into account. Whether, for example, we are over land or water, mountains or plains, deserts or tropical forests, has an effect on the local climate. Putting all of these together allows for a much more realistic simulation of the global climate and how it is distributed.

Early versions of these three-dimensional models were mainly atmospheric models: Their purpose was to simulate the primary fields of motion in the atmosphere and they were referred to as **General Circulation Models** (GCMs). Similar developments then took place in efforts to model the circulation of the oceans. To differentiate between the two types of models we labeled them *Atmospheric General Circulation Models* (AGCMs) and *Ocean General Circulation Models* (OGCMs). When they were later combined into a single model it became an *Atmosphere–Ocean General Circulation Model* (AOGCM). Then, as if to confuse everybody, as these models became even more sophisticated and attempted to capture the whole climate system (we talk further about this later in the chapter), they were again referred to as **Global Climate Models** (GCMs)! In this book we use the names AOGCM and GCM interchangeably. If a model includes only the atmosphere or only the ocean, we will refer to it as an AGCM or an OGCM.

Figure 6-3 gives an impression of the information that is included in the model and the types of calculations made by the model. You can imagine that solar radiation enters a grid box at the top of the atmosphere. The model calculates how much is reflected back to space by air molecules, how much is absorbed in the cell, and how much is transmitted down to the next level. The same calculations are made as the radiation is transmitted down through each level until it reaches the surface. There, radiation is either reflected back upward, or absorbed. The absorbed radiation heats the surface, which then emits longwave radiation upward as a function of its temperature. This is an application of the Stefan-Boltzmann law again (Chapter 3). Some of this radiation will be absorbed by clouds, water vapor, and other greenhouse gases, which raises the temperature of the atmospheric grid boxes. Each box will, in turn, emit longwave radiation both upward and back down toward the surface (i.e., the greenhouse effect).

The radiation absorbed at the surface also results in evaporation and the transfer of *sensible* and latent heat upward into the atmosphere. Recall from Chapter 3 that latent heat is the energy associated with phase transitions of water (ice to liquid and liquid to vapor). **Sensible heat** is the energy associated with the thermal motion of air molecules (which is to say with air temperature). As the model develops differences in temperature, vertically or horizontally, this causes changes in air density that result

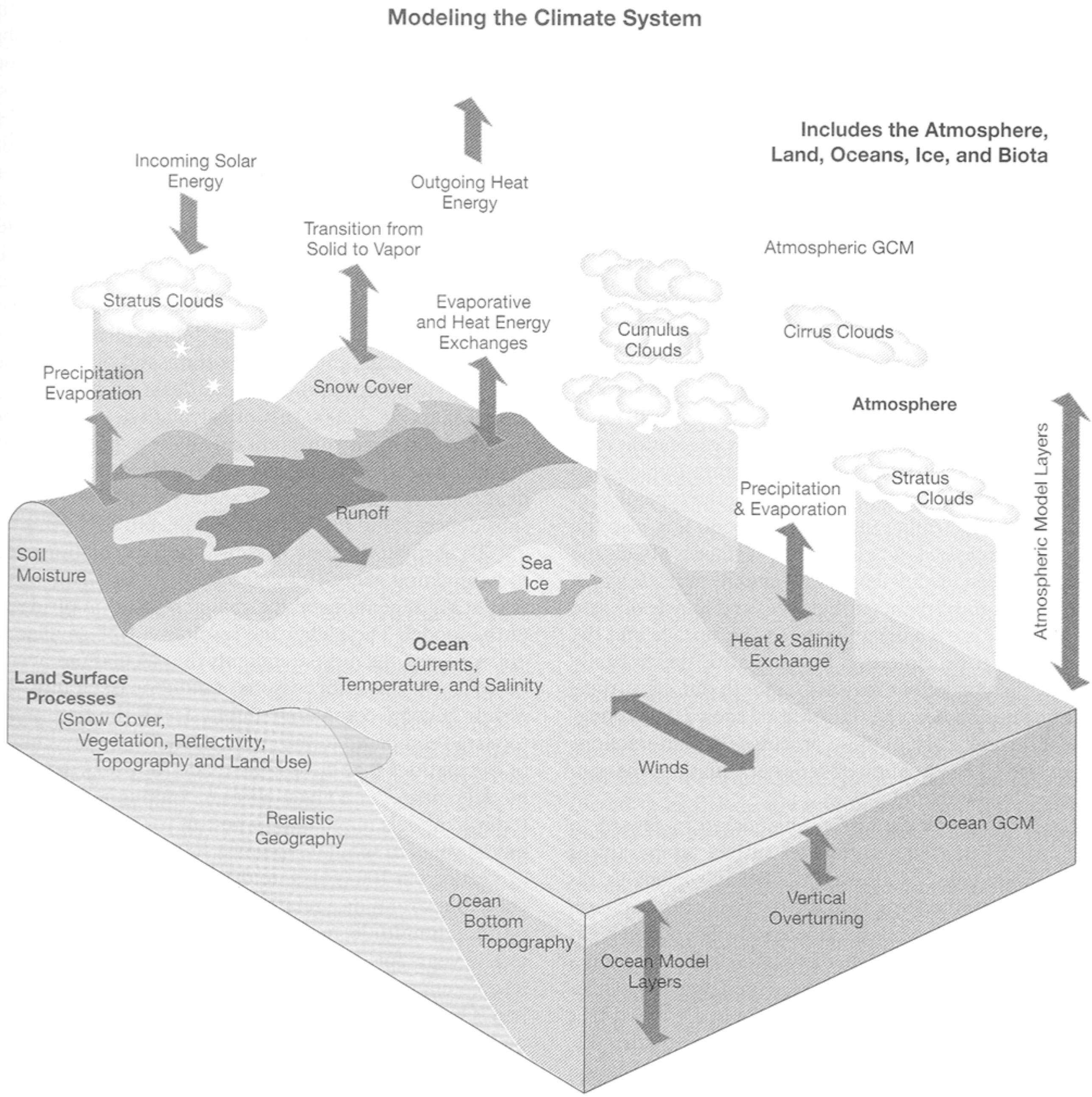


FIGURE 6-3

Some of the processes included in Global Climate Models. (From the National Assessment Synthesis Team, *Climate Change Impacts on the United States: The Potential Consequences of Climate Variability and Change,* U.S. Global Change Research Program.) (*www.usgcrp.gov*)

in the movement of air between adjacent boxes on each side (horizontal winds) or vertically (uplift or convection, and subsidence). As the air moves between these boxes it carries with it energy, mass (including water vapor and aerosols), and momentum. Depending on the temperature and humidity of the box, the water may stay as a gas or it may condense to form clouds. The clouds produce rain that falls through the atmosphere. It may evaporate as it falls or reach the surface as precipitation, which will occur as rain or snow depending on the temperature. The changing characteristics of the box (in terms of water vapor, aerosols, temperature, clouds) all change the radiative properties of the box, which further changes the transmission, absorption, and reflection of the solar and terrestrial radiation fluxes and the transfer of sensible and latent heat.

