Climate System Modeling

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Introduction to climate modeling

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1.1 Context of global climate change

The Earth's climate changes. It is vastly different now from what it was 100 million years ago, when dinosaurs dominated the planet and tropical plants thrived at high latitudes; it is different from what it was even 18,000 years ago, when ice sheets covered much more of the Northern Hemisphere. In the future it will surely continue to evolve. In part the evolution will be driven by natural causes, such as fluctuations in the Earth's orbit. But future climatic change, unlike that of the past, will probably have another source as well: human activities. We may already be feeling the climatic effects of having polluted the atmosphere with gases such as carbon dioxide. Many other activities associated with human economic development driven by growing populations using technology and organization to improve their standard of living have altered the physical and chemical environment in ways that modify the natural stocks and flows of energy and materials in the environment. When these modifications become large enough it is natural to expect significant global changes. Indeed, human activities that release carbon dioxide, chlorofluoromethanes, nitrous oxide, methane, atmospheric particles (aerosols), and heat, or that use land for urbanization, agriculture or deforestation, are examples of such modifications to the stocks and flows of processes or materials related to the maintenance of environmental services. (Chapter 2 discusses human impacts on the climate system in more detail.) Table 1.1 (modified from Schneider and Londer, 1984) summarizes briefly a range of such activities, their potential climatic effects and an estimate of the scale and importance of the potential effects. The list is comprehensive but not exhaustive, and considerable uncertainty surrounds the predicted effects in specific cases. Nevertheless as Table 1.1 suggests, growing human numbers using technology and organization to increase per capita levels of consumption could have a substantial impact on climatic or other environmental systems. The fact that such global changes could be of considerable importance to human and natural systems is what motivates the need for quantitative evaluation of the potential impact of human activities in creating global change, and that evaluation of such

Table 1.1. Summary of principal human activities that can influence climate (modified after Schneider and Londer, 1984).

Activity	Climatic effect	Scale and importance of the effect
Release of carbon dioxide by burning fossil fuels.	Increases the atmospheric absorption and emission of terrestrial infrared radiation (greenhouse effect), resulting in warming of lower atmosphere and cooling of the stratosphere.	Global: potentially a major influence on climate and biological activity.
Release of methane chlorofluoromethanes, nitrous oxide, carbon tetrachloride, carbon disulfide etc.	Similar climatic effect as that of carbon dioxide since these, too, are infrared-absorbing and fairly chemically stable trace gases.	Global: potentially significant influence on climate.
Release of particles (aerosols) from industrial and agricultural practices. Sulfur dioxide is of primary concern since it photochemically converts to sulfuric acid particles.	These sunlight scattering and absorbing particles (especially soot) could decrease albedo over land, causing a warming and could (especially sulfate) increase albedo over water causing a cooling; they also change stability of lower atmosphere; net climatic effects still speculative, although net cooling effect seems more likely.	Largely regional, since aerosols have an average lifetime of only a few days, but similar regional effects in different parts of the world could have nonnegligible net global effects; stability increase may suppress convective rainfall, but particles could affect cloud properties with more far-reaching effects.
Release of aerosols that act as condensation and freezing nuclei. Again, released soot or sulfur dioxide by industrial activities is of primary concern.	Influences growth of cloud droplets and ice crystals; may effect amount of precipation or albedo of clouds in either direction.	Local (at most) regional influences on quantity and quality of precipitation, but unknown and potentially important change to Earth's heat balance if cloud albedo is altered. Some calculations suggest SO_2 released between 1950 and 1980 opposed much of the Northern Hemispheric warming trend that otherwise would have been experienced from rapid buildup of greenhouse gases during those decades.
Release of heat (thermal pollution).	Warms the surface layers directly.	Locally important now; could become significant regionally; could modify circulation.

Table 1.1 (Continued)

Activity	Climatic effect	Scale and importance of the effect
Upward transport of chlorofluoromethanes and nitrous oxide into the stratosphere.	Photochemical reaction of their dissociation products probably reduces stratospheric ozone.	Global but uncertain influence of ozone depletion on climate; less total stratospheric ozone allows more solar radiation to reach the surface but compensates by reducing greenhouse effect as well; however, if ozone concentration decreases at high altitudes, but increases comparably at lower altitudes, this would lead to potentially large surface warming; could cause significant biological effects from increased exposure to ultraviolet radiation if total column amount of ozone decreases.
Release of trace gases (e.g., nitrogen oxides, carbon monoxide, or methane) that increase tropospheric ozone by photochemical reactions.	Large atmospheric heating occurs from tropospheric ozone, which enhances both solar and greenhouse heating of lower atmosphere.	Local to regional at present, but could become a significant global climatic warming if large-scale fossil fuel use leads to combustion products that significantly increase tropospheric ozone. Contact with ozone also harms some plants and people.
Patterns of land use, e.g., urbanization, agriculture, overgrazing, deforestation. etc.	Changes surface albedo, evapotranspiration and runoff and causes aerosols.	Largely regional: net global climatic importance still speculative.
Release of radioactive Krypton-85 from nuclear reactors and fuel reprocessing plants.	Increases conductivity of lower atmosphere, with possible implications for Earth's electric field and precipitation from convective clouds.	Global: importance of influence is highly speculative.
Large-scale nuclear war.	Could lead to very large injections of soot and dust causing transient surface cooling lasting from weeks to months, depending on the nature of the exchange and on how many fires were started.	Could be global, but initially in midlatitudes of Northern Hemisphere. Darkness from dust and smoke could disrupt photosynthesis for weeks with severe effects on both natural and agricultural ecosystems of both combatant and noncombatant nations. Transient freezing outbreaks could eliminate some warm season crops in midlatitudes or weakening of monsoon rainfall could be devastating to any vegetation in tropics or subtropics. Details still speculative.

change is central to any potential policy responses to mitigate those potential changes (e.g., Schneider, 1990).

How can human societies prepare for so uncertain a climatic future? Clearly it would help to predict that future in some detail, but therein lies a problem: the processes that make up a planet's climate are too large and too complex to be reproduced physically in a laboratory. Fortunately they can be simulated mathematically with the help of a computer. In other words, instead of building a physical analogue of the land–ocean–atmosphere system, one can devise mathematical expressions for the physical principles that govern the system – laws of thermodynamics and Newton's laws of motion – and then allow the computer to calculate how the climate will evolve in accordance with these laws. For a variety of reasons to be detailed in this chapter and volume, mathematical climate models cannot simulate the full complexity of reality. They can, however, reveal the logical consequences of plausible assumptions about how the climate system operates. The critical scientific task is to formulate, build and then validate the models.

1.2 Mechanisms of climatic change

1.2.1 The climate system

Climate is typically the average state of the atmosphere observed as the weather over a finite time period (e.g., a season) for a number of different years. Thus, we can speak of the climate of a day–night cycle, month, season, year, decade, or even longer period. Climate is usually defined by the mean state together with measures of variability or fluctuations such as the standard deviation or autocorrelation statistics for the period (e.g., Mearns et al., 1990).

Although the same physical laws usually are applied to the most comprehensive tools for both climate and weather prediction, the climatic prediction is complicated by considering complex interactions between, as well as changes within, all the components of the climate system – the atmosphere, oceans, land, ice and snow, and terrestrial and marine biota. Plate 1 (Earth System Science, 1986. See also Plate 2 for a simpler version) is an attempt to represent schematically the interacting physical, chemical and biological processes which, on time scales up to centuries, control global changes. Although a weather forecaster need not consider, for example, the small day-to-day changes in ice, temperature, or circulation of the sea, such changes affecting the lower atmosphere must be considered by the predictor of atmospheric changes from one season or one decade to another. On the other hand, change of the Earth's orbit occurs during thousands of years and is negligible when considering climate changes during less than a few millennia (Shackleton and Imbrie, 1990).

