Chapter II - Description of Global Climate System Models

Modern climate models are comprised of a system of model components, each of which simulates a different part of the climate system, and usually can be run independently for certain applications. Nearly all of the CMIP3 class of models are composed of four primary components, the atmosphere, land surface, the ocean and sea ice. The atmospheric and ocean components are known as “general circulation models” or GCMs, because they explicitly simulate the large scale global circulation of the atmodphere and ocean. Sometimes, climate models are referred to as coupled atmosphere-ocean GCMs, which may be misleading, because a coupled GCM model can be employed to simulate aspects of weather and ocean dynamics, without being a climate model. What follows in this chapter is a description of the major components of a modern climate model, and how they are coupled together and tested for climate simulation.

Atmospheric General Circulation Models

Atmospheric general circulation models (AGCMs) are numerical programs that calculate the state variables of the atmosphere, such as temperature, pressure, humidity, kinetic energy, etc, as a function of space and time. . The set of model equations is formulated by using geophysical fluid dynamics theory and physical laws governing the exchanges of the mass and energy. because of the various assumptions and approximations that are made to more complete equations of classical fluid dynamics. The atmosphere can be thought of as a thin spherical shell of air that envelopes the earth. For climate simulation, typically only the lowest 20-30 km or so of the atmosphere, the troposphere and part of the stratosphere, are explicitly simulated. Within this volume all weather occurs because it contains over 95 % of the mass and virtually all of the water vapor. Because of disparity between the scales of the horizontal and vertical motions resolved in tyrpical global models, the horizontal motions are treated differently than vertical motions.by the model algorithms. The resulting basic set of equations is often referred to collectively as the primitive equations (Haltiner and Williams, 1980), Although nearly all AGCMs use the same primitive dynamical equations, they use different numerical algorithms to solve.them. In all cases, the atmosphere is divided into discrete vertical layers, which are then overlaid with a two dimensional horizontal grid, producing a three dimensional mesh of grid elements. The set of primitive equations is then solved as a function of space and time on this mesh. The portion of the model code governing the fluid dynamics explicitly simulated on this mesh is often referred to as the model's "dynamics." Computational solutions of the model dynamics can be grouped into four categories: spectral methods, finite difference methods, semi-Lagrangian methods, (Washington and Parkinson, 2005) and finite volume methods (Lin and Rood, 1996). The majority of the climate models use the first two approaches, Even with the same numerical approach, AGCMs differ in spatial resolutions and configuration of model grids. Some models have few layers above the troposphere (the moving boundary between the troposphere and stratosphere), while others could have as many layers above the troposphere as in it. AGCMs all use transformed equations to treat the Earth's surface as a constant coordinate surface so that the specification of heat, moistrure, trace substances and momentum exchanges between the earth's surface and the atmosphere can be simplified. Numerical algorithms of AGCMs should preserve the basic conservation of mass and energy of the atmosphere. Typical AGCMs have spatial resolution of 200 kilometers in the horizontal and 20 levels below the altitude of 15 km. Because numerical errors often depend on flow patterns, there are no simple ways to assess the accuracy of numerical discretization of AGCMs. Therefore, AGCMs are tested using a series of both idealized and realistic test cases (e.g. Held and Suarez ,1994) before being included in a climate model. Table 1 lists the specifications of numerical approaches and resolutions of some AGCMs.

All GCMs use parameterizations, or approximate sub-models, to simulate many processes that are too small, or operate on time scales too fast, to be resolved on the grid of the model dynamics. Some of the most important parameterizations are those that calculate radiant energy (or "radiative") transfer, cloud formation and dissipation, the vertical motions on small scales caused by thunderstorm clouds (cumulus convection), and turbulence and subgrid scale mixing. The radiative transfer code computes the absorption and emission of electrcomagnetic waves by air molecules and atmospheric particles. Most atmospheric gases absorb and emit radiation at discrete wavelengths, but the computational costs are too high to perform this calculation at individual wavelengths.

AGCMs use approximations, which differ among models, to group bands of wavelengths together in a single calculation. Most models have separate radiation codes to treat solar or visible, radiation differently from the much longer wavelength terrestrial, or infrared, radiation. The radiation calculation includes the effects of water vapor, carbon dioxide, ozone, and clouds. Many models also include aerosols and trace gases such as methane. Validation of the AGCM radiation codes is often done offline against resolved wavelength model calculations which, in turn, are compared against laboratory and field observations

For cloud calculations, AGCMs treat ice and liquid water as part of the atmospheric state variables. Some models also separate cloud particles into ice crystals, snow, graupel, cloud water, and rainwater. Empirical relationships are used to calculate conversions between different particle types. The representation of these processes on the scale of model grids is particularly difficult. It relies heavily on empirical formulations because of the lack of sub-grid scale information. This includes the calculation of cloud amount, which greatly affects radiative transfer and model sensitivity. Current models use one of the following two methods to calculate cloud amount: statistical distribution of thermodynamic and hydrological variables within a grid box, or prognostic cloud amount calculation. The statistical method may use simple model diagnostics, such as relative humidity, or more sophisticated calculations with higher order of moments of moisture contents. A sample of cloud schemes used in AGCMs is listed in Table II 1. None of the current AGCMs calculates size-resolved cloud particles nor do they treat the effects of and non-spherical ice particles.

Cumulus convective transports, which are important in the atmosphere but cannot be explicitly resolved at GCM scale, are calculated using convective parameterization algorithms. Most current models utilze a cumulus mass flux scheme patterned after that proposed by Arakawa and Schubert (1974), in which the upward motion is the convection is envisioned as occurring in very narrow plumes that takes up a negligible fraction of the area of a grid box. Schemes differ in the techniques used to determined the amount of mass flowing through these plumes, and the manner in which air is entrained and detrained by the plume as it rises. Most models do not separately calculate the area and vertical velocity of convection, but try to predict only the product of the mass and the area, or the covnective mass flux. Most current schemes do not account for the differences of convection between organized mesoscale systems and simple plumes. The turbulent mixing rate of updrafts and downdrafts with the environments, and the phase changes of water vapor within the convective systems with a mix of empiricism and constraints due to the moist thermodynamics of rising air parcels.. Some models also include a separate calculation of shallow, non-preciiptiating convection (or “fair-weather cumulus cloud) with different assumptions from those for deep convections. . Cloud genrated by cumulus convection should therefore be thought as based in large-part on empirical relationships. Convection schemes used in AGCMs are listed in Table II 1.

All AGCMs compute turbulent transport of momentum, moisture, and energy in the atmospheric boundary layer (ABL) near the surface. A long-standing theoretical framework, “Monin-Obukhov Similarity theory” is used to calculate the vertical distribution of turbulent fluxes and state variables in a thin air layer of tens of meters adjacent to the surface. Above that, turbulent fluxes are calculated based on covariances and closure assumptions for the ABL which differ among AGCMs. Some models use high order closures in which the fluxes or second order moments are prognostically calculated. Other models calculate the fluxes diagnostically. Turbulent ABL fluxes heavily depend on surface conditions such as roughness, soil moisture, and vegetation. Besides explicit calculation of boundary layer turbulence, all models use additional diffusion schemes to either calculate the impact of “clear air turbulence”, or to damp artificial numerical modes introduced in the discretization of the model. Table II.A lists turbulent schemes in AGCMs.

Table II. 1. Physical parameterization schemes in a sample of AGCMs.

	Resolution	Convection	ABL	Stratiform Clouds	Convective Clouds	Cloud Microphysics
CAM3	T85L26 (1.4^{o}x1.4^{o}) Spectral	Mass Flux [Hack 1994; Zhang and McFarlane, 1995]	1st order non-local [Holtslag and Boville, 1993]	Diagnostic (RH based) [Kiehl *et al.*, 1996]	Diagnostic [Rasch and Kristjansson, 1998]	Rasch and Kristjánsson [1998]
GFDL	2.5^{o}x2.0^{o}L24 Finite Difference	Mass flux (RAS) [Moorthi and Suarez, 1992]	Cloud entrainments [Lock *et al.*, 2000; GFDL GAMDT, 2004]	Prognostic [Tiedtke, 1993; GFDL GAMDT, 2004]	Prognostic [Tiedtke, 1993; GFDL GAMDT, 2004]	Rotstayn [1997], GFDL GAMDT [2004]
GISS	4^{o}x5^{o} L12 Finite Difference	Mass flux [Del Genio and Yao, 1993]	2nd order [Cheng *et al.*, 2002]	Diagnostic (RH based) [Del Genio et al., 2004]	Diagnostic [Del Genio *et al.*, 2004]	Del Genio *et al.* [2004]
GSFC	2.5^{o}x2^{o} L40 Finite Volume	Mass flux (RAS) [Moorthi and Suarez, 1992]	2.5 order [Helfand and Labraga, 1988]	Diagnostic (RH based) [Del Genio *et al.*, 2004]	Diagnostic [Del Genio *et al.*, 1996]	Del Genio et al. [1996], Sud and Walker [1999]
HadAM4	3.75^{o}x2.5^{o}L30 Finite Difference	Mass flux [Gregory and Rowntree, 1990; Gregory and Allen, 1991]	1st order with cloud entrainment [Lock *et al.*, 2000; Martin et al., 2000]	Diagnostic statistical [Smith, 1990; Pope et al.].	Diagnostic [Gregory and Rowntree, 1990]	Wilson and Ballard [1999]
ECHAM5	T63L31 (1.9^{o}x1.9^{o}) Spectral	Mass flux [Tiedtke, 1989; Nordeng, 1994]	1st order, [Brinkop and Roeckner, 1995]	Prognostic statistical [Tompkins, 2002],	Diagnostic [Roeckner *et al.*, 1996]	Lohmann and Roeckner [1996]
LMD	3.75^{o}x2.5^{o}L19	Emanuel	1st order [Li,	Statistical	Statistical	Le Treut and Li