All of this sounds complicated enough, but present-day Global Climate Models are much more sophisticated than this. The variables seem endless. The surface cells can be land or water. If they are land cells they have an elevation, so the model takes into consideration the surface topography, as well as soil and vegetation cover. When rain falls on the surface, some evaporates, some infiltrates the soil, and some runs off over the surface. How much depends on the soil characteristics, the surface temperature, how much water is already present, and the surface slope. The soil is divided into layers, allowing for water and heat to be transferred up and down in the soil and for water to flow laterally between adjacent boxes below the surface. Each model grid cell has a characteristic surface cover that includes a predominant vegetation type.

The vegetation type determines how deeply roots penetrate the soil (which influences water infiltration rates and evapotranspiration), and each vegetation type has a characteristic stem and leaf structure (which also affects evapotranspiration). In addition, the type of vegetation helps determine the surface albedo and also the surface roughness. How "rough" the surface is affects the friction between the low-level wind and the surface, which influences air movement (turbulence) and affects the rate at which heat is exchanged between the surface and the atmosphere. If precipitation falls as snow, this not only changes the albedo, but may also smooth out the surface as the vegetation becomes covered by snow (thus changing heat fluxes). Each of these processes and interactions is described by an equation or series of equations that can be solved by the model.

Early models were less sophisticated in regard to oceanic data and used observed sea surface temperatures for the ocean cells, or allowed the ocean surface to exchange water and energy with the overlying atmosphere. In this case, many models included a 50–65m mixed layer that allowed for energy exchange at the surface and some seasonal heat storage. Some of these models allowed heat to be transported across adjacent cells in a pattern that matched the present-day distribution of ocean currents. Today, Global Climate Models include ocean circulation models that are as detailed as the atmospheric component. The ocean is divided into layers. Energy, mass, and momentum are exchanged at the ocean surface. The wind drives ocean currents and mixes the surface layers, and the model tracks temperature and salinity changes that determine water density and drive the deep ocean (thermohaline) circulation. There is convergence and divergence resulting in downwelling and upwelling water that connects the deep ocean and the surface layers. Sea ice forms where the surface ocean temperature is at or below the freezing point. Again, within each cell, the ice is divided into layers, heat is transferred through the ice, and the ice thickens and thins through the seasons as the water and air temperatures change.

This still does not give a full description of all the processes and calculations that are included in a Global Climate Model. Hopefully, however, this description does illustrate the point that these are very sophisticated models that take into account all of the processes described so far in Chapters 3, 4, and 5, plus others that go beyond the level of detail we have discussed in the text. Let's think about the number of calculations involved in running such a model. A model with a $2.5° \times 2.5°$ longitude/latitude grid size will have over 20,000 cells distributed around the globe. If we divide the atmosphere into 20 layers, that gives 400,000 grid boxes in which all of the atmospheric calculations have to be made (including all of the transfers with the adjacent cells). Six thousand of the surface cells are over land and include at least a five-layer soil model (another 30,000 sets of calculations). Remember that there are many different calculations going on within and between each cell: calculations of radiation fluxes, heat transfers, momentum transfers winds, convection, subsidence, cloud formation, rainfall, and so on.

To account for time passing, these calculations need to be updated every 10 to 30 model minutes (the model time step) depending on the spatial resolution (grid size) of the model. The higher the resolution (the smaller the grid squares), the more frequently the calculations need to be updated. This is for reasons of numerical stability—the results become completely unphysical if the time step is too large compared to the grid spacing. There are millions of calculations that have to be made for every day of the model climate, and given current computer capabilities, today's models may need to be run for several hundred model years to simulate the past 100 years of climate and projecting out 100 years into the future. The latest *parallel* computers, running a relatively high resolution $2.8° \times 2.8°$ latitude/longitude model, may simulate three or four model years for each 24 hours of computing time. Parallel computers run a task on multiple processors—the problem is broken down into pieces and each processor works on a separate piece. The world's more powerful and expensive parallel computers with 32 processors might take 50 days of continuous processing to do a single model run. As you might expect, there are only a relatively small number of these models in the world. They are extremely expensive to develop, and costly to run. The Yokohama Institute for Earth Sciences took delivery of the world's fastest supercomputer in March 2002. With 5,120 processors, the computer is expected to achieve peak performances of 40TFLOPS (40 trillion floating-point operations per second). This computer is being dedicated to developing an "Earth Simulator" and the system is currently testing new atmospheric and oceanic circulation models.

There is one more wrinkle we must discuss here. In the past, these models were constructed essentially the way we described above—by dividing the globe into boxes and calculating all of the transfers across the adjacent cell boundaries. All of the model equations were expressed as finite difference equations; that is, the transfers of energy, mass, and momentum between boxes were calculated as some function of the difference in various quantities (e.g., temperature) between the boxes. These models are, therefore, referred to as **finite difference models**. Because the atmosphere is unbounded at the sides (i.e., it is continuous around the globe) it is possible to express all of these equations very differently, as wave functions. In a model that uses wave functions, the number of waves resolved determines the spatial resolution of the model—the equivalent to the grid size in the finite difference models. These models are referred to as **spectral models**. Most current GCMs are spectral models; their output, however, is still usually presented as a gridded product.

