Coupled Ocean-Atmosphere Models: Physical Processes

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Synopsis

This article describes the concepts and parameterization methodologies of the major sources and sinks of heat, momentum, and constitutes in the coupled ocean–atmosphere models, along with their limitations and uncertainties. The article also gives examples of the current performances of these models and discusses their major common biases.

The basis of coupled ocean–atmosphere models is a set of physical laws that govern the motion, temperature, pressure, density of the atmosphere and ocean and the concentrations of constituents in them: The Newton's second law governs the atmospheric winds and ocean currents; the first thermodynamic law describes the temperature change as a function of the heat sources and sinks; the law of mass conservation expresses the changes of concentration of various constituents as functions of their respective sources and sinks. These laws are written on a set of discretized grids and are averaged in the individual domains corresponding to this discretized set of grids. All sources and sinks of heat and constituents, and the effects of unresolved processes on the revolved scale fields, are referred to as physical processes in coupled ocean–atmosphere models. This article describes these processes. The differences in the parameterizations of these processes among the models are believed to be the main cause of differences in their simulations and predictions of weather, climate variability, and future climate change.

Radiative Transfer

Absorption and emission of radiation are internal sources and sinks of heat within the atmosphere and oceans. The wavelengths of radiation from the Sun span the range from alpha, gamma, and X-rays, to UV radiation, visible light, infrared radiation, extending to microwaves and radio waves. At the Sun's temperature of about 6000 K, the majority of the radiative energy from the Sun lies in the range from the UV to the near-infrared, with the peak in the visible lights of 0.4–0.7 μm wavelengths. Radiation emitted from the atmosphere, the oceans, and the land surface is primarily in the range of infrared radiation with peaks in the band from 10 to 20 μm due to their lower temperatures. Because of the large difference of the temperatures of the Sun and the Earth and the separation of their respective radiation wavelength spectra, ocean–atmosphere models often calculate radiations from the Sun and from the Earth separately with different approximations as solar radiation and infrared radiation. They are also referred to as shortwave (solar) and longwave (infrared) radiations. Figure 1 shows the separation of radiation spectra from two emitting bodies of 6000 and 285 K representing the Sun and the terrestrial systems.

Radiative transfer in gases is highly sensitive to wavelengths due to the interaction of electromagnetic waves with the quantized molecular energy structure of gases. Absorption and emission of radiation differ for different types of gas molecules. Figure 2 shows portions of the absorption spectra in the shortwave for water vapor molecules and in the longwave for carbon dioxide molecules. These spectral properties are obtained from standard databases constructed from spectroscopy theory and laboratory measurements. Almost all atmospheric radiative transfer models use data from the HITRAN database (High-resolution Transmission molecular absorption database). HITRAN is a long-running project started by the Air Force Cambridge Research Laboratories in the late 1960s in response to the need for detailed knowledge of the infrared properties of the atmosphere. It has been continuously developed at the Harvard–Smithsonian Center for Astrophysics with participation of a large community. The HITRAN database includes spectroscopic parameters for 39 atmospheric molecules. These

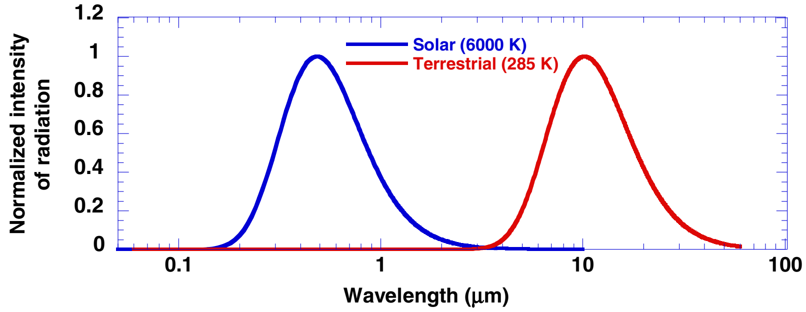

Figure 1 Wavelength dependence of normalized blackbody radiation at 6000 K (Sun) and 285 K (Earth).

