Climate Modeling*

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Key Words

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Abstract

Climate models simulate the atmosphere, given atmospheric composition and energy from the sun, and include explicit modeling of, and exchanges with, the underlying oceans, sea ice, and land. The models are based on physical principles governing momentum, thermodynamics, cloud microphysics, radiative transfer, and turbulence. Climate models are evolving into Earth-system models, which also include chemical and biological processes and afford the prospect of links to studies of human dimensions of climate change. Although the fundamental principles on which climate models are based are robust, computational limits preclude their numerical solution on scales that include many processes important in the climate system. Despite this limitation, which is often dealt with by parameterization, many aspects of past and present climate have been successfully simulated using climate models, and climate models are used extensively to predict future climate change resulting from human activity.

Contents

1. INTRODUCTION

Climate models simulate the atmosphere, given atmospheric composition and energy from the sun, and include explicit modeling of, and exchanges with, the underlying oceans, sea ice, and land (1). A major application of societal interest has been the variation of climate with changes in atmospheric composition (especially anthropogenic) and solar output (2). Climate models are currently undergoing further development into "Earth-system models," which typically include carbon and nitrogen cycles, atmospheric chemistry, ocean biogeochemistry, and ice sheets on land. Links between Earth-system models and economic models to study human dimensions of climate and climate change are envisioned.

IPCC: Intergovernmental Panel on Climate Change

Solar (shortwave) radiation: radiation emitted by the sun; maximum intensity at wavelengths around 0.4 to 0.7 micrometers

The Intergovernmental Panel on Climate Change (IPCC) has documented in detail developments in climate modeling to about 2004, and comprehensive texts are available describing the principles and applications of climate modeling (1, 2). The complementary goal of this concise review is to describe for a broad readership the emergence of the climate system from fundamental principles, the capabilities and limitations of climate models in simulating the climate system, and the major ways in which climate models can be applied to larger scientific and societal issues.

This review focuses on physical climate modeling, where the primary goal is to simulate the multiyear behavior, both natural and forced, of the annual cycles of atmospheric temperature, water content, and motion (winds). Physical climate models, at higher resolution and with data assimilation, are also used for global numerical weather prediction. The physical processes in these models are nonlinear, so their detailed, time-dependent solutions (weather forecasts) depend strongly on initial conditions and uncertain details in the mathematical formulations in the models. Thus, weather predictions with a high degree of temporal detail are possible for only limited periods (of about a couple of weeks) (3, 4). However, time-averaged conditions (climate) are generally independent of initial conditions and depend instead on boundary conditions, i.e., solar (shortwave) radiation and its disposition in the Earth-atmosphere system. The latter depends on atmospheric and oceanic composition and surface characteristics. Climate models have played a major role in understanding the mechanisms that determine climate but are best known outside the scientific community as tools for estimating climate change, particularly anthropogenic climate change.

This review describes the basic structure of climate models and the physical principles that underlie them. Key research issues in model development are indicated, and major applications are summarized below. The goal is to promote the judicious use of model results for a wide array of scientific and policy applications by providing users with a sense of the capabilities and limitations of current models, the active research to advance their scientific basis, and the applications possible with the present generation of climate models. We do not attempt to describe in detail the mathematical approaches to representing physical processes in models but try to indicate where understanding of these processes remains limited and the implications of these limitations.

2. CLIMATE MODEL COMPONENTS AND THEIR COUPLING

Climate models consist of state-of-the-art component models representing the atmospheric general circulation, ocean general circulation, sea-ice dynamics and thermodynamics, and relevant land processes (5, 6). Emerging Earth-system models will also contain component models representing vegetation and its dynamic evolution, cycles of carbon and nitrogen over land, ocean biogeochemistry (7), atmospheric chemistry, and continental ice sheets. The atmospheric, vegetation, land, ocean, and ice interfaces exchange energy, momentum, and mass, and the empirical rules governing these exchanges are key ingredients of climate models that couple the components. Critical inputs to these rules include the sea surface temperature (SST), sea-ice surface temperature, land surface temperature, and soil moisture.

SST: sea surface temperature

Figure 1 illustrates the component models that comprise current climate models. The atmospheric model is most responsible for the fluxes of energy and moisture that drive the land, ocean, and sea-ice models, e.g., solar radiation and precipitation reaching the surface. In turn, the other component models strongly influence energy and moisture fluxes at the base of the atmospheric model, e.g., water vapor fluxes into the atmosphere (related to soil moisture and SST). River and groundwater runoff into the oceans is a driver of the ocean model by the land model. Energy, water, and salt are exchanged between the ocean and sea ice. The land and sea ice are solids, but because the latter floats on the ocean, it can redistribute

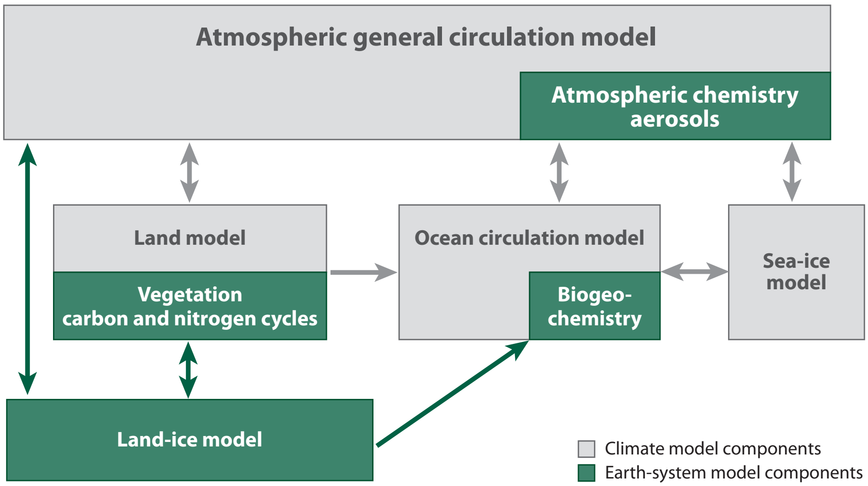

Figure 1

Schematic illustration of the components of climate and Earth-system models. The components of current climate models are gray. The additional components required to construct Earth-system models are shown in green. The connecting arrows indicate the exchanges that couple the components. They show that all components interact directly with the atmosphere and that most of the component models interact with each other.

climate properties from one geographical region to another, though only at polar latitudes. Except on very long timescales, and as modeled in present climate models, the continents and their ice sheets are stationary, so transports are restricted to the vertical, and negative feedbacks oppose the storage of large amounts of heat and water, apart from the seasonal snowpack.