As mentioned above, the following components interact to make the observed climate state the result: atmosphere, oceans, cryosphere, and land/biosphere. The atmosphere and the oceans are two fluid components of the system, each containing organized circulation, chaotic motions, and random turbulence. They react to perturbations on very different time scales. Interactions between and within them occur on many scales and tend to be concentrated close to their boundary as well as internally where gradients of physical properties, such as temperature or density, can be large. These interactions will be introduced in this chapter briefly, but discussed in depth in subsequent chapters.

The chemical composition of the atmosphere also affects climate. Aerosols, water vapor, carbon dioxide, and ozone directly affect the atmosphere's absorption and transmission of solar radiation, which provides almost all the energy for the entire system. Further, aerosols (e.g., dust or sulfate particles) may cause clouds to form and precipitation to fall. For example, Twomey et al. (1984) have argued that increased sulfuric acid aerosols (e.g., from SO_2 injections from coal or oil burning or possibly photoplankton emissions) could increase cloud brightness in unpolluted areas (see Chapter 7 for more details). Wigley (1989) updated the old suggestion (e.g., SMIC, 1971) that this may have offset some fraction of anticipated CO_2-induced Northern Hemispheric warming since 1950, see also Charlson et al. (1991). However, any such effect would be highly regional, would have diminished as SO_2 controls were applied in the 1970s to combat acid rain, and would saturate as background pollution or other aerosol levels increase.

Other complex processes in the climate system include the salinity of oceans (see Chapter 4 for more details), which affects water density, and thus circulation of the oceans. The exchange between air and surface of such absorbers of radiation as water vapor, carbon dioxide, methane and nitrogen and sulfur oxides is another example, and is determined by such physical processes as winds, rainfall, or runoff and biological processes in forest or from photoplankton productivity.

The third component of the climate system is the cryosphere (see Chapters 12 and 13), which includes the extensive ice fields of Antarctica and Greenland, other continental snow and ice, and sea ice. Continental snow and sea ice vary seasonally and interannually, causing large annual variations in continental heating and upper ocean mixing and in energy exchange between the surface and atmosphere. Although the large continental ice sheets do not change rapidly enough to cause seasonal or annual climatic anomalies, they play a major role in climatic changes during hundreds to thousands of years such as the glacial and interglacial cycles that have occurred repeatedly for at least the past one million years (Shackleton and Imbrie, 1990).

The land and its biomass constitute a fourth component of the climate system, as depicted schematically on Plate 1. This component includes the slowly changing extent, position, and orography of the continents and the more rapidly varying characteristics of lakes, rivers, soil moisture, and vegetation. Thus, the land and its biomass are variable parts of the climate system on all time scales. Proper inclusion of the biophysics of energy and materials exchange between the atmosphere and land biosphere is important to simulation of the effects of deforestation (e.g., Henderson-Sellers et al., 1988)

The entire climate system involves the interaction of the biota, air, sea, ice and land, with solar radiation providing the energy that drives it. Variations of gaseous and particulate constituents of the atmosphere, along with changes in the Earth's position relative to the Sun, vary the amount and distribution of sunlight received. The temperature of the oceans has a marked influence on the heating and moisture content of the atmosphere. The unreflected radiant energy drives the atmospheric circulation, and by wind stress and heat transfer it drives the circulation of the oceans. The atmosphere and oceans are both influenced by the extent and thickness of the ice covering the land and sea as well as by the shape and composition of the land surface itself. Since each of these components has a different range of response times, the whole system evolves continuously, with some parts lagging or leading other parts.

The system also contains feedback loops between the interacting components, as illustrated in Plate 1. These amplify (positive feedback) or damp (negative feedback) perturbations. For example, any increase in the area of polar ice or snow from a forced cooling reflects more of the incoming solar radiation, leaving less to be absorbed by the surface. If snowfall is adequate, this further lowers surface temperature, increasing ice and snow cover in a positive feedback loop. One might expect, however, that increasing snow cover and associated coldness of a continental interior could gradually limit the overlying atmosphere's ability to import moisture into the region. This eventually decreases snowfall and limits growth of the snow cover in a negative feedback loop.

1.2.2 Radiation balance and the greenhouse effect

The Sun radiates energy corresponding roughly to an ideal (black) radiator with a temperature of about 5,800 K. This implies that 90% of the radiant energy lies in the interval with wavelengths from 0.4 to 4 μm, with a maximum intensity in the green portion of the visible spectrum at 0.48 μm.

About 30% of the incoming solar energy is reflected back to space and is unavailable to warm the Earth. This reflected fraction is called the planetary albedo. Reflection occurs from the clouds, the Earth's surface, and from molecules and particles present in the atmosphere. The clouds contribute the largest part of the albedo, reflecting about 25% of the incoming radiation when averaged over a long period of time, but due to the natural variability of cloudiness over the globe, the Earth's albedo can change substantially from day to day and also from season to season.

The cloudless part of the Earth comprises the remaining 5% of the global albedo. The albedo of the cloudless part of the Earth is determined by the surface albedo and by reflection from atmospheric molecules and suspended particles. The latter, though contributing at most a small percentage to the total albedo, can be of practical importance, since such particles are a factor that can be biased by human activities. Of the incoming radiation, about 25% is absorbed by gases, clouds, and particles in the atmosphere, 30% is reflected to space as discussed above, and the remainder is absorbed at the Earth's surface. This identifies another factor that can be affected by human activities. Since humanity has significantly altered the character of the Earth's surface, it has indirectly affected the climate (at least in limited regions) by disturbing the heat and water budgets through changes in the character and albedo of the surfaces.

To maintain equilibrium, the incoming solar energy that is absorbed by the Earth–atmosphere system must be balanced by an equal amount of outgoing radiant energy. Otherwise, the temperature of the Earth would undergo a continuous change until the "energy balance" is restored. The Earth emits radiant energy, as do all physical things, in proportion to its absolute temperature. But, since the wavelength of maximum radiant energy is inversely proportional to the temperature of the radiator, the Earth emits radiation primarily in the longwave or infrared region, with most of the energy residing in the wavelengths from 4 to 100μm. Figure 1.1 (Goody, 1964) shows radiant energy spectra for solar and terrestrial "blackbody" radiators, along with a representation of the absorption of radiation by gases in the atmosphere.

If we calculate the total solar energy absorbed in the Earth–atmosphere system and equate this to the escaping infrared radiation, then we can determine an "effective radiation temperature" of the planet from the Stefan–Boltzmann law relating the flux of radiant energy to temperature. This has been observed from space to be about $-18°$C (255 K) for the effective radiative temperature of the Earth, whereas we know the average surface temperature to be about $+15°$C.

The 33°C difference in the two temperature values is, of course, due to the presence of our atmosphere. The optically active gases, principally water vapor, carbon dioxide, methane and ozone, absorb and re-emit infrared radiation in selective "bands" of the infrared spectrum. Clouds and particles also affect the infrared radiation, with the clouds (except thin cirrus clouds) absorbing nearly all the infrared radiation they receive throughout the infrared spectrum, and the particles absorbing or scattering *relatively* little infrared radiation, depending upon the character of the particulate material.

