Chapter 4
Oceanic General Circulation Models

Jin-Ho Yoon and Po-Lun Ma

Glossary

Bathymetry	The depth of ocean floor. Bathymetric map represents the terrain of the seafloor.
Biogeochemical cycle	A pathway that is described by the physical, biological, and chemical processes that control the evolution of elements found in the Earth system.
Boussinesq approximation	In Boussinesq approximation, a fluid parcel is assumed to maintain the same volume or density during its transport because of near incompressibility. This approximation was named after French physicist and mathematician, Joseph Valentin Boussinesq. By adopting this approximation, sound waves that propagate through a density change can be eliminated in numerical model.
Diapycnal mixing	Mixing of a fluid across different density surfaces. To be contrasted with isopycnal mixing which occurs along the same density surfaces.
El Niño-Southern Oscillation (ENSO)	A quasi-periodic change of the ocean and atmospheric conditions along the equatorial Pacific Ocean. The change in the sea surface temperature (SST) can be as large as $\pm 2°C$ during its extreme phases: anomalously warm over the tropical Pacific (El Niño) and cold (La Niña). Surface

This chapter was originally published as part of the Encyclopedia of Sustainability Science and Technology edited by Robert A. Meyers. DOI:10.1007/978-1-4419-0851-3

J.-H. Yoon (✉) • P.-L. Ma
Atmospheric Science and Global Change Division, Pacific Northwest National Laboratory,
902 Battelle Boulevard, P. O. Box 999, MSIN K9-24, Richland, WA 99352, USA
e-mail: Jin-Ho.Yoon@pnnl.gov; Po-Lun.Ma@pnnl.gov

P.J. Rasch (ed.), *Climate Change Modeling Methodology: Selected Entries from the Encyclopedia of Sustainability Science and Technology*,
DOI 10.1007/978-1-4614-5767-1_4,

	air pressures measured at both ends of the tropical Pacific basin vary closely with the change of SST.
Geostrophic approximation	The angular momentum is balanced by the Coriolis force and the horizontal pressure gradient force. It is generally true when the spatial and temporal scales are large, roughly over 100 km and a few days in the deep ocean.
Hydrostatic approximation	The equation describing vertical motion of the ocean column is simplified to assume that the vertical pressure at any level is due to the weight of the air and water above it. Variation of density is considered only in vertical direction when gravitational acceleration term (g) exists. This is valid when the vertical scale of a feature is small compared to the horizontal scale for both the atmosphere and ocean.
Isopycnal coordinate	Vertical coordinate that follows a constant density surface.
Meridional overturning circulation (MOC)	This has often been assumed to be the same as the thermohaline circulation. However, the MOC explicitly describes the ocean circulation system with the upwelling/downwelling and associates the northward/southward transport.
Ocean gyre	A large-scale rotating circulation in the ocean primarily forced by the atmospheric wind and the Coriolis force. These include the North Atlantic Gyre, South Atlantic Gyre, Indian Ocean Gyre, North Pacific Gyre, and South Pacific which tend to be more elongated in the east-west direction. There are also other types of Gyre forced by different mechanisms such as baroclinicity.
(Oceanic) Mesoscale eddy	A vigorous rotational circulation or vortex at spatial scales roughly 100 km and smaller, existing for weeks to months.
Rossby radius (of deformation)	The horizontal scale at which rotational effect becomes as important as buoyancy or gravity wave effects. Mathematically, this can be computed in terms of potential temperature, temperature, wind speed, or the depth of the boundary layer. This radius is important in determining the phase speed and wavelength of Rossby waves.
Salinity	Dissolved content of the salt in the ocean. Traditionally, salinity is represented in the unit of either g/Kg or PSU (Practical Salinity Unit).
Shallowness approximation	This approximation can be applied when the vertical-to-horizontal aspect ratio is very small.
Structured (regular)/ unstructured (irregular) grid	A structured (unstructured) grid has regular (irregular) connectivity with neighboring points. In structured (unstructured) grid, its connectivity can (cannot) be easily represented with a two- or three-dimensional array.
Thermocline	A distinct ocean layer where the temperature changes greatly with its depth compared to the layers above and

	below. It is often thought of as a boundary separating the well-mixed upper ocean and the deep ocean.
Thermohaline circulation	The global oceanic circulation driven by the density gradients, primarily determined by salinity and temperature.
(Atmospheric) Wind stress	The horizontal force exerted by the atmospheric wind on the ocean surface. This can also be interpreted as the vertical transfer of horizontal momentum from the atmosphere to the ocean surface. Wind stress is a function of the square of the wind speed.

