Part II The State of the Art in Environmental Modelling

9 Climate and Climate-System Modelling

L.D. Danny Harvey

Department of Geography and Planning, University of Toronto, Canada

9.1 The complexity

Climate and climatic change are particularly difficult to model because of the large number of individual components involved in the climate system, the large number of processes occurring within each component, and the multiplicity of interactions between components. The climate system consists of the atmosphere, oceans, cryosphere (glaciers, ice caps, sea ice and seasonal snow cover), biosphere, and lithosphere (the Earth's crust). All of these components affect, and are affected by, the other components, so they form a single system. For example, the atmosphere and oceans influence each other through the exchange of momentum (by winds), heat and moisture. The growth and decay of glaciers, and the formation of sea ice and land-snow cover, depend on atmospheric temperature and the supply of moisture, but ice and snow surfaces in turn influence the absorption of solar energy and hence influence temperature, and modulate the flow of water vapour and heat between the land or ocean and the atmosphere. Continental-scale ice sheets, such as have periodically occupied most of Canada and northern Europe, reflect sunlight, deflect winds and depress the crust, leading to lower ice-sheet elevations that contribute to their periodic collapse. Meltwater during the rapid demise of ice sheets is likely to have altered ocean circulation. The biosphere is affected by atmospheric and/or oceanic conditions but also influences climate through the effect of land vegetation on surface roughness and hence on winds, and through its effect on evaporation and the reflection or absorption of solar energy. Micro-organisms in the upper ocean affect the reflectivity of the ocean surface to sunlight. Both the marine and terrestrial biosphere play an important role in determining the atmospheric concentration of a number of climatically important trace gases, the most important being carbon dioxide (CO_2) and methane (CH_4). Finally, the Earth's crust influences the climate through the influence of land topography on winds and the distribution of rain, through the role of the ocean-continent configuration and the shape of the ocean basin on ocean currents, and through the role of chemical weathering at the Earth's surface and the deposition of marine carbonates in modulating the atmospheric concentration of CO_2 at geological time scales. Even this thumbnail sketch of the climate system leaves out many important considerations. The occurrence of clouds, which are part of the atmosphere, is strongly influenced by conditions at the land surface. Their optical properties are influenced in part by micro-organisms in the upper few metres of the ocean, which emit sulphur compounds that ultimately become cloud-condensation nuclei. The occurrence of clouds dramatically affects the flow of both solar and infrared radiation between the atmosphere, land surface and space. Many chemical reactions occur in the atmosphere and they determine the concentrations of climatically important trace gases and aerosols; many of the important reactants in atmospheric chemistry are released from the terrestrial or marine biosphere. Cloud droplets serve as important sites for many important chemical reactions, and the optical properties of clouds are themselves influenced to some extent by the chemical reactions that they modulate. A more thorough

Environmental Modelling: Finding Simplicity in Complexity, Second Edition. Edited by John Wainwright and Mark Mulligan.

discussion of the interactions between and within the various components of the climate system can be found in Chapter 2 of Harvey (2000).

In building computer models of the climate system, as with any system, there are a number of basic considerations. The first involves the number of components to be included. The various components of the climate system change at different time scales. For example, at the decade to century timescale, changes in the extent of the large ice caps (Greenland, Antarctica) can be ignored. There would thus be no need to include models of these components for simulation of 100 years or less; rather, the observed present-day distribution can be simply prescribed as a lower boundary condition on the atmosphere. Similarly, changes in the geographical distribution of the major terrestrial biomes can be ignored at this time scale. However, at longer time scales, changes in biomes and in ice sheets, and their feedback effects on the atmosphere and oceans, would need to be considered. Thus, the comprehensiveness of a climate model (the number of components retained) depends in part on the timescale under consideration. The comprehensiveness is also dictated in part by the particular purpose for which one is building the climate model.

The flip side to comprehensiveness is model complexity. Generally, more comprehensive models tend to be less complex – that is, each component represents fewer of the processes that occur in reality, or represents them in a more simplified manner. This simplification is because the more comprehensive models tend to be used for longer time scales, so limitations in computing power require that less detailed calculations be performed for a given period of simulated time. Furthermore, the addition of more climate system components also tends to increase the overall computational requirements, which can be offset by treating each component in less detail.

9.2 Finding the simplicity

Nature varies continuously in all three spatial dimensions, thus comprising an infinite number of infinitesimally close points. However, due to the finite memory capacity of computers, it is necessary to represent variables at a finite number of points, laid out on a grid of some sort. The calculations are performed only at the grid points. The spacing between the grid points is called the model resolution. In global atmospheric models the typical horizontal resolution is 200 to 400 km. In ocean models the resolution can be as fine as tens of kilometres. Many important elements of the climate system (such as clouds, land-surface variation) have scales much smaller than this. Detailed models at high resolution are available for such processes by themselves, but these are computationally too expensive to be included in a climate model, and the climate model has to represent the effect of these sub-grid-scale processes on the climate system at its coarse grid-scale. A formulation of the effect of a small-scale process on the large scale is called a parameterization. All climate models use parameterizations to some extent. Some parameterizations inevitably include constants that have been tuned to observations of the current climate, and which might not be entirely valid as the climate changes (see Chapter 2 for further approaches to model parameterization and Chapter 5 for issues of parameter scaling).