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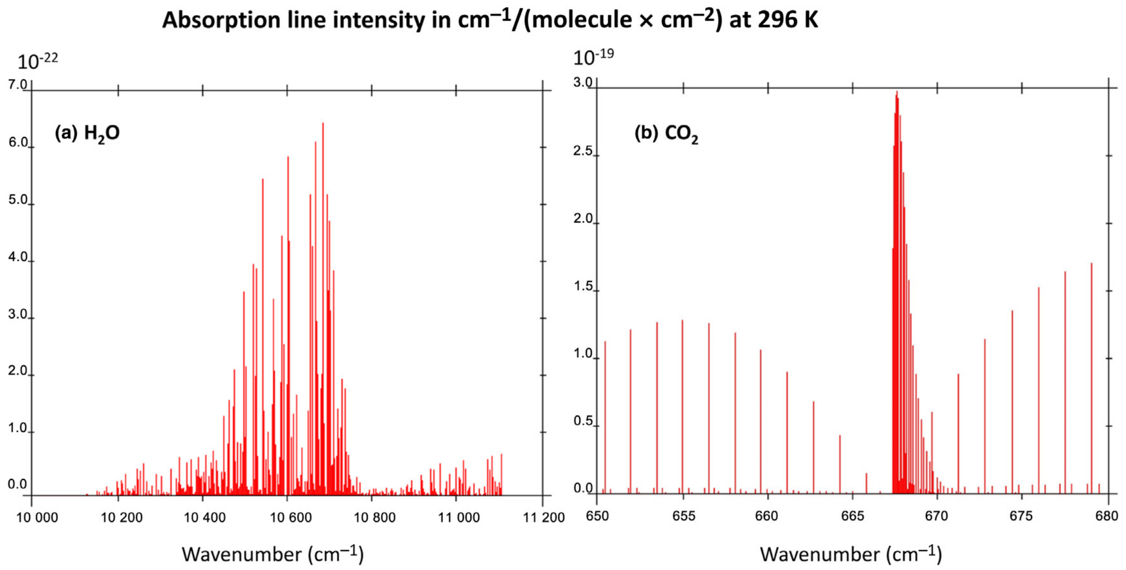

Figure 2 (a) Intensity of absorption of shortwave radiation by a water vapor molecule in the spectral range of 10 000 cm^{-1} (1 μm) to 11 200 cm^{-1} (0.89 μm). (b) Intensity of absorption of longwave radiation by a carbon dioxide molecule in the spectral range of 650 cm^{-1} (15.38 μm) to 680 cm^{-1} (14.71 μm).

include all molecule types that need to be considered for atmospheric radiative transfer, such as water vapor, carbon dioxide, ozone, methane, and other atmospheric trace gases.

The most comprehensive atmospheric radiative transfer models use the line-by-line calculations. They compute radiative transfer at individual wavelengths, and sum all wavelengths together to obtain the radiative energy. These models are computationally prohibitive to use in ocean–atmosphere models. Therefore, drastic simplifications are used to divide the entire shortwave and longwave spectra into no more than two dozen broad bands. These bands are selected based on the absorption behavior of all gas molecules in the atmosphere. Within each band, the atmospheric absorption, emission, and scattering are calculated by using various approximations. The most widely used approach is the so-called correlated-*k* method, in which the wavelength dependence of the absorptions is sorted according to the magnitudes (or *g*-points) into different bins; radiative transfer is then calculated in these bins, so that the wavelength dependence is minimized. The bins are summed to obtain the radiative fluxes in the bands; the bands are summed to obtain values for the whole spectra.

Radiative transfer calculation in the ocean is much simpler than that in the atmosphere. Radiation is absorbed in very short distances from the ocean surface. Only reflection and scattering at the ocean surface and absorption within about 30 m from the surface are calculated. The specification of the reflection and scattering contains uncertainties because the ocean surface is not perfectly flat. The atmospheric winds continuously induce waves of different scales at the surface. The absorption of radiation within the top layer of the ocean is also affected by the abundance of chlorophylls. The chlorophylls concentration is often specified from observational climatology with large uncertainties. Some ocean–atmosphere models do not include the treatment of chlorophylls in radiation calculation.

Radiative transfer in sea ice is treated the same as in the ocean but with different reflection, scattering, and absorption properties. Sophisticated models consider air bubbles and melted water that are trapped in the ice; they also consider the roughness of the ice surface. However, uncertainties of these variables are large.

Radiative transfer over land is calculated based on surface types. These include vegetation, snow, lake, bare soil, and urban structure. Vegetation is often categorized into various plant functional types. For each vegetation type, the areas and shapes of the leaves and the height of plants are assumed. For snow, the granule size, snow age, and area concentration of dark particle matters are needed as input. For bare soil, the color properties are specified as input. The radiative transfer calculation gives the amount of radiation that is reflected, scattered, and absorbed by the surface. Radiation in the vegetation influences the photosynthesis and canopy temperature, thus affecting water vapor evaporation and transpiration from the leaves. Uncertainties in the input data of radiative transfer over land can be large. The spatial heterogeneity of the input data is often difficult to parameterize.

Solar Radiation

Solar radiation reaching to the top of the atmosphere is calculated based on the Sun–Earth distance, the eccentricity (orbital shape of the Earth), the procession (the rotation of the tilted axis of the Earth), and the obliquity (the angle between the Earth's self-rotation axis and the orbital axis). Solar activities such as the 11-year solar cycle are considered in most models. The absolute amount of the output of energy from the Sun is not precise. Current satellite measurements give a range of mean incoming radiation to the top of the atmosphere, or the solar constant, from about 1370 to