The average surface temperature is higher than the effective radiative temperature primarily because the atmosphere is semitransparent to solar

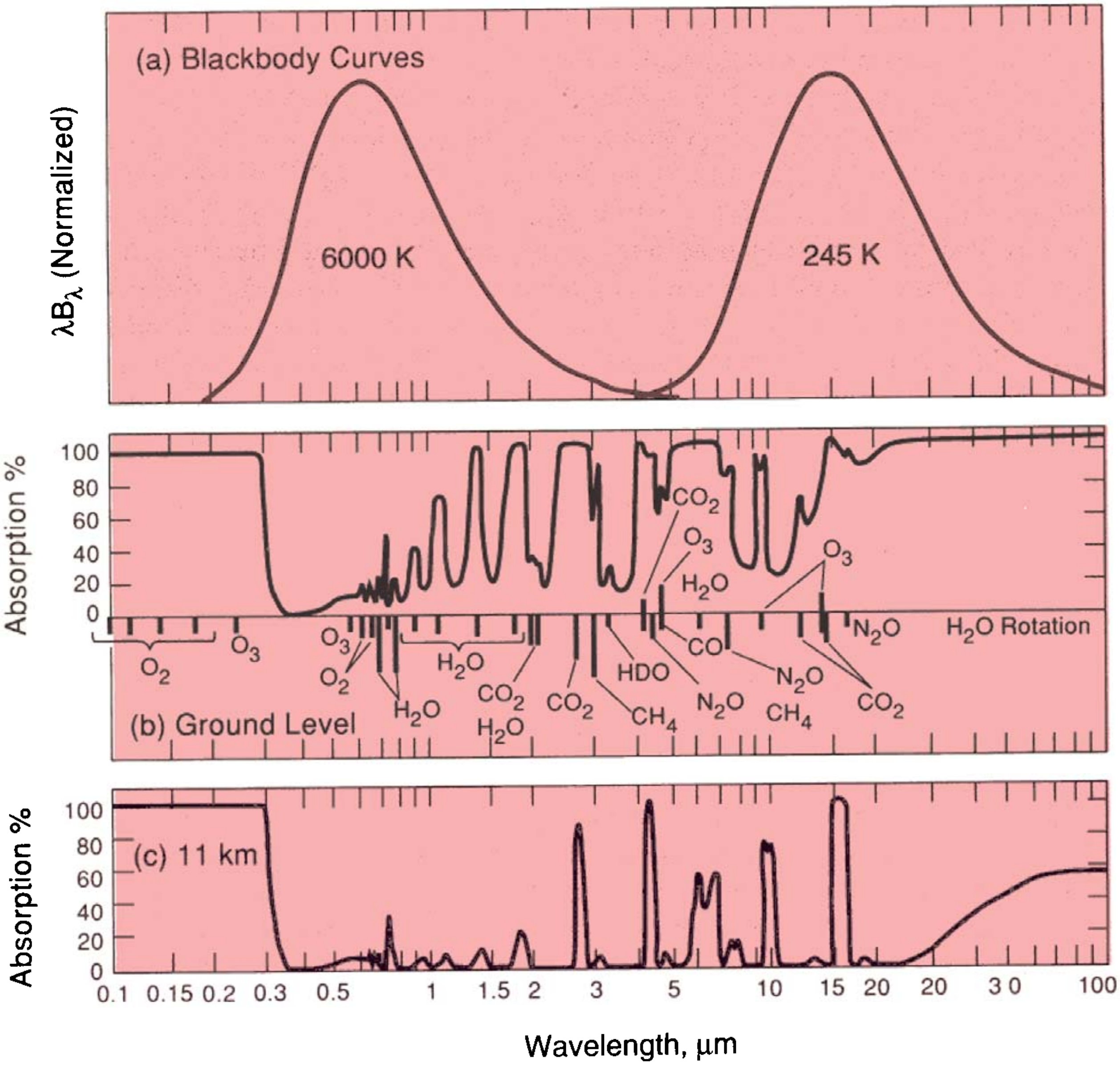

Fig. 1. (a) Spectral distribution of longwave emission from blackbodies at 6,000 K and 245 K, corresponding approximately to the mean emitting temperatures of the Sun and Earth, respectively, and (b) atmospheric absorption spectrum for a beam of radiation reaching the ground; (c) the same for a beam reaching the tropopause in temperate latitudes. Notice the comparatively weak absorption of the solar spectrum and the region of weak absorption from 8 to 12 μm in the longwave spectrum.

radiation but nearly opaque to infrared radiation as a result of absorbing gases and clouds. Thus, the surface, which absorbs nearly half (e.g., see Fig. 1.2; Schneider, 1990) the solar radiation, becomes a heat source for the lower atmosphere, which on the average cools steadily with increasing altitude to about 10 km. This part of the atmosphere is called the troposphere. The tropospheric vertical temperature *lapse rate*, $-\partial T/\partial z \approx 6.5$ K km^{-1}, is affected by both radiative heating and vertical convective

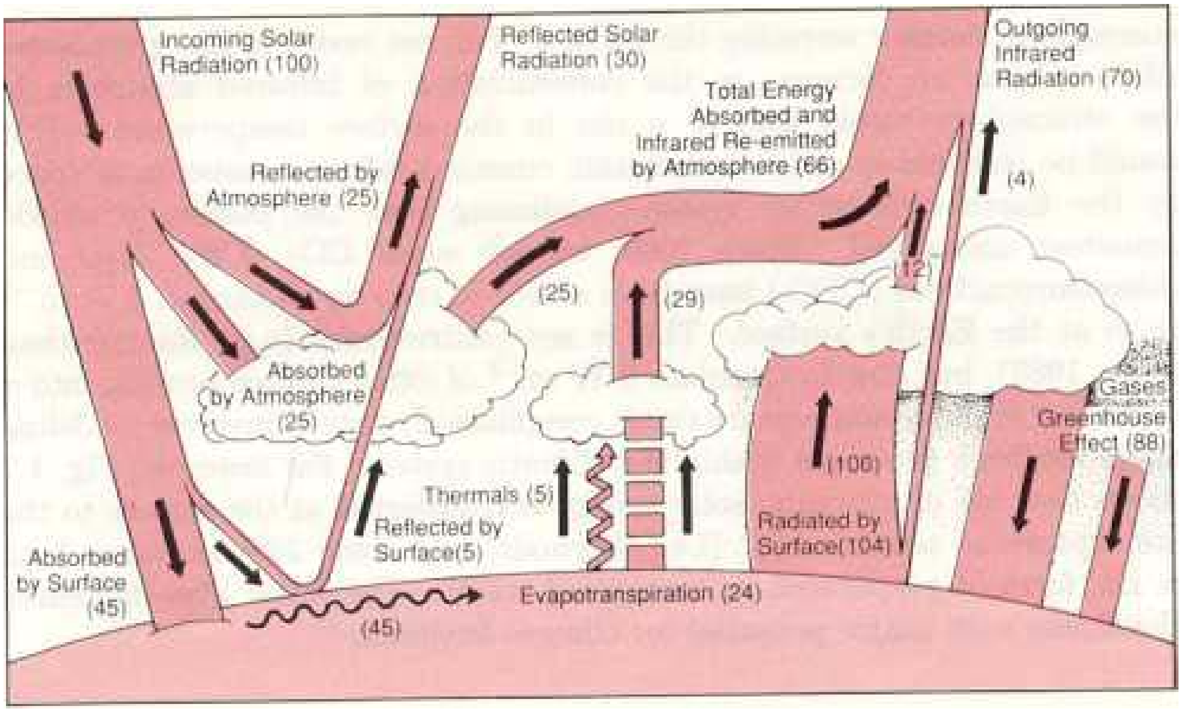

Fig. 1.2 The Earth's radiation energy balance, which controls the way the greenhouse effect works, can be seen graphically here. The numbers in parentheses represent energy as a percentage of the average solar constant – about 340 W m^{-2} – at the top of the atmosphere. Note that nearly half the incoming solar radiation penetrates the clouds and greenhouse gases to the Earth's surface. These gases and clouds re-radiate most (i.e., 88 units) of the absorbed energy back down toward the surface. This is the mechanism of the greenhouse effect.

processes (e.g., Manabe and Wetherald, 1967).