Another kind of simplification used in climate models is to average over a complete spatial dimension. Instead of, for instance, a three-dimensional longitude-latitude-height grid, one might use a two-dimensional latitude-height grid in models of the atmosphere or oceans, with each point being an average over all longitudes at its latitude and height (examples include Peng *et al.*, 1982; Yao and Stone, 1987, and Stone and Yao, 1987 for the atmosphere; and Wright and Stocker, 1991, for the ocean). Another choice is to average in both horizontal dimensions, retaining only the vertical dimension, as in one-dimensional radiative-convective models that have been used in the detailed simulation of the vertical transfer of solar and infrared radiation and in studies of the effects of changes in the composition of the atmosphere (examples include Manabe and Wetherald, 1967; Lal and Ramanathan, 1984; Ko *et al.*, 1993) and the one-dimensional upwelling-diffusion ocean model that has been used to study the role of oceans in delaying the surface-temperature response to increasing greenhouse-gas concentrations (Hoffert *et al.*, 1980; Harvey and Schneider, 1985). A third choice is to average vertically and in the east-west direction but to retain the north-south dimension, as in the classical energy-balance climate models. These models have provided a number of useful insights concerning the interaction of horizontal heat transport feedbacks and high-latitude feedbacks involving ice and snow (e.g. Held and Suarez, 1974). When the dimensionality is reduced, more processes have to be parameterized but less computer time is required.

The equations used in climate models are a mixture of fundamental principles that are known to be correct (such as Newton's laws of motion and the First Law of Thermodynamics), and parameterizations. Parameterizations are empirical relationships between model-resolved variables and subgrid scale processes that are not rigorously derived from physical principles (although the overall form of the parameterization might be derived from physical considerations), but are derived from observed correlations. An example of a parameterization would be to compute the percentage of cloud cover in a given atmospheric grid box based on the grid-averaged relative humidity and whether the vertical motion is upward or downward. However, because parameterizations are not rigorously derived, the observed correlation might not be applicable if the underlying conditions change – that is, as the climate changes.

9.2.1 Radiative forcing, radiative feedbacks and climate sensitivity

The climate of the Earth automatically adjusts itself so as to achieve a balance (in the global average) between the absorption of solar radiation and the emission of infrared radiation to space. An increase in atmospheric CO_2 concentration, or the addition of aerosols to the atmosphere, changes climate by upsetting this balance; any change in this balance is called the radiative forcing [and has units of $W\,m^{-2}$ – the change in net radiation (absorbed solar minus emitted infrared) averaged over the entire area of the Earth]. If there is a radiation surplus, the climate begins to warm, but as the climate warms the emission of infrared radiation to space increases, reducing the surplus. However, as soon as the climate begins to warm, other things that affect the radiative balance also change. Any process that adds to an initial imbalance is a positive feedback, as a larger temperature is then required in order to restore radiative balance (an example of a positive feedback would be the increase in atmospheric water vapour – a greenhouse gas – or the decrease in the area of highly reflective snow and sea ice as the climate warms). Conversely, any process that subtracts from the initial imbalance is a negative feedback, as it results in a smaller temperature change in response to the initial radiative forcing (an example of a negative feedback would be a hypothetical increase in the amount of low-level cloud as the climate warms). The ratio of the long-term change in global mean temperature to the radiative forcing is called the climate sensitivity. However, the term is now used by climatologists to refer to the eventual global mean warming specifically for a fixed doubling in the concentration of atmospheric CO_2. The consensus for the past 30 years is that the climate sensitivity is very likely to fall between 1.5 °C and 4.5 °C.

The feedbacks that determine the climate sensitivity are referred to as fast feedbacks. These involve things such as changes in the amount and vertical distribution of water vapour in the atmosphere, changes in the extent of seasonal ice and snow, changes in the rate of variation of temperature with height in the atmosphere (the so-called lapse rate) and changes in the amount, location and properties of clouds (Harvey, 2000: Chapter 3; Randall *et al.*, 2007: Section 8.6). These characteristics of the climate system all respond to changes in temperature over a period of days to months. However, other changes would occur over a period of decades to centuries, which would lead to further changes in the radiative balance. These include changes in the distribution and extent of different plant types, changes in the extent of ice caps, and changes in the concentration of CO_2 itself. These are referred to as slow feedbacks. Most prominent are a variety of positive feedbacks between climate and the carbon cycle, including:

1. The likely transition of the terrestrial biosphere from a net sink (absorber) to a net source (emitter) of CO_2 due to eventual adverse effects on increasing temperature on photosynthesis and an increase in the rate of respiration with increasing temperature (Matthews *et al.*, 2007).
2. The catastrophic transition of carbon-rich biomes (such as the Amazon rainforest) to grasslands due to the transition to a dryer climate (Friedlingstein *et al.*, 2006).
3. Substantial releases of CO_2 and CH_4 from thawing permafrost (Khvorostyanov *et al.*, 2008; Schuur *et al.*, 2009; Dorrepaal *et al.*, 2009).