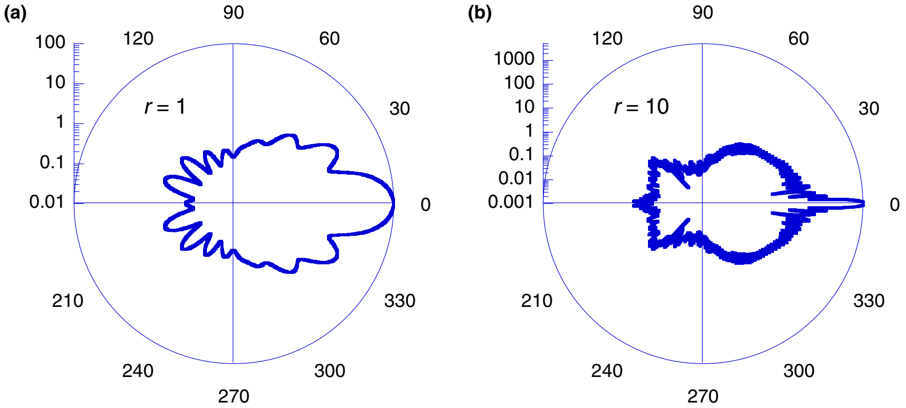

Figure 3 Angular distribution of scattered radiation of an incident visible light at 0.7 μm wavelength from the left by a liquid water particle of radius (a) 1 μm, (b) 10 μm.

1365 W m^{-2}. Different models may use slightly different values of the solar constant.

In the transfer of solar radiation, internal emitting sources of radiation within the atmosphere can be neglected. This is a convenient simplification. However, because the wavelengths of solar radiation are prone to strong interactions with atmospheric aerosol particles and cloud particles with sizes of 1–10 μm, solar radiation is strongly scattered by these particles. The scattered radiation is further scattered by the particles, leading to multiple scattering processes until they are absorbed by the particles, atmosphere, the surface, or are scattered to outer space.

Scattering of radiation by particles is highly dependent on the size of the particles and the wavelengths of the radiation as well as the electromagnetic properties of the particles. This affects the total amount of scattered and absorbed radiation, and the directional distribution of the scattered radiation. Figure 3 shows an example of the normalized directional distribution of scattered radiation by a cloud particle in liquid phase at two wavelengths. Because of these wavelength dependences and because of the multiple scattering processes, the calculation of solar radiation involves integration across space angles. Fortunately, the multiple scattering processes smear out some of the angular inhomogeneity, so many of the bulk simplifications, such as the use of two streams to describe the directional distribution, and the averaged optical properties of the particles, can provide accurate calculation of scattered radiations.

The largest uncertainties in the calculation of the solar radiation are due to the input data, particularly cloud particles and atmospheric aerosols. Cloud processes are crudely parameterized in current ocean–atmosphere models (see later Section on Clouds). Atmospheric aerosols include many types; their concentration, sizes, shapes, and time evolution represent some of the largest uncertainties in radiative calculations.

Current ocean–atmosphere models all use the plane-parallel assumption in which the radiation and the atmosphere within a grid are horizontally homogeneous. Three-dimensional radiative transfer calculations have indicated that lateral radiative fluxes may reach tens of watts per square meter at the cloud scales. It is not clear whether these lateral fluxes are important in high-resolution climate models. Three-dimensional radiative transfer calculations will incur significantly more computational costs. It is unlikely that they will be implemented in climate models in the foreseeable future.

Infrared Radiation

The surface and the atmosphere emit and absorb infrared radiation. Even though these are internal sources, calculation of infrared radiative transfer is much simpler than that of solar radiation. This is because at the wavelengths of infrared radiation, scattering by cloud particles and the majority of aerosols can be neglected.

Scattering of infrared radiation can occur on large aerosol particles. These are typically not considered in current ocean–atmosphere models. The large particles have short residence time in the atmosphere due to fast gravitational sedimentation, so the errors of neglecting them are small. Scattering of infrared radiation can also occur on cloud particles, but because absorption of infrared radiation by liquid or ice particles is large in this wavelength range, radiation is absorbed after a few scattering events, so scattering can be safely ignored.

Clouds

Clouds strongly regulate the energy balance of the atmosphere–earth system. Cloud particles reflect solar radiation to space, acting to cool the planet; they also strongly absorb infrared radiation emitted by the surface and by the below-cloud atmosphere, acting to trap infrared radiation and warm the planet. How clouds vary in response to climate changes, including variations in their area coverage, thickness, altitude, and liquid and ice water content, is the subject of intensive research in the last three decades as the cloud-climate feedback problem. Clouds are part of the hydrological cycle of the atmosphere–earth system. They are a manifestation of the phase changes of water between vapor, liquid, and ice in the atmosphere.

Cloud processes in ocean–atmosphere models are typically represented by two components. One is the cloud macrophysics; the other is cloud microphysics. The cloud macrophysical component calculates the fractional area coverage of clouds, spatial inhomogeneity of cloud properties, grid-scale condensation and evaporation, sublimation and vaporization when only a fraction of the grid domain is occupied by clouds. The macrophysical parameterization is often based on empirical relationships and assumptions on the subgrid scale distribution of total water. These empirical relationships or assumptions are approximate. They are often made differently for different types of clouds. They should be resolution dependent, but most current models are not designed to account for their dependences on resolutions.