Using Models for Climate Experiments

For the rest of this chapter we focus on learning more about conducting climate change experiments with these models, but it is worth noting that GCMs have many other uses besides projecting greenhouse warming. These types of models also allow us to test our understanding of how the climate system works. We can perform analyses of climate sensitivity by changing some of the parameters and seeing what effect it has on the outcome. For example, we can have a fixed cloud cover or let the cloud cover vary to see how important clouds are (we'll talk more about this further on). We can change the vegetation cover, what happens to global climate if we replace the Amazon rainforest with grassland? You'll find out in Chapter 18. In Chapters 4 and 5 you saw how mountain chains and the distribution of land and ocean affects climate, and in Chapter 7 you will see how the distribution of continents has changed with time. We can enter previous land–ocean distributions into the GCM and determine what effect this had on past climates (*paleoclimate modeling*), or we can model the effects of future and past volcanic eruptions on climate (Chapter 15). The science and research applications are numerous, and we use versions of these models everyday for weather forecasting.

The first attempts at numerical climate modeling did, in fact, begin as experiments in weather forecasting. Lewis Fry Richardson, a British ambulance driver in France during World War I, spent his off-duty time solving the basic equations of atmospheric motion using a mechanical calculator. His objective was to develop a numerical weather forecast for parts of Europe. He published the results in 1922 but, as you might expect, there was little further progress in numerical modeling until the development of electronic computers in the 1940s. Weather forecasting was indeed one of the first challenges worked on with the new computer at the Princeton Institute for Advanced Studies in the late 1940s. Today, almost all weather forecasts are made with numerical models of one sort or another.

Efforts to transform weather prediction models into climate models began in the late 1950s, and the development of ocean circulation models began in the 1960s. Since then, Global Climate Models have steadily grown in size and complexity in response to the ever-increasing computing power that has become available. Figure 6-4, taken from the Third Assessment Report (TAR) of the International Program on Climate Change (IPCC), published in 2001, shows the history of climate model development. It is interesting to note that different parts of the model were developed independently and later incorporated into the GCMs. We see that these models now include atmosphere, ocean, land, and sea ice components, as well as various aerosol and carbon cycle models. Looking ahead, the next developments will likely be the incorporation of dynamic vegetation and atmospheric chemistry models (i.e., as climate changes and those changes result in land use and vegetation changes, these, in turn, will modify atmospheric chemistry and surface characteristics, which will further change the climate).

Originally, researchers typically ran GCMs by inputting present conditions such as the solar constant, the rotation rate of the Earth, the distribution of continents and oceans, topography, vegetation cover, and gas composition of the atmosphere (these fixed conditions are referred to as the model **boundary conditions**) and then putting the model into motion until it reached an equilibrium state. This involved running it for about 15 years (model time) until the model developed a stable climate (in which the mean climate didn't change with time) and then running it for 15 to 30 more years to obtain some measure of the climate variability. Once the models reached the steady state point, scientists would change one of the boundary conditions (e.g., the concentration of greenhouse gases in the atmosphere) and run the model again. The difference in the two model runs indicated the climate change that would be expected for the boundary condition, in this case the change in greenhouse gases. Most experiments doubled or quadrupled the greenhouse gas concentrations. Because the models are run out to equilibrium conditions each time before boundary conditions are changed, these are referred to as **equilibrium climate experiments**.

Equilibrium Climate Change Experiments Early atmospheric GCM experiments during the 1970s held the cloud cover constant (the observed global cloud distribution was set as a boundary condition of the model). A climate change experiment was then conducted to see what the effect of four times the (then) present level of atmos-