Parameterization: mathematical representation of a physical or chemical process, using variables resolved by a general circulation model and obtained from observation, theory, and optimization

The additional components required to construct Earth-system models are also shown in **Figure 1**. In Earth-system models, one goal is to predict atmospheric composition from specified emissions. Models for atmospheric chemistry and aerosols would be included in the atmospheric model. Chemical species in the atmosphere are advected by the wind field. Chemical reactions are dependent on temperature and atmospheric radiation, and aqueous chemistry occurs in liquid and ice in clouds. Removal of chemical species by wet deposition is a prominent example of the latter. Atmospheric chemistry can be linked to dynamical, radiative, and cloud processes in the atmospheric component of climate models to study interactions between atmospheric chemistry and atmospheric circulation. Global chemical transport models with atmospheric chemistry have been extensively developed (8, 9), but these models use wind fields analyzed from observations and are not able to simulate feedbacks between atmospheric chemistry and climate change. The land and ocean components would include physical and biological processes relating the uptake and release of carbon to the physical state of the overlying atmosphere and ocean. Dynamic land-ice models, capable of simulating the evolution (including melting) of continental glaciers, would become a new component in Earth-system models. A major goal of developing these models (10) is to study sea-level changes related to changes in land-ice sheets.

The fundamental computational challenge for climate models is the wide range of time and space scales over which the components evolve and vary. The ocean and atmosphere are both geophysical fluids, meaning that large-scale (up to global) flows feel Earth's rotation in proportion to the projection of Earth's rotation axis onto the local vertical, through the latitudinally dependent Coriolis force. There are also important smaller-scale motions, down to the millimeter scales of molecular viscosity, which ultimately convert kinetic energy into heat. For a given range of scales, governing equations have been developed from the basic Navier-Stokes equations and have been shown to represent nature faithfully (11). However, present computational capability allows at most the largest three of the ten-decade range in scales to be explicitly resolved, leaving a majority of scales neglected, specified in various ways, or parameterized.

Unfortunately, unresolved small scales can have important impacts, particularly on land processes and societal interests. Therefore, downscaling, either statistically or through nested regional models, is a vital, though difficult, aspect of some climate modeling. Conversely, unresolved small scales, such as mountain and ocean topography, can produce first-order effects on the atmospheric and oceanic flows, so upscaling parameterizations and nested regional models are becoming more integral in this aspect of climate modeling.

Because climate change involves long timescales and variations, which are often small relative to higher frequency phenomena, such as the seasonal cycle, requirements for accuracy and balance in fluxes across model interfaces are quite exacting. For example, spurious lack of conservation of water substance in the atmosphere, or in the lakes, wetlands, and aquifers of a land model, can generate large changes in sea level over model-simulated decades. Oceans and sea ice are driven by fluxes of energy and momentum at their surfaces, which are functions of the state of both the ocean or ice and the atmosphere, as well as rather uncertain empirical flux laws. If fluxes across the surfaces of the component models are insufficiently accurate, the climate will behave in unrealistic ways (12).

3. ATMOSPHERIC GENERAL CIRCULATION MODELS

The atmosphere circulates as a result of instabilities arising from its pattern of heating and cooling. To a large degree, this heating and cooling pattern, as well as the flows of energy through the atmosphere, is controlled by the distributions of clouds and radiatively active gases and aerosols. The distributions of clouds, gases, and aerosols are themselves controlled to a large degree by the circulation. Atmospheric general circulation models are thus highly nonlinear with extensive feedbacks among processes. The key processes in atmospheric general circulation models are described in this section.

3.1. Radiative Transfer

The terms solar (or shortwave) and terrestrial (or longwave) radiation are used to refer to radiation emitted by the sun and Earth, respectively. The former is the fundamental energy source for the atmosphere; the Earth loses energy mostly through the latter and less via reflected solar. Shortwave radiation, whose maximum intensity is at wavelengths around 0.4 to 0.7 micrometers, is transferred through Earth's atmosphere, suffering reflection and absorption by clouds, aerosols, and gases as it does so. The fraction that reaches Earth's surface is reflected and absorbed there. Earth subsequently emits longwave radiation, with maximum intensities around 10 to 20 micrometers, which is then subject to absorption and emission by clouds, aerosols, and gases. Water vapor, carbon dioxide (CO_2), and ozone are especially important gases for radiative transfer (13). The angle, and therefore the intensity, at which solar radiation strikes Earth's surface depends on latitude, with more striking Earth in the annual mean at the equator. Variation with latitude of terrestrial radiation emitted by Earth is much less than of solar radiation. This difference produces an imbalance with respect to latitude in net energy received by the Earth-atmosphere system. Detailed radiation codes in the atmospheric component of climate models calculate the absorption, emission, and scattering (reflection) of solar and terrestrial radiation (14–17). Radiation codes are based on well-established principles and measurements in spectroscopy and are by themselves very robust. As will be seen in subsequent sections, the distribution of atmospheric constituents depends on several processes, which are much less understood than the principles of radiative transfer. Because radiative transfer can ultimately be modeled only as realistically as the composition of the transferring media is known, radiative energy flows in the atmosphere are much less certain than the radiation codes themselves.

3.2. Atmospheric Dynamics and Convection

Motions in the atmosphere transport energy, water, and other constituents. The motions arise in response to variations in pressure (pressure gradients), gravity, friction (viscosity), and apparent forces evident in a reference frame fixed to the rotating Earth (centrifugal and Coriolis forces). Motions are often associated with instabilities in atmospheric flows. Two basic instabilities are consequences of radiative transfer. First, the unequal receipt of solar radiation between the equator and poles generates a temperature gradient. A horizontal temperature gradient of sufficient magnitude is unstable (baroclinic instability) (11). The resulting motions reduce the gradient by transporting heat from the equator toward the poles and thus play a critical role in establishing the equilibrium temperature distribution with respect to latitude. Second, the vertical distribution of absorption and emission of radiation in the atmosphere tends toward a density distribution that is unstable (18) and results in concentrated upward and downward motions (convection). Convection distributes heat upward, resulting in less dense air farther from the surface and restoring stability.