The cloud microphysical component calculates the time evolution of the condensed cloud mass, particle number concentration, and their sources and sinks. Some cloud microphysical models only calculate the mass amount of condensed water. These models are called one-moment models. Models that include both the mass amount and number concentration are called two-moment models. To calculate the impact of aerosol on cloud microphysics, two-moment schemes are needed. Therefore, most current generation ocean–atmosphere models use two-moment schemes. Sources and sinks of cloud mass are calculated based on grid-scale condensation or sublimation, evaporation or vaporization, conversion between cloud drops and raindrops and snow, scavenging by rain and snow, breakout of raindrops or snowflakes. Sources and sinks of number concentration are calculated based on nucleation, evaporation, and vaporization of raindrops and snowflakes, and their breakup and removal by rain and snow. Figure 4 shows an example of the processes in a two-moment scheme.

Current models treat ice clouds and mixed phase clouds very crudely. Ice clouds rely heavily on the presence of ice nuclei, whose concentration and nucleation properties are very uncertain. Ice crystals form in different shapes as functions of the ambient air conditions and ice nuclei. Different shapes have different density and falling speed. The most common practice in current models is to use a ramp function of temperature to calculate whether the clouds are liquid, ice, or mixed phase.

Precipitation

Precipitation processes include the formation and changes of raindrops, snowflakes, and their evolution in the falling process to the ground. Some models also include categories of graupels and hails. Similar to cloud microphysical models, if a model only calculates the precipitation mass, it is called a one-moment scheme; if a model calculates both the mass and number concentration of precipitation particles, it is called a two-moment scheme. The separation of cloud particles and precipitation particles is based on observational support that the size distributions of these two types of particles are typically well separated, and so they have very different falling speeds.

Formation of precipitation is calculated from conversion of cloud drops and ice crystals, and collection of cloud particles by the larger falling precipitation particles. The conversion occurs because cloud particles have different sizes and different sedimentation velocities. Large particles fall faster than smaller particles, and so the small particles are captured by large particles. Since the size distribution of cloud particles in current models is approximate, the formulation of the cloud-to-precipitation conversion contains many empirical parameters.

One efficient mechanism for precipitation particles to grow occurs in mixed phase clouds. The saturation vapor pressure over water is greater than over ice. In mixed phase clouds, water vapor is saturated with respect to liquid droplets, but supersaturated over ice particles. As a result, liquid particles can

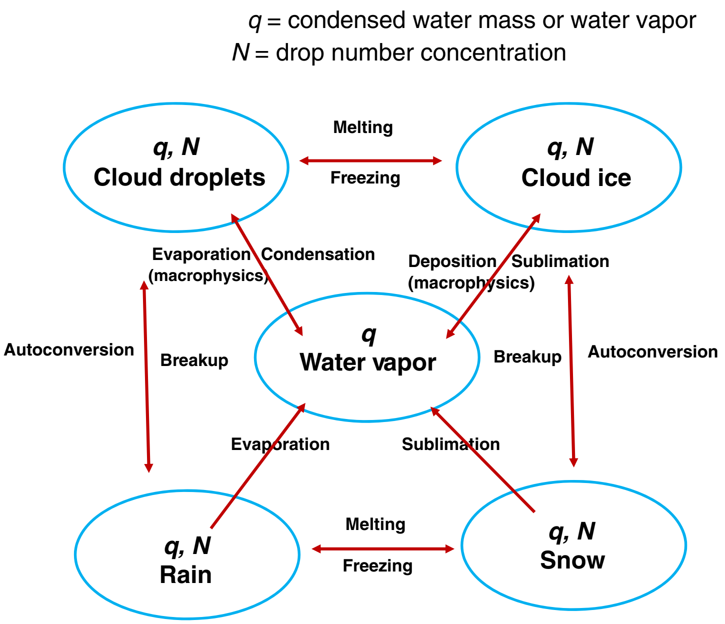

Figure 4 Schematics of some of the cloud and precipitation microphysical processes.

rapidly evaporate to sublimate to the ice particles to make them grow. This process, also called Bergeron process, is calculated in the precipitation parameterization. It also applies to ice crystals falling into supercooled liquid clouds.

Besides their formation, precipitation particles evolve along their falling path before reaching the ground. They accumulate cloud particles or smaller precipitation particles; they also experience evaporation or vaporization and breakup. The accumulation depends on the falling speed of the precipitation particles, which require information of their sizes and shapes. For raindrops, spherical shape can be safely assumed, but for snowflakes, since their shapes are highly uncertain and are crudely parameterized, the falling speed is approximate. The evaporation and vaporization of falling precipitation particles depend on the relative humidity of the ambient air, often taken as the mean condition of the grid domain, which may not be accurate.

Turbulent Mixing

Turbulent transports or mixing of heat, momentum, and constituents refer to the effects of subgrid scale processes on the resolved-scale fields. Because these subgrid scale processes operate at all scales smaller than resolved-scale fields, even in high-resolution models with grid spacing of several kilometers, turbulent transports still need to be parameterized. Turbulent fluxes are an essential part of any ocean–atmosphere model. For example, the turbulent transport of water vapor from the surface to the atmosphere is equal to the global precipitation.