Thus, the long-term climate response to a doubling of the CO_2 concentration caused by human emissions could be substantially greater than the climate sensitivity based on fast feedbacks alone because the CO_2 concentration would not be limited to the initial CO_2 doubling (that is, the assumption of a fixed CO_2 increase that underlies the concept of climate sensitivity would be violated).

9.2.2 Early coupled atmosphere-ocean general circulation models (AOGCMs)

The most detailed and realistic climate models are coupled three-dimensional atmospheric and oceanic general circulation models (AOGCMs). These models divide the atmosphere and ocean into a horizontal grid with a typical resolution of 2–4° latitude by 2–4° longitude in the latest models, with typically 10 to 20 layers in the vertical in both the atmosphere and ocean. They directly simulate winds, ocean currents, and many other features and processes, as illustrated in Figure 9.1. AOGCMs compute the transfer of solar and infrared radiation through the atmosphere, and so automatically compute the radiative forcing associated with changes in greenhouse-gas or aerosol concentrations, as well as the fast feedback processes that collectively determine the climate sensitivity. Atmospheric and oceanic general circulation models also compute the absorption of heat by the oceans, which delays and distorts the surface temperature response to increasing greenhouse-gas concentrations. McGuffie and Henderson-Sellers (2005: Chapter 5) provide a compact introduction to AOGCMs, while their strengths and weaknesses in simulating the present climate are discussed in Randall *et al*. (2007) and Reichler and Kim (2008).

The early AOGCMs computed the transfer of heat and moisture between the land surface and atmosphere using prescribed vegetation types (such as rainforest, grassland or savannah) with fixed properties for each vegetation type. A key parameter controlling the rate of evapotranspiration is the stomatal resistance r_s, and in early AOCGMs r_s was computed as some simple function of the amount of moisture in the model soil.

9.2.3 Adding the terrestrial biosphere component of the carbon cycle to AOGCMs

The most recent models compute the distribution of ten or so plant functional types (PFTs), with some allowing only one PFT in a given grid cell and others allowing for a mixture of PFTs in each grid cell. Some recent models also include a simple representation of the terrestrial carbon cycle in each grid cell, as illustrated in Figure 9.2. Carbon fluxes (photosynthesis, respiration and transfers) are computed several times per hour, then summed annually and used to compute the change from one year to the next in the amount of carbon as leafy and woody plant biomass, as detritus, and as soil carbon. The photosynthetic carbon flux depends in part on the stomatal resistance, with the stomatal resistance now dependent not only on soil moisture but also on atmospheric CO_2 concentration