Dynamical cores in atmospheric general circulation models advect momentum, heat, moisture, and other substances and employ advanced numerical methods in their computer codes. The cores are based on fundamental principles in classical dynamics and thermodynamics governing momentum, forces, and energy (19). These cores explicitly resolve the motions

General circulation model: a set of equations solved by computer algorithm, representing the time-dependent global circulation

Terrestrial (longwave) radiation: radiation emitted by Earth; maximum intensity at wavelengths around 10 to 20 micrometers

Baroclinic instability: instability associated with a large horizontal density gradient in the ocean or atmosphere

Cloud macrophysics: processes governing the distribution of moisture, temperature, and motion related to cloud formation

Cloud microphysics: processes related to the formation, growth, and precipitation of cloud droplets and ice crystals

Boundary layers: layers in ocean and atmosphere near their surfaces, often characterized by turbulence and rapid variations in properties

Gravity wave: a wave set up by a localized disturbance within a fluid or at its boundaries; the displaced fluid is restored by gravity

associated with baroclinic instability. The horizontal sizes of convective updrafts are smaller than the motions than can be explicitly resolved in current climate models (20). (The scales of motion that can be resolved by dynamical cores are limited by available computational power.) The effects of convection are included in atmospheric components of climate models using theories that relate heating and moisture changes by convection to properties of the atmospheric flow resolved by the dynamical cores. The theories are referred to as "convective" or "cumulus parameterizations" (21, 22). These parameterizations are highly uncertain and the subject of active research.

3.3. Clouds

Clouds, consisting of liquid, ice, or a mixture of phases, reflect and absorb solar radiation, and they emit and absorb terrestrial radiation. In general, their effect on solar radiation is to cool the Earth-atmosphere system, whereas their effect on terrestrial radiation is to warm it (23, 24). The magnitude of clouds' influence depends on their height in the atmosphere, their areal extent, the mass of ice or liquid they contain, and the sizes and shapes of their liquid drops and ice crystals. Clouds are often below the resolvable scale of dynamical cores (25). Cloud macrophysics parameterizations determine unresolved distributions of moisture, temperature, and motion related to cloud formation (26), whereas cloud microphysics parameterizations describe the formation, growth, and precipitation of cloud droplets and ice crystals (27). For climate change, a particularly important aspect of cloud microphysics is the formation of cloud droplets and ice crystals. The number of cloud droplets and ice crystals in a cloud system is related to the sizes of these crystals for a given water content, and the size in turn is a key determinant of the reflection and absorption of radiation by the clouds. The number of cloud droplets and ice crystals formed depends on atmospheric composition. Parameterizations linking droplet and ice nucleation to atmospheric composition have been developed and incorporated into the most recently developed atmospheric general circulation models (28–31). Cloud macrophysics parameterizations, like cumulus parameterizations, are highly uncertain even at fundamental, conceptual levels. Many aspects of cloud microphysics, especially those related to liquid, are well understood, but some basic questions in ice microphysics, e.g., nucleation of ice crystals, require additional theoretical, laboratory, and field study (32).

3.4. Turbulence

Small-scale turbulence plays an important role in governing the vertical fluxes of momentum, heat, moisture, and chemicals in the atmosphere. Particularly important is turbulence in the planetary boundary layer, which is destabilized by heating at the surface and rapid variation with height of the wind (shear) near the surface. These processes are represented by boundary layer parameterizations (33) in the atmospheric component of climate models. Convection and interaction between winds and orographic disturbances generate waves (gravity waves), whose impacts are treated by gravity wave parameterizations (34, 35) in atmospheric components of climate models. Diffusion parameterizations treat other forms of unresolved turbulence. Some dissipation also occurs as an artifact of numerical methods used in dynamical cores. Boundary layer and gravity wave parameterizations require further development to advance the state of climate models.

4. OCEAN GENERAL CIRCULATION MODELS

In a coupled climate model, the ocean provides both a long-term memory and large storage capacity for the overall climate system. Compared to the atmosphere, it has much longer timescales (millennia versus years) and a much larger storage capacity for heat, water, and radiatively important atmospheric constituents such as CO_2. For example, the oceans hold more than 97% of Earth's water with the capacity to hold it all; the entire heat capacity of the atmosphere is equivalent to that of only the upper 3 m of the ocean (36), and the oceans contain about 50 times more CO_2 than the atmosphere, with current uptake about one third of that released from human activity (37).

Like the atmosphere, ocean density decreases with temperature and increases with pressure, but instead of decreasing with humidity, it increases with the ocean salt content, or salinity. All three dependencies are highly nonlinear and modeled with a full equation of state.

4.1. Radiative Transfer

The modeling of radiative transfer in the ocean is much simpler than in the atmosphere. The absorption and emission of longwave radiation occurs within centimeters of the surface, and both can be treated as a surface heat flux, with the emission governed by the blackbody radiation formula. A large fraction of incoming solar radiation is absorbed within the upper meter, and the absorption of the remainder is well described as a double exponential decay (38), with absorption coefficients depending on water clarity, which in the open ocean depends on biological activity. Options used in ocean models include specified clarity, specified monthly varying chlorophyll (39), or explicitly modeled biology. The chlorophyll is inferred from satellite ocean color observations and serves as a proxy for the biology. Even though only about 1% of the incoming solar penetrates below 50 m in depth, modeling of the absorption is important to the evolution of SST (40), and hence to interaction with the atmosphere.

4.2. Ocean Dynamics and Convection

Motions in the ocean generally follow the same laws as those in the atmosphere (Section 3.2), and the discussion of turbulence and convection (Section 3.4) generally applies to the ocean too. Virtually every night all over the ocean the absence of the only stabilizing surface heat flux (solar radiation) produces shallow (order 100 m) convection. Deep-ocean convection (order 1000 m) is associated with late-winter cooling and is confined to a few high-latitude regions, such as the Labrador Sea, the Greenland Sea, and the Weddell Sea (41). It provides a direct conduit to the deep ocean, where millennia may pass before contact is reestablished with the atmosphere.

However, fundamental differences arise in large-scale flows because only the ocean has full-depth lateral boundaries imposed by the continents. Such boundaries permit the buildup of unbounded horizontal pressure gradients throughout the water column in response to wind forcing. Off the equator, the result is a characteristic gyre circulation, with strong western boundary currents providing the mass and vorticity balances to the broader, but weaker, interior Sverdrup circulation (42). These currents are responsible for most of the meridional transport of heat and other ocean properties, so that, unlike the atmosphere, very little is left to be transported by the mesoscale eddy field, spawned by baroclinic instability. Therefore, for many climate purposes, it has been found adequate to parameterize the effects of mesoscale ocean eddies (43), but it is no longer acceptable to neglect them in modern ocean models. This situation has fortunate modeling consequences, because explicit eddy resolution requires at least an order of magnitude finer horizontal ocean grid (<10 km) than is needed in the atmosphere (order 100 km).

Equatorial ocean dynamics (44) are unique because of the loss of the Coriolis force (geostrophy) and the presence of full-depth lateral boundaries. In the Pacific and Atlantic oceans, the zonal winds are predominantly westward, and they pile water in the west until the resulting pressure gradient balances the surface wind stress. At depths between 200 m (in the west) and 80 m (in the east), this surface pressure gradient loses the influence of the wind and drives eastward Equatorial Undercurrents that have no atmospheric analogue. These flows are meridionally convergent and supply the equatorial upwelling that balances the divergent westward surface currents. Thus, all these flows contribute to the upper-ocean heat budget, and perturbations to this budget

ENSO: El Niño-Southern Oscillation

Albedo: fraction of radiation striking a surface that is reflected

give rise to SST variations seen by the atmosphere, including those associated with El Niño-Southern Oscillation (ENSO) variability in the eastern equatorial Pacific.