	Finite Difference	[1991]	1999]	[Le Treut and Li, 1991]	[Bony and Emanuel, 2001]	[1991]

Ocean General Circulation Models

General overview: The ocean (Ocean General Circulation Models: OGCM) component of the current generation of climate models can be placed into one of two general categories. All the models are fully four dimensional primitive equation models and are coupled to the atmosphere and ice models through the exchange of fluxes of heat, temperature, and momentum at the boundary between components.. TableII 2 gives a brief summary of the major differences between the models described in the next paragraphs. Like the atmosphere, the horizontal dimensions of the ocean are much larger than the vertical dimension, again resulting in separating the processes that occur in the vertical from those that occur in the horizontal. Unlike the atmosphere, which only has to deal with terrain differences at the lower boundary, the ocean has a much more complex, three-dimensional boundary, with continents and submarine basins and ridges. Further, the fluid behavior of sea water is very different than that of air, resulting in a slightly different set of equations controlling ocean fluid dynamics.

The models utilized by the three US climate modeling groups that contributed models to the CMIP3 archive are used here to illustrate some of the choices made by ocean modelers.

An important category of OGCMs are referred to as Z-level models in which the model's vertical levels are calculated at fixed distances below the surface. (Many of these models are based on the early efforts of Bryan and Cox (1967) and Bryan (1969a, b). The GFDL and CCSM ocean components fall into this category (Griffies *et al*., 2005, Smith and Gent, 2002The models are similar in that the fundamental physical quantities advancing in time are the same. These quantities are velocity, potential temperature, salinity, sea surface height, and any number of specific passive tracers that maybe included for a given simulation. The two modeling efforts use similar horizontal resolution at about the same order: 1 degree or 110 km for most of the Earth and about 1/3 of a degree at the equator. Usually the models have increasing resolution between 5°N and 5°S to increase their ability to simulate important equatorial processes.

The vertical and horizontal structure of the models can also differ and are listed in TableII 2. The CCSM OGCM’s horizontal grid has its north pole displaced onto a land coordinate (a so-called stretched grid) and the GFDL models use a grid that has three poles (Murray, 1996). There is an explicit treatment of the bottom boundary and overflow regions in the GFDL models (Beckman and Doscher, 1997) to improve the down-slope flow of water. Such treatment of the overflows should improve the representation of deep ocean waters (Roberts and Wood, 1997), but problems remain (Griffies *et al*., 2005).

The second category of OGCMs includes those developed by GISS. There are two different ocean models that are used in the GISS simulations: the "Russell Ocean" (GISS-ModelE-R and GISS-AOM: Russell *et al*., 1995, Russell *et al*., 2000) and the "HYCOM Ocean" (GISS-ModelE-H: Sun and Bleck, 2001; Bleck 2002; Sun and Hansen, 2003; **Hy**brid **C**oordinate **O**cean **M**odel). The fundamental (prognostic) variables for the E-R and AOM simulations are potential enthalpy (rather than potential temperature), salt, mass, vertical gradients of potential enthalpy and salt, in addition to velocity. At this time, these models are run at a resolution much lower than the models of the first category (see Table 2 The vertical coordinate is defined in units of mass/unit area (while in category 1, the unit is meters).

The HYCOM OGCM (GISS-EH) fundamental variables include temperature, salinity, layer thickness, and velocity,. The horizontal grid is different from the others described. It is two grids, with one a Mercator grid to 60°N with a resolution of 2°and it is patched (i.e. boundary values exchanged at each time step) to a North Pole grid defined as 1° at 60°N to 0.5° at the North Pole. The vertical grid is a complex or "hybrid" with a z-level grid (units meters) to represent the mixed upper ocean and layers below represented as density layers (Bleck, 2002).

The analyses of the simulations, in most cases, are performed on the model fields that are interpolated to a common grid. This interpolation may introduce small inaccuracies (AchutaRao *et al.*, 2006) in the results of analyses of a model, but is not considered significant. For example, no more than 3% of heat content change can be associated with regridding errors at the end of a simulation.

Table II 2 Ocean CGM Characteristics

Model	Resolution Long x Lat L = Levels	Diabatic Mixing	Adiabatic Mixing	Primary Variables	Other Comments
CCSM3 POP	320x395 L40	KPP	GM	Velocity, T, S, SSH, ideal age	z-level vertical coordinate
GFDL: CM2: OM3	360x200 L50	KPP	GM	Velocity, T, S, SSH, ideal age	z-level vertical coordinate
GISS: AOM	90x60 L16	KPP	none	Potential Enthalpy, velocity, salt, mass	z*vertical coordinate
GISS: ER	72x46 L16	KPP	GM	See AOM	See AOM
GISS EH:	180x90	Kraus-Turner	No special treatment	T, S, SSH, mass flux, velocity	Isopycnal Vertical coordinate

Parameterization

Ocean Mixing: At the interface of the atmosphere and the ocean, the sea surface temperature plays a critical role in the climate problem. Processes that control mixing in the ocean are complicated and take place on small scales (order of centimeters) in the turbulent regime near the surface (the mixed layer). Within the stratified, adiabatic interior of the ocean, mixing is influenced by the exchange of water on scales on the order of meters to kilometers (**Figure II.A**). The current ocean components of climate models are at resolutions that are greater than either of these scales. The mixing of the ocean contributes to the ocean's stratification and heat uptake. This stratification, in turn, affects the circulation patterns on temporal scales of decades and longer. It is also generally felt (Schopf *et al.,* 2003) that the mixing schemes in the ocean modeling components contribute significantly to the uncertainty in the estimates of the ocean's contribution to the predictions of climate change.

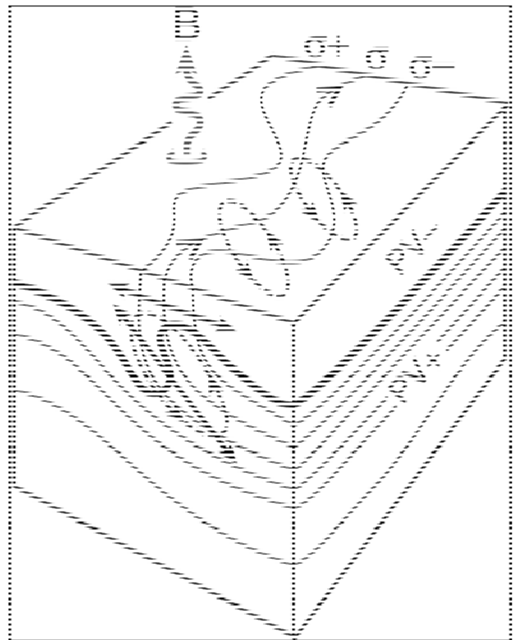


Figure II.A. Schematic showing the interaction of a mixed layer (low Potential Vorticity: PV) with the stratified interior (high PV) in a strong frontal region with outcropping isopycnal surfaces, , undergoing cooling, "B" indicates where eddies forming along the front play a central role in controlling horizontal fluxes through the mixed layer and quasi-adiabatic exchange between the mixed layer and the interior. This process is poorly observed, understood and modeled and must be parameterized in large-scale models. (from *Coupling Process and Model Studies of Ocean Mixing to Improve Climate Models - A Pilot Climate Process Modeling and Science Team*, a US CLIVAR white paper by Schopf, Gregg, Ferrari, *et al*., (2003).

For turbulent mixing of the upper ocean at the boundary with the atmosphere, the current generation of climate models (resolutions on the order of degrees) parameterizes the processes primarily through the use of several different approaches. Large *et al*., (1994) also provides a more complete comparison of these mixing schemes. While not all international AOGCMs use the K-profile Parameterization (KPP; Large *et al*., 1994) scheme, most of the major US climate models incorporate a version of the scheme. Li *et al*., (2001) showed that in the tropical Pacific, the use of the KPP scheme for handling the mixed layer of the upper ocean reduced the error in the simulation as compared to observations over a simulation that used a more simplified method (Pacanowski and Philander, 1981).

The adiabatic mixing, related to the interactions of eddy motions, generally is handled through the incorporation of the methods of Gent and McWilliams (GM) (1990) and Griffies (1998). Eddies will generally mix the ocean on constant density surface. The GM method incorporates various separate parameters that include the scale of the process to be considered and a parameter related to the ability of a parcel to move up and down. For any model the parameters are set so that coefficient related to diffusivity is high in the boundary currents and low in the interior of the ocean (Griffies *et al*., 2006). The ocean's flow is effected by the eddies, leading to adjustments in how much heat is moved through the oceans, and thus impacts the climate characteristics of the ocean.