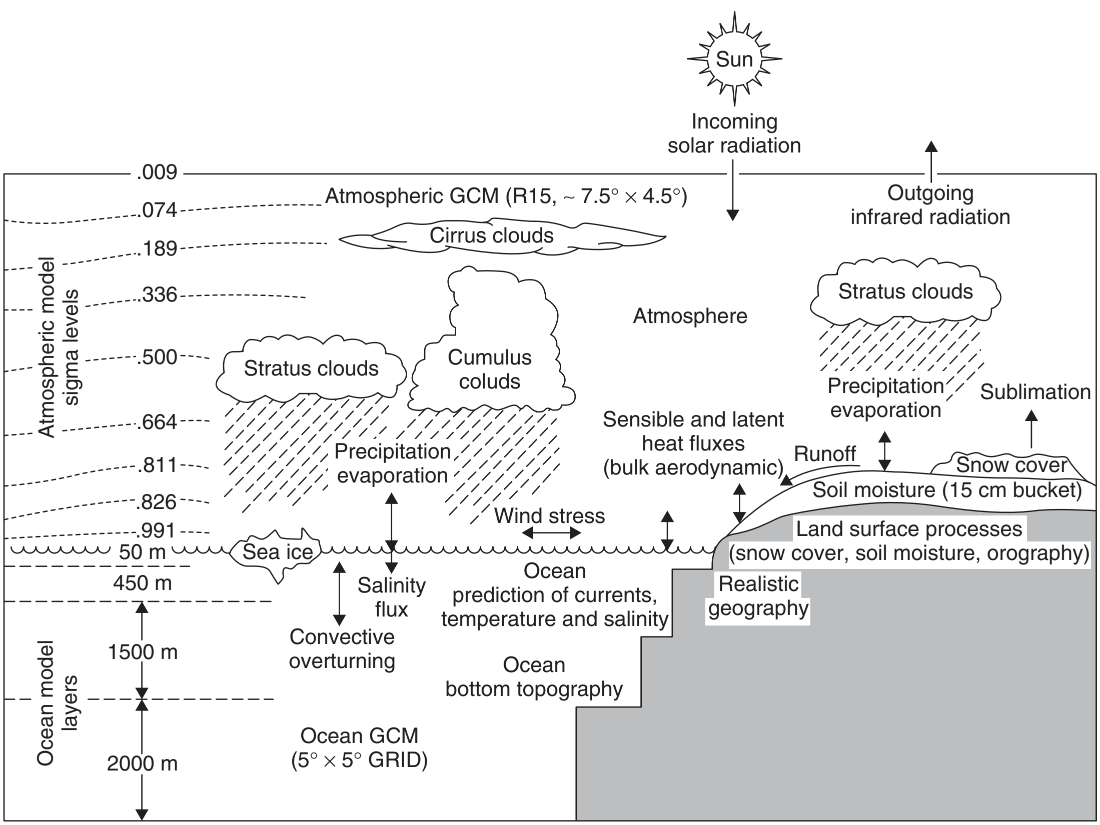

Figure 9.1 Illustration of the major processes occurring inside a typical horizontal grid cell of an atmosphere-ocean general circulation model (AOGCM). Reproduced from Washington and Meehl (1989), where a full explanation can be found. 'Sigma level' refers to the ratio of pressure at a give height to the surface pressure.

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Figure 9.2 The structure of the terrestrial biosphere model that is applied at each land grid square in the Canadian Climate Centre AOGCM (Reproduced with permission from Arora, V.K., Boer, G.J., Christian, J.R., *et al.* (2009). The effect of terrestrial photosynthesis down regulation on the twentieth century carbon budget simulated with the CCCma Earth System Model. *Journal of Climate*, 22, 6066–88. © American Meteorological Society).

and photosynthetic demand. Fluxes of moisture and CO_2 between the land and atmosphere are thereby coupled in the latest models. In this way, the terrestrial biosphere component of the global carbon cycle has been incorporated into the latest AOGCMs. This coupling permits the simulation of carbon fluxes to or from the atmosphere in response to changes in the distribution of biomes (including potential collapse of the Amazon rainforest) and due to changes in the carbon balance of individual biomes due to the direct physiological effects of higher atmospheric CO_2 on photosynthesis and due to changes in climate. Thus, some of the potential slow positive climate-carbon cycle feedbacks can be included.

9.2.4 Adding the oceanic component of the carbon cycle to AOGCMs

The exchange of CO_2 between the atmosphere and ocean depends on the difference between the partial pressure of CO_2 in the atmosphere and in the surface water that is in contact with the atmosphere. The latter in turn depends on the temperature, salinity, alkalinity and concentration of total dissolved carbon $(CO_2 + HCO_{3-} + CO_3^{2-})$ in the surface water. The ocean GCM component of an AOGCM computes ocean currents and turbulent mixing processes that determine, in part, the distribution of temperature and salinity in the ocean. The OGCM can also be used to compute the distribution of dissolved carbon, alkalinity and nutrients in the ocean. Marine micro-organisms are critical to the distribution of carbon, alkalinity and nutrients in the oceans, as photosynthesis in the surface layer leads to the incorporation of dissolved carbon into organic tissue, some of which settles to the deep ocean, where it is released through respiration. The rate of photosynthesis depends in part on the supply of nutrients in the surface layer, which in turn depends on the intensity of removal in sinking organic material and the upwelling of nutrient-rich deep water. Some micro-organisms build skeletal material out of calcium carbonate ($CaCO_3$), which also settles to the deep ocean, carrying carbon and alkalinity with it. Figure 9.3 illustrates a typical marine biosphere model as used in AOGCMs. By embedding a model of the marine ecosystem in each surface grid cell of an ocean GCM, and combining this with calculations of the distribution of carbon, alkalinity and nutrients in the ocean and of the partial pressure of CO_2 in the surface water, much of the oceanic component of the carbon cycle can be included in an AOGCM

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Figure 9.3 The structure of the marine ecosystem model that is applied at each ocean surface layer grid cell in the Canadian Climate Centre AOGCM (Reproduced with permission from Arora, V.K., Boer, G.J., Christian, J.R., *et al.* (2009). The effect of terrestrial photosynthesis down regulation on the twentieth century carbon budget simulated with the CCCma Earth System Model. *Journal of Climate*, 22, 6066–88. © American Meteorological Society).