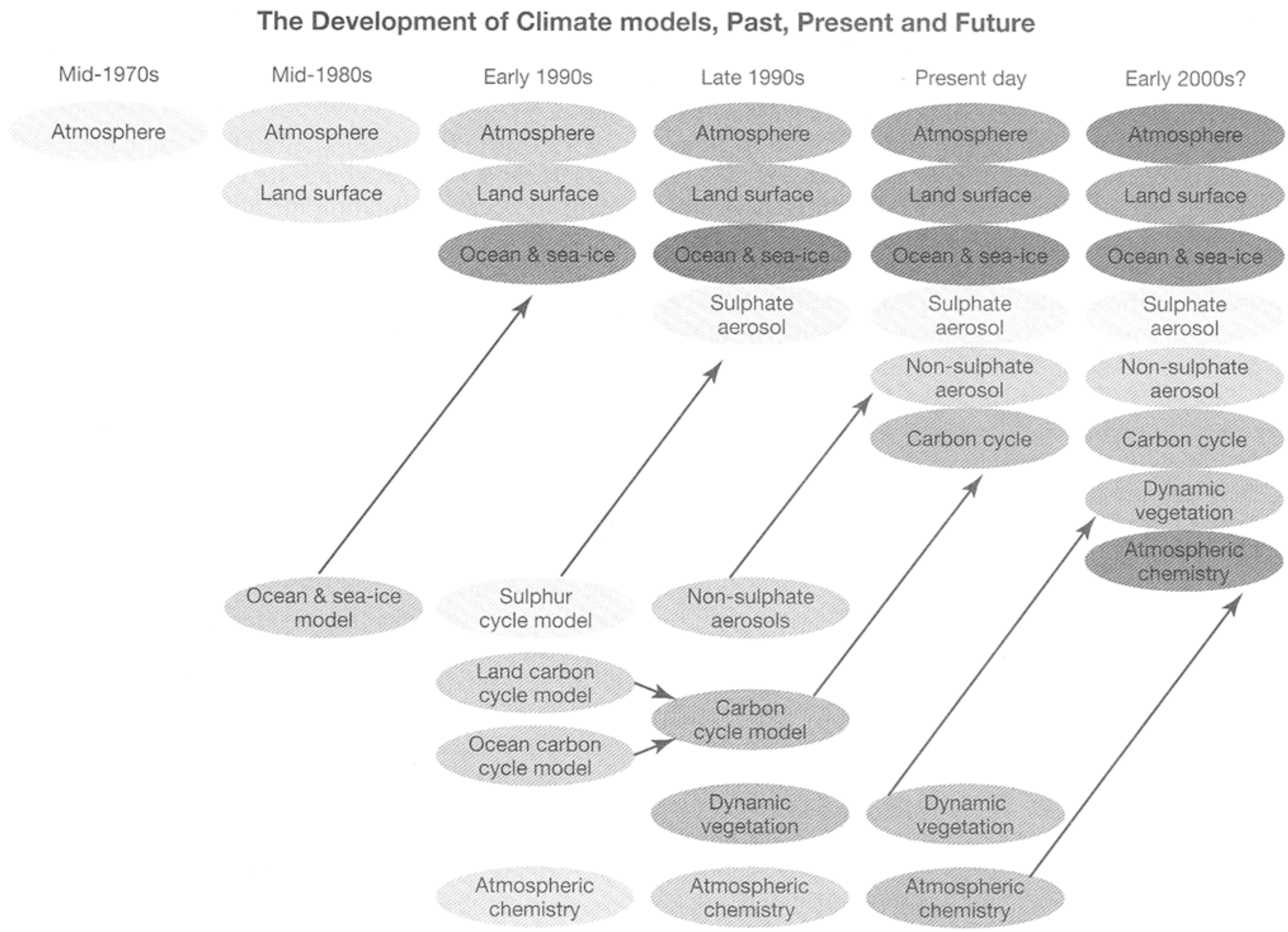


FIGURE 6-4

History of climate model development. (From the Intergovernmental Panel on Climate Change, Climate Change 2001, *The Scientific Basis: Technical Summary of the Working Group I Report.*) (*http://www.grida.no/climate/ipcc_tar/wg1/010.htm*)

pheric CO_2 would be on the climate. The result was a 4°C increase in global temperature. A later version of the same model with variable cloud cover produced a 4°C mean global temperature increase when the amount of atmospheric CO_2 was only doubled. The difference between the two experiments illustrates the important role that clouds play in the climate system and further illustrates the need for numerical climate models. One might expect that surface warming due to increased levels of atmospheric CO_2 would increase evaporation and increase cloud cover. You learned in Chapter 3 that clouds act to both warm and cool the system, and in the present climate these two effects almost cancel each other out—with the cooling (albedo) effect being slightly greater than the warming effect. One might think, therefore, that increasing cloud cover might have little impact, or might enhance slightly the negative feedback that dominates at present. In fact, the results were exactly opposite to this—the clouds had a strong positive feedback that enhanced the CO_2 warming. The increased temperatures produced deeper (higher) moist convection, which produced an increase in the amount of high-level cirrus clouds. You will also remember from Chapter 3 that the greenhouse effect for high clouds is larger than the albedo feedback, and so high clouds cause surface warming.

A number of model experiments were carried out in the 1980s with model resolution varying from 5° × 7.5° to 8° × 10° latitude/longitude grids, variable cloud cover, a mixed-layer ocean, and a prescribed oceanic latitudinal heat transport. These models were all equilibrium models that produced mean global temperature increases of approximately 2.5°C to 5°C when atmospheric CO_2 was doubled. Toward the end of this time period there were also several higher resolution model runs that produced temperature changes of about 3.5°C. If we compare these to the one-dimensional model results presented in Chapter 3 (which were all for equilibrium conditions) we see that the direct radiative forcing (with no climate feedbacks) resulted in a 1.2°C surface temperature increase. When the water vapor feedback was added to the one-dimensional model, it doubled the response to about 2.5°C. The three-dimensional models that include geography and additional feedback processes increased the response still further. The indication is that as we make the models more realistic they appear to be dominated by feedback processes that amplify the initial change in climate. The more complex and realistic the model, the greater the resulting changes in climate.

Transient Climate Experiments. More recent projections have used **transient model experiments**. In this case, rather than running the model to equilibrium conditions before changing the boundary conditions, we start the model at the present or at some point in the past and then run it forward to assess how the climate changes with time. To do this we need to input data that will indicate how the greenhouse forcing will change with time. Various scenarios have been developed assuming a wide range of future socioeconomic states and the resulting changes in greenhouse forcing. These are discussed further in Chapter 16.

These transient experiments require a more sophisticated treatment of the ocean than the simple mixed layer models discussed earlier. We have seen that the atmosphere responds quickly to any change, while the oceans respond much more slowly. Time-scales of change are on the order of weeks in the atmosphere, seasons for the land surface and upper ocean, and hundreds of years for the deep ocean. Ocean models thus take considerably longer to reach equilibrium than atmospheric models. To save computational time, the atmosphere and ocean components are usually run separately before they are coupled together. Most techniques for coupling the two models, however, result in some degree of climate "drift" in the model away from the observed climate. Modelers are faced with the choice of letting the drift occur (and therefore not reproducing the observed climate), or adding in some corrective adjustment (referred to as a *flux adjustment*). Recently this problem may have been solved in some coupled model runs that did not exhibit this sort of climate drift.

A common approach with early transient model experiments was to increase atmospheric CO_2 by 1% per year. This results in a doubling of atmospheric CO_2 in about 70 years. The model would be run until it reached present-day equilibrium and then atmospheric CO_2 would be increased at a compounded rate of 1% per year. A comparison of 19 of these runs produced a warming of between 1.1°C and 3.1°C with a mean of 1.8°C (most were in the 1.5°C to 2.5°C range) by the time atmospheric CO_2 doubled. For many of these models, only about 60–65% of the equilibrium change had taken place by the time the atmospheric CO_2 doubled. This reflects the large thermal inertia introduced into the systems by the oceans. However, the equilibrium climate response remains unchanged. In other words, once the atmospheric CO_2 has doubled in the transient experiment, if the model calculation is extended forward in time with that level of atmospheric CO_2, it develops an equilibrium climate similar to the equilibrium models reported earlier. This is illustrated in Figure 6-5, which shows the results of a coarse resolution Atmosphere–Ocean General Circulation Model (AOGCM) running with a 1% per year increase in atmospheric CO_2. The CO_2 increase is stabilized at the doubling point (lower curves), and when it is quadrupled (upper curves). The colored lines show the results with a simplified model that allows no energy exchange with the deep ocean (so the difference shows the effects of ignoring these ocean exchanges). The transient surface temperature change is about 2°C. If a doubled CO_2 level is maintained, there would still be a further increase of approximately 1.5°C in global mean temperature as time passes in the model run.