5. SEA-ICE MODELS

Sea-ice effects that need to be well modeled include insulation of the relatively warm underlying ocean from the potentially much colder polar atmosphere and reflection of a high fraction of solar radiation. The former is highly dependent on ice thickness, which has motivated models with multiple (typically 3 to 5) thickness categories. An important negative feedback is that increases in thickness, and hence in heat deficit and freshwater storage, are in turn dependent on the transfer of heat from the ocean through the ice, which is retarded by the greater insulating effect of thicker ice. The albedo depends on snowfall from the atmosphere, very strongly on how a model distributes the snow over the sea ice, on the presence of (parameterized) melt ponds during the melt season, and less on air temperature. Most solar radiation that does not reflect is absorbed and heats the ice, but a small amount can reach the ocean. All other things being equal, more snow and ice increase the albedo, which cools the ocean and sea ice and creates more sea ice. This positive ice-albedo feedback is largely controlled by changes in fractional sea-ice cover over the ocean, which is constrained to lie between 0 (ice free) and 1 (fully ice covered).

Climatically important sea-ice motions are explicitly resolved and are largely a result of a balance between wind forcing, ocean drag, and Coriolis force. These motions transport heat (mostly as a latent heat of fusion), water, and a small amount of salt. The most uncertain dynamic is the rheology (45), but happily, the resulting internal ice stress is usually small. The exceptions are in regions where ice convergence, especially along coastlines, increases the ice thickness through complicated ridging processes. These processes and wind forcing are major factors in producing the very thick ice observed north of Greenland and east of the Antarctic Peninsula.

There are a number of thermodynamic sea-ice processes that pose serious modeling challenges. In addition to the albedo and distribution of snow, there is frazil ice formation as the ocean temperature falls below freezing, basal ice formation at the base of a cooling ice column, surface melting and the fate of the resulting liquid water, lateral melting, and radiative transfer through snow and ice.

6. LAND SURFACE MODELS

The major task of land surface models is to interact with the atmosphere by extracting momentum, by reflecting solar radiation, and by exchanging heat and atmospheric constituents, such as moisture and CO_2 (46, 47). There are negative feedbacks that limit the heat and moisture uptake by the land, although a significant amount of water can be isolated in deep reservoirs. Carbon storage in both living and decaying plants can potentially change atmospheric CO_2 and thus affect Earth's radiation balance appreciably (48). This storage capacity depends on plant type, which is often specified, but dynamic vegetation models that predict the vegetation type are becoming more common. The fidelity of such models is very dependent on the realism of the atmospheric model, especially the seasonal cycles of precipitation and surface temperature, which cannot be guaranteed.

The dynamical components of land surface models are the hydrological processes, including very fast river runoff and very slow motions of land ice held in glaciers and ice sheets. Only very long timescale climate models (e.g., of glacial transitions) require that the slow motions be explicitly modeled, so most often land-ice distributions are specified. Recent observations have raised concerns about the possibility of relatively rapid changes in the stability of land-ice sheets and the ability of current land-ice models to simulate them (49).

7. ATMOSPHERIC COMPOSITION AND RADIATIVE FORCING

As discussed in Section 3.1, absorption, transmission, and reflection of radiation emitted by the sun and Earth are the primary drivers of the atmospheric circulation and determinants of its mean structure, including such basic characteristics as the temperature and moisture fields. The composition of the atmosphere changes on short and long timescales, owing to both natural (e.g., volcanoes) and anthropogenic processes, such as fossil fuel and biomass burning. Incorporating changes in atmospheric composition because of anthropogenic processes is an important part of climate modeling and has as its goals both understanding recent observed multidecadal changes in climate and predicting possible future climate changes associated with possible emission scenarios. The impact of composition changes on climate is measured by *radiative forcing*, which is the change in the energy balance at the top of the atmosphere produced by a composition change [with all other factors held constant (50)].

The radiative forcing components with the largest magnitudes arise from CO_2 and aerosols (51). Aerosols produce radiative forcing both through their *direct effect*, the scattering and absorption of atmospheric radiation by aerosol particles, and their *indirect effect*, changes in cloud microphysics stemming from the effect of aerosols on nucleation of cloud droplets and ice crystals (cf. Section 3.3).

Current climate models specify atmospheric composition, with future composition generally determined from chemical transport models, using plausible scenarios for future emissions. Among the most recent developments in climate modeling has been the coupling of models for vegetation and ocean chemistry to atmospheric and oceanic general circulation models. The atmospheric component of climate models provides radiation, temperature, and precipitation fields for vegetation, which can evolve dynamically with changes in climate. In turn, the roles of vegetation as sources or sinks for radiatively active gases (e.g., CO_2) can be inferred. Similarly, ocean chemistry and dynamics determine the role of the oceans as a carbon sink for the atmosphere (7). Feedback loops both in the ocean and on land (e.g., effect on plants by nitrogen on land and iron in the ocean) introduce additional complexity into the modeling of biogeochemical cycles.

The direct effects of aerosols have been incorporated into atmospheric general circulation models using specified aerosol distributions, which can be specified from chemical transport models. Coupled climate-chemistry models, which are under development, will link the evolution of aerosol concentrations to other aspects of the climate system and emissions. Incorporating indirect effects of aerosols requires the parameterization of complex links between aerosol composition, cloud microphysics, and cloud macrophysics (52). Early efforts at incorporating indirect effects in atmospheric general circulation models bypassed this complexity by using empirical relationships between cloud drop number and aerosol composition (53).

8. CLIMATE SIMULATIONS

In addition to the mean states of all its components, a climate model is designed to simulate variability over decades to centuries. Unlike weather forecasts, high frequency (daily to monthly) variability is important only in how it projects (upscales) onto longer climate scales. Important scale interactions also involve the annual cycle. Simulations of the annual cycle also provide an important means of evaluating a model's ability to change properly in response to a large external solar forcing because there is an abundance of observations for comparison. Interannual variability has also been well observed, but it is mostly internally generated because there is little interannual signal in incoming radiation at the top of the atmosphere. Its simulation, therefore, provides a stringent test of the instabilities and feedbacks of a climate model. The climate of the twentieth century has exhibited variability on decadal timescales, which realistically formulated climate models must reproduce. These tests and comparisons with the observed mean state are the basis for having any confidence in future climate projections, where there are no observations for either guiding or comparing the model. However, confidence is increased when a climate model is shown to be successful in reproducing and explaining the paleoclimatic record. Although the degree of success in simulating paleoclimate, the seasonal cycle, interannual variability, and twentieth-century climate provides an important gauge of the fidelity of a climate model to reality, the climate system will be subject to anthropogenic forcings in the future, which are not exact replicas of these. As a consequence, it is difficult to know categorically how much confidence should be accorded predictions of future climate using the degree of success from simulations of past and present climate.