(this description omits a sub-module for the build-up or dissolution of carbonate-rich sediments on the sea floor, which is relevant at times scales of 1000 years and longer). The combination of photosynthesis in the surface layer and downward settling of carbon and its release at depth is referred to as the biological pump. By transferring carbon to the deeper waters, a much lower CO_2 partial pressure is maintained in the ocean surface layer, which in turn produces a much lower atmospheric CO_2 concentration than would otherwise be the case (namely, a preindustrial concentration of about 280 ppmv rather than 450 ppmv or higher in the absence of the biological pump).

Coupled AOGCMs with the carbon-cycle components embedded in the ocean GCM can simulate the absorption of anthropogenic CO_2 by the oceans in response to increasing atmospheric CO_2 concentration, as well as the impact of change in climate on the oceanic absorption or release of CO_2. The biological pump, although crucial to the present-day distribution of carbon in the oceans and to the pre-industrial atmospheric CO_2 concentration, plays no role in the absorption of the additional CO_2 added to the atmosphere by humans unless its strength changes for some reason. Rather, the oceanic absorption of CO_2 occurs through the inorganic processes of air-sea gas exchange and the generally slow downward mixing of carbon-rich water. However, the biological pump is likely to change as the climate itself changes, both due to changes in ocean circulation altering the supply of nutrient to the surface layer and due to changes in marine ecology (and hence in the proportions and vitality of different kinds of micro-organisms). These changes constitute a climate-carbon cycle feedback. Another climate-carbon cycle feedback arises through the decrease in the solubility of CO_2 as the surface layer of the ocean warms, causing an outgassing of CO_2 that is estimated to be about 10% of the oceanic absorption of CO_2 that would otherwise occur (i.e. an outgassing of about 0.2 Gt C yr^{-1} compared to a background absorption of about 2 Gt C yr^{-1} at present) (Goodwin and Lenton, 2009).

9.2.5 Models of atmospheric chemistry and aerosols

Atmospheric chemistry is central to the distribution and amount of ozone in the atmosphere. The dominant chemical reactions and sensitivities are significantly different for the stratosphere and troposphere. These processes can be adequately modelled only with three-dimensional atmospheric models (in the case of the troposphere) or with two-dimensional (latitude-height) models (in the case of the stratosphere). Atmospheric chemistry is also critical to the removal of CH_4 from the atmosphere and, to a lesser extent, all other greenhouse gases except H_2O and CO_2. In the case of CH_4, a change in its concentration affects its own removal rate and, hence, subsequent concentration changes. An accurate simulation of changes in the removal rate of CH_4 requires specification of the concurrent concentrations of other reactive species, in particular NO_x (nitrogen oxides), CO (carbon monoxide) and the VOCs (volatile organic compounds); and use of a model with latitudinal and vertical resolution. However, simple globally averaged models of chemistry-climate interactions have been developed. These models treat the global CH_4-CO-OH cycle in a manner that takes into account the effects of the heterogeneity of the chemical and transport processes. They provide estimates of future global or hemispheric mean changes in the chemistry of the Earth's atmosphere. An even simpler approach, adopted by Osborn and Wigley (1994), is to treat the atmosphere as a single well-mixed box but to account for the effects of atmospheric chemistry by making the CH_4 lifetime depend on CH_4 concentration in a way that roughly mimics the behaviour of globally averaged models or of models with explicit spatial resolution.

Atmospheric chemistry is also central to the distribution and radiative properties of small suspended particles in the atmosphere referred to as aerosols, although chemistry is only part of what is required in order to simulate the effects of aerosols on climate. The primary aerosols that are affected by atmospheric chemistry are sulphate (SO_4^{3-}) aerosols (produced from the emission of SO2 and other S-containing gases), nitrate aerosols (produced from emission of nitrogen oxides), and organic carbon aerosols (produced from the emission of a variety of organic compounds from plants and gasoline). The key processes that need to be represented are the source emissions of aerosols or aerosol precursors; atmospheric transport, mixing, and chemical and physical transformation; and removal processes (primarily deposition in rainwater and direct dry deposition onto the Earth's surface). Since part of the effect of aerosols on climate arises because they serve as cloud condensation nuclei, it is also important to be able to represent the relationship between changes in the aerosol mass input to the atmosphere and, ultimately, the radiative properties of clouds. Establishing the link between aerosol emissions and cloud properties, however, involves several poorly understood steps and is highly uncertain.