To accurately represent ocean mixing at scales important to climate, other processes may need to be represented explicitly or parameterized in the model. These include incorporation of tidal mixing and more accurate representation of interactions with the ocean's bottom. Some of the models also include a scheme for handling tidal mixing (Lee *et al*., 2005). The limited study of *Lee et al.* (2005) shows that the tidal mixing enhanced the ventilation of the surface waters and increased the formation of deep water in the Labrador Sea by homogenizing the salinity distribution but did not have a major effect on the overturning circulation. It is still an open discussion on the importance of tidal mixing in ocean in relationship other larger scale changes occurring in the ocean related to climate. A few OGCMs also explicitly treat the bottom boundary and sill overflows (Beckman and Dosher, 1997).

***Other parameterizations*:** Another aspect of the model that is available to climate modelers when running the simulations is the explicit treatment for handling the penetration of sunlight (and thus, affecting chlorophyll distributions) into the upper ocean (e.g.. Paulson and Simpson, 1977: Morel and Antoine, 1994: Ohlmann, 2003). All of the US models include such capability. The inclusion of river input (which, in turn, effects ocean mixing locally) in OGCMs is also handled by the models in a variety of ways. The models' low resolution results in the smaller seas of the Earth being isolated from the large ocean basins. This requires that there be a method to exchange water between an isolated sea and the ocean to simulate what in nature involves a channel or strait. The various modeling groups have chosen different methods to handle the mixing of the water between these seas and the larger ocean basins, and is one potential source of model differences in climate simulations.

Evaluation of OGCMs: Like the atmosphere, ocean components of climate models are separately evaluated, in addition to the evaluation of coupled ocean-atmosphere GCMs discussed in Chapter V below. Ocean model evaluation requires specification (as input to the computer models) of boundary conditions at the air-sea interface. Typically, these are specified to match observations of the recent decades, and the OGCM simulation is then evaluated by comparison with observations of the ocean from the same time period. OGCM experiments with specified sea surface boundary conditions are at present less robost and generally exhibit more uncertainty in model performance than similar experiments for the atmosphere.

Land Surface Models

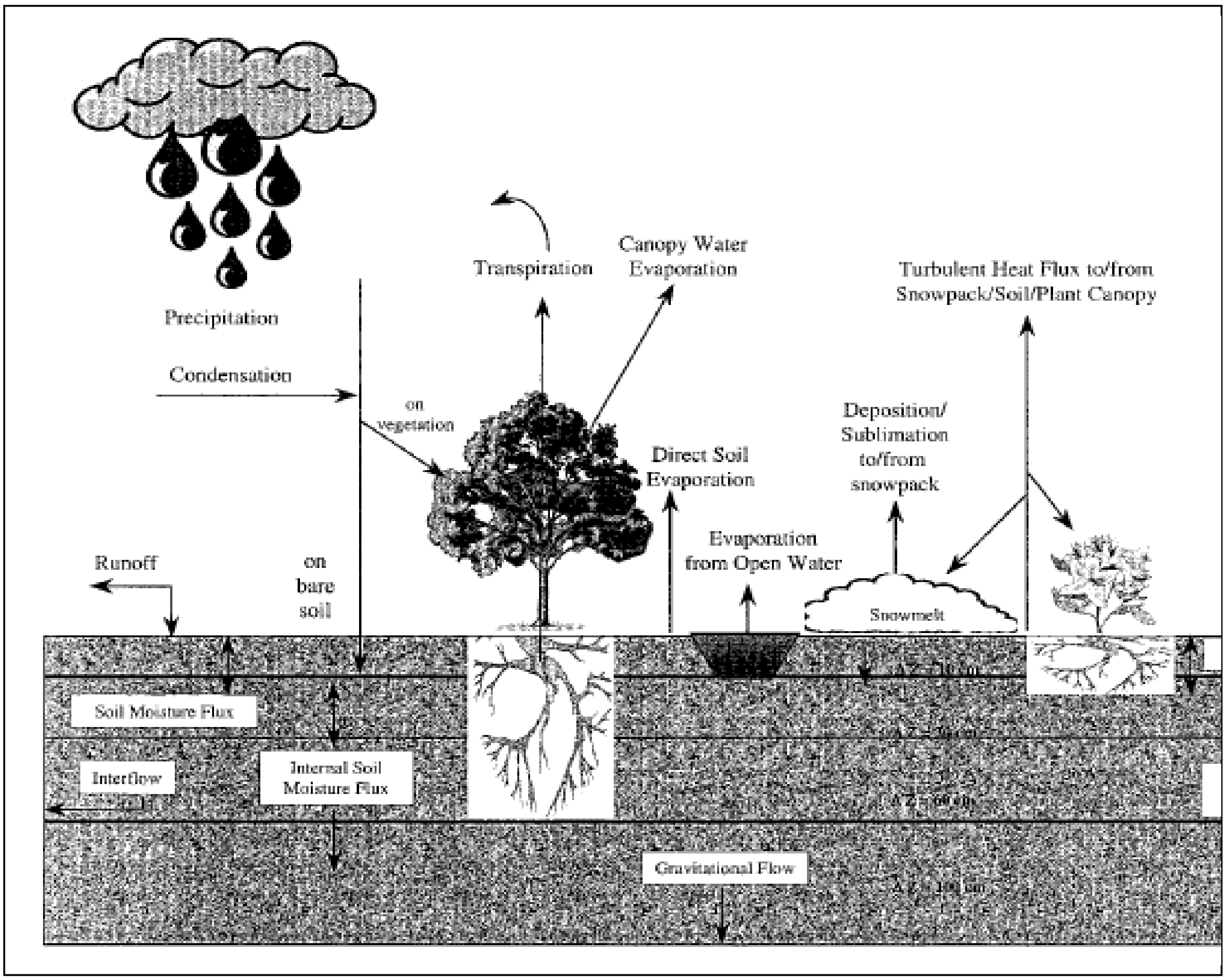
The interaction of the Earth's surface with the atmosphere is an integral aspect of the climate system. At the interface, there are exchanges (fluxes) of mass and energy, notably heat, water vapor, and momentum. Feedbacks between the atmosphere and the surface affecting these fluxes have important effects on the climate system (Seneviratne *et al*., 2006). Modeling the processes over land is particularly challenging because the land surface is very heterogeneous and biological mechanisms in plants are important. Climate model simulations are very sensitive to the choice of land parameterizations (Irannejad *et al*., 2003).

In the earliest global climate models, the land surface modeling occurred in large measure to provide a lower boundary to the atmosphere that was consistent with energy, momentum and moisture balances (e.g., Manabe 1969). The land surface was represented by a balance among incoming and outgoing energy fluxes and a "bucket" that received precipitation from the atmosphere and evaporated moisture into the atmosphere, with a portion of the bucket's water draining away from the model as a type of runoff. The bucket's depth equaled soil field capacity. There was little attention given to the detailed set of biological, chemical and physical processes linked together in the terrestrial portion of the climate system. From this simple starting point, land surface modeling for climate simulation has increased markedly in sophistication, with increasing realism and inclusiveness of terrestrial surface and subsurface processes.

Although these developments have increased the physical basis of land modeling, the greater complexity has at times contributed to greater differences between climate models (Gates *et al*., 1995). However, the advent of systematic programs comparing land models, such as the Project for Intercomparison of Land Surface Parameterization Schemes (PILPS; Henderson-Sellers *et al*., 1995; Henderson-Sellers, 2006) has gradually led to greater agreement with observations and among land models (Overgaard *et al*. 2006), in part because more observations have been used to constrain their behavior. However, choices for adding processes and increasing realism have varied between land-surface models (e.g., Randall et al. 2007), so convergence of simulations by current models should not be expected. This section reviews the range of developments that have led to contemporary simulation of land processes in climate models.

Figure II. B shows schematically the types of physical processes included in typical land models. It is noteworthy that the schematic in Figure II. B describes a land model used for both weather forecasting and climate simulation, an indication of the increasing sophistication demanded by both. The figure also hints at important biophysical and biogeochemical processes that have gradually been added to land models used for climate simulation (and continue to be added), such as biophysical controls on transpiration and carbon uptake.

Figure II. B. Schematic of physical processes in a contemporary land model (from Chen and Dudhia, 2001).



***Vegetation*:** Some of the most extensive increases in complexity and sophistication have occurred with vegetation modeling in land models. An early generation of land models (Wilson *et al*., 1987; Sellers *et al*., 1986) introduced biophysical controls on plant transpiration by adding a vegetation canopy over the surface, thereby implementing vegetative control on the terrestrial water cycle. These models included exchanges of energy and moisture between the surface, canopy and atmosphere, along with momentum loss to the surface. Further developments included improved plant physiology that allowed simulation of carbon dioxide fluxes (e.g., Bonan 1995; Sellers *et al*., 1996), which lets the model treat the flow of water and carbon dioxide as an optimization problem balancing carbon uptake for photosynthesis against water loss through transpiration. Improvements also included implementation of model parameters that could be calibrated with satellite observation (Sellers *et al*., 1996), thereby allowing global-scale calibration.

Continued development has included more realistic parameterization of roots (Arora and Boer, 2003; Kleidon, 2004) and adding multiple canopy layers (e.g., Gu *et al*., 1999; Baldocchi and Harley, 1995; Wilson *et al*., 2003). However, the latter has not been used in climate models as the added complexity of multi-canopy models renders unambiguous calibration very difficult. An important ongoing advance is the incorporation of biological processes that produce carbon sources and sinks through vegetation growth and decay and cycling of carbon in the soil (e.g., Li *et al*., 2006), although considerable work is needed to determine observed magnitudes of carbon uptake and depletion.