A comprehensive treatment of all these aspects of climate model behavior far exceeds the limitations placed on this review, but the following few examples are illustrative of the salient points. Sections 8.1 through 8.5 provide examples of each.

8.1. Simulations of the Annual Cycle

As an example of an important and complex annual cycle, **Figure 2** shows the seasonal change in the extent of Arctic sea ice in two different ways. First, the mean annual cycle of Arctic area covered by sea ice is shown in **Figure 2*a***, both as observed from satellites (54) between 1979 and 2004 and from a 20-year climate simulation (55). The maximum ice coverage in March is the largest discrepancy, but it is less than 10%. The ice growth from a minimum of 6 million km^2 in September through December is very well simulated. However, the model gives a more rapid contraction from the maximum ice extent in March through July, with likely contributors being the parameterization of low-albedo melt ponds and the atmospheric forcing, especially the down-welling longwave radiation.

The geographical distribution of this seasonal melting is presented in **Figure 2*b*** (observed) and **Figure 2*c*** (modeled), as the difference in fractional sea-ice coverage between March and September. The central Arctic is nearly fully ice covered all year, so the plotted ice concentration difference is small, but there is less summer ice in the simulation, which is more akin to observations of more recent years, especially 2007 (56). Well simulated are regions such as Hudson Bay, Baffin Bay, Kara Sea, and Sea of Okhotsk, which are nearly fully ice covered in March and ice free in September, so that the concentration difference approaches 1. There are other ice-free regions in September (e.g., Gulf of Saint Lawrence, Labrador Sea, Greenland Sea, and Bering Sea), where the differences shown in **Figure 2** are the March ice fractions. There is a tendency for the simulations to produce more winter ice in these marginal seas in the Pacific sector and less in the Atlantic. Without this compensation, the agreement in **Figure 2*a*** would not be as good.

The seasonal freeze/melt cycle is important because sea ice is a very effective insulator between the cold polar atmosphere and the warmer ocean and because of the positive ice albedo feedback. An ice-free Arctic in summer has become a distinct possibility because of the record minimum ice extent observed in September 2007 (56) and simulations showing abrupt Arctic sea-ice changes in the near future (57).

Simulating this cycle is complex because the ice response to seasonal solar forcing depends on the interaction of a variety of processes in the ocean, ice, and atmosphere. The ocean general circulation produces a warm, salty subsurface Atlantic layer over much of the Arctic, but the amount of heat transferred to the ice is limited by the overlying cold fresh layer that is fed by continental runoff, especially from Siberian rivers. The largest term in the surface heat budget of Arctic sea ice is the longwave radiation emitted from low-level clouds. The wind forcing is the dominant driver of ice motion and takes newly formed ice in the western basin counterclockwise around the Arctic basin and out the Fram Strait between Greenland and Svalbard. During this journey, the ice transforms into thicker multiyear ice, with different albedo, strength, and insulating capacity. The comparisons of **Figure 2** indicate that this system can be represented with some fidelity in modern climate models but that there is more to be done.

In contrast, and despite considerable effort, climate models continue to struggle to generate realistic intraseasonal Madden-Julian Oscillations (MJOs). An MJO (58) is an important mode of variability in the coupled ocean and atmosphere in the tropics. Its spatial extent is around 1000 km, and it propagates eastward from the Indian Ocean to the western Pacific over a period of 30 to 60 days. In the ocean, associated equatorial Kelvin waves are observed as deep as 1500 m, less than simulated by an ocean model forced with intraseasonally varying winds (59). In the atmosphere, the MJO is characterized by interactions among scales from small-scale convective clouds to the planetary scale. As discussed in Sections 3.2 and 3.3, scale interactions related to clouds and convection remain poorly understood. Indeed, an experimental atmospheric general circulation model with a resolution fine enough to resolve individual convective clouds realistically simulates the MJO variability (60).

A critical aspect of climate simulation that climate models can do well is the ocean's seasonal modulation of atmospheric temperature through storage of summertime solar heat in the upper ocean, mixing of this heat into the seasonal thermocline in fall, and wintertime release to the atmosphere. A key process is subgrid-scale vertical mixing in both components, so the fidelity of its parameterization is a priority.

Seasonal cycles on the land include changing vegetation characteristics and the storage of water as snow. A primary influence of both is through the land surface albedo, which has a first-order effect on land surface temperatures and hence on interactions with the overlying atmosphere.

8.2. Simulations of Interannual Variability

The best-known example of internally generated interannual variability is the ENSO cycle. The overall lack of success in its prediction underscores the complexity of the underlying processes and interactions. Nevertheless, climate models are demonstrating an increasing capability of capturing many aspects of the phenomenon. One necessary condition for success is an ocean component that reproduces the observed upper equatorial Pacific temperature and velocity structure when forced with observed atmospheric conditions. It is satisfied by some, but not all, ocean climate models (S.M. Griffies, C. Böning, A. Biastoch, F. Bryan, G. Danabasoglu, et al., submitted), with vertical viscosity and diffusion, lateral viscosity, diurnal rectification, and resolution of tropical instability waves, all likely contributing factors. Another necessity is a realistic mean wind field, which depends on many aspects of the atmospheric model. Some of these involve uncertain parameterizations, such as reevaporation, convective heat and momentum transport, boundary layer turbulence, and cloud formation.

MJO: Madden-Julian Oscillation

ENSO variability has recently been described (61) as a delicate balance between the coupled rectification of atmospheric noise and an ocean oscillatory mode. The latter is complex, involving wind anomalies across the equatorial Pacific, an oceanic response that warms the eastern basin SST, a negative feedback that reduces the air-sea heating of the ocean, an atmospheric response with global teleconnections, a Pacific ocean wave response with propagation first westward then equatorward along the western boundary and then eastward along the equator, and finally a reversal of the oceanic SST warming. Adding to the modeling challenge is the fact that a major component of atmospheric noise in the equatorial Pacific are westerly wind events associated with the MJO. This intraseasonal variability is notoriously difficult to generate in models (62).