Transient GCM experiments have also been run with a range of other forcing scenarios. These use time-dependent forcing (forcing that changes over time) in which the simulations start at some point in the past (usually the middle of the 19th century) and are run with estimates of the actual forcing during the 20th century. These models can be validated against the observed climate record, and are then extended into the future using a projection of future atmospheric radiative forcing. Some of these models have also been run with projections of future sulfate aerosol concentrations (see Chapter 16), which in several ways act to cool the surface by increasing the albedo. In these experiments, CO_2 doubling occurs by about 2060. Comparing 2021–2050 with 1961–1990, models that just include the CO_2 forcing (and no future sulfate aerosol concentrations) show an increase of 1.6°C with a range from 1.0°C to 2.1°C. (This actually is only a little lower than the numbers presented in Figure 6.5 when you take into account they are from a time interval well before doubling occurs.) When sulfate aerosols are included, the mean temperature increase is 1.3°C with a range from 0.8°C to 1.7°C.

Can We Trust What the Models Tell Us?

GCMs compute the three-dimensional exchanges of energy, mass, and momentum within the atmosphere and oceans and at the interface between the atmosphere and the land or ocean surface. The rules that govern these exchanges can be expressed as equations derived from the basic physical laws that describe these processes. Deal-

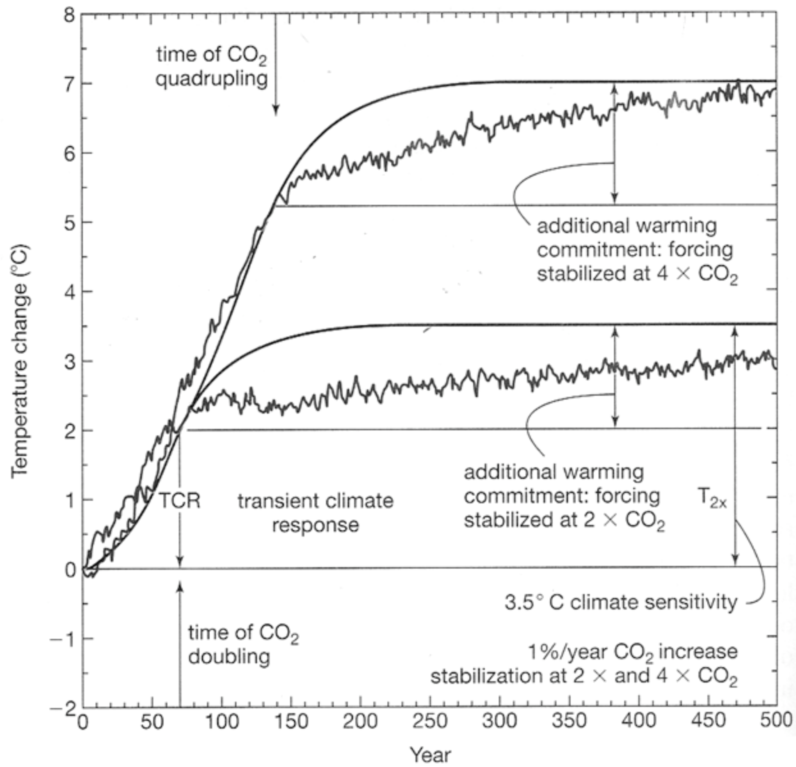


FIGURE 6-5

Global mean temperature change for 1%/yr CO_2 increase with subsequent stabilization at 2 × CO_2 and 4 × CO_2. (From the Intergovernmental Panel on Climate Change, Climate Change 2001, *The Scientific Basis: Projections of Future Climate Change.*) (*http://www.grida .no/climate/ipcc_tar/wg1/338.htm*)

ing as we are with basic principles of physics that are reasonably well understood, it would seem that there should be no inherent reason why these models should not produce an accurate representation of the climate system. As we will see below, they do in fact produce a reasonable simulation of the present climate. However, there are problems, most of which are due to the resolution at which the calculations are made.

Subgrid-Scale Processes

Many of the physical processes that take place in the climate system occur on spatial scales that are too small for the grid-scale of the model. A good example of this is cloud cover. A typical grid cell in today's models may be 250 km across, so the cell may represent an area of over 62,000 km^2. In effect, the model has to treat this as a uniform region with a single temperature, rainfall rate, cloud cover, and so on. However, you only have to look upward to see that clouds do not form as rectangular boxes, 250 km on a side—which is what the model produces. The physical processes that give rise to localized convection in the atmosphere, cause clouds to form and dissipate, and cause ice crystals to grow and raindrops to fall, all occur on spatial scales that are much smaller then the model grid cell. While we can develop mathematical models of cloud formation, we cannot include them in GCMs because the computing power simply isn't available to do the calculations at the *resolution* needed.

Resolution in this context refers to the effective grid size of the model (spatial resolution) and the model time step (temporal resolution). Refer back to the earlier discussion of the number of calculations made by GCMs. If we take the 2.5° grid cell we used as an example before, that gave us 400,000 grid boxes for a 20-layer model. If we doubled the resolution to 1.25°, that means four times the number of grid boxes for which these computations have to be made, plus four times the number of land cells so four times the number of soil calculations, and so on. In actual fact, the number of extra calculations will be even greater because, for any given model, increasing the grid's resolution usually means you also have to increase the number of vertical layers and shorten the time step, so more calculations for each day of model climate. A model run that took 50 days at the previous resolution may now take 250 days at the new resolution—and you still do not have a model that can resolve the cloud cover. As another example, think of the landuse/vegetation cover. At present, most models can only be assigned one type of vegetation in a grid cell, so we use the predominant vegetation type. It may, for example, be deciduous forest. In reality there will be a mix of landuse or vegetation cover in any 62,000-km^2 region. If nothing else, there are likely to be variations in forest density, plus there may be roads, lakes, maybe some towns, and so on—all of which have very different effects on the energy and water exchange at the surface. Even if the vegetation cover was completely uniform, the actual energy and moisture exchanges occur at the level of individual trees, roots, and leaves, which certainly can't be resolved by the model.