To simulate the increase in the amount of a given aerosol in the atmosphere in response to the increase in emissions of the precursors to that aerosol requires the simultaneous simulation of all the major aerosols in the atmosphere due to coupling between the different aerosols (Stier *et al*., 2006). For example, a reduction in sulphur emissions, while reducing the sulphur aerosol loading, leads to an increase in the lifetime and loading of other aerosol species, especially at high latitudes and especially for black carbon and particulate organic matter. The geographical distribution of emissions is also important. Between 1985 and 2000, global emissions fell by 12% but the atmospheric sulphate aerosol loading is estimated to have fallen only 3% because the locus of emissions shifted southward to latitudes where in-cloud processing of sulphur oxides (SO, SO_2) into sulphate is more effective (Manktelow *et al*., 2007).

9.3 The research frontier

A long-term goal of the climate-research community is the development of increasingly sophisticated models that couple more and more components of the climate system. A large number of modelling groups have created three-dimensional models that couple the atmospheric and oceanic components of the climate system, and that include increasingly realistic representations of sea ice and land surface processes (in particular, the buildup and melting of snow cover, runoff generation, and the coupled fluxes of water, CO_2 and between vegetation and the atmosphere). Vegetation-atmosphere fluxes of CO_2 and water are coupled through the partial dependence of both on plant stomata. The size of the stomatal openings depends on the photosynthetic demand for carbon, as well as on the simulated leaf temperature and soil-moisture content. The most recent land-surface models that have been incorporated into AOGCMs compute photosynthetic and respiration fluxes every 20 to 60 minutes of simulated time. This is referred to as a biogeochemical land-surface model. The annual net carbon gain or loss is used to update the amounts of carbon in a three- or five-box model of the terrestrial biosphere at each grid point. From the carbon mass, a leaf-area index might be computed that, in turn, would be used in the calculation of the amount of solar radiation absorbed by the plant canopy, which in turn affects the calculated annual photosynthesis (see, for example, Arora *et al*., 2009). The PFTs present in each grid cell are prescribed in some AOGCMs, while in others the biogeochemical module might be used to determine which PFTs are present in each grid cell based on the carbon balance as simulated by the biogeochemical module for each potentially present PFT (see, for example, Foley *et al.*, 1996; Winter *et al.*, 2009; and Notaro *et al.*, 2007). In addition to the above, the entire seasonal phenological cycle is internally computed in the vegetation model of Krinner *et al.* (2005). A further advance involves computing the cycling of nitrogen between soil carbon and plant biomass, as C-N interactions are important for the response of vegetation to increasing CO_2 (Plattner *et al.*, 2008; Gerber *et al.*, 2010). Sensitivity studies have shown that the change in vegetation type as the climate changes can significantly modify the final climate change (at least locally) compared to the change that would occur if the vegetation was fixed. The physiological plant response to higher CO_2 (changes in stomata) can also be important local compared to the radiative effect on climate (Sellers *et al.*, 1996; Bounoua *et al.*, 1999) (see also Chapter 12).

Another major area of intensive research is in the representation of clouds in AGCMs. Differences in the way in which clouds are modelled are the main reason for the wide uncertainty in the long-run warming resulting from a doubling atmospheric CO_2 concentration (with global mean responses ranging from 2.1 °C to 4.4 °C in recent models, as summarized by Randall *et al.*, 2007, Table 8.2). The climatic effects of clouds depend on the details: exactly when, where, and how high clouds are; their detailed microphysical properties, such as cloud droplet radius, ice crystal radius and shape, and the extent of impurities (such as soot) in cloud droplets. As the climate changes, the various cloud properties will change in hard-to-predict ways, with a range of strong but competing effects on climatic change (some cloud changes will tend to amplify the change, thereby serving as a positive feedback, whereas others will serve to diminish the change, thereby serving as negative feedback). Increasingly sophisticated cloud schemes are being developed for and implemented in AGCMs (e.g. Del Genio *et al.*, 1996; Kao and Smith, 1999) that explicitly compute more and more of the cloud processes of climatic importance. However, greater sophistication and detail do not guarantee that the predicted changes in climate will be more reliable because even the most detailed schemes still require parameterizations to represent the effects of subgrid scale processes, and the nature of the parameterizations can strongly influence the results.

Other areas of ongoing research include improved treatment of sea ice, the realistic routing of runoff to the oceans, and the incorporation of atmospheric chemistry within AOGCMs used for climate-simulation purposes. With regard to the latter, a typical atmospheric chemistry model might track the concentration of over 100 chemical species in three dimensions, including aerosol species in a dozen or more size categories, and would include over 300 different chemical and phytolysis reactions (Jacobson and Streets, 2009). Due to the large amount of computer time required for atmospheric chemistry, simulations with fully coupled three-dimensional atmospheric chemistry models embedded in AOGCMs have not yet been performed. Rather, a recent approach has been to run the atmospheric chemistry model for two years with winds, temperature and other meteorological parameters prescribed from a specific two-year period from an AOGCM simulation, along with the corresponding anthropogenic emissions, in order to compute the concentrations of aerosols and short-lived GHGs and the associated radiative forcing. This radiative forcing is then used, along with the forcing from long-lived GHGs, to simulate another two to three decades of transient climatic change with the AOGCM, at which point the chemistry model is rerun for another two years with the latest meteorological variables and emissions (Shindell *et al.*, 2007). Climatic change also affects changes in the natural emissions of the precursors to various aerosols and ozone, although the effects are rather small (Jacobson and Streets, 2009).