Observed and simulated (61) global ENSO teleconnections are shown in **Figure 3**, as the correlation of SST anomalies at a point with the anomalies averaged over the the white box in **Figure 3**, which is commonly known as the Niño 3 region. The positive comparison is not a trivial result because many climate models show much less agreement. Some similarities to note in the two patterns include the following: (*a*) the westward and meridional extent of the area of positive correlation in the eastern tropical Pacific; (*b*) the surrounding horseshoe-shaped region of negative correlation; (*c*) the large area of positive correlation in the Pacific sector of the Southern Ocean; (*d*) the positive correlations all along the coasts of North and South America, which are believed to be caused by poleward-propagating coastal Kelvin waves; and (*e*) the positive correlations in the Indian and Atlantic Oceans. However, the Indian Ocean signal is too confined to the west, where it is too strong, and in the Atlantic Ocean the correlation is too widespread and too strong. Another factor indicating that the climate model is working well is the frequency distribution of ENSO SST variability (61, 63). As observed, there is a broad peak at periods between 3 and 5 years, but the amplitude is sensitive to a number of parameterizations, especially of ocean mesoscale eddies and of the atmospheric and oceanic boundary layers, so this is an area of active research.

Less well-studied challenges for simulations of interannual variability are the regime shifts in the North Pacific, such as those observed in the late 1970s. It is encouraging that analysis of a climate model has found multidecadal, regime-shift-like behavior in North Pacific SST (T. Powell, personal communication, 2007). A sequence of 8 warm years, followed by 13 cold years, and then 29 warm years is reminiscent of the observed Pacific Decadal Oscillation of the latter twentieth century.

8.3. Simulations of Twentieth-Century Climate

The climate of the twentieth century, including its evolution with time, has been simulated by models developed by numerous major research centers. Forced by changes in solar output, greenhouse gases, and aerosols, these models can reproduce with reasonable fidelity basic features of observed climate, such as global mean temperature in the lower atmosphere. However, it should be noted that there is great uncertainty in forcings associated with aerosols and that different climate models use different estimates of aerosol forcing in producing these simulations (2, 64). **Figure 4**, from Reference 51, shows that only by including anthropogenic forcings can twentieth-century temperatures be simulated realistically. These results underlie the attribution of recent global warming to human activity. It is noteworthy that only some of these models include effects of past volcanic eruptions, and these all show less warming than those that do not.

8.4. Simulations of Anthropogenic Climate Change

The same class of models that has been employed to simulate the twentieth century has also been used to assess future climate change associated with anthropogenic modifications of atmospheric composition and surface characteristics, notably increasing CO_2 (65). An example is shown in **Figure 5**, which shows precipitation changes from 2090 to 2099 and 1980 to 1999 predicted by an ensemble of models for an intermediate increase in anthropogenic emissions (51). A rough characterization of the results is that both wet and dry regions are predicted to become more so. The models are more consistent in their predictions about precipitation changes at higher latitudes during winter and are at variance in several critical land areas, including the United States, Australia, and equatorial Africa during northern summer, as well as in eastern South America and Australia during southern summer.

These simulations have played a critical role in discussions of public policy response to climate change and have been the subject of extensive assessment at both national and international levels, with the IPCC assessments particularly prominent. As can be seen in **Figure 5**, showing that even the sign of the precipitation change varies in some regions, the models used in these assessments vary somewhat in their predicted responses to anthropogenic change. Not only do uncertainties arise owing to uncertain future emissions of greenhouse gases and aerosol formation, but also owing to uncertainties in parameterizations of key physical processes. Certainly future volcanic eruptions and their effects are not known. Parameterizations related to clouds and convection are especially large sources of uncertainty. Active research in model development focuses on improved parameterizations and understanding of cloud processes in the climate system. Increasingly powerful computers enable climate models to be integrated at finer resolution, enabling a broader spectrum of motions to be simulated. However, the range of scale from cloud droplets and ice crystals to phenomena of climatic importance is enormous, and understanding atmospheric scale interactions remains one of the basic challenges in atmospheric science (2).

8.5. Simulations of Paleoclimate

Coarse-resolution versions of some of the models used for twentieth-century and anthropogenic change simulations have been used to simulate climate under conditions associated with ice ages and altered configurations of continents. Generally, climate models are able to maintain Ice Age climates if integrated from appropriate boundary conditions for ice. Also, conditions at the Permian-Triassic boundary (ca. 251 Mya) have been successfully simulated using the paleogeography of that period (66). Although these models can simulate conditions favorable for the onset of glaciation, i.e., perennial snow cover in ice-sheet nucleation regions, when subject to Milankovich (astronomical) forcing hypothesized to trigger glaciation (67–69), they currently lack the slow feedback mechanisms involving the carbon cycle and ice dynamics required to simulate glacial-interglacial transitions.

SUMMARY POINTS

1. Climate models simulate the atmosphere, given atmospheric composition and energy from the sun, and include explicit modeling of, and exchanges with, the underlying oceans, sea ice, and land.
2. Climate models are based on physical principles governing momentum, thermodynamics, cloud microphysics, radiative transfer, and turbulence.
3. Although the fundamental principles on which climate models are based are quite robust, computational limits preclude their numerical simulation on scales that include many processes important for the determination of climate.
4. Processes that are too small to be resolved in climate models are represented by parameterizations, and some of these parameterizations are highly uncertain.
5. Limitations on resolution (imposed by available computational power) and uncertainty in parameterizations bound the realism with which past and current climate can be simulated and introduce uncertainty in predictions of future climate.
6. In spite of the limitations posed by resolution and parameterization uncertainty, many aspects of past and present climate, as well as recent climate change, have been successfully simulated by climate models.
7. Climate models are evolving into Earth-system models, which simulate chemical and biological processes and afford the prospect of links to studies of human dimensions of climate and climate change.
8. Climate models are providing important guidance for policy formulation related to human-induced climate change.

FUTURE ISSUES

1. Additional computational power is needed to increase the resolution of climate models.
2. More realism is required in parameterizations for processes that are important for climate but too small to be resolved by climate models, especially for processes that relate to interactions among clouds, aerosols, and climate.
3. Climate models (presently including atmosphere, oceans, and sea ice) are evolving to Earth-system models (also including continental ice sheets, atmospheric chemistry, and carbon and nitrogen cycles).
4. Climate and Earth-system models will interface with studies on human dimensions of climate and climate change.
5. Understanding and quantifying uncertainty in predictions of future climate remain key challenges, both scientifically and for their policy implications.
6. Improved representation of climate and climate change is needed on regional scales.
7. Expanded observational inference of recent and distant past climates will be important for validation of climate model simulations and better understanding of long-timescale climate change, such as the last millennium and glacial to interglacial transitions.
8. Using climate models to predict the evolution of climate from decadal to centennial scales is an emerging research goal.

DISCLOSURE STATEMENT

The authors are not aware of any biases that might be perceived as affecting the objectivity of this review.