Definition of the Subject

The purpose of this text is to provide an introduction to aspects of oceanic general circulation models (OGCMs), an important component of Climate System or Earth System Model (ESM). The emerging need for understanding the Earth's climate system and especially projecting its future evolution has encouraged scientists to explore the dynamical, physical, and biogeochemical processes in the ocean. Understanding the role of these processes in the climate system is an interesting and challenging scientific subject. For example, a research question exploring how much extra heat or CO_2 generated by anthropogenic activities can be stored in the deep ocean is not only of scientific interest but also important in projecting future climate of the Earth. Thus, OGCMs have been developed and applied to investigate the various oceanic processes and their role in the climate system.

Coupled climate models incorporating some representation of ocean circulations have been used since the pioneering work of Manabe and Bryan [38]. Because of computational limits at that time, an extremely simplified Earth geography – composed of only one continent and ocean basin (Fig. 4.1) – was used. Despite its simplicity, this was the first numerical model to include reasonably complex formulations for the atmosphere, land, ocean, and more importantly feedbacks among these components. Later it was demonstrated that deep ocean circulation in this simplified coupled model was not at an equilibrium even after a century-long integration [64]. Like this case, processes that control the adjustment of deep ocean circulation toward an equilibrium remain an important issue in contemporary climate models. It took another 6 years for Manabe and Bryan to produce a coupled climate model with more realistic geography of the Earth [39]. Now contemporary OGCMs can resolve complex regional flows in a much more realistic way. However, there are still many issues to be solved and large room for improvement [48].

The goal in this text is to outline the basics of Ocean General Circulation modeling, reviewing the mathematical representation of the ocean circulation

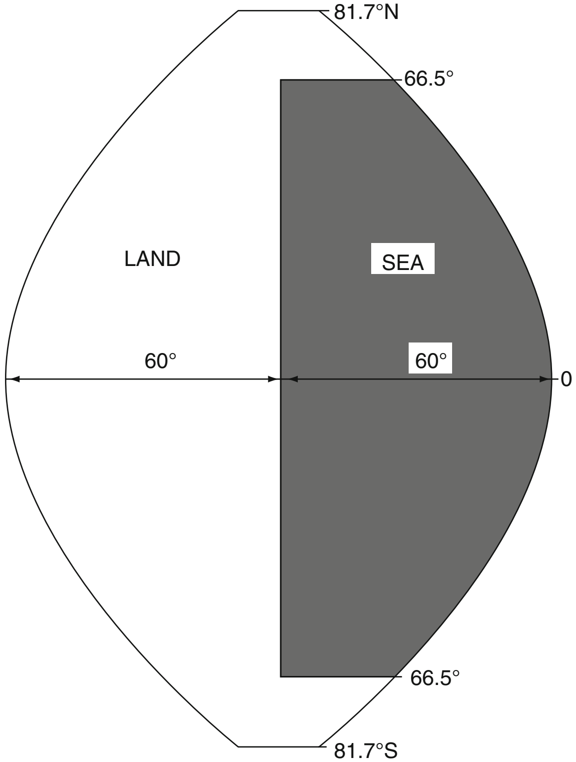

Fig. 4.1 Land-ocean configuration of the first coupled climate model of [38] (From Manabe and Bryan, J. Atmospheric Sciences 26 (1969): 786–789. © American Meteorological Society. Reprinted with permission)

(section Equations of Motion), and the discretization method with horizontal and vertical coordinate systems (section Horizontal and Vertical Grid System). A brief discussion is provided on how the sub-grid scale processes are represented in relatively coarse-resolution OGCMs (section Sub-grid Scale Parameterization). Then, using a couple of relatively simple examples, the coupling between the ocean and other climate components, especially the atmosphere by various physical processes will be explained (section Simple Conceptual Models). Also, how the ocean is coupled in terms of biogeochemical cycle will be illustrated in the section Biogeochemical Cycle Modeling. The entry concludes with a brief description of contemporary OGCMs and its future development (section Future Directions).