9.4 Online material

Rather than delving into the details of selected climate-model components, the online material that accompanies this chapter is designed to illustrate basic principles governing changes in climate in response to a radiative forcing, and governing changes in the terrestrial biosphere in response to changes in atmospheric CO_2 and temperature. The change in temperature in response to a given radiative forcing, once temperatures everywhere have had time to fully adjust to the radiative forcing and so are no longer changing, is referred to as the equilibrium temperature change. The variation in temperature over time, as the climate system approaches the equilibrium change, is referred to as the transient temperature change. Similarly, one can speak of equilibrium and transient changes in the terrestrial biosphere. Of course, if the radiative forcing (for climate) or climate and CO_2 concentration (for the terrestrial biosphere) are themselves continuously changing, then the climate and terrestrial biosphere will never get a chance to reach an equilibrium.

The following outlines the equilibrium and transient calculations for both climate and the terrestrial biosphere that are performed in the online Excel package. A much more detailed explanation of and justification for the calculations is contained in the online explanatory material that accompanies the Excel package.

9.4.1 Equilibrium and transient climate response

The first two worksheets use a zero-dimensional model (in which a single point, with a single temperature, represents the entire global average) to illustrate how the change in global mean temperature in response to a radiative forcing depends on the rate of change of net radiation with temperature, dN/dT, and how dN/dT can be decomposed into terms involving individual feedback processes. Worksheet 3 introduces linear feedback analysis for fast-feedback processes, shows how to quantify the inherent strength of a given feedback, and shows that the impact on temperature of adding a feedback of given strength depends on the overall strength of the pre-existing feedbacks. Worksheet 4 extends the analysis to take into account climate-carbon cycle feedback.

The next two worksheets explore the transient (time-varying) approach to the final equilibrium temperature change. Worksheet 5 presents the transient response for the simplest possible case, in which only a single temperature is computed, which can be thought of as representing a single box. This box represents some combination of the atmosphere, land surface and upper layer of the ocean. Worksheet 6 presents the transient response for a two-box model, where the second box represents the deep ocean as a single, well-mixed thermal reservoir (that is, having only a single temperature). Although this representation is highly simplified, it does permit the elucidation of a number of important conceptual points, including: the fact that the first two thirds or so of the transient response is relatively rapid, being governed by the heat capacity of the ocean surface layer, while the final approach to the equilibrium change is much slower.

9.4.2 Globally aggregated terrestrial biosphere model

Worksheet 7 presents a three-box representation of the global terrestrial biosphere that is used to represent basic modelling principles and to illustrate some features of how simple models respond to temperature and CO_2 perturbations. The three boxes are: above-ground vegetation, detritus on the soil, and carbon in the soil. Each box is globally aggregated; that is, it represents that particular type of carbon worldwide. Similar box models have been applied to each land grid cell in the global grid of AOCGMs. The key parameters in Worksheet 7 govern how the rate of net primary production (gross photosynthesis minus plant maintenance respiration) varies with temperature (first increasing, then decreasing as temperature increases), how respiration of detritus and soil carbon increase with temperature, and how net primary production and the steady-state increase in above-ground equilibrium biomass vary with the atmospheric CO_2 concentration. The control parameters can be adjusted to replicate the aggregate global behaviour of complex, spatially resolved models, and the interaction between temperature- and CO_2-driven changes in net primary production and in temperature-driven increases in respiration can be explored.

9.4.3 Future greenhouse gas emissions, climatic change, and climate-carbon cycle feedbacks

Worksheets 8 to 10 present a simple framework for projecting future global energy demand and the mix of energy supplies into the future, and from that, generating a scenario of future fossil fuel CO_2 emissions. In Worksheet 8 the world is divided into two regions (roughly, developed and developing countries), and in each region total primary energy demand is computed as the product of the following factors: population × GDP (gross domestic product) per year per capita × primary energy per unit GDP or, in terms of units,

$$\text{Energy demand (MJ) per year} = P \times (\$/\text{yr})/P \times \text{MJ}/\$ \tag{9.1}$$

where MJ/$ (primary energy use per unit of GDP) is referred to as the energy intensity of the economy. Worksheet 9 contains data on the age distribution of existing nuclear power plants as of 2009; these plants are assumed to continue operating until the end of a 40-year lifespan and are not replaced as they retire, leading to a gradual phaseout of nuclear energy. Worksheet 10 contains information for specifying the timing of peaks in the global supply oil and gas and in the subsequent decline using logistic functions. Logistic functions are also used to specify the growth in power supply from C-free energy sources (biomass, hydro-electric, and other). The difference between energy supply from oil, natural gas, nuclear power and the C-free power supplies is assumed to be met by coal. The annual amounts of oil, natural gas and coal that are used along with their CO_2 emission factors are used to compute the global fossil fuel CO_2 emissions. Worksheet 11 contains an interface for optionally using CO_2 emissions from the more detailed calculations in the online material associated with Harvey (2010a, b), in place of the emissions from Worksheet 10.