Soils: The spatial distribution of soils, at least for the contiguous U.S. appears to be fairly well mapped (Miller and White 1998). Most land models include only inorganic soils, generally composed of mixtures of loam, sand and clay. However, high-latitude regions may have extensive zones of organic soils (peat bogs), and some models have included organic soils topped by mosses, which has led to decreased soil heat flux and increased surface sensible and latent heat fluxes (Berringer et *al*., 2001).

Snow and ice: Climate models initially treated snow as a single layer that could grow through snow fall or deplete though melt (e.g., Dickinson et *al*., 1993). More recent land models for climate simulation include sub-grid distributions of snow depth (Liston, 2004) and blowing of snow (Essery and Pomeroy, 2004). Snow models now may use multiple layers to represent fluxes through the snow (Oleson *et al*., 2004). Effort has also gone into including and improving effects of soil freezing and thawing (Koren *et al*., 1999; Boone *et al*, 2000; Warrach *et al*., 2001; Li and Koike, 2003; Boisserie *et al*., 2006) though permafrost modeling is more limited (Malevsky-Malevich *et al*., 1999; Yamaguchi *et al*., 2005).

Vegetation interacts with snow by covering it, thereby masking snow's higher albedo (Betts and Ball, 1997) and retarding spring snowmelt (Sturm et al., 2005). The net effect is to maintain warmer temperatures than would occur without vegetation masking (Bonan et al., 1992). Vegetation also traps drifting snow (Sturm et al., 2001), insulating the soil from subfreezing winter air temperatures and potentially increasing nutrient release and enhancing vegetation growth (Sturm et al., 2001). The albedo masking is included in some land surface models, but it requires accurate simulations of snow depth to produce accurate simulation of surface-atmosphere energy exchanges (Strack et al., 2003).

Ice Sheets Global sea level is rising at a rate of 30 cm/century, thanks to a combination of ocean thermal expansion, melting of mountain glaciers and small ice caps, and retreat of the large ice sheets of Greenland and Antarctica (Cazenave and Nerem, 2004; Church and White 2006). The rise in sea level provides a common disruption and challenge to nearly every country, and the 400 million inhabitants who live within roughly 20 meters of elevation above sea level (Small et *al.*, 2000). By far the greatest uncertainty in sea level rise is associated with ice sheets. Complete melting of the Greenland and West Antarctic ice sheets, which are believed vulnerable to climate warming, would raise sea level by about 7 m and 5 m, respectively. During the last interglacial period, roughly 125,000 years ago, these ice sheets were smaller and sea level was a few meters higher than its present-day value (McCulloch and Ezat 2000, Siddall et *al.* 2003). Given the potentially catastrophic impacts of sea level rise, it is essential to be able to predict how fast ice sheets will melt and whether that melting, once begun, can be reversed. This is not yet possible because key ice sheet dynamical processes are poorly understood and are not included in current climate models. The recent IPCC assessment report (IPCC, 2007) underscores the need for improved ice sheet models, but because of the early stage of model development, specifically excluded rapid changes in ice flow from its 21st century sea level projections.

Ice sheets were once thought to be too sluggish to respond to climate change on time scales of less than a century. However, analysis of coral reefs at several locations indicate periods, including around 14,000 years ago, when sea level rose by as much as a few meters per century (Bard *et al.*, 1990). Recent observations suggest that ice sheets are already responding to warming. Outlet glaciers in Greenland have accelerated and thinned (Rignot and Kanagaratnam, 2006), driven by ocean warming and possibly by increased basal sliding. Ice shelves in the Amundsen Sea embayment of West Antarctica have thinned and retreated, giving rise to acceleration of glaciers tens of km upstream (Payne et *al*., 2004). Satellites provide near-complete spatial coverage and recent instruments have measured changes in total ice volume with precision that is unprecedented. Surface altimetry and synthetic aperture radar interferometry measure the height of the ice surface, and can be used to estimate changes in ice volume with additional information or assumptions about depth (Rignot and Kanagaratnam, 2006). Surveys of the changing gravitational field provide direct measurements of ice mass (Velicogna and Wahr 2006). Both indicate that the Greenland and Antarctic ice sheets are losing mass. Shepherd and Wingham (2007) estimate a net loss of about 125 Gt/yr (which includes losses of 100 Gt/yr for Greenland and 50 Gy/yr for West Antarctica, offset by a gain of 25 Gt/yr from increased snowfall in East Antarctica). The resulting contribution to sea level rise is currently a modest 3.5 cm/century, but this contribution will likely increase in a warming climate.

Most global climate models to date have been run with prescribed, immovable ice sheets, but several modeling groups are now incorporating dynamic ice sheet models. Scientists are coupling GLIMMER, an ice sheet model originally developed at the University of Bristol, to the Community Climate System Model. GLIMMER will be forced with temperature, precipitation, and other climate fields, and will return a modified surface elevation profile along with meltwater *freshwater* fluxes. As the ice sheet thins, melting will likely increase because the surface descends to a lower elevation where the temperature is higher *temperature-elevation feedback*. Meanwhile, meltwater *freshwater fluxes* will freshen the upper ocean and possible modify the thermohaline circulation. GLIMMER will initially be used to model the Greenland ice sheet and later will be used for simulations of the Antarctic ice sheet as well as paleo ice sheets (e.g., the Laurentide ice sheet that covered much of North America during the last glacial period).

Like most current-generation models, GLIMMER is based on the shallow-ice approximation, which assumes that ice flow is dominated by vertical shear. This approximation is valid in slow-moving ice sheet interiors but is insufficient to model fast dynamic changes near the ice sheet margin. A number of physical, numerical and computational improvements are needed to provide realistic projections of 21st century ice sheet changes. Among the major challenges are the following.

- Incorporate a unified treatment of stresses: both the vertical shear stresses that dominate in the ice sheet interior and the longitudinal stresses that are important in ice shelves and ice streams.
- Decrease grid spacing to 5 km or less to resolve small-scale features such as ice streams and outlet glaciers. This may require nested or unstructured grids, as well as parallel codes that scale efficiently with large numbers of processors.
- Develop improved methods of downscaling atmospheric fields, which are typically at a grid spacing of 100 km or more, to the finer ice sheet grid, making sure that energy is conserved in the process.
- Develop realistic parameterizations of surface and subglacial hydrology. Fast dynamic processes are largely controlled by the pressure and extent of water at the base of the ice sheet.
- Model the interaction of ice shelves with the ocean circulation. Ocean models, which usually assumed fixed topography, must be modified to include flow beneath advancing and retreating ice.

Meeting these challenges will require increased interaction between the glaciological and climate modeling communities, which until recently have been largely isolated from one another.

Hydrology: The initial focus of land models was vertical coupling of the surface with the overlying atmosphere. However, horizontal water flow through river routing has been available in some models for some time (e.g., Sausen *et al.*, 1994; Hagemann and Dümenil, 1998), with spatial resolution of routing in climate models increasing in more recent versions (Ducharne *et al.*, 2003).

However, freezing soil poses additional challenges for modeling runoff (Pitman *et al*., 1999), with more recent work showing some skill in representing its effects (Luo *et al*., 2003; Rawlins *et al*., 2003; Niu and Yang, 2006).

Work is also underway to couple ground-water models into land models (e.g., Gutowski *et al*., 2002; York *et al*., 2002; Liang *et al*., 2003; Maxwell and Miller, 2005; Yeh and Eltahir, 2005). Ground water potentially introduces longer time scales of interaction in the climate system in places where it has contact with vegetation roots or emerges through the surface.

Scale considerations: Land models encompass spatial scales ranging from the size of the model grid box down to biophysical and turbulence processes operating on scales the size of leaves. Explicit representation of all these scales in a climate model is beyond the scope of current computing systems as well as observing systems that would be needed to provide adequate model calibration for global and regional climate. As indicated above, land models have been developed to increase the sophistication of their climate-system simulation without becoming so complex as to be intractable. Thus, for example, typical land models in climate simulation do not represent individual leaves but the collective behavior of a canopy of leaves, and multiple canopy layers are generally represented by a single, effective canopy.

Although model fluxes are primarily in the vertical direction, they do not represent a single point but behavior in a grid box that may be many tens or hundreds of kilometers across. Initially, these grid boxes were treated as homogeneous units, but starting with the pioneering work of Avissar and Pielke (1989), many land models have tiled a grid box with patches of different land-use and vegetation types. Although these patches may not interact directly with their neighbors, they are linked by their coupling to the grid box's atmospheric column. This coupling does not allow possible small-scale circulations that might occur because of differences in surface-atmosphere energy exchanges between patches (Segal and Arritt, 1992; Segal *et al*., 1997), but under most conditions, the imprint of such spatial heterogeneity on the overlying atmospheric column appears to be limited to a few meters above the surface (e.g., Gutowski *et al*.,1998).