Introduction

An oceanic general circulation model (OGCM) is used to simulate the physical and dynamical features of the global ocean using equations representing the conservation of momentum, energy, and mass for water and salt. These equations are usually called the equations of motion, and they describe how the temperature, salinity, and currents evolve with time in the ocean. OGCMs also need appropriate boundary and initial conditions that describe how the ocean interacts with the atmosphere, land and sea ice and the state of the ocean when the model is started. An OGCM has a great similarity with its atmospheric counterpart in the entry Atmospheric General Circulation Modeling. However, there are important and non-negligible differences. First of all, the ocean is primarily forced at its surface boundaries by both thermal and mechanical forcings, while the atmosphere is forced throughout its entire volume mainly by thermal forcing. The global ocean is more strongly constrained by its complex terrain which results in oceanic gyres confined by lateral boundaries (continents and shelves), while the atmosphere is relatively less restricted. Also, very narrow boundary layers exist on almost every ocean surface and within its interior. Finally, observing the ocean is much more difficult than its atmospheric counterpart. The amount of ocean observation data is far less available than in the atmosphere. Thus, boundary and initial conditions are difficult to determine and the lack of long term and uniform observation imposes a challenge in validating the OGCMs' performance. Although a great deal of additional measurements has become available since satellite and buoys which have been deployed in the 1980s, it is estimated that data available to oceanic scientist in the 1990s is still an order of magnitude fewer than in the atmosphere [21].

The role played by the ocean in the global climate system is relatively well understood. For example, the ocean regulates the Earth surface heat budget through its large heat capacity and heat transport by the oceanic circulation from tropics to high latitudes. Large portions of this heat transport is carried out by both strong western boundary currents and oceanic mesoscale eddies (a dominant dynamical feature of the global ocean circulation at spatial scales from roughly 100 km and smaller with time scales of weeks to months). One example of a mesoscale eddy can be seen near the Atlantic coast of the USA in Fig. 4.2. The figure shows satellite-retrieved sea surface temperature from the NASA Aqua/MODIS mission over the western Atlantic where gray areas represent the land and color indicates temperature over the ocean surface. One can observe that generally warmer (colder) water exists in the south (north). Along the boundary that warm and cold waters are interwined, small-scale meanders and eddies are evident. These eddies are generated by strong shear instabilities in ocean currents, occurring particularly from currents along western boundaries and by horizontal density gradients. Oceanic kinetic energy is largely redistributed and transported by these eddies, and they are very important in the poleward transport of energy in the Earth's climate system.

In order for OGCMs to properly simulate these mesoscale eddies, they must operate at a high spatial resolution, resolving features that can be a few tens of kilometers and smaller. OGCMs divide the ocean up into small "cells" where temperature, salinity, and water velocity are calculated (discussed in more detail later). Ocean models that are designed to reproduce the behavior of eddies accurately (called Eddy-Permitting models) contain many cells and are extremely expensive in terms of computation. As of 2007, only a couple of the coupled

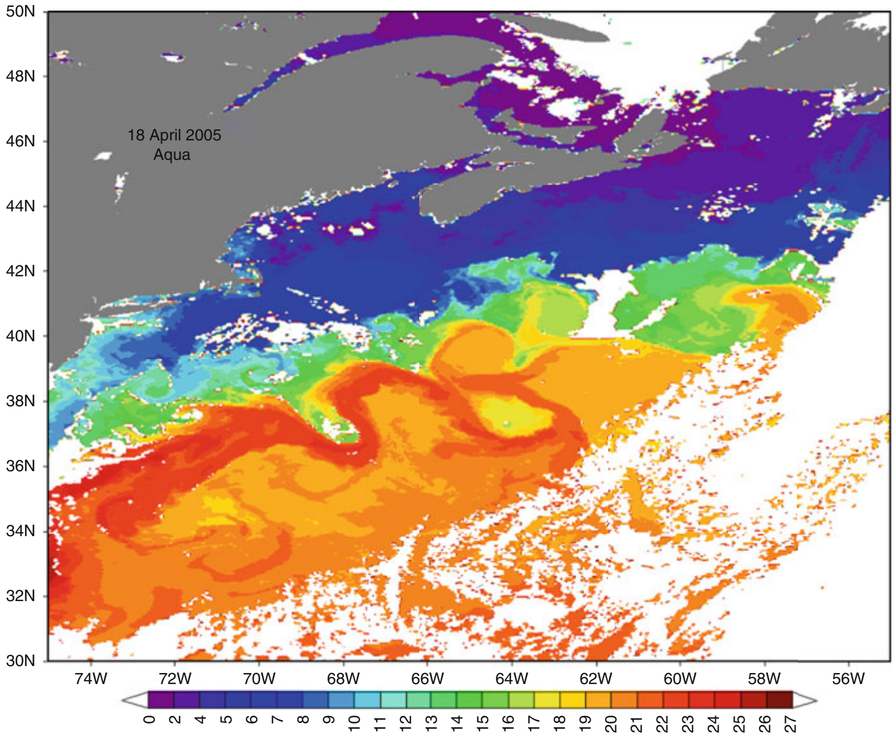

Fig. 4.2 Sea surface temperature at 18 April 2005 retrieved from Aqua MODIS [63]. Level 3 mapped data of MODIS SST was obtained from NASA JPL

climate models with eddy-permitting resolution (around 1/6° to 1/3°) successfully simulated these features explicitly [48].