The warm surface layer emits infrared radiation, most of which is intercepted by optically active atmospheric gases, clouds, and particles. These constituents re-emit radiation both up to space and back down to the surface, the latter reducing dramatically the net loss of heat from the surface. Since the atmospheric emitters are colder than the surface, they emit proportionally less radiant energy. Because of this, the total outgoing infrared radiation from the Earth–atmosphere system is less than the radiant energy emitted by the surface alone, and the effective radiation temperature of the Earth is influenced more by the temperature of the cooler atmospheric gases and cloud tops (which emit radiation roughly like a blackbody with the temperature of the atmosphere at the cloud tops) than the warmer surface below. This phenomenon has often been called the *greenhouse effect.*

Should the amount of an infrared-absorbing gas in the atmosphere be increased, it would then intercept a larger fraction of the infrared energy coming upward from the warm layers near the surface. Thus, the outgoing infrared flux to space would be reduced by adding infrared-absorbing gases to the atmosphere. Furthermore, as seen on the right-hand arrows on Fig. 1.2, this would also increase the downward infrared flux in the lower atmosphere, further warming the surface. The net result of the greenhouse effect is that an increase in the concentration of infrared absorbers in the atmosphere would lead to a rise in the surface temperature. This would be required in order to maintain constant infrared emission to space by the Earth–atmosphere system, assuming that the planetary albedo remained unchanged. Since 1900, enough extra CO_2, CH_4, N_2O and chloroflurocarbons (CFCs) have been added to trap an additional 3 W m^{-2} or so at the Earth's surface. This is not controversial (e.g., Ramanathan et al., 1985), but how to translate 3 W m^{-2} of extra surface heating into x degrees of surface temperature rise is complicated, since it involves modeling many feedback processes within the climatic system. For example, Fig. 1.2 shows that 5% of incoming solar energy is transferred at the surface to the atmosphere as sensible heat (i.e., thermals) and some 24% as latent heat in the form of evaporated or transpired water. The latter also influences cloudiness with major potential for climate feedback.

1.2.3 Climate models of the radiation balance

Consider the simplest possible climate model, a radiation balance for the Earth.

$$S = F \tag{1.1}$$

is an energy balance which requires that the amount of solar energy input to the system S is, in equilibrium, exactly balanced by the amount of terrestrial radiation energy F that leaves the system.

Translating this radiation balance into a simple climate model involves the energy balance equation (1.1) the Stefan–Boltzmann law, $F = \sigma T_p^4$, where σ is the Stefan–Boltzmann constant, and T_p is the effective blackbody radiative temperature of the Earth–atmosphere system in K. Rewriting (1.1) into the simplest possible climate model takes the form

$$Q(1 - \alpha) = \sigma T_p^4 \,. \tag{1.2}$$

In the left-hand term, $Q = S_o/4$ represents the incoming solar radiation of approximately 343 W m^{-2} averaged over the Earth's surface (i.e., this is S_o the "solar constant" divided by 4 since the area of the Earth's surface is 4 times the area of the Earth's disk which intercepts sunlight); α is the albedo of the planet. We can define a sensitivity parameter (see Chapter 10, Sec. 10.2.1 for details)

$$\lambda = \frac{1}{\frac{dF}{dT} - \frac{dS}{dT}} \tag{1.3}$$

which represents how much temperature change the planet would undergo if either the solar output S or the longwave radiation back to space F were to change. If one differentiates both sides of (1.2) with respect to T and rearranges terms, one can show that for the simple climate model

$$\lambda_p = \frac{T_p}{4S}. \qquad (1.4)$$

Observations by infrared sensors on satellites suggest that the effective planetary temperature for Earth infrared radiation to space is equal to that of a blackbody with absolute temperature of approximately 255 K. Therefore, λ_p from (1.4) is approximately equal to 0.27 K W^{-1} m^2. In simpler terms this means that if the Sun were to somehow increase its energy output by 1%, the Earth's radiative temperature would eventually (i.e., in equilibrium) warm up by 0.64 K.

1.2.4 Climatic feedback mechanisms

Let us consider next what would happen if there were feedback in the system. That is, $\alpha_p = \alpha(T_p)$. Substituting α_p into (1.2) and differentiating both sides with respect to T one can obtain λ_p^*, a modified expression for λ_p in the case of feedback

$$\lambda_p^* = \frac{\lambda_p}{1 + \lambda_p \left(\frac{-}{1-\alpha_p}\right) \frac{\partial \alpha_p}{\partial T_p} S} \qquad (1.5)$$

where λ_p is the sensitivity parameter obtained when albedo is constant (i.e., Eq. 1.4). It is an instructive exercise to provide some intuitive feeling for what this equation means. Supposing for example $\alpha_p = a - bT_p$ where a is constant and b is 0.01. What that implies is that if the Earth's temperature were to increase by 1°C then albedo would decrease by 0.01. Thus, for $\alpha_p = 0.3$ at $T_p = 255$ K and $b = 0.01$, this implies $\lambda_p^*/\lambda_p \approx 10$, indicating that a seemingly "small" feedback ($b = 0.01$) can have dramatic impact on the sensitivity of the system!

Earth satellites have suggested that for temperatures observed on Earth a linear relationship between outgoing infrared radiation, F_{ir} and surface temperature, T_s, is not a bad first approximation (e.g., Warren and Schneider, 1979). Let us rewrite our simple climate model then with this linearized form assuming α is a constant,

$$Q(1-\alpha) = A + BT_s. \qquad (1.6)$$

Consequently

$$\lambda_s = \frac{1}{B}. \qquad (1.7)$$

Empirically B has been found from satellite observations to have a value of 1.83 W m^{-2} K^{-1} (Warren and Schneider, 1979). For $Q = 340$ W m^{-2} and $\alpha = 0.3$ then $\lambda_s = 0.55$ K W^{-1} m^2. This is twice λ_p, which suggests that empirically derived values of outgoing infrared radiation as a function of surface temperature lead to amplifying feedbacks.

The most often postulated feedback that could describe this amplification is the *water vapor–greenhouse effect* feedback. It is well known that increasing surface temperature increases evaporation, because evaporation increases nonlinearly with surface temperature through the Clausius–Clapeyron relationship between vapor pressure of water and temperature (e.g., see Chapter 3). In the midlatitudes, for example, although relative humidity is fairly constant from one season to the next, absolute humidity can increase by a large factor from winter to summer owing to this water vapor pressure–temperature relationship. Some recent empirical information from satellites strongly suggests that the water vapor–greenhouse feedback is indeed positive (e.g., Raval and Ramanathan, 1989), and may very well account for the substantially enhanced sensitivity of the linearized semi-empirical model (1.6) relative to the original blackbody model (1.2).

However, water vapor–greenhouse effect positive feedback is not the only potentially important feedback in the system. For example, consider

$$\alpha(T_s) = \delta + \gamma T_s \,, \tag{1.8}$$

where δ and γ are empirical constants. Plugging (1.8) into (1.6) and using (1.3) gives

$$\lambda_s = \frac{1}{B + \gamma Q}. \tag{1.9}$$

What does this mean? If γ is a positive number it means that when surface temperature increases albedo increases. This means that whatever causes surface temperature to increase would cause an increase in the reflectivity of the planet which would limit that original surface temperature rise, and is a negative feedback. If γ is negative then (1.9) implies positive feedback, since B is a positive number and the denominator of (1.9) would be reduced. If, in addition, the absolute value of γQ is greater than B, this system would become unstable. The greatest challenge in climate modeling (demonstrated here in linearized, zero-dimensional formalism) is to determine what the sum $B + \gamma Q$ is, based upon the many processes that interact in the climate system. Moreover, these processes which affect the feedback parameter γ do not occur uniformly across the globe; rather the global average values represented by the simple expressions derived so far are the manifestations of many local or regional changes that could be larger, smaller or even of opposite sign to the net global effect. For example, Raval and Ramanathan (1989) have shown empirically strong water vapor–greenhouse effect feedback, whereas Ramanathan and Collins (1991) have shown empirically strong negative cirrus cloud feedback in the part of the world with tropical cumulus clouds and ocean surface temperatures greater than 303 K. Nevertheless, this exercise given by (1.9) does show how sensitive the Earth's climate can be to seemingly small changes in these feedback processes.