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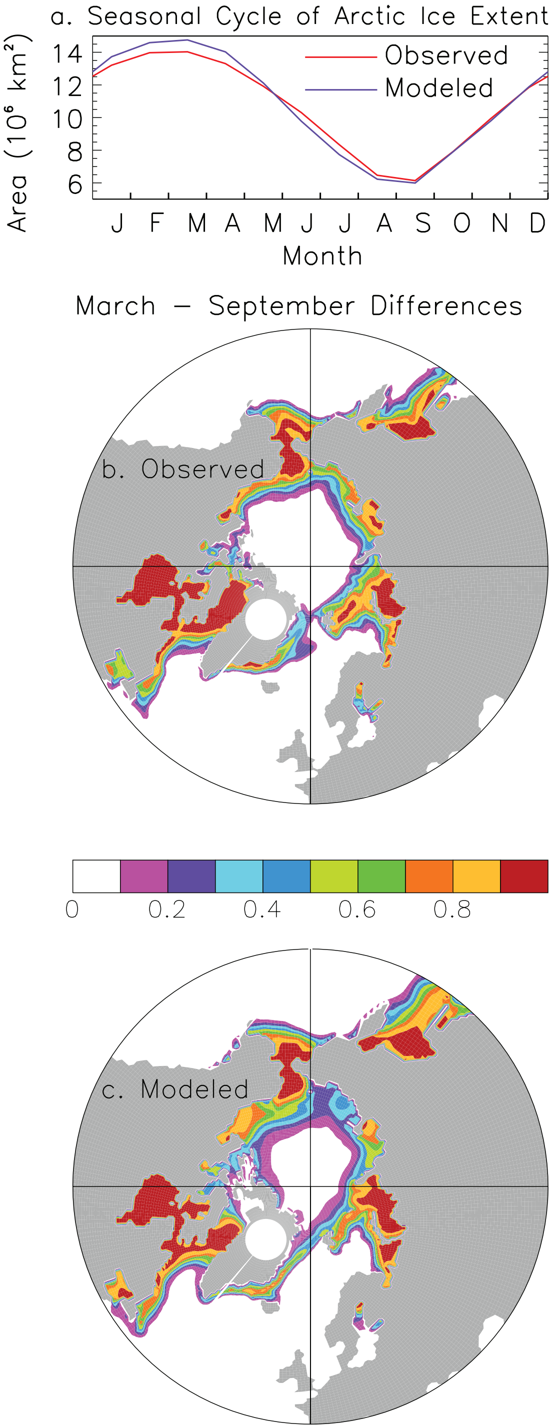

Figure 2

(*a*) Observed and modeled seasonal cycle showing the extent of Arctic sea ice. (*b*) Observed (54) and (*c*) modeled (55) March-minus-September differences in fractional Arctic sea-ice coverage.

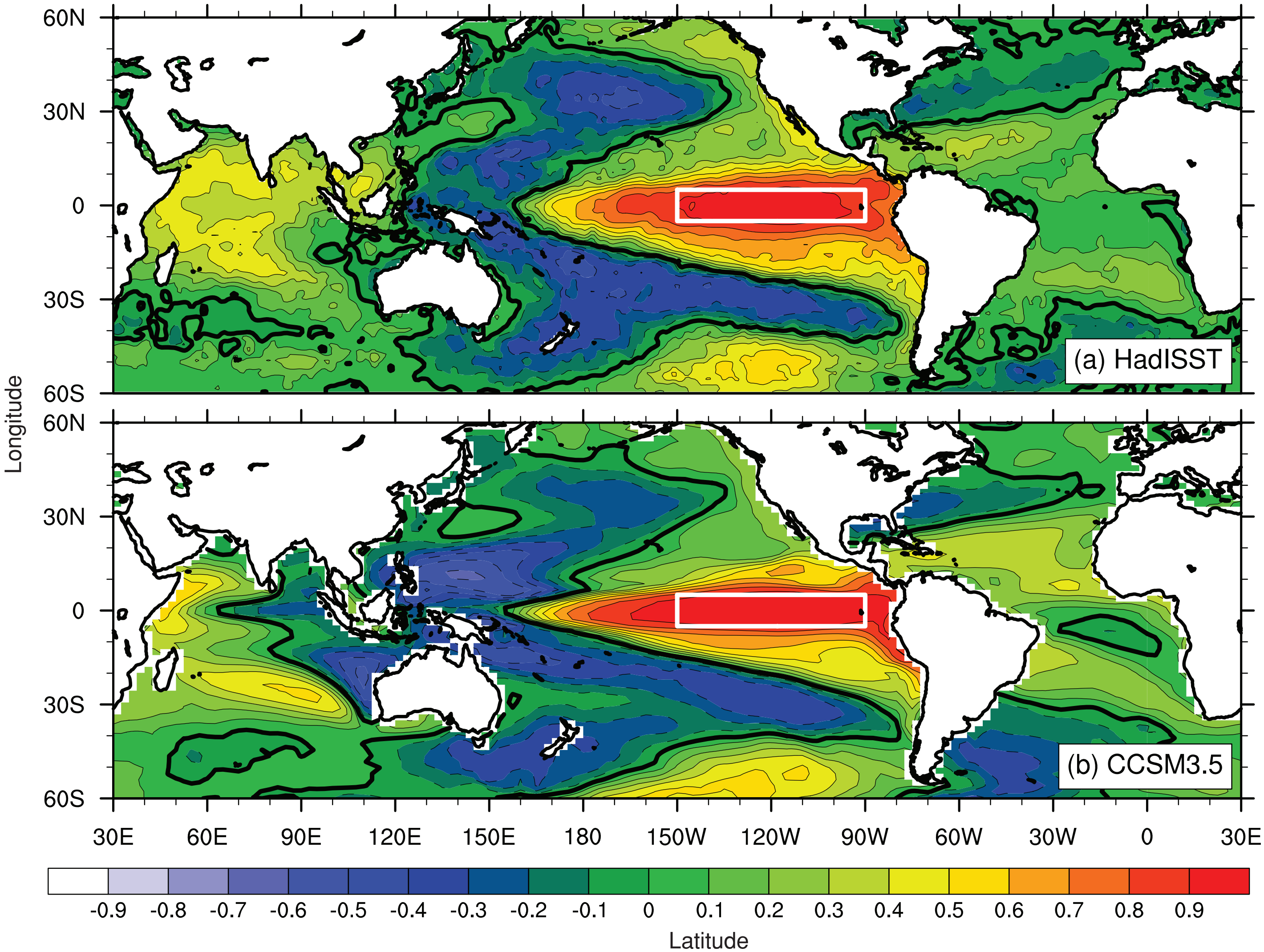

Figure 3

Spatial distributions of correlations of sea surface temperature (SST) anomalies with anomalies averaged over the white box that defines the Niño 3 region from (*a*) the Hadley Centre Sea Ice and SST data set (HadISST) and (*b*) simulations (61) of the Communtity Climate System Model version 3.5 (CCSM3.5).