OGCMs have historically been run at a much lower resolution to reduce the high computational burden. But since these lower resolution models are not able to explicitly represent these eddies, they must use other ways to represent the eddy effects. These effects are frequently treated through "parameterization" (more information on parameterization is provided below Sub-grid Scale Parameterization).

Recently ocean models now include representations of other important ocean components such as nutrient transports, ocean chemistry, and ecosystem evolution in order to calculate the evolution of biogeochemical cycles of the planet, including CO_2. The ocean regulates the amount of CO_2 stored through both biological and physical processes, which in general are similar to their terrestrial counterpart. The ocean absorbed about half of the CO_2 released in the last few centuries (e.g., [31]). And it becomes more important to understand how much of extra carbon can the ocean absorb in the near future.

This document has outlined the basis of the dynamical and physical properties of the ocean. More information on ocean processes can be found in various textbooks or other review articles (e.g., [24, 29]). In the next section, more emphasis will be given on the numerical formulation, and discretization of these processes will be provided, along with some information on the coupling between atmosphere and the ocean in a numerical form.

Equations of Motion

Ocean models solve the equations governing oceanic flow using computational fluid dynamics (CFD) methods with specific assumptions and approximations appropriate to the ocean. The equations for global ocean models are, like atmospheric models, based on Newtonian mechanics and irreversible thermodynamics applied to a particular fluid. Conservation of momentum, heat, and mass of constituents comprise the equations used for an OGCM. However, a wide range of mathematical formulation has been used to represent the numerous oceanic processes and phenomena occurring at various spatial and temporal scales, in the ocean. Only the primitive equations of motion that are the core of OGCM numerical formulations are described. A more complete derivation can be found in Griffies [24] and Vernois [62].

Hydrostatic Primitive Equations

A number of approximations are used in simplifying the fluid equations when used to represent ocean circulations, namely, the hydrostatic, shallowness, Boussinesq, and rigid lid approximations. The hydrostatic approximation assumes that the vertical momentum equation can be simplified by neglecting all terms except those resulting in the vertical pressure gradient being balanced by the gravitational acceleration (equivalently that the vertical pressure at any level is due to the weight of the air and water above it). The shallowness approximation assumes the ocean is thin compared to the Earth's radius. The Boussinesq approximation neglects variations in density except where density appears multiplied with gravity. The advantage of the Boussinesq approximation is to eliminate acoustic waves [7]. Although this approximation has been applied in many OGCMs, a couple of limitations seem to be found. One of the most noticeable limitations is that it cannot account for steric effect, i.e., thermal expansion and salinity-density compensation of sea water. Thus, sea-level change due to global warming cannot be properly predicted by this kind of model. Finally, the surface elevation of the global ocean is assumed static in the rigid lid approximation. With these approximations, the temporal and spatial variations of density are small compared to its mean value. Thus, it can be described in the following form:

$$\rho_{\text{tot}} = \bar{\rho} + \rho(x, y, z, t), \tag{4.1}$$

where $\bar{\rho}$ represents the mean density of the ocean and $\bar{\rho} \gg |\rho|$ is true and ρ can be replaced by $\bar{\rho}$ except the buoyancy term, ρg and in the thermodynamic density conservation equation.

With these approximations, the hydrostatic primitive equation in Cartesian coordinates is described in the following format:

$$\frac{du}{dt} = fv - \frac{1}{\bar{\rho}}\frac{\partial p}{\partial x} \tag{4.2}$$

$$\frac{dv}{dt} = -fu - \frac{1}{\bar{\rho}}\frac{\partial p}{\partial y} \tag{4.3}$$

$$\frac{\partial p}{\partial z} = -\rho g \tag{4.4}$$

$$\frac{dT}{dt} = 0 \tag{4.5}$$

$$\frac{dS}{dt} = 0 \tag{4.6}$$

and

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \tag{4.7}$$

where (x, y, z) represents Cartesian coordinates and (u, v, w) as current in the direction of x, y, and z, respectively. $\bar{\rho}$ is the mean density of the ocean as in Eq. 4.1, f is the Coriolis frequency, and T and S are temperature and salinity of the ocean. And the total derivative and local derivatives are related in the following way:

$$\frac{d}{dt} = \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y} + w\frac{\partial}{\partial z} \tag{4.8}$$

To close the set of equations (Eqs. 4.2–4.6), an equation of state, another equation of state to link density to ocean temperature, salinity, and pressure is added:

$$\rho = \rho(T, S, p) \tag{4.9}$$

It is also noted here that these sets of equation are written for spherical geometry in a global ocean model. Details can be found in Haidvogel and Bryan [29] or Bryan [5].