1.2.5 Transient response

But what if we are interested in the transient response of the system to some global change forcing where the balance in (1.1) is perturbed? In this case our energy equation is rewritten to include an extra term for energy storage rate. That is, the rate of energy storage equals solar energy input to the Earth minus infrared radiant energy out. In symbolic terms

$$R\frac{\partial T_s}{\partial t} = Q(1-\alpha) - F_{ir} \tag{1.10}$$

where R is the "thermal inertia" of the system, i.e., the effective heat capacity of the atmosphere, oceans, land, etc. This, in turn, is proportional to the mass of each of the components of the climate system times their respective specific heats. If α is constant, or (to first order) its effects are lumped into the linear term B of (1.6), then (1.10) can be rewritten

$$\begin{aligned} R\frac{\partial T_s}{\partial t} &= Q(1-\alpha) - (A + BT_s) \\ &= \hat{Q}(t) - BT_s\,. \end{aligned} \tag{1.11}$$

The solution to such an ordinary differential equation is well known from classical theory, and takes the form $T_s = T_{\text{inhomo}} + T_{\text{homo}}$. The homogeneous solution, which we could call T_{pert}, would be of the form

$$T_{\text{pert}} \quad \propto \quad \exp(-t/\tau)\,. \tag{1.12}$$

where

$$\tau = R/B. \tag{1.13}$$

τ is the response time of the system to a step function forcing in Q, and is simply the heat capacity of the system divided by the radiative damping. B, the radiative damping coefficient, is also intimately involved in the sensitivity of the system. Recall, from (1.7) and (1.9), that the larger B is the less sensitive the system is to external forcing Q. What does this large damping mean physically? If B is a large number then the outgoing infrared radiation to space increases by a substantial amount if surface temperature increases. That is, the input of additional heat to the system would be damped out very effectively to space if B is a large number. That would limit the sensitivity of the system to an input of energy of any kind. For example, if a few W m^{-2} of additional energy were input by an increase in CO_2 or sunlight, then a large B would mean that only a small temperature change would be necessary to damp that extra heat back out to space. On the other hand a small B would imply very high sensitivity to small amounts of energy input, since the temperature would have to go up a great deal to damp that extra few W m^{-2} of heating back to space.

For the response time, τ, of the system, a large B implies a short response, since rapid damping means the system would approach its reduced equilibrium response more quickly than if there were little damping associated with a small B, as (1.13) shows. Thus, the feedback factors, which in linear form aggregate into B, not only affect inversely the overall sensitivity of the planet to forcing, but also its response time.

It is also required by (1.13) that the response time would be larger if the heat capacity were larger. This is intuitively obvious since a more thermally massive planet (i.e., one with a large portion of oceans or with oceans mixed deeply) would take longer to respond to global forcing than would one with relatively little heat capacity. The time-dependent model derived so far (i.e., (1.10)) is for an Earth with a fixed-heat capacity R.

The real world, of course, does not have a fixed-heat capacity, but consists of a mix of multiple-heat capacities, capacities which in fact change with the climatic state (e.g., Thompson and Schneider, 1979). For example, the heat capacity of the climate system over land is relatively small, consisting largely of the atmosphere itself and a few centimeters of soil. The response time is thus on the order of a month or so. The effective heat capacity of the middle of tropical oceans is largely governed by the depth of the oceanic mixed layer which is in contact with the atmosphere. Although that mixed layer is slowly ventilated from below, the dominant term in the heat capacity is that 50–70 m deep mixed layer. Nevertheless, the thermal inertia of the center of tropical oceans is at least an order of magnitude or so larger than that of the center of continents. However, polar seas, such as the Norwegian or Weddell Seas, in which oceanic convection causes the mixed layer to penetrate thousands of meters in depth, have effective thermal inertias another order of magnitude or so larger than that in the tropical oceans. Thus, to predict the transient response of the climate to global change forcings (e.g., CO_2 or other greenhouse gas buildups) will require not simply a global average model extended to include time dependence, but a model that has enough spatial and temporal resolution to capture the important nonglobal nature of the regionally heterogeneous effects of physical processes and heat capacities (e.g., Schneider and Thompson, 1981); this is explored in more detail in Chapter 17.

1.2.6 Hierarchy of models

The complexity of the climate system means that a hierarchy of models is necessary for studying the full response of the climate system to external forcing. The zero-order models we have described here are at the simpler end of the hierarchy, and three-dimensional, coupled atmospheric–oceanic–soil–vegetation–ice and chemistry models are at the more complex, more comprehensive end of that hierarchy. Many models in between the simple and complex ends can and have been constructed (see Chapter 10), with the virtue of the simpler models being their capacity to help us to understand the relative importance of interacting processes. The strength of the more comprehensive models is (hopefully) fidelity of simulation skill as well as their capacity to make regional, time-evolving projections of the response of the climate system to changing global forcing (simple models usually cannot resolve regional changes).

Many scientists believe that the ultimate goal of climate modeling should be fully comprehensive, three-dimensional models of all elements of the climate system including very high resolution and as much detail as possible. While such a goal is clearly appropriate for the distant future, practical considerations require compromises, as discussed in the next section.

1.3 Climate predictions

1.3.1 Empirical statistical versus "first principles"

Climate prediction, like most other forecasts of complex systems, can involve extrapolation. We attempt to determine the future behavior of the climate system from knowledge of its past behavior and present state, basically taking two approaches. One, the "empirical statistical", uses empirical statistical methods, such as regression equations with past and present observations, to obtain the most probable extrapolation. The other uses "first principles": equations believed to represent the physical, chemical, and biological processes governing the climate system for the scales of interest. The latter approach is usually called "climate modeling." Since the statistical approach depends on historical data, it is obviously limited to predicting climates that have been observed or are not caused by new processes. The statistical method cannot easily answer "what if?" questions, such as the effects of rapidly increased atmospheric carbon dioxide. Thus, the more promising approach to climate prediction for conditions or forcings different from the present or from historic precedent is climate modeling. Then, the validation of the predictions of such models becomes a chief concern.

Climate models vary in their spatial resolution, that is, the number of dimensions they simulate and the spatial detail they include. A simple model calculates only the average temperature of the Earth, independent of time, as an energy balance among the Earth's average reflectivity and the average "greenhouse" properties of the atmosphere. Such a model is zero-dimensional: it reduces the real temperature distribution on the Earth to a single point, a global average. In contrast, three-dimensional climate models reproduce the variation of temperature with latitude, longitude, and altitude. The most complex models, the General Circulation Models (GCMs), predict the time evolution of temperature plus humidity, wind, soil moisture, sea ice, and other variables through three dimensions in space (e.g., Washington and Parkinson, 1986).

Although GCMs are usually more complex than simpler models in their physical, chemical or biological detail, they are also more expensive to design, run, and interpret. The optimal level of complexity for a model depends on the problem and the resources available. More is not necessarily better. Often it makes sense to attack a problem first with a simple model and then employ the results to guide research at higher resolution. In other words, deciding how complicated a model to use for a task and whether to trade completeness and accuracy for tractability and economy is more an intuitive judgment than a scientific choice subject to explicit, logical criteria (e.g., Land and Schneider, 1987).