Equations of turbulent fluxes can be formulated if higher order subgrid transport terms are known. The calculation of these high-order terms requires even higher order terms. The equations are therefore not closed and approximations are made to parameterize some terms. The simplest type of turbulence parameterization is a diffusion scheme, in which the turbulent flux is represented by the vertical gradient of a field multiplied by a diagnostically calculated diffusivity. These schemes are called first order model. The diffusivity is typically a specified function of vertical static stability and wind shear, or the dimensionless Richardson number. In some models it is calculated as a prescribed function of height, with a peak in the middle of the boundary layer to mimic results from large-eddy simulation models. Some models also add a counter-gradient term to the down-gradient diffusion term to account for large-eddy transport. This term is parameterized based on the source of large eddies such as surface buoyancy fluxes.

Some models prognostically calculate the turbulent kinetic energy (TKE) based on buoyancy flux, shear production, transport, and dissipation of eddy energy. This energy is then used in the calculation of the diffusivity. Other models may diagnostically calculate TKE. If the TKE is diagnostically calculated, the scheme is still first order; if it is prognostically calculated, the scheme is often called one-and-half order, since not all of second order moments are prognostically calculated. Very few models use second order or higher order closure schemes.

In the atmosphere, two separate types of turbulent transports are typically parameterized. One is for the free atmosphere that is away from the atmospheric boundary layer (ABL); the other is for the ABL. There are no fundamental differences in the philosophy of their parameterizations, except for the degree of simplifications. In the ABL, most models first judge whether the layer is a stable boundary layer or unstable boundary layer. Different closure assumptions are used based on the stability. Convective boundary layer, in which the surface upward buoyance flux is positive, is the most common type ABL.

In the ocean, the boundary layer turbulence is driven by wind mixing, shear of ocean currents, and static instability. The static instability can be caused by cooling of ocean surface and salinity change. The parameterization of oceanic turbulent fluxes in the vertical direction uses similar approaches as that for the atmosphere. In the horizontal direction, however, turbulent mixing in the ocean has two more complications than that in the atmosphere. First, because of the long timescale of ocean circulations and the large role of mesoscale eddies, horizontal turbulent mixing cannot be neglected. Earlier models parameterize the horizontal diffusivities of small and mesoscale eddies on the native model horizontal surfaces. Current generation models formulate them along the isopycnal or density surfaces. A second complication is the turbulent mixing along the lateral boundaries. The same equations as for the vertical boundary layer are often used, in which the turbulence is primarily driven by horizontal shear of ocean currents.

Surface Fluxes

Surface fluxes include shortwave and longwave radiative fluxes, and the heat, momentum, constituent fluxes across the interfaces between the atmosphere and the ocean or land. The radiative fluxes are calculated from the radiative transfer parameterizations described in previous sections. The turbulent heat, momentum, and constituent fluxes across the surface are calculated using similarity theory instead of the eddy diffusivities because the turbulence at the surface is strongly confined by the surface geometry.

In the similarity theory, a length scale – the Monin–Obukhov length scale – is used to normalize the height in the surface boundary layer. Empirical relationships of the vertical profiles of the winds, temperature, and constituents are obtained with respect to this normalized height based on observations. The surface fluxes appear as constant parameters in the profile relationships and in the Monin–Obukhov length scale. Vertical integration of these profiles from the surface to a reference height gives the formula to calculate the surface fluxes. The most widely known forms are

$$\tau = -\rho_{\mathrm{air}} C_{\mathrm{D}} |V_{10}| v_{10}$$

$$H = \rho_{\mathrm{air}} C_{\mathrm{H}} |V_{10}| (T_{\mathrm{s}} - T_2)$$

$$E = \rho_{\mathrm{air}} C_{\mathrm{E}} |V_{10}| (q_{\mathrm{s}} - q_2)$$

where τ, H, E are surface momentum stress acting on the atmosphere, surface heat flux, and constituent (such as water vapor) fluxes respectively; ρ_{air} is the density of air; V_{10} is the wind at 10 m height; T_s, q_s are temperature and the constituent concentrations at the surface; T_2, q_2 are temperature and constituent concentrations at 2 m; C_D, C_H, C_E are the bulk transfer coefficients of momentum, heat, and constituents. These coefficients are dependent on the fluxes themselves, so iterations are needed. In simple calculations or applications, however, they are often specified as constants.

At the ocean–atmosphere interface, net flux of freshwater is needed to calculate the salinity. The freshwater flux to the ocean is calculated by simply taking the difference of precipitation and surface evaporation.

Convection in the Atmosphere and Oceans

Convection is one type of turbulence, but because it is highly anisotropic in the vertical and horizontal directions, it is always separately parameterized. Convection schemes are used to calculate the convective transport of heat, momentum, constituents, and their sources and sinks within the convective portion of a grid cell.