Vertical fluxes linking the surface, canopy and near-surface atmosphere generally assume some form of down-gradient diffusion, though counter-gradient fluxes can exist in this region much like in the overlying atmospheric boundary layer, so there has been some attempt to replace diffusion with more advanced, Lagrangian random-walk approaches (Gu *et al*., 1999; Baldocchi and Harley, 1995; Wilson *et al*., 2003).

Digital Elevation Models: Topographic variation within a grid box is usually ignored in land modeling. However, implementing detailed river-routing schemes will require accurate digital elevation models (e.g., Hirano *et al*., 2003; Saraf *et al*., 2005). In addition, some soil water schemes also include effects of land slope on water distribution (Choi *et al*., 2007) and surface radiative fluxes (Zhang *et al*., 2006).

Validation: Validation of land models, especially globally, remains a problem, due to lack of measurements for relevant quantities such as soil moisture and energy, momentum, moisture and carbon fluxes. PILPS (Henderson-Seller *et al*., 1995) has provided opportunity to make detailed comparisons of multiple models with observations at point locations around the world with differing climates, thus providing some constraint on the behavior of land models. Global participation in PILPS has led to a greater understanding of differences among schemes and improvements. The latest generation of land surface models exhibit relatively smaller differences (Henderson-Sellers *et al*., 2003) compared to previous generations. River routing can provide a diagnosis versus observations of the spatially distributed behavior of a land model (Kattsov *et al*., 2000). Remote sensing has been useful for calibration of models developed to exploit it, but it has not generally been used for model validation. The development of regional observing networks that aspire to give Earth-system observations, such as some of the mesonets in the United States, offers promise of spatially distributed observations of important fields for land models that resolve some of the spatial variability of land behavior.

Future: Land modeling has developed in other disciplines roughly concurrently with the advances implemented in climate models. Applications are wide ranging and include detailed models used for water resource planning (Andersson *et al*. 2006), managing ecosystems (e.g., Tenhunen *et al*., 1999), estimating crop yields (e.g., Jones and Kiniry, 1986; Hoogenboom *et al*.; 1992), simulating ice sheet behavior (Peltier, 2004), and projecting land-use, such as for transportation planning (e.g., Schweitzer; 2006). As suggested by this list, there are widely disparate applications, which have developed from differing scales of interest and focus processes. Land-model development in some of these other applications has informed advances in land models for climate simulation, as in representation of vegetation and hydrologic processes. Because land models do not include all climate system processes, they can be expected in the future to engage other disciplines and encompass a wider range of processes, especially as resolution increases.

Sea Ice Models, including parameterizations and evaluation

General overview: All the considered climate models have sea ice components that are both dynamic and thermodynamic. That is, the models include the physics for ice movement as well as the physics that is related to energy and heat within the ice. The differences in the various models relate primarily to how complex the code for the dynamics is in determining the representation of ice rheology and their use of parameters.

Two dynamical codes are in common use in ice models, the standard Hibler viscous-plastic (VP) rheology (Hibler, 1979; Zhang and Rothrock, 2000) and the more complex elastic-viscous-plastic (EVP) rheology of Hunke and Dukowicz (1997). The EVP method explicitly solves for the ice stress tensor, while the VP solution uses an implicit iterative approach. The solutions are similar (Hunke and Zhang, 1997). The NOAA-GFDL models [Delworth *et al*., 2005] and the NCAR-CCSM3 (Collins *et al*., 2005) use the EVP rheology, while the NASA-GISS models use the VP implementation. The EVP is more efficient, especially when using multiple processors.

The thermodynamics portions of the codes also vary in their implementation. Previous climate models generally used the thermodynamics code of Semtner (1976). This classic sea ice model includes one snow layer and two ice layers with constant heat conductivities and a simple parameterization of the brine (salt) content. The NOAA-GFDL models continue to use the Semtner structure with three layers but extend the code relating to brine content in the upper ice layer to be represented by variable heat capacity (Winton, 2000). The NCAR-CCSM3 and NASA-GISS models use variations of the Bitz and Lipscomb (1999) thermodynamics (Briegleb *et al*., 2002).

The code accounts for more of the physical processes within the ice, including the melting of internal brine regions and conserves energy.

The prognostic variables of the sea ice components of the separate climate models are similar to their ocean counterpart, that is the NOAA-GFDL and NCAR-CCSM use velocity, temperature and volume while the NASA-GISS models use velocity, enthalpy, and mass. The amounts of snow and ice for the layers are also computed with each model defining the number of ice layers and ice categories differently. The NOAA-GFDL models use a snow layer, two ice layers and five ice-thickness categories. The NCAR-CCSM3 model has a snow layer, four ice layers, and six ice categories. The NASA-GISS model includes one snow layer, three ice layers, and two ice categories. There is variation among the models on how ice categories are defined, but all include a "no ice" category. The resolution of the sea-ice component is the same as the ocean components of a specific climate model: NASA-GISS is at a relatively low resolution of 4°x5°, while the NOAA-GFDL and NCAR-CCSM models are on the order of 1°.

Parameterizations

Albedo: As an important feedback to the atmosphere, the albedo (the proportion of incident radiation reflected off a surface) of the snow and ice plays a significant role in the climate system. All the sea ice component models parameterize the albedo to some extent. Figure II. C from Curry *et al*. (1995**)** illustrates the interrelations of the sea-ice system and how the albedo is a function of the snow or ice thickness, ice extent, open water, and the surface temperature, along with other factors, including the spectral band of the radiance. The various models treat the different contributions to the total albedo in similar ways, but vary on the details. For example, the NCAR-CCSM3 sea-ice component does not include dependence on the solar elevation angle (Briegleb *et al.*, 2002), while the NASA-GISS model does (Schmidt *et al*., 2006). Both of these models include the contribution of melt ponds (Ebert and Curry, 1993; Schramm *et al*., 1997) The NOAA-GFDL model follows Briegleb *et al*. (2002), but accounts for the differences in spectral contributions using fixed ratios (Delworth *et al.* 2006).

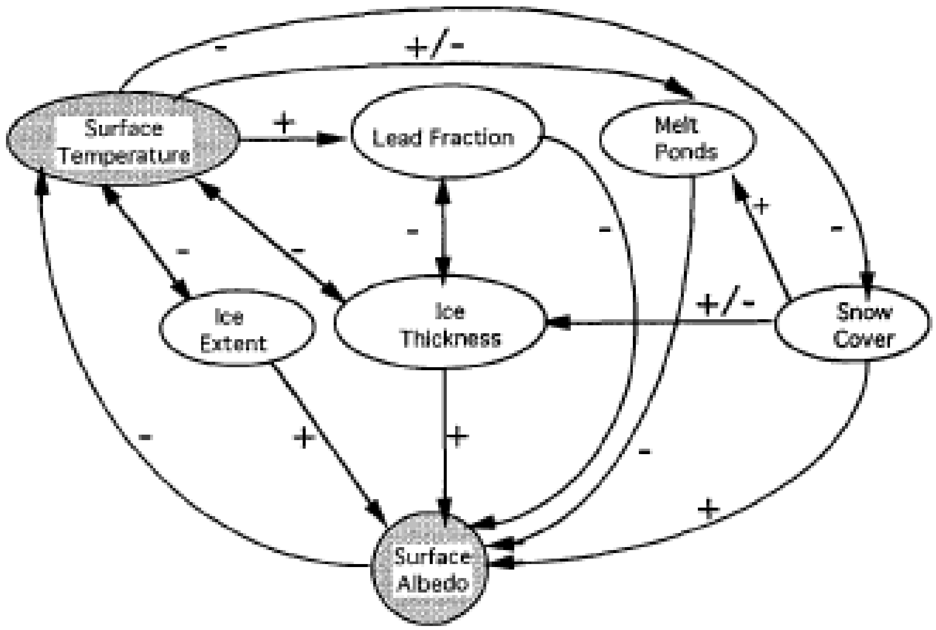


FIG. 6. Schematic diagram of the sea ice–albedo feedback mechanism. The direction of the arrow indicates the direction of the interaction. A "+" indicates a positive interaction (an increase in the first quantity leads to an increase in the second quantity) and a "−" indicates a negative interaction (an increase in the first quantity leads to a decrease in the second quantity). A "±" indicates either that the sign of the interaction is uncertain or that the sign changes over the annual cycle.

Figure II. C (from Curry *et al*. (1995)).

Other parameterizations: Additional parameters include reference values for defining ice salinities, strengths, roughness, and drag coefficients. Details of these parameters can be found in the references listed above which describe the basic sea-ice models of the various groups.

Component coupling and coupled model evaluation

We describe in the following some of the key aspects of the model development process at the three U.S. groups that contributed models to the Fourth Assessment of the IPCC, with particular focus on those aspects most relevant for simulation of the 20th century global mean temperature record on the one hand, and the model's climate sensitivity on the other hand. We begin with some general comments on the model development process.

The complexity of the climate system, and our inability to resolve all relevant processes in our models, result in a host of choices for development teams to make. Differing expertise, experience, and interests result in distinct development pathways for each climate model. While we eventually expect to see model convergence, forced by increasing insights into the working of the climate system, we are still far from that limit today in several important aspects of the models. Given this level of uncertainty, multiple modeling approaches are clearly needed. Models differ in their details primarily because development teams have differing ideas concerning the underlying physical mechanisms relevant for the less well-understood aspects of the system.