1.3.2 Grids and parameterization

Even the most complex GCM is limited in the spatial detail it can resolve. No computer is fast enough to calculate climatic variables everywhere on the Earth and in the atmosphere in a reasonable time. Instead, calculations are executed at the widely spaced points of a three-dimensional grid at and above the surface. For a typical example, divide the surface of the Earth into a grid of 1,920 squares, each 4.5° latitude by 7.5° longitude. At 40° latitude each square is 500 by 640 km. Then divide the atmosphere above each square into nine strata. The calculation of a year of simulated "weather" in these 17,280 grid boxes by 30 minute increments takes some 10 hours on a Cray X-MP computer.

Wide spacing creates a problem. Many climatic phenomena occur over smaller scales than an individual "box" of the grid. For example, clouds reflect much incident sunlight back to space, and they also block the escape of infrared radiation from below, thus influencing the greenhouse effect. Therefore, they help to determine the temperature on the Earth. Predicting changes in cloudiness is, therefore, an essential part of climate simulation. No GCM now available or likely to be available in the next few decades, however, has a grid fine enough to resolve individual clouds, which tend to shade a few kilometers rather than a few hundred kilometers.

Subgrid-scale phenomena like clouds are represented collectively by parameterization (short for parametric representation) rather than individually. A parameterization could, for example, be based on climatological data to derive statistical relations between variables that are resolved by the grid and ones that are not. For instance, the average temperature and humidity over, say, the large area beneath one box, can be related to the average cloudiness over the same area; to make the equation work one introduces a parameter or proportionality factor derived empirically from the cloudiness, temperature, and humidity data. Since a model can calculate the temperature and humidity over a box from physical principles, the semi-empirical parameterization predicts the average cloudiness in the box even though it cannot predict individual clouds. Modelers, of course, strive to keep their parameterizations as physical and nonempirical and scale-independent as practical. Thus, the validity of parameterization and overall model performance as well, depends ultimately on empirical tests, not only on the inclusiveness of the first principles. In other words, even our most sophisticated "first principles" models contain "empirical statistical" elements within the model structure.

Climate sensitivity and scenarios

Uncertainty about parameterizations of feedback mechanisms like clouds or sea ice is one reason the goal of climate modeling – reliable, verified forecasting of key variables such as temperature and rainfall on a regional, time-evolving basis – is not attainable yet. Another source of uncertainty external to the models is human behavior. Forecasting, for example, the impact of carbon dioxide on climate requires knowing how much carbon dioxide is going to be emitted (e.g., Nordhaus and Yohe, 1983) and how that emission will be distributed or removed by the physical, chemical, and biological processes of the carbon cycle as already noted on Plate 1.

What the climate models can do well is analyze the sensitivity of the climate to uncertain or even unpredictable variables. In the case of carbon dioxide, one could construct plausible scenarios of economic, technological and population growth to project growth of CO_2 emission and model the climatic consequences (e.g., as done by IPCC, 1990a). Such uncertain climatic factors as cloud-feedback parameters could be varied over a plausible range or alternative parameterization formulations can be used. The calculations would indicate which uncertain parameter or formula are most important in making the climate more or less sensitive to carbon dioxide increase. One could then concentrate research on narrowing the uncertainty surrounding those factors. The results of such sensitivity tests would also suggest the range of climatic futures that ecosystems and societies may be forced to adapt to and at what rates. How to respond to such information, of course, is a value-laden issue, which is examined in other chapters and has been addressed by the author elsewhere (Schneider, 1990).

Theoretical issues

Although the atmosphere–earth–ice–ocean system is complex, we can describe the known physical laws mathematically, at least in principle. In practice, however, solving these equations in full, explicit detail is impossible. First, the possible scales of motion in the atmospheric and oceanic components range from the submolecular to the global. Second are the interactions of energy transfers among the different scales of motion. Finally, many scales of disturbance are inherently unstable; small disturbances, for example, grow rapidly in size if conditions are favorable. Thus, seemingly small differences between two similar atmospheric or oceanic states cause later divergence. Meteorologists have found from theoretical considerations and from experience that useful detailed weather prediction of beyond about 10 days is impossible using current observations (e.g., Somerville, 1987). Is the longer-term climate prediction thus a hopeless task?

Several reasons, however, make prediction of climate, in contrast to weather, feasible for comparatively long periods. For one thing, although day-to-day weather is not predictable far in advance, some success can be obtained in predicting average conditions for an extended period. The situation is approximately analogous to the statistical-mechanical theory of gases: although we cannot predict the behavior of individual molecules, we can accurately predict the expected mean state and variance of an ensemble of molecules for some sets of conditions.

Another reason that climate predictions for longer periods may be possible is that the climate system is subject to forcing processes that may be of overriding importance for some time or space scales. An obvious example is the annual variation in the global distribution of solar radiation. The strength of this forcing causes the seasons to follow each other predictably, although differences (anomalies) in seasons are, of course, important from year to year (e.g., Namias, 1972; Trenberth et al., 1988).

Some forcing mechanisms are predictable whereas others, such as volcanic activity, are largely unpredictable today. Also, the atmosphere is forced by such mechanisms as oceanic surface temperatures that themselves respond gradually to atmospheric forcing, a complicating feedback. These feedbacks are commonly referred to as "internal forcings", in contrast to the straightforward "external forcings" from outside such as solar radiation. What is internal on long-time scales may be external on shorter ones, depending on what processes are included in the climatic system defined for an investigation.

In any event, the presence of forcing implies that some aspects of climate may be predictable on those time scales where the forcing and its response are important. In fact, one would expect the degree of climate response, and hence predictability, to be related to both the amplitude and the period of forcing, or, for aperiodic cases, the time scale of the forcing. Although this is true in general for external forcing, nonlinear feedbacks within the system can produce unexpected results such as periodic or aperiodic oscillations that come or go with small changes in the values of internal model parameters (e.g., see Ghil, 1981; and Harvey and Schneider, 1984). For instance, internal interactions can cause internal stochastic oscillations (e.g., Hasselmann, 1976) on different time scales or even chaotic behavior (e.g., Lorenz, 1968).

For many longer time scales, it is widely believed the global-scale climate system responds deterministically to current "boundary conditions" and has little memory of its history. In other words, the climate system will move to a unique equilibrium after a transient adjustment for these longer times, as Imbrie and Imbrie (1979) assumed when modeling glacial–interglacial cycles – just as others typically assume in CO_2 doubling experiments. Thus, deterministic models provide an equilibrium "snapshot" appropriate for the time scale of the external forcing. The validity of such equilibrium experiments, however, depends on the existence of a unique equilibrium for the given boundary conditions, which is still a debated assumption for the Earth's climate system (e.g., Lorenz, 1968; Schneider and Gal-Chen, 1973; North, 1975).

1.3.5 Scale transition

Finally, within the foreseeable future even the highest resolution three-dimensional GCMs will not have a grid much less than 100 km. They will not, therefore, be able to resolve individual thunderstorms, or the important local or mesoscale effects of hills, coastlines, lakes, vegetation boundaries, or heterogeneous soils. For regions that have relatively uniform land surface characteristics, such as a thousand kilometer scale savannah or a tropical forest with little elevation change, GCM grid-scale parameterizations of surface albedo, soil type, and evapotranspiration could adequately be used to estimate local changes. Alterations in climate predicted within a box would probably apply fairly uniformly across such nicely behaved, homogeneous areas. On the other hand, steep topography over watersheds smaller than GCM grids can mediate regional climate. Therefore, even if GCM predictions were accurate at grid scale, they would not necessarily be appropriate to local conditions.