Most atmospheric convection schemes use mass-flux plume models. These schemes parameterize the triggering condition of convection that determines when and where convection is activated. They also parameterize the entrainment and detrainment rates of the plume mass, the vertical distribution of the convective mass fluxes, and the cloud microphysical processes within the convective plumes.

The triggering condition of convection always includes atmospheric conditional instability as one criterion. However, models differ in using additional factors, such as the origination level of convection, whether the instability is calculated for undiluted or diluted air parcels, whether convection can be inhibited by a layer of negative buoyancy, and whether resolved-scale dynamics are considered. Lateral entrainment and detrainments can also differ greatly among different models: some use specified rates; others calculate them as functions of humidity of the ambient air; some even ignore them altogether. Convection mass fluxes are often calculated using closure assumptions, which are subject to considerable uncertainties. Cloud microphysical processes within convection are typically crudely parameterized such that when the condensed water exceeds a threshold value, the excess amount is treated as precipitation. These assumptions represent some of the largest uncertainties in current ocean–atmosphere models.

Most current convective parameterizations are diagnostic. They are not passed from one time step to another time step. Almost all schemes are designed for models with grid sizes that encompass an ensemble of cumulus clouds. As high-resolution models become more popular, these convection schemes may need structural improvements to correctly reflect their temporal and spatial effects on resolved-scale motions.

Atmospheric convective parameterizations often include separate schemes for shallow convection and for deep convection. The essential elements of these two types of schemes are the same. The differences are due to the different depth of the instability layer so that different assumptions are used for the entrainment and detrainment rates. In addition, because deep convection is typically thicker and is with larger instability, it is associated with strong precipitation, whose evaporation causes downdrafts.

Convection parameterization in the ocean typically uses simple adjustment schemes. Vertically unstable water column is adjusted to neutral stratifications. Convection is often initiated because of density anomaly at the ocean surface, which can be caused by radiative or evaporative cooling.

Gravity Waves

Most models parameterize the effects of internal gravity waves on the resolved-scale flow. These waves cannot be resolved by the large-scale models because of their small vertical and horizontal wavelengths. These gravity waves are calculated by using information of terrain, jets, or frontogenesis in highly idealized forms. The calculation also considers saturation of gravity waves for which the vertical wavelength is small enough to cause static instability. Momentum transport by the gravity waves is then obtained and vertically differentiated to obtain the momentum wave drag.

It has been shown that the momentum wave drag has a large impact on the mean zonal wind of the stratosphere and the temperature in polar regions. Large gravity wave drag slows down the westerly jet in the stratosphere and warms the polar stratosphere. Gravity wave parameterizations often contain many empirical parameters that are subject to large uncertainties.

Aerosols

Modern ocean–atmosphere models all include aerosols as prognostic variables. Aerosols not only affect radiation directly but also impact the cloud condensation nuclei and therefore clouds. These two effects are called aerosol direct and indirect effects on climate. The role of anthropogenic aerosols – aerosols caused by human activities – in past and future climate change has been a subject of intensive research in the last two decades.

Aerosols in the models are typically categorized into groups. These often include sulfate, ammonia, sea salt, dust, black carbon, and organic carbon. The mass amount and number concentration of each group are calculated. Assumptions are made on the size distributions of aerosols and how they are mixed, since these determine the radiative properties, water activity, and chemical reactions of the aerosols.

The parameterizations of aerosols include emissions of both natural and anthropogenic sources, such as dust and sea salt from natural emissions, sulfate and ammonia from anthropogenic sources. Aerosols are also calculated to evolve with time according to gas-particle conversion, coagulation, and chemical aging in addition to transport. The gas-particle conversion involves many types of precursor gases that originated from organic sources, many of which are not well known. Sinks of aerosols are removals by dry and wet depositions.

Superparameterizations

Superparameterization refers to the use of cloud-resolving models as a substitute of some physical parameterization components (cloud macrophysical and convection schemes) in large-scale atmospheric models. Since cloud-scale dynamics are resolved and the clouds are calculated based on these resolved-scale fields, superparameterization is superior to traditional parameterizations of cumulus convection and cloud processes. The disadvantage is the more expensive computational cost.

Current superparameterizations use spatial grid sizes of about 1 km in the cloud-resolving models. Boundary-layer turbulences and cloud microphysics still rely on parameterizations. Superparameterizations have been only recently used in coupled ocean–atmosphere models for limited amount of simulations.

Performances of Ocean–Atmosphere Models

When forced with incoming solar radiation, current ocean–atmosphere models are able to simulate earthlike global distribution of surface temperature, vertical distribution of atmospheric winds and ocean currents. They can also simulate cyclones, storm tracks, stationary waves, and interannual variability of tropical ocean temperature that resembles the observed El Niño events. As two examples, Figure 5 shows the simulated annually averaged global distribution of sea-surface temperature by the Community Earth System Model Version 1 and its comparison with observation. Figure 6 shows the simulated atmospheric zonally averaged eastward winds in the December–February months and its comparison with observation. The model is able to reproduce most of the important observational features in the ocean surface temperature and in the atmospheric winds.