The NOAA Geophysical Fluid Dynamics Laboratory Model Development Path

The Geophysical Fluid Dynamics Laboratory of NOAA conducted a thorough restructuring of its atmospheric and climate models over more than five years prior to its delivery of a model to the CMIP-3/IPCC database in 2004. This was performed partly in response to need for modernizing the software engineering, and partly in response to new ideas in modeling the atmosphere, ocean, and sea ice. The differences between the resulting models and the previous generation of climate models at GFDL are sufficiently varied and substantial, that mapping out exactly why climate sensitivity and other aspects of the climate simulations differ between these two generations of models would be very difficult and has not been attempted. Unlike the earlier generation, the new models do not use flux adjustments.

The new atmospheric models developed at GFDL for global warming studies are referred to as AM2.0 and AM2.1 (GFDL Atmospheric Model Development Team, 2006). A key point of departure from previous models at GFDL was the adoption of a new numerical core for solving the fluid dynamical equations for the atmosphere. Much of the atmospheric development was based on running the model over observed seas surface temperature and sea ice boundary conditions over the period 1980-2000, with a focus on both the mean climate and the response of the atmosphere to ENSO variability in the tropical Pacific. Given the basic model configuration, several subgrid closures were varied to optimize aspects of the climate. Modest improvements in the midlatitude wind field were obtained by adjusting a part of the model referred to as "orographic gravity wave drag" which accounts for the effects of the force exerted on the atmosphere by unresolved topographic features ("hills"). Substantial improvements in tropical rainfall and its response to ENSO resulted from an optimization of parameters as well, especially the treatment of vertical transport of horizontal momentum by moist convection.

The ocean model chosen for this development was the latest version of the Modular Ocean Model developed over several decades at GFDL, notable new features in this version being a grid structure better suited to simulating the Arctic ocean and a framework, that has been nearly universally accepted by ocean modelers in recent years, for sub-gridscale mixing that avoids unphysical mixing between oceanic layers of differing densities (Gent and McWilliams, 1990). A new sea ice model includes the large-scale effective rheology that has proven itself in the past decade in several models, and multiple ice thickness/lead classes in each grid box. The land model chosen was relatively simple, with vertically resolved soil temperature but retaining the "bucket hydrology" from the earlier generation of models.

The resulting climate model was studied, restructured, and tuned for an extended period, with particular interest in optimizing the structure and frequency of the model's spontaneously generated EL Nino events, minimizing surface temperature biases, and maintaining an Atlantic overturning circulation of sufficient strength. During this development phase, climate sensitivity was monitored by integrating the model to equilibrium with doubled CO2 when coupled to a "flux-adjusted slab" ocean model A single model modification reduced the model's sensitivity from a value of 4.0–4.5 K to values between 2.5 and 3.0 K. The change responsible for this reduction was the inclusion of a new model of mixing in the planetary boundary near the Earth's surface. It was selected for inclusion in the model because it generated more realistic boundary layer depths and near surface relative humidities. The reduction in sensitivity resulted from modifications to the low level cloud field; the size of this reduction was not anticipated.

Aerosol distributions used by the model were computed off-line from the MOZART-II model as described in Horowitz, *et al*., (2003). No attempt was made to simulate the indirect aerosol effects (interactions between clouds and aerosols) as the confidence in the schemes tested was deemed insufficient for inclusion in the model. In the 20th century simulations, solar variations followed the prescription of Lean *et al*., (1995), while volcanic forcing was estimated from obervations. Stratospheric ozone was prescribed, with the Southern Hemisphere ozone hole prescribed, in particular, in the 20th century simulations. A new detailed land-use history provided a time-history of vegetation-types.

Final tuning of the global energy balance of the model, using two parameters in the cloud prediction scheme, was conducted by examining control simulations of the fully coupled model using fixed 1860 and 1990 forcings. The IPCC-relevant runs of the resulting model (CM2.0) were provided to the CMIP-3/IPCC archive under considerable time pressure.

The simulations of the 20th century with time-varying forcings provided to the database were the first simulations of this kind generated with this model. There was no retuning of the model, and no iteration of the aerosol or any other time-varying forcings, at this point.

Model development efforts proceeded in the interim, and a new version of the model emerged rather quickly in which the numerical core of the atmospheric model was replaced by a "finite-volume" code (Lin and Rood, 1996), substantially improving the wind fields near the surface. These improved winds in turn resulted in improved extratropical ocean circulation and temperatures. ENSO variability increased in this model, to unrealistically large values. But the efficiency of the ocean code was also improved substantially, and with a retuning of the clouds for global energy balance, the new model, CM2.1, was deemed to be a substantial enough improvement to warrant generating a new set of runs for the database. CM2.1 when run with a slab ocean model was found to have a somewhat increased sensitivity, (3.3K). However, the transient climate sensitivity, the global mean warming at the time of CO_2 doubling in a fully-coupled model with 1%/yr increasing CO_2, is actually slightly smaller than in CM2.0.

The solar, aerosol, volcanic, and greenhouse gas forcings are identical in the two models.

The Community Climate System Model Development Path

A new version of the Community Climate System Model, version 3 (CCSM3) has been developed, and was released to the climate community in June, 2004. CCSM3 is a coupled climate model with components representing the atmosphere, ocean, sea ice, and land surface connected by a flux coupler. CCSM3 is designed to produce realistic simulations over a wide range of spatial resolutions, enabling inexpensive simulations lasting several millennia or detailed studies of continental-scale dynamics, variability, and climate change. Twenty six papers documenting all aspects of the CCSM3, and runs performed with it, were published is a *Journal of Climate Special Issue*, Vol 19, No 11, June 2006. Three different resolutions of the model are supported. The highest resolution is the configuration used for climate-change simulations, with a T85 grid for the atmosphere and land, and a grid with approximately 1º resolution for the ocean and sea-ice, but finer meridional resolution near the equator. The second resolution is a T42 grid for the atmosphere and land, with the 1ºocean and sea-ice resolution. There is also a lower resolution version, designed for Paleoclimate studies, that has T31 resolution for the atmosphere and land, and a 3ºversion of the ocean and sea ice.

The new version of the CCSM3 incorporates several significant improvements in the physical parameterizations. The enhancements in the model physics are designed to reduce or eliminate several systematic biases in the mean climate produced by previous versions of CCSM. These include new treatments of cloud processes, aerosol radiative forcing, land-atmosphere fluxes, ocean mixed-layer processes, and sea-ice dynamics. There are significant improvements in the sea-ice thickness, polar radiation budgets, tropical sea-surface temperatures, and cloud radiative effects. CCSM3 produces stable climate simulations of millennial duration without ad hoc adjustments to the fluxes exchanged among the component models. Nonetheless, there are still systematic biases in the ocean-atmosphere fluxes in coastal regions west of continents, the spectrum of ENSO variability, the spatial distribution of precipitation in the tropical oceans, and continental precipitation and surface air temperatures. Work is underway to produce the next version of the CCSM, which will reduce these biases further, and to extend the CCSM to a more accurate and comprehensive model of the complete Earth's climate system.

The climate sensitivity of the CCSM3 has a weak dependence on the resolution used. The equilibrium temperature increase due to a doubling of carbon dioxide, using a slab ocean model, is 2.71C, 2.47C, and 2.32C, respectively, for the T85, T42, and T31 atmosphere resolutions. The transient climate response to doubling carbon dioxide in fully coupled integrations is much less dependent on resolution, being 1.50C, 1.48C, and 1.43C, respectively, for the T85, T42, and T31 atmosphere resolutions, see the Kiehl *et al*. paper in the *Journal of Climate Special Issue,* Vol 19, No 11, June 2006, 2584–2596.

For the IPCC Fourth Assessment Report, the following CCSM3 runs were submitted for evaluation, and to PCMDI for dissemination to the climate scientific community. Long, present day and 1870 control runs, an ensemble of eight 20th century runs, and smaller ensembles of future scenario runs for the A2, A1B, and B1 scenarios, and for the 20th century commitment run, where the carbon dioxide levels were kept at their 2000 values.
The control and 20th century runs are documented and analysed in several papers in the *Journal of Climate Special Issue*, and the future climate change projections using the CCSM3 are documented by Meehl *et al* (2006).

The GISS Development Path

The most recent version of the GISS atmospheric GCM, modelE, resulted from a substantial reworking of the previous version, model II'. While the model physics has become more sophisticated, execution by the user is simplified as a result of modern software engineeering and improved model documentation embedded within the code and accompanying web pages. The model can be downloaded from the GISS website by outside users, and is designed to run on myriad platforms ranging from laptops to a variety of multi-processor computers, partly as the result of the rapidly shifting computing environment at NASA. The most recent (post-AR4) version can be run on an arbitrarily large number of processors.

Historically, GISS has eschewed flux adjustment. Nonetheless, the net energy flux at the top of atmosphere and surface have been reduced to near zero, by adjusting the threshold relative humidity for water and ice cloud formation, two parameters that are otherwise weakly constrained by observations. Near-zero fluxes at these levels are necessary to minimize drift of either the ocean or the coupled climate.