The absorption of CO_2 by the oceans involves a sequence of processes occurring at successively slower rates: initial rapid (within one year) air-to-sea transfer of gaseous CO_2, then gradual mixing of dissolved inorganic carbon progressively deeper into the ocean. A pulse of CO_2 that is injected into the atmosphere can be divided into a series of fractions, each of which decays (decreases in concentration) with its own time constant. This mathematical representation is referred to as the impulse response for CO_2.

A continuous emission of CO_2 can be represented by a series of annual emission pulses, each of which decays according to the impulse response. The amount of CO_2 in the atmosphere at any given time is the sum of the amounts remaining from each of the preceding annual pulses going back to the beginning of human emissions. Due to the nonlinear carbon chemistry of ocean water, the rate of decay of successive pulses becomes slower as the cumulative emission (and hence, absorption by the oceans) increases. This slowing can be represented by adjusting the coefficients in the impulse response as a function of the cumulative emission. Worksheet 12 contains the impulse responses that are to be used in successive time intervals as the cumulative emission increases.

Worksheets 13 to 17 carry out the calculation of the increase in atmospheric CO_2 concentration, total radiative forcing and the change in global mean temperature. The total CO_2 emission is summed up in Worksheet 13 and used with the impulse functions from Worksheet 12 to compute the increase in atmospheric CO_2 concentration. The total emission involves fossil-fuel emissions (from Worksheets 10 or 11), emissions due to land-use changes (such as deforestation) and the production of cement, emissions from the terrestrial biosphere other than through land-use changes (where absorption of CO_2 is a negative emission), direct emissions of CO_2 from thawing permafrost soils, and emissions CO_2 from the oxidation of methane from fossil fuel sources (such as leaks in natural gas distribution systems) or from thawing permafrost soils. Emissions from land-use changes and the production of cement are derived from parameters that are specified in the worksheet. The CO_2 emission from oxidation of fossil fuel methane depends on the user-specified fossil fuel methane emissions and the methane lifespan in the atmosphere, which depends in part on the methane concentration itself. Net CO_2 emissions from the terrestrial biosphere depend on the CO_2 concentration (through the stimulation of photosynthesis by higher CO_2) and on change in temperature (which affects both photosynthesis and respiration), but the CO_2 concentration and temperature change depend in part on the emissions, so there is a climate-carbon cycle feedback loop. Another feedback loop exists through the dependence of CO_2 and CH_4 emissions from yedoma soils (Crich soils in Siberia) on the temperature change. Emissions from (or absorption by) the terrestrial biosphere are computed in Worksheet 14 (which uses a four-box terrestrial biosphere model rather than a three-box model), while emissions of CO_2 and CH_4 from thawing yedoma soils are computed in Worksheet 15. Worksheet 13 also contains calculations for the buildup of atmospheric methane and nitrous oxide (N_2O) concentrations and for emissions by sulphur aerosols, all based on a handful of parameters that can be altered by the user and which are fully explained in the online supporting information.

Worksheet 16 contains the calculation of the radiative forcing (heat trapping or, in the case of aerosols, reflection of solar radiation) due to the buildup of CO_2, CH_4 and N_2O, as well as due to tropospheric ozone, stratospheric water vapour, and aerosols. Also included up to 2000 are estimated radiative forcings due to changes in the solar luminosity and due to volcanic eruptions. Worksheet 17 uses the total radiative forcing to compute the change in surface temperature using the two box model that is featured in Worksheet 6.

9.4.4 Conservation of energy and mass

The final worksheet presents a series of model diagnostics that illustrate conservation of mass and conservation of energy. Research by Sterman and Sweeney (2007) indicates that the public's concepts of the relationship between changes in emissions and changes in CO_2 concentration, and between changes in CO_2 concentration and changes in temperature violate the principles of conservation of mass and energy. This perception arises because people tend to assume that changes in concentration immediately track changes in emissions (rather than depending on the difference between sources and sinks) and the changes in temperature in turn immediately track changes in CO_2 concentration (rather than depending on the current radiation balance). This tendency was found to be true of graduate MBA and engineering students at Harvard even after first explaining the difference between stocks and flows, something one would expect both groups of students to understand already. The final worksheet and supporting online material have been prepared in response to these conceptual problems.

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