Where there are processes for which we can't explicitly write and solve a set of physical equations, that is, all of these processes that take place at a finer time or space scale than the model can resolve (**subgrid-scale processes**), we have to **parameterize** the relationships involved. Parameterizations are simply empirical or statistical functions that relate two variables based on observed relationships between those variables (like the statistical models referred to earlier). Clouds, for example, are generated in the model as some function of humidity and temperature, based on what we observe in today's climate. Different models may use different parameterization schemes, and how models parameterize these subgrid-scale processes accounts for much of the variability in the output from different models. This is why we present a range of results for the different types of climate experiments described earlier. The various ways models parameterize cloud cover accounts for a large part of the differences in model results, and is probably the most important factor leading to uncertainties and errors in the atmospheric component of the GCMs.

Dependence on Initial Conditions

Starting the climate model from almost identical, but still slightly different, conditions can also result in differences in the model output. This is illustrated in Figure 6-6 from the IPCC Third Assessment Report. In the figure there are three realizations of temperature distributions made with the same model and same greenhouse gas and aerosol forcing. The maps show the temperature difference between the period 1975–1995 and the first decade of the 21st century. Each of the simulations began with slightly different initial conditions a century earlier. You can see that the model results are essentially the same when looking at broad global patterns, but the sensitivity to initial conditions does result in some differences in regional detail. This example demonstrates that any single

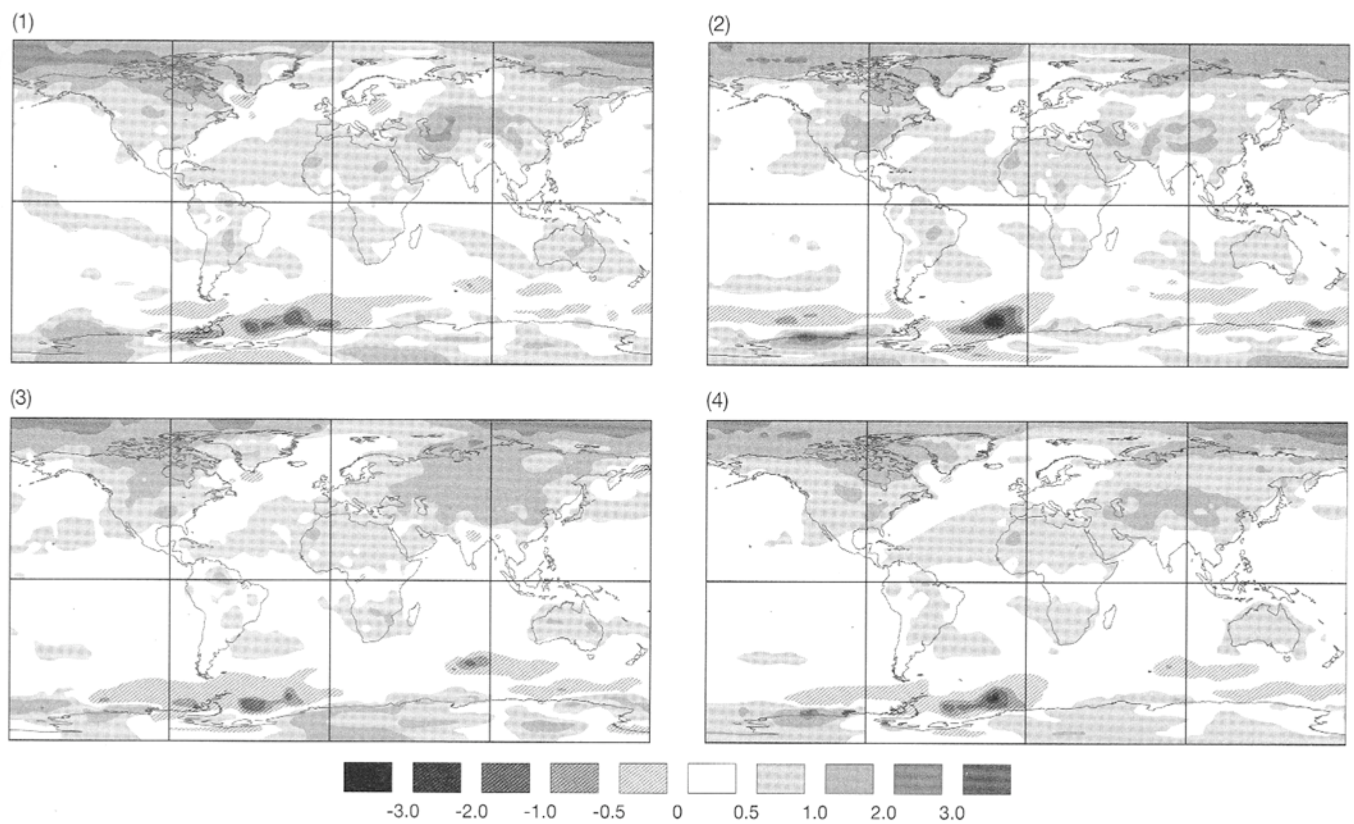


FIGURE 6-6

Ensemble model runs showing three realizations of a transient experiment with identical atmospheric forcing, but with slightly different initial conditions. The temperatures show the differences from the first decade of the 21st century to the period 1975–1995. The fourth map is the ensemble mean. (From the Intergovernmental Panel on Climate Change, Climate Change 2001, *The Scientific Basis: Projections of Future Climate Change.*) (*http://www.grida.no/climate/ipcc_tar/wg1/338.htm*) run of the climate model simply represents only one of a range of possible outcomes. This problem is addressed in present climate experiments by running models several times from slightly different starting points and presenting the model runs as an ensemble. The ensemble runs can then be averaged to give a single ensemble mean. In Figure 6-6 the fourth map represents the ensemble mean.

Are the Model Results Usable?