Large-scale observed climatic anomalies are translated to local variations in Fig. 1.3 (Gates, 1985). This analysis of the local climatic variability for the state of Oregon was based on several years of data using a technique known as *empirical orthogonal functions*[1]. The north–south Cascade Mountains translate a simple change in the frequency or intensity of westerly winds into a characteristic climatic signature of typically wetter on the west slope and drier on the east or vice versa. In other words, a GCM producing altered westerlies in response to say, CO_2 doubling, could be applied to the map on Fig. 1.3 to determine the impact in a local watershed. Such a map, constructed from variations of climate observed over several years, seems an ideal way to translate information at a GCM grid scale to the local or mesoscale. Because empirical data have been used, however, such a relation would only be valid if the causes of recent climatic variations or oscillations which gave rise to Fig. 1.3 carried forward and included the effects of climatic changes forced by trace gases. It is not obvious that the signature of climatic change from increases in trace gases will be the same as

[1] Empirical orthogonal functions (EOFs) are widely used to analyze spatial fields into their principal patterns of temporal variability, see Morrison, 1976, Chapter 8.

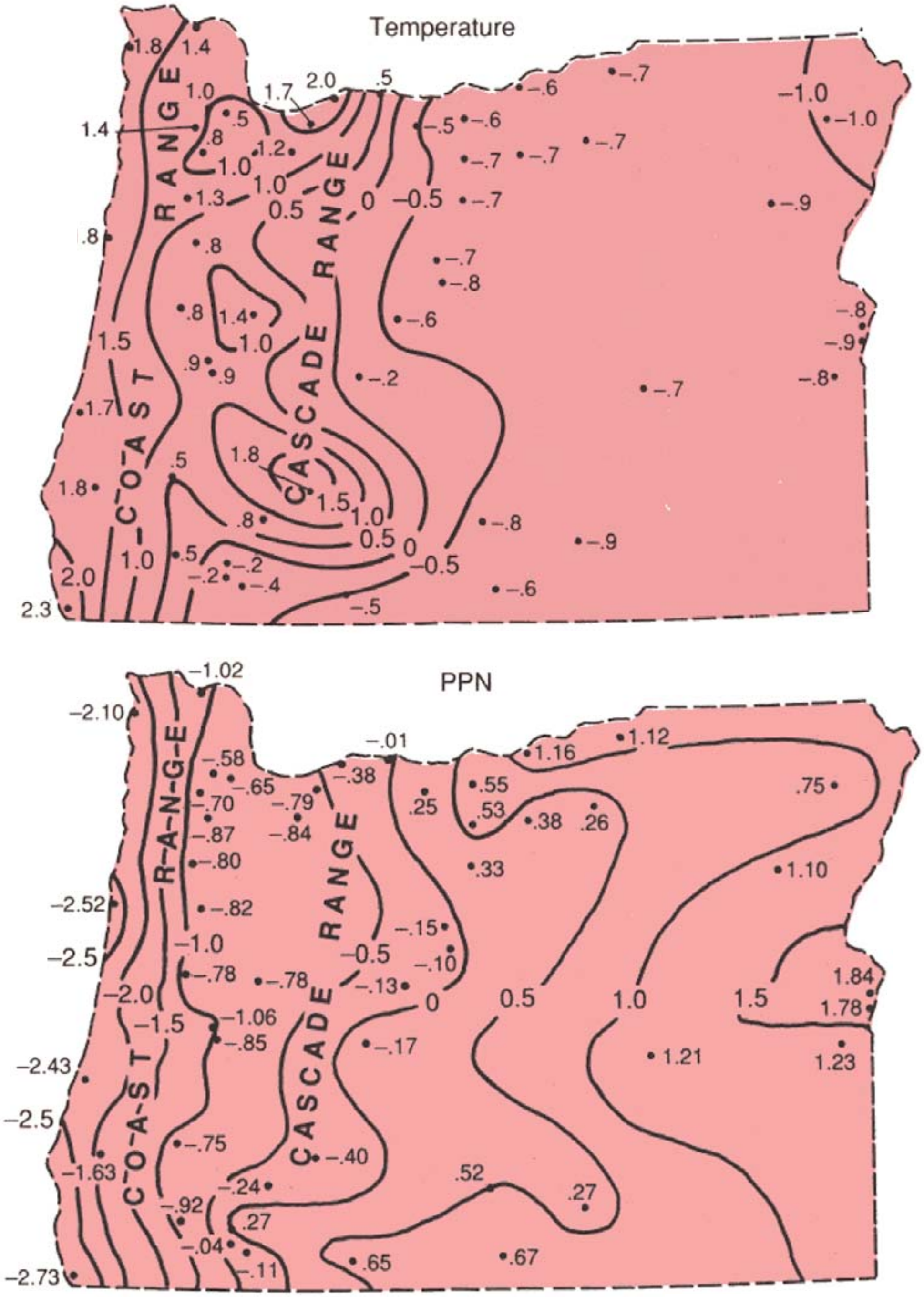

Fig. 1.3 The distribution of the relationship between large-scale (area-averaged) and local variations of the monthly mean surface air temperature (above) and precipitation (below), as given by the first empirical orthogonal function determined from thirty years of observational monthly means at 49 stations in Oregon in comparison with the state-wide average.

past vacillations, many of which could have been internal oscillations within the climate system, not created by external forcing such as changes in trace gases. Thus other translations of scale need to be considered.

One might embed a high resolution "mesoscale model" within a few boxes of a GCM, using as boundary conditions for the mesoscale model the wind, temperature, and so forth predicted by the GCM at the mesoscale model's boundaries. The mesoscale model could then account for local topography, soil type, lakes and vegetation cover, and translate GCM forecasts to local scale. Such embedding techniques have shown considerable early promise (e.g., Giorgi, 1990). But for such a method to have any reasonable hope of success, however, the GCM must produce accurate climatic statistics for the special grid area. To return to the Oregon case in Fig. 1.3, for example, if the climatic average of the GCM's winds in the unperturbed case (i.e., the "control" case) has the wrong westerly component, the local climate change in a region where topography amplifies any such error in the wind direction will probably be misrepresented. A prerequisite to performing scale transition through embedding of local or regional models, therefore, is assurance that the control climate of the GCM for the important variables is accurate enough that it makes sense to take the next step of imposing a scenario of trace gas increase on the GCM to estimate how the local-scale climate might change.

Practically, while testing scale transitions in steep topography and other rapidly varying local features proceeds, modelers should examine the behavior of their models using grid boxes that are much less pathological. That is, examine boxes where local features are relatively homogeneous and where translation of local-to-grid scales should prove a less serious obstacle.

1.4 Validation

The most perplexing question about climate models is whether they can be trusted to provide grounds for altering social policies, such as those governing carbon dioxide emissions. How can models so fraught with uncertainties be verified? There are actually several methods. Although none is sufficient alone, they can together provide significant, albeit circumstantial, evidence of a model's credibility.

The first verification method is checking the model's ability to simulate today's climate. The seasonal cycle is one good test because the temperature changes are larger, on a hemispheric average, than the change from an ice age to an interglacial period (i.e., 15°C seasonal range versus 5°C glacial–interglacial cycle, respectively, in the NH). GCMs map the seasonal cycle well, which suggests that their surface temperature sensitivity to large-scale radiative forcing is not way off. The seasonal test, however, does not indicate how well a model simulates such slow processes as changes in deep ocean circulation or ice cover, which may have an important effect on the decade to century time scales over which CO_2 is expected to double.