To assess the sensitivity of the climate response to the treatment of the ocean, modelE has been coupled to a slab-ocean model with prescribed horizontal heat transport, along with two ocean GCMs. One GCM, the Russell ocean (Russell *et al.*,1995), has 13 vertical layers and horizontal resolution of 4º latitude by 5º longitude, and is mass conserving (rather than volume conserving like the GFDL MOM). Alternatively, ModelE is coupled to the Hybrid Coordinate Ocean Model (HYCOM), an isopycnal model developed originally at the University of Miami (Sun and Bleck, 2006). HYCOM has 2º latitude by 2º longitude resolution at the equator, with the latitudinal spacing decreasing poleward with the cosine of latitude. A separate rectilinear grid is used in the Arctic to avoid the polar singularity, and joins the spherical grid around 60 N.

Climate sensitivity to doubling of CO_2 depends upon the ocean model due to differences in sea-ice. For the slab-ocean model, the climate sensitivity is 2.7 C, and 2.9 C for the Russell ocean (Hansen et *al* 2005). As at GFDL and CCSM, no effort is made to match a particular sensitivity, nor is the sensitivity or forcing adjusted to match 20th century climate trends (Hansen et *al* 2007). Aerosol forcing is calculated from prescribed concentration, computed offline by a physical model of the aerosol life cycle. In contrast to the GFDL and NCAR models, modelE includes a representation of the aerosol indirect effect. Cloud droplet formation is related empirically to the availability of cloud condensation nuclei, which depends upon the prescribed aerosol concentration (Menon and Del Genio 2005).

Flexability is emphasized in model development (Schmidt *et al.*, 2006). ModelE is designed for a variety of applications, ranging from simulation of stratospheric dynamics and the middle atmosphere response to solar forcing, to projection of twenty-first century trends in surface climate. Horizontal resolution is typically 4º latitude by 5º longitude, although twice the resolution is more often used for studies of cloud processes. The model top has been raised from 10 mb (as in the previous model II') to 0.1 mb, so that the top has less influence upon the stratospheric circulation. Coding emphasizes “plug-and-play” structure, so that the model can be easily adapted for future needs, such as fully interactive carbon and nitrogen cycles.

Model development is devoted to improving the realism of individual model parameterizations, such as the planetary boundary layer, or sea ice dynamics. Because of the variety of applications, relatively little emphasis is placed upon optimizing the simulation of specific phenomena such as El Nino or the Atlantic thermohaline circulation; as noted above, successful reproduction of one phenomena usually results in a sub-optimal simulation of another. Nonetheless, some effort was made to reduce biases in previous versions of the model that emerged from the interaction of various features of the model, such as subtropical low clouds, tropical rainfall, and variability of the stratospheric winds. Some of the model adjustments were structural, as opposed to the adjustment of a particular parameter: for example, the introduction of a new planetary boundary layer parameterization that reduced the unrealistic formation of clouds in the lowest model level (Schmidt *et al.*, 2006).

Because of their uniform horizontal coverage, satellite retrievals are emphasized for model evaluation, like Earth Radiation Budget Experiment fluxes at TOA, Microwave Sounding Unit channels 2 (troposphere) and 4 (stratosphere) temperatures, and International Satellite Cloud Climatology Project (ISCCP) diagnostics. Comparison to ISCCP is through a special algorithm that samples the GCM output to mimic data collection by an orbiting satellite. For example, high clouds may include contributions from lower levels in both the model and the downward looking satellite instrument. This satellite perspective within the model allows a rigorous comparison to observations. In addition to satellite retrievals, some GCM fields like zonal wind are compared to in situ observations adjusted by the ERA-40 reanalyses. Surface air temperature is taken from the Climate Research Unit (Jones *et al.,* 1999).

Common problems

The CCSM and GFDL Development Teams met several times during this period to compare experiences and discuss common biases in the two models. A topic of considerable discussion and concern, for example, was the tendency for too strong an equatorial cold tongue in the Eastern Equatorial Pacific and associated problems with the pattern of precipitation (often referred to as the "double ITCZ problem"). It was noted in these meetings that the climate sensitivities of the two models had converged to some extent from an earlier generation in which the NCAR model was on the low end of the canonical sensitivity range of 1.5–4.5K, while the GFDL model had been near the high end. This convergence in the global mean was considered by the teams to be coincidental; it was not a consequence of any specific actions taken so as to engineer convergence, and did not reflect convergence either in the specifics of the cloud feedback processes that resulted in these sensitivity changes, nor in the regional temperature changes than make up these global mean values.

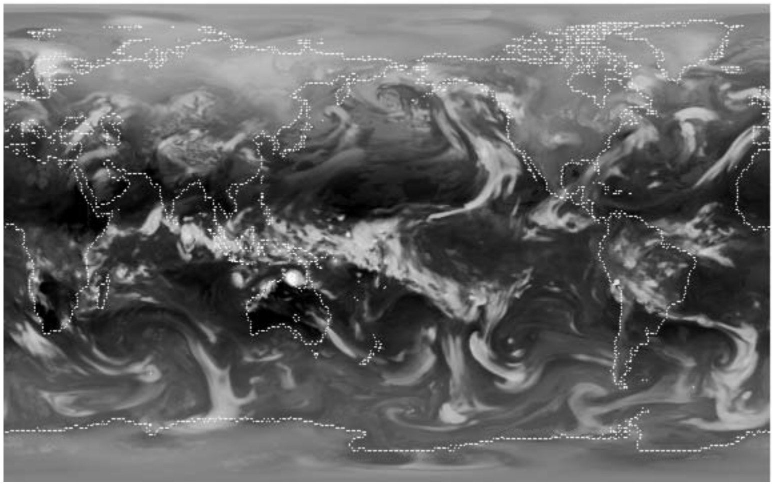
A procedure common to each of these three models, and to all other comprehensive climate models, is a tuning of the global mean energy balance. A climate model must be in balance at the top of the atmosphere and globally averaged, to within a few tenths of a W/m^2 in its control (pre-1860) climate if it is to avoid temperature drifts in 20th and 21st century simulations that would obscure the response to the imposed changes in greenhouse, aerosol, volcanic, and solar forcings. Especially because of the difficulty in modeling clouds, but even in the clear sky, untuned models do not currently possess this level of accuracy in their radiative fluxes. The imbalances are more typically range up to 5 W/m^2 or more. Parameters in the cloud scheme are then altered to create a balanced state, often taking care that the individual components of this balance, the absorbed solar flux and emitted infrared flux, are individually in agreement with observations, since these help insure the correct distribution of the heating between atmosphere and ocean. This is occasionally referred to as the "final tuning" of the model, to distinguish it from the various choices made with other motivations while one is configuring the model.

The need for this final tuning does not preclude the use of these models for global warming simulations, in which the radiative forcing is itself of the order of several W/m^2. Consider for example, the study of Ramaswamy *et al*., (2001) of the effects of modifying the treatment of the "water vapor continuum" in a climate model. This is an aspect of the radiative transfer algorithm in which there is significant uncertainty. While modifying the treatment of the continuum can change the top-of-atmosphere balance by more than 1 W/m^2, the effect on climate sensitivity is found to be insignificant. The change in radiative transfer in this instance alters the outgoing infrared flux by roughly 1% , and it affects the sensitivity (by altering the derivative of the flux with respect to temperature) by roughly the same percentage. But a change in sensitivity of this magnitude, say from 3K to 3.03K, is of little consequence given uncertainties in the cloud feedbacks. It is some aspects of the models that affect the strengths of temperature-dependent feedbacks that are of particular concern, not errors in mean fluxes *per se*.

<u>Reductive vs. holistic evaluation of models</u>*:*

In order to evaluate models, appreciation is needed of their structure. For example, the discussion of the climatic response to increasing greenhouse gases is intimately related to the question of how the infrared radiation escaping to space is controlled. When summarizing the results from climate models, one often speaks and thinks in terms of a simple energy balance model in which the global mean infrared energy escaping to space is a single number that has a simple dependence on global mean surface temperature. Water vapor or cloud feedbacks are often incorporated into such global mean energy balance models with simple empirical relationships that can easily be tailored to generate a desired result. In contrast, Figure II D shows a snapshot at an instant in time of the infrared radiation escaping to space in the kind of atmospheric general circulation model discussed in this report. The detailed distributions of clouds and water vapor simulated by the model, transported by the model's evolving wind fields, create complex patterns in space and time that, if the simulation is sufficiently realistic, resemble the images seen from satellites viewing the Earth at infrared wavelengths.

Figure II D



A snapshot in time of the infrared radiation escaping to space in a version of the atmospheric model AM2 (GAMDT, 2004) constructed at NOAA's Geophysical Fluid Dynamics Laboratory. The energy emitted is largest in the darkest areas and smallest in the brightest areas. (This version of the atmospheric model has higher resolution than that used for the simulations in the CMIP3 archive (50 km rtrher than 200km)but other than resolution it uses the same numerical algorithm.)

This class of model evolves the state of the atmosphere/land system forward in time, starting from some initial condition. It consists of rules that generate this state (temperature, winds, water vapor, clouds, rainfall rate, water storage in the land, land surface temperature) from the preceding state, in this case one half hour earlier. By this process it evolves the "weather" over the Earth. To change the way in which this model's infrared radiation reacts to increasing temperatures, one would need to modify these rules.