There are other problems with the models and their projections of present and future climates that we do not need to go into here. We simply want to demonstrate that the models are not perfect and to show that one of the main reasons why there is uncertainty in the results is due to the fact that many important climate processes operate on time and space scales that the models cannot resolve. This does not mean that the model results are invalid. These models do a very good job of projecting ahead the mean global climate. There are differences between models primarily because of the different ways they treat subgrid-scale processes. These differences account for the spread (the differences) in the model results that we examined earlier in the chapter. However, there is also extensive agreement between the models. Positive feedback processes dominate in all of the global warming projections; all show similar circulation patterns and all agree, in a broad sense, on the global pattern of temperature changes. The changes are greatest at high latitudes and least in the tropics and are greater over the continents compared to the oceans at the same latitude.

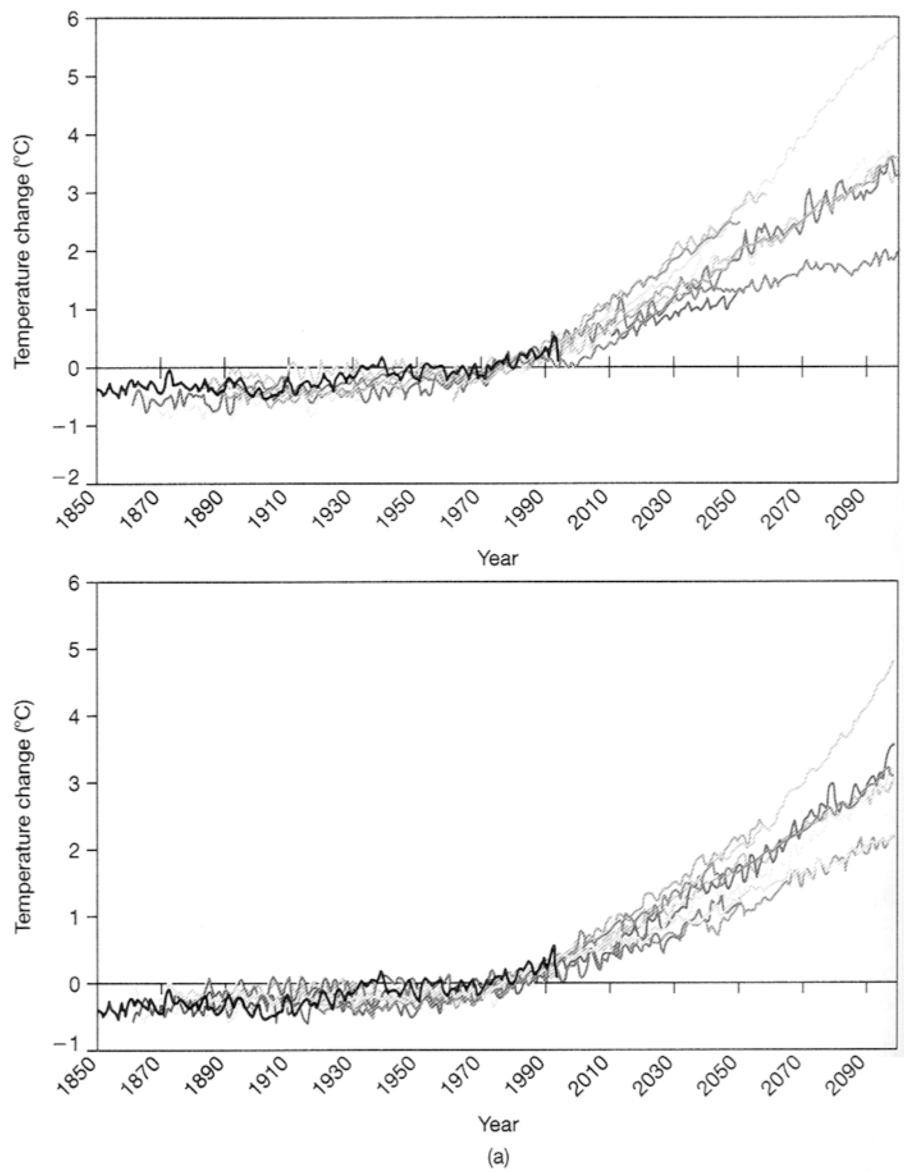


FIGURE 6-7
Time evolution of globally averaged temperature change relative to the period 1961–1990. The top graph shows the results of greenhouse gas forcing, the bottom graph shows the results of greenhouse gas forcing plus aerosol forcing. (From the Intergovernmental Panel on Climate Change, Climate Change 2001, *The Scientific Basis: Projections of Future Climate Change.*) (*http://www.grida.no/climate/ipcc_tar/wg1/338.htm*)

Figure 6-7 from the IPCC Third Assessment Report shows the time evolution (change over time) of the temperature change produced for 10 different model climates—all using the same atmospheric (greenhouse and aerosol) forcing. The temperatures presented represent the difference between the annual globally averaged value of each year and the global mean for the period 1961–1990. Each wavy line shows the results of a different model run. The solid black line shows the observed temperature change. The top graph shows the results with just the greenhouse forcing, while the bottom graph includes both greenhouse gas and aerosol forcings. Figure 6-8 shows the same results, but for precipitation. Notice that the model spread (the differences between individual models results) increases into the future as the forcing becomes larger, but all of the models are very close to each other and to the observations for the 150 years of the observed record. This indicates that the models (even if their results differ slightly from each other) do a reasonable job of simulating the present-day climate system. The present generation of GCMs, which are coupled ocean–atmosphere climate models, are effective at simulating the climate system at subcontinental scales (i.e., for large regions like Western Europe) and over time periods of seasons to decades. However, there are still large uncertainties when it comes to using these models to assess regional climate change over briefer periods of time.