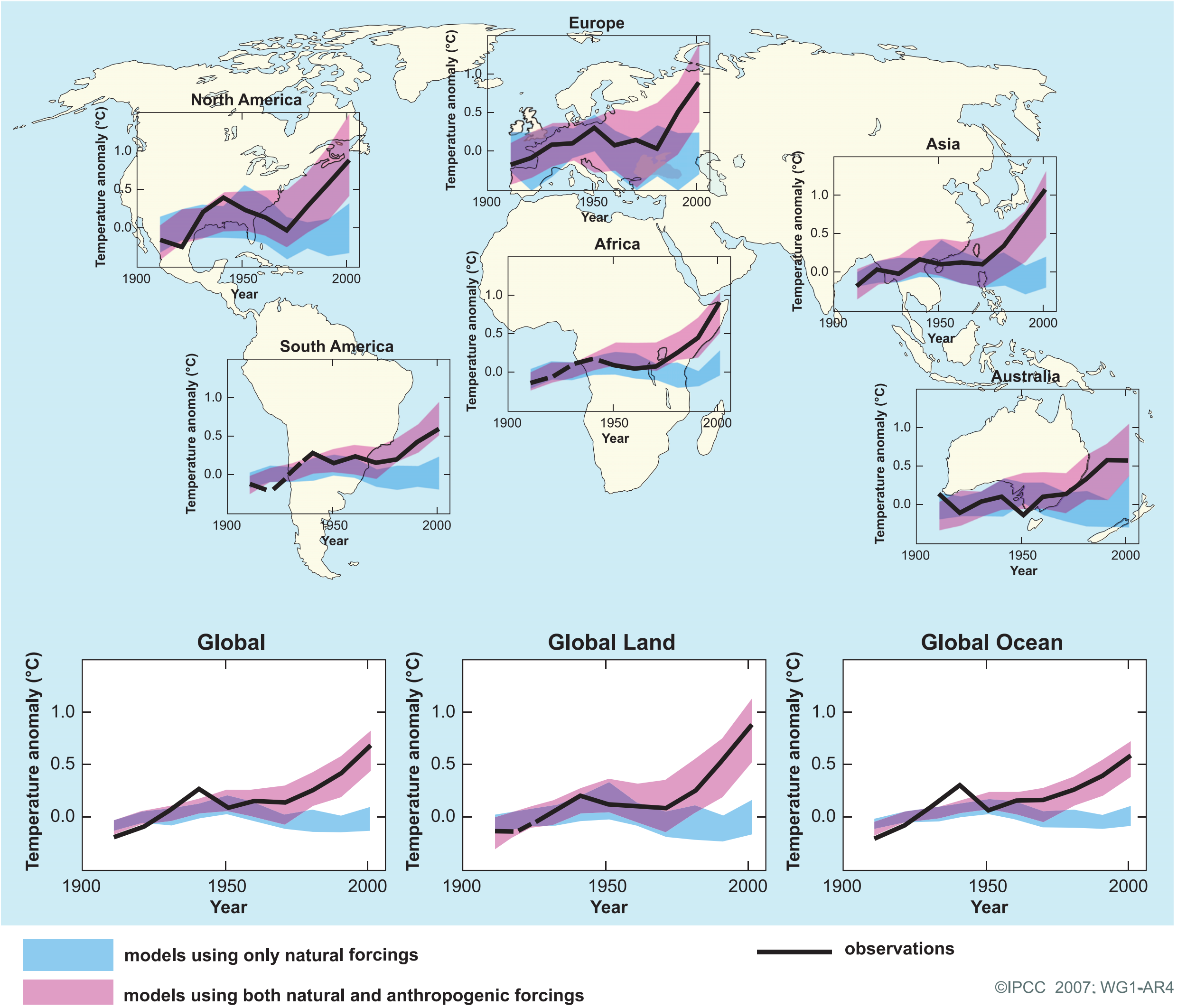

Figure 4

Temperatures during the twentieth century in 19 simulations by five climate models. Reprinted with permission from Reference 51, Figure SPM.4, p. 11.

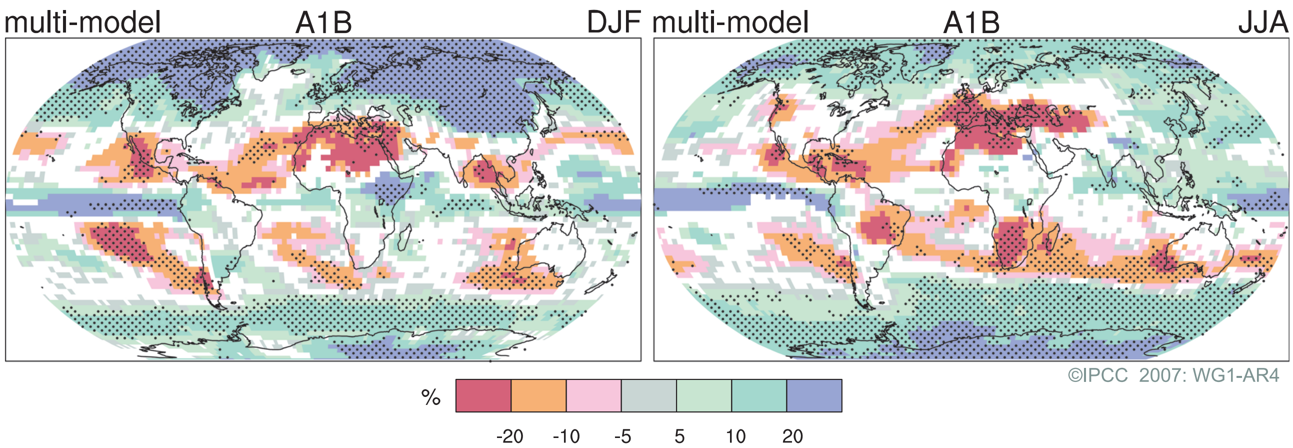

Figure 5

Relative changes in precipitation for the Intergovernmental Panel on Climate Change (IPCC) emissions scenario A1B for 2090–2099 relative to 1980–1999 (51). The left panel shows changes for December-February (DJF), and the right panel shows changes for June-August (JJA). Values are multimodel averages, with white areas indicating that less than 66% of the models agreed on the sign of the change and stippled areas indicating that more than 90% of the models agreed on the sign of the change. Reprinted with permission from Reference 51, Figure SPM.7, p. 16.

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