The goal of the climate modeling enterprise is to decrease the level of empiricism and to base models as much a possible on well-established physical principles. This goal is pursued primarily by decomposing the climate system into a number of relatively simple processes and interactions, and by focusing on the rules governing the evolution of these individual processes, rather than working with more holistic concepts such as the global mean infrared radiation escaping to space, the average summertime rainfall over Africa, or the average wintertime surface pressure over the Arctic. These are all outcomes of the model, determined by the set of reductive rules that govern the model's evolution.

Suppose one is interested in how ocean temperatures affect rainfall over Africa. One can develop an empirical, holistic, model, using observations and standard statistical techniques, in which one "fits" the model to these observations. Alternatively, one can try to use a general circulation model of the sort pictured above, which does not deal directly with a high level climate output such as African rainfall averaged over some period, but rather attempts to simulate the inner workings, or dynamics, of the climate system at a much finer level of granularity. To the extent that the simulation is successful and convincing, with analysis and manipulation of the model one can hope to uncover the detailed physical mechanisms underlying this causal connection. The resulting fit may or may not be as good as the fit obtained with the explicitly tuned statistical model, but a reductive model ideally provides a different level of confidence in its explanatory and predictive power. See, for example, Hoerling, et al 2006 for an analysis of African rainfall/ocean temperature relationships in a set of atmospheric GCMs.

Our confidence in the explanatory and predictive power of climate models grows based on their ability to simulate many aspects of the climate system *simultaneously* with the same set of physically based rules. When one evaluates a models ability to simulate the evolution of the global mean temperature evolution over the 20th century, it is important to try to make this evaluation in the context of the model's simultaneous capacity to simulate the seasonal cycle of the Asian monsoons, for example, and it ability to generate the poleward shift of the jet stream in the Southern hemisphere over the past 30 years that has impacted rainfall over southern Australia, and its ability to spontaneously generate El-Nino's of the correct frequency and spatial structure and to capture the effects of El Nino on rainfall and clouds. The quality of the simulation in all of these respects adds confidence in the reductive rules being used to generate the simultaneous simulation of all of these phenomena.

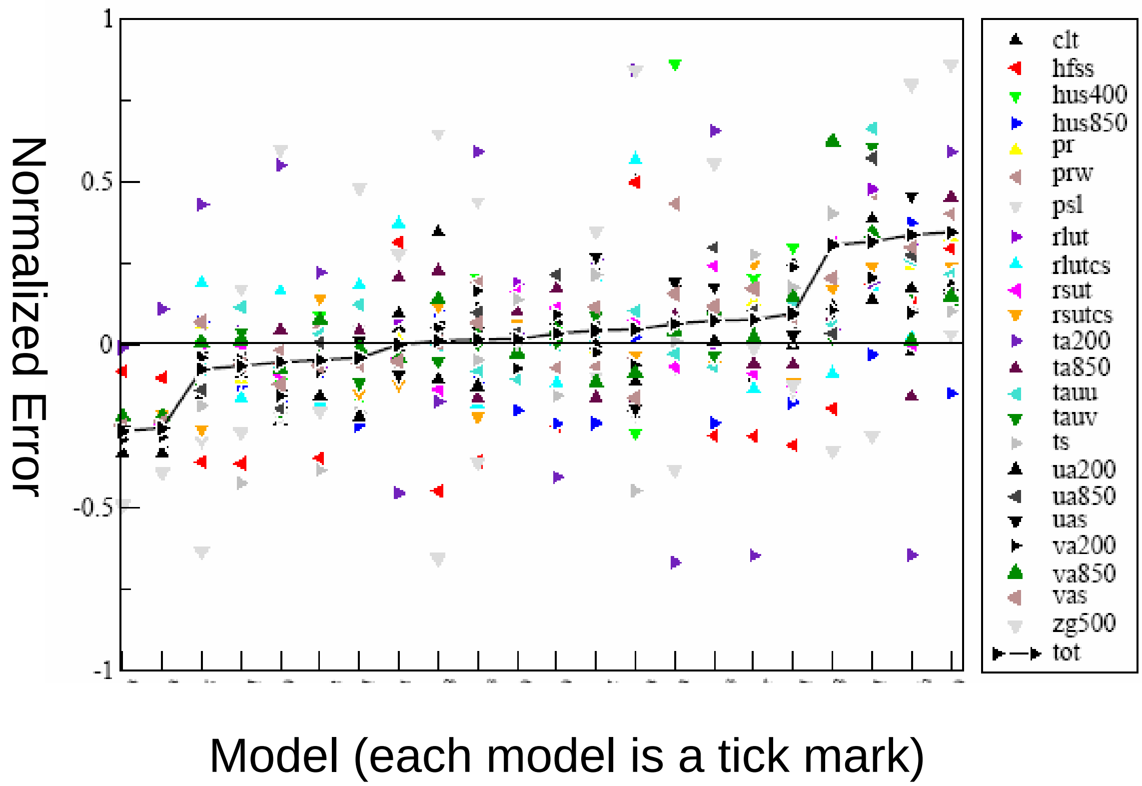
A difficulty that we will return to frequently in this report is that of relating the qualities of a climate simulation to a level of confidence in the model's ability to predict climate change.

The use of model metrics

Recently, objective evaluation of models has exploded with the wide availability of model simulation results in the CMIP3 model database (Meehl, et al, 2006). One important area of research is in the design of of metrics to test the ability of models to simulate well observed climate features (Reichler and Kim, 2007; Gleckler, et al,, 2007). It is unclear which aspects of observed climate must be simulated to ensure reliable future predictions. For example, it is not clear that the most realiable climate projections for temperature over North America are obtained from models that simulate the most realistic present-day temperatures for North America. The projected climate changes in North America may depend strongly on the changes in ocean temperature in the tropical Pacific Ocean, and the manner in which the jet stream over the Pacific responds to these changes in temperature. The quality of a models simulation of atmosphere-ocean cpoupling over the Pacific could potentially be a more relevant metric of quality in this instance. However, metrics can provide guidance about the overall strength and weaknesses of individual models as well as the general state of modeling.

The use of metrics can also inform the community as why it is impossible at this time to determine which is the "best" climate model. In Figure II E below, each of the colored triangles represents a different metric for which each model was evaluated, for example, "ts" represents surface temperature. The figure displays the relative error value for a variety of metrics, for each model, represented by a vertical column above each tick mark on the horizontal axis. Values less than zero represent a better than average simulation of a particular field measured by the metric, while values greater than zero show models with errors greater than the average. The black triangles connected by the dashed line represent the normalized sum from the errors of all 23 fields. The models were then ranked from left to right based on the value of this total error. As can be seen, the models with the lowest total error, tend to also score better than average in most individual metrics, however, the "best" models do not score the best for every metric. For an individual application, the model with the lowest total error may not be the best choice.

Figure II E – Model metrics for 23 different climate fields. Values less than 0 indicate an error less than the average CMIP3 model, while values greater than 0 show values greater than the average. The black triangles connected by the black line is a total score obtained by averaging all 23 fields.



Climate simulations discussed in this report

Three types of climate simulation are discussed in this report. They differ according to the climate forcing factors used as input to the models:

Control runs use constant forcing. (The name "control runs" originated in comparing them with the other simulation types discussed below.) The Sun's energy output and the atmospheric concentrations of carbon dioxide and other gases and aerosols do not change in control runs. As with the other types of climate simulation, day-night and seasonal variations occur, as well as internal "oscillations" such as ENSO (see below). Other than these variations, the control run of a well-behaved climate model is expected to reach a steady state eventually.

Values of control-run forcing factors are typically set to match present-day conditions, and model output is then compared with present-day observations. Actually, the present climate is affected not only by current forcing but also by the history of forcing over time—in particular past emissions of greenhouse gases—but present-day control run output and observations are expected to agree fairly closely if models are reasonably accurate. We compare model control runs with observations in Chapter V below.

Idealized climate simulations are aimed at understanding important processes in models and in the real world. They include experiments in which the amount of atmospheric carbon dioxide increases at precisely 1% per year (about twice the present rate of increase) or doubles instantaneously. The carbon dioxide doubling experiments are typically run until the simulated climate reaches a steady state in equilibrium with the enhanced greenhouse effect. Until the mid-1990's, idealized simulations were often employed to assess possible future climate changes including human-induced global warming. Recently, however, the more realistic time-evolving simulations defined immediately below have been used for making climate predictions. We discuss idealized simulations and their implications for climate sensitivity in Chapter IV below.

Time-dependent climate forcing simulations are the most realistic, especially for eras in which climate forcing is changing rapidly such as the 20th and 21st centuries. Input for the 20th century simulations includes observed time-varying values of solar energy output, atmospheric carbon dioxide, and other climate-relevant gases and aerosols including those produced in volcanic eruptions. Each modeling group uses its own best estimate of these factors. There are significant uncertainties in many of them, especially atmospheric aerosols, so that different models use somewhat different input for their 20th century simulations. We discuss these simulations in Chapter V after comparing control runs with observations.

Time-evolving climate forcing is also used as input for modeling future climate change. This subject is discussed in CCSP Synthesis and Assessment Product 3.2. Finally, we mention for the record simulations of the distant past (various time periods ranging from the early Earth up to the 19th century). These simulations are not discussed in this report, but some of them have been used to loosely “paleocalibrate” simulations of the more recent past and the future (Hoffert and Covey, 1992; Hansen *et al*., 2006; Hegerl *et al*., 2006).