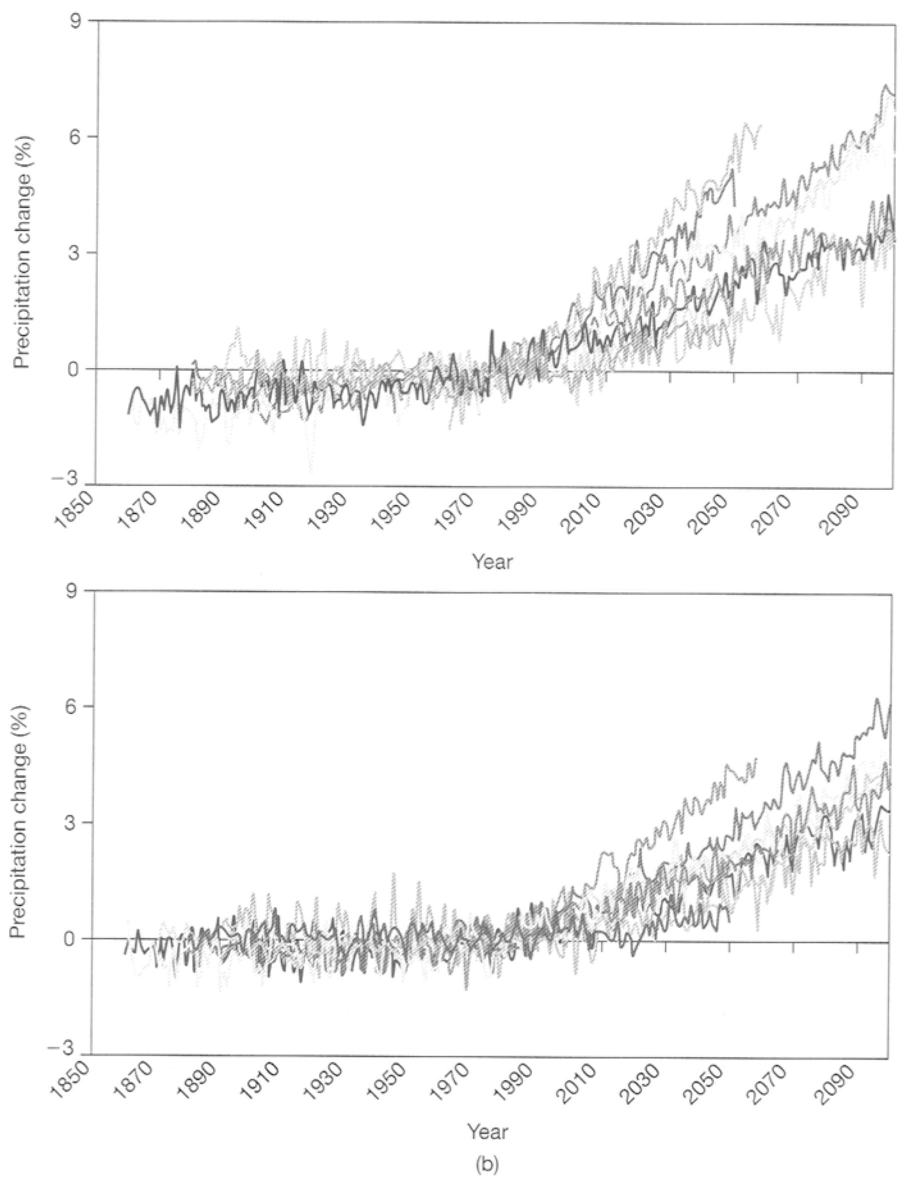


FIGURE 6-8

Time evolution of globally averaged precipitation change relative to the period 1961–1990. The top graph shows the results of greenhouse gas forcing, the bottom graph shows the results of greenhouse gas forcing plus aerosol forcing. (From the Intergovernmental Panel on Climate Change, Climate Change 2001, *The Scientific Basis: Projections of Future Climate Change.*) (*http://www.grida.no/climate/ipcc_tar/wg1/338.htm*)

Chapter Summary

1. The climate system is inherently complex. Multiple climate variables interact over a broad range of time and space scales through interconnected positive and negative feedback processes. The complexity is such that mathematical models of the physical processes are the only feasible approach to projecting climate change.
2. Researchers employ a range of models from simple one-dimensional globally averaged radiative-convective models (described in Chapter 3), to more complex three-dimensional general circulation models (GCMs). GCMs include linked models of the atmosphere, ocean, soils, vegetation, and sea ice cover. These models are used for a variety of applications—to study how the climate system works, to forecast weather, to recreate past climates, and to assess the potential for future climate change. Modern GCMs run in a transient mode and reflect the time history of atmospheric forcing and projections of future greenhouse gas emissions.
3. The biggest weakness of these models is their relatively coarse spatial resolution. The inability to resolve subgrid-scale processes, many of which are important to the climate system, produces some uncertainty in the model results. The way in which different models parameterize these subgrid-scale processes is one of the primary reasons for the spread in climate change projections produced by different GCMs.
4. Despite these weaknesses, the models do an effective job of simulating the observed global climate record. They produce climate change projections that are valid at the subcontinental level and for time periods of seasons to decades.

Key Terms

boundary conditions
energy-balance climate models
equilibrium climate experiments
finite difference models
general circulation models
global climate models
parameterize
resolution
spectral models
subgrid-scale processes
transient model experiments

Review Questions

1. What is meant by a conceptual model? Give an example.
2. What is meant by a statistical model?
3. What is the difference between a one-dimensional radiative-convective model and a two-dimensional energy balance model?
4. What is a general circulation model?
5. What are some of the science and research applications for GCMs?
6. What is meant by the term *model boundary conditions*? Give examples of the boundary conditions that might be used in a climate model.
7. What is meant by the term *equilibrium climate experiment*?
8. What is the difference between an equilibrium experiment and a transient experiment?
9. Why are subgrid-scale processes a problem for global climate models?
10. Explain what is meant by parameterization.
11. What is the purpose of running an ensemble model?
12. Why are ensemble model runs necessary?

Critical Thinking Problems

1. Construct a systems diagram of a global climate model. Show each subcomponent (atmosphere, oceans, etc.) as a separate box. Include the ocean as two boxes: the mixed layer ocean and the deep ocean. Use arrows to connect all of the boxes and describe the types of interactions that occur between the different subcomponents. Divide the interactions into short-term and long-term processes.
2. Obtain some topographic maps for your local region. If possible, find one that has a scale of about 1:50,000. Also find one that shows about 250 km on each side of the same region. Divide the large area map into 250-km squares and visually estimate the average elevation in each square. Then divide the squares into quarters and repeat the exercise. Compare the results and then compare them to the topography in the higher resolution map. Discuss what the implications of this exercise are for climate models.

Further Reading

Advanced

Trenberth, K. (ed.). 1992. *Climate System Modeling*. Cambridge: Cambridge University Press.