Ocean–atmosphere models have been also used to simulate past climate change and to make projections of future climate. To calculate past changes, the models are given time-dependent forcing fields of greenhouse gases such as CO_2, solar variability, volcanic forcing, anthropogenic aerosol changes, and land use and land cover changes. Most models are able to simulate the global temperature increase in the twentieth century. These results have been summarized in the past reports of the Intergovernmental Panel on Climate Change.

Common Biases

There are several common biases in virtually all coupled ocean–atmosphere models that researchers have been trying to solve in the last several decades. One is the so-called double ITCZ syndrome. It refers to a double intertropical convergence zone (ITCZ) simulated by the models in the annual mean precipitation in the central Pacific that is not present in observations (Figure 7). Higher spatial resolutions do not seem to improve this aspect of the model simulations. The second long-standing problem is the overall failure of most models in simulating the Madden–Julian oscillation, which is an

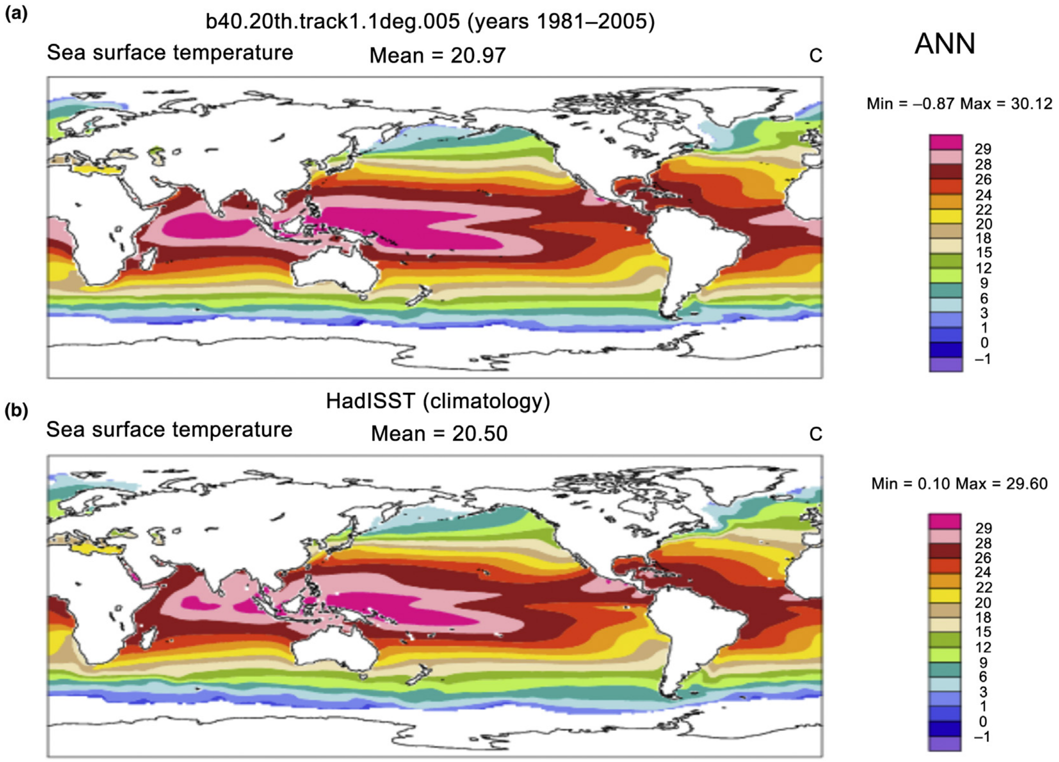

Figure 5 Annual mean sea surface temperature: (a) simulation by the Community Earth System Model Version 1; (b) observation. ANN refers to annual averages.