A second verification technique is isolating individual physical components of the model, such as its parameterizations, and testing them against reality. For example, one can check whether the model's parameterized cloudiness statistics matches the observed cloudiness statistics of a particular box. But this technique cannot guarantee that the complex interactions of individual model components are properly treated. The model may be good at predicting average cloudiness but bad at representing cloud feedback. In that case, simulation of overall climatic response to, say, increased carbon dioxide is likely to be inaccurate. A model should reproduce to better than, say, 10% accuracy the flow of thermal energy between the atmosphere, surface, and space (see Fig. 1.2). Together, these energy flows comprise the well-established *greenhouse* effect on Earth and constitute a formidable and necessary test for all models. A model's performance in simulating these energy flows is an example of physical verification of model components. As another example, Raval and Ramanathan (1989) compared the enhanced infrared heat trapping with increasing surface temperature by using satellite observations (e.g., see Chapter 10). They compared observed water vapor–greenhouse effect feedback calculations in GCMs against satellite observations.

A third method for determining overall, long-term simulation skill is the model's ability to reproduce the diverse climates of the ancient Earth (e.g., see Chapter 21) or even of other planets (e.g., Kasting et al., 1988). Paleoclimatic simulations of the Mesozoic Era, glacial–interglacial cycles, or other extreme past climates help in understanding the coevolution of the Earth's climate with living things. As verifications of climate models, however, they are also crucial to estimating both climatic and biological future (Schneider, 1987).

Overall validation of climatic models thus depends on constant appraisal and reappraisal of performance in the above categories. Also important are a model's response to such century-long forcings as the 25% increase in carbon dioxide and other trace greenhouse gases since the Industrial Revolution. Indeed, most climatic models are sensitive enough to predict that warming of 1°C should have occurred during the past century. The precise "forecast" of the past 100 years also depends upon how the model accounts for such factors as changes in the solar constant or sulfur dioxide emissions or volcanic dust (e.g., Schneider and Mass, 1975, or Hansen et al., 1981). Indeed as Fig. 1.4 (Wigley and Raper, 1990) shows using a highly simplified one-dimensional climate model of atmosphere and oceans, the typical prediction of an 0.5 to 1°C warming over the twentieth century is broadly consistent with, but somewhat larger than, observed. Possible explanations for the discrepancy include (see Gilliland and Schneider, 1984): (a) the sensitivity of state-of-the-art models to trace gas greenhouse increases have been overestimated some twofold; (b) modelers have not properly accounted for such competitive external forcings as volcanic dust or changes in solar energy output; (c) modelers have not accounted for other external forcings such as regional tropospheric aerosols from biological, agricultural

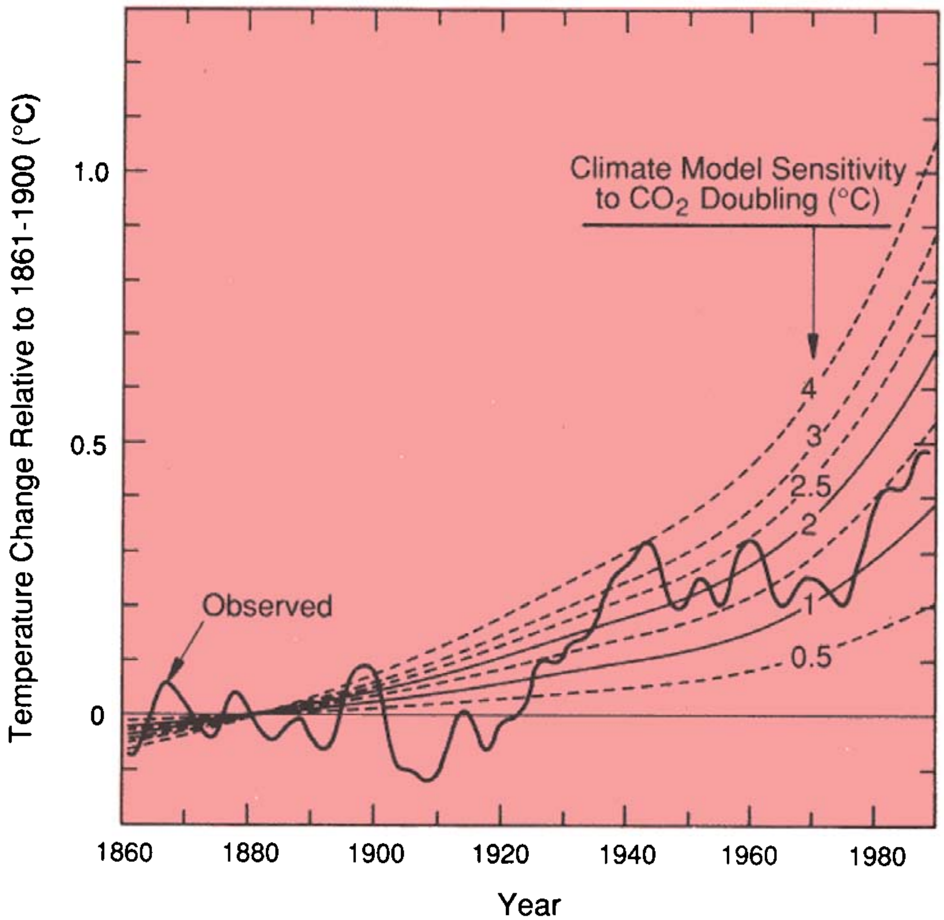

Fig. 1.4 Observed global-mean temperature changes (1861–1989) compared with predicted values. The observed changes are given in Sec. 7 of the IPCC (1990a). The data have been smoothed to show the decadal and longer time scale trends more clearly. Predictions are based on observed concentration changes and concentration–forcing relationships as given in Sec. 2 of the IPCC (1990a) and have been calculated using the highly simplified upwelling-diffusion climate model of Wigley and Raper (1990).

and industrial activity; (d) modelers have not properly accounted for the large heat capacity of the oceans taking up some of the heating of the greenhouse effect and delaying warming of the atmosphere; (e) both present models and observed climatic trends could be correct, but models are typically run for equivalent doubling of carbon dioxide whereas the world has only experienced a quarter of this increase and nonlinear processes have been properly modeled and produced a sensitivity appropriate for doubling but not for 25% increase; (f) the incomplete and inhomogeneous network of thermometers has underestimated warming; and (g) there may have been a natural cooling trend of up to 0.3–0.5°C during the twentieth century.

Despite this litany of excuses why observed global temperature trends in the past century and those anticipated by most GCMs (i.e., +2 to 5°C for a CO_2 doubling) disagree somewhat, the approximately twofold discrepancy is not fundamental. Most climatologists do not yet proclaim the observed temperature records to have been caused beyond doubt by the greenhouse effect. Depending upon what assumptions one makes [e.g., (a)–(g) above], the twentieth century surface temperature trend could be consistent with an equivalent CO_2 doubling, equilibrium temperature response of 0.5–5.0°C (Wigley and Raper, 1990)! Thus, a greenhouse effect signal cannot yet be said to be unambiguously detected in the record. It is still possible that the observed trend and the predicted warming could be chance occurrences. One cannot easily rule out that other factors, such as natural fluctuation or solar constant variations or anthropogenic dust, simply have not been adequately accounted for over the past century – except during the past decade when adequate instruments have been measuring the last two. Nevertheless, this empirical test of model predictions against a century of observations certainly is consistent with a rough factor of 3. This test is reinforced by the good simulation by most climatic models of the seasonal cycle, diverse ancient paleoclimates, hot conditions on Venus, cold conditions on Mars (both well simulated), and the present distribution of climates on Earth. When taken together, these verifications provide a strong circumstantial case that the modeling of sensitivity of temperature to greenhouse gases is probably valid within threefold (as also was suggested by IPCC, 1990a).

Another decade or two of observations of trends in Earth's climate, of course, should produce signal-to-noise ratios sufficiently obvious that almost all scientists will know whether present estimates of climatic sensitivity to increasing trace gases have been predicted well or not. Unfortunately, the global change "experiments" now underway are not merely academic exercises in the microchips of supercomputers, but are being performed (as noted nearly four decades ago by Revelle and Suess, 1957) on the "laboratory" that we and all other living things share – Earth.