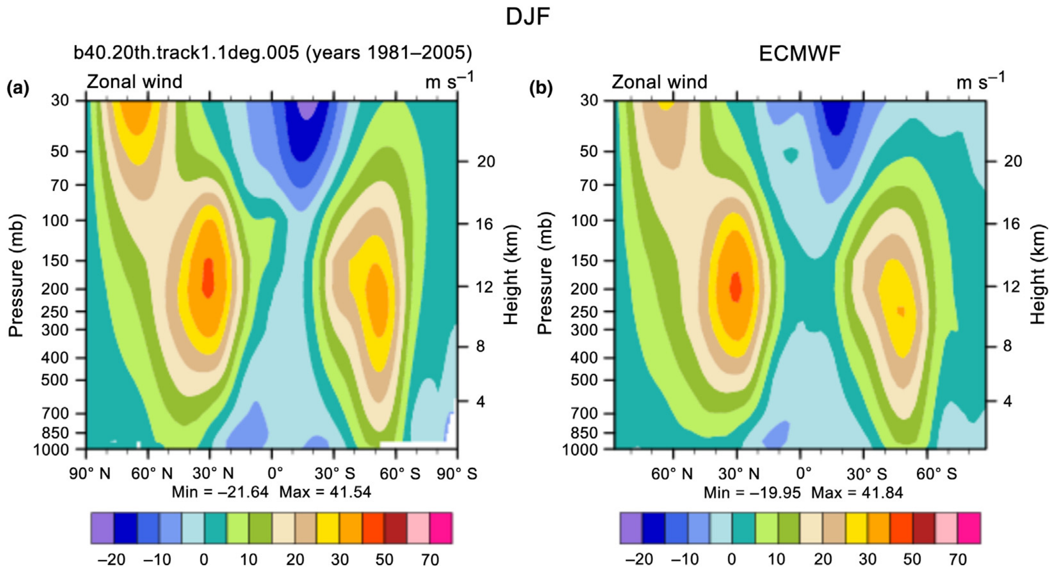

Figure 6 Height–latitude cross-section of eastward atmospheric wind averaged over all longitudes: (a) simulation by the Community Earth System Model Version 1; (b) observational estimates. DJF refers to December–January–February season. ECMWF refers to observational reanalysis at the European Center for Medium Range Weather Forecasting.

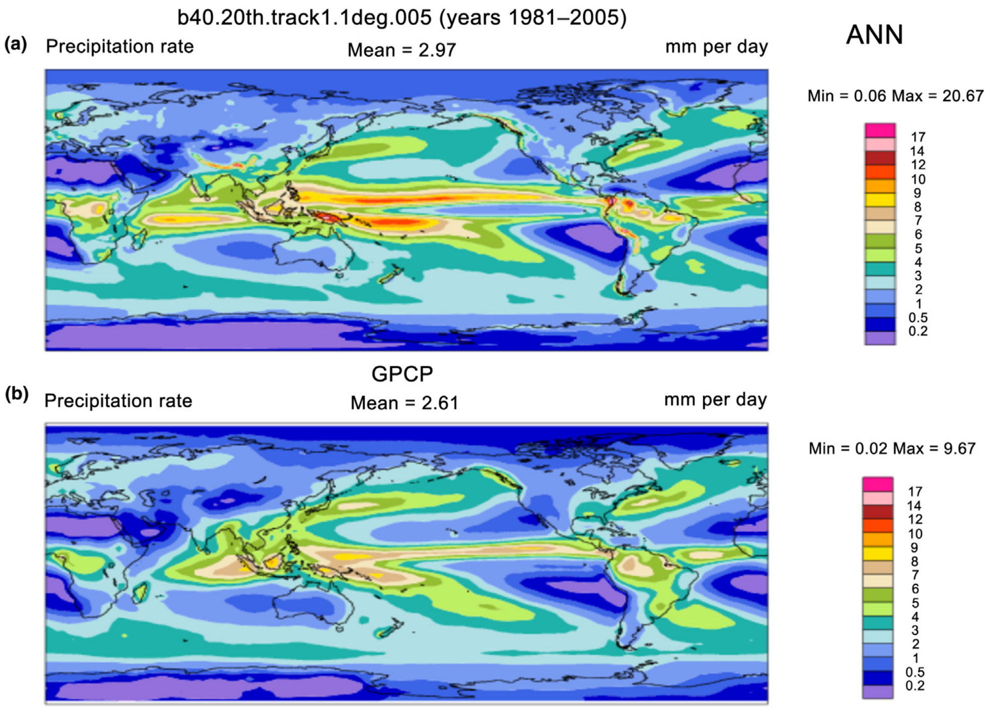

Figure 7 Annual mean precipitation: (a) simulation by the Community Earth System Model Version 1; (b) observation from the Global Precipitation Climatology Project (GPCP).

intraseasonal propagation of atmospheric circulation anomalies and convection from the equatorial eastern Indian Ocean to the western Pacific with a period of about 30–60 days. This intraseasonal oscillation has been shown to affect weather in many regions of the globe. The third common bias is the misrepresentation of the diurnal variation of precipitation over the oceans. Models simulate peak precipitation at noontime, but observations show peak precipitation during the night over most regions of the oceans. All these biases are believed to be due to inaccurate physical parameterizations in the models.

Individual models may have other significant biases, including the simulations of El Niño and its impact, monsoon rainfall, the deep ocean circulation, and land surface temperature. All these biases are being actively studied in the modeling centers.

See also: **Aerosols:** Role in Radiative Transfer. **Air Sea Interactions:** Momentum, Heat, and Vapor Fluxes. **Boundary Layer (Atmospheric) and Air Pollution:** Modeling and Parameterization; Ocean Mixed Layer. **Clouds and Fog:** Cloud Microphysics. **Land-Atmosphere Interactions:** Overview. **Numerical Models:** Cloud System Resolving Modeling and Aerosols; General Circulation Models; Model Physics Parameterization; Parameterization of Physical Processes: Clouds; Parameterization of Physical Processes: Gravity Wave Fluxes; Parameterization of Physical Processes: Turbulence and Mixing. **Radiation Transfer in the Atmosphere:** Absorption and Thermal Emission; Cloud-Radiative Processes; Scattering. **Turbulence and Mixing:** Overview.

Further Reading

Randall, D.A., 2000. General Circulation Model Development, Past, Present and Future. Academic Press, p. 807.