The above sets of equation are derived from the fluid dynamics with the help of approximations or assumptions. Some of the approximations, like Boussinesq approximation, have been still used in OGCM. Others, like the rigid lid approximation are not used any longer. There is a trend toward more explicitly representing the ocean status and processes. More realistic simulation of the ocean with better computational efficiency is now being achieved in contemporary OGCMs (see [27]).

Boundary Conditions

To solve the equations of motions, boundary conditions are needed. Upper boundary conditions are governed by interactions with the atmosphere or sea ice. The surface boundary condition on momentum below an ice-free atmosphere is governed by the wind stress:

$$\tau_x = \bar{\rho} v_x \frac{\partial u}{\partial z}\bigg|_{z=0}, \tag{4.10}$$

$$\tau_y = \bar{\rho} v_y \frac{\partial v}{\partial z}\bigg|_{z=0}, \tag{4.11}$$

where v_x and v_y are the turbulent eddy viscosity coefficients. The wind stress is sometimes provided by observational estimates (from a measurement climatology) and other times is provided by an atmospheric model. More discussion will be given later in Atmospheric forcing section.

The boundary condition for temperature is also treated in a variety of ways. When ocean models are coupled with a climate model, they calculate turbulent fluxes of heat and radiative fluxes based upon the state of the atmosphere and ocean. When ocean models are driven with observations, other choices are possible. Sometimes sea surface temperature (SST) is specified. In other situations surface fluxes are calculated in a similar fashion to the way they are done when coupled to an atmospheric model.

Salinity boundary condition depends upon evaporation, and the freshwater flux from precipitation, river runoff, and sea ice melt. In simple models where feedback between salinity and its forcing is not considered, one can use the prescribed surface salinity flux. However, it is found that there is important feedback between salinity and its forcing even in the case when ocean models are not coupled to atmosphere and sea ice models (e.g., [6]).

The Surface Mixed Layer

In the ocean, the heating term can be neglected generally except in the regions where solar radiation can play a role and heat exchange with the overlaying atmosphere is active, which is called the surface mixed layer. Both vigorous mixing and surface heating play important roles. Therefore, the surface heating term has to be included in the temperature equation:

$$\frac{\partial T}{\partial t} = -\frac{\partial}{\partial z}(Tw) + \frac{1}{\bar{\rho} c_p} Q(z, t), \tag{4.12}$$

where the first term in the right hand side represents the heat exchange due to vertical mixing, the second term is the solar heating absorbed at depth in the water column, and c_p is the specific heat of water. Although the equation looks simple, complication arises how to close. Two general approaches are taken: differential and bulk models [33]. The Mellor Yamada level-2 model is an example of the former [43], in which the stress terms now include both eddy coefficients for both within and below the mixed layer. On the other hand, the Niler model [41] uses a bulk model in which the mixed layer is treated as a well-defined homogeneous layer.

Horizontal and Vertical Grid System

To solve the governing equations of the oceanic general circulation, these equations have to be discretized with appropriate horizontal and vertical grid systems. Various methodologies have been developed and applied in the different OGCMs. In this section, both horizontal and vertical grid systems will be reviewed.

Horizontal Grid System

OGCMs have widely adopted the finite volume and finite elements methods. These methods will be discussed here for brevity, but a few distinguishing characteristics be mentioned. Discretized governing equations are applied in flux form so scalars such as temperature, density, salinity, etc., are updated through fluxes at the boundaries of each grid, and therefore, these quantities are conserved. Structured meshes with regular distributions of neighboring grids throughout the entire domain and unstructured meshes where the number of neighboring grids varies throughout the domain are both utilized with finite volumes and finite element discretizations. Successful model implementations include orthogonal meshes (e.g., [58]), cubed sphere meshes (e.g., [1]), icosahedral meshes (e.g., [47, 51, 52]), and unstructured triangular meshes (e.g., [10, 12]). Figure 4.3 shows an example of a structured and unstructured grid.

Structured grids are computationally efficient, accurate, and convenient. However, the major shortcoming with structured grids is that the grid-spacing is not uniform. Lack of flexibility to increase resolution in coastal regions or regions with complex topography make the computation expensive if the resolution of the entire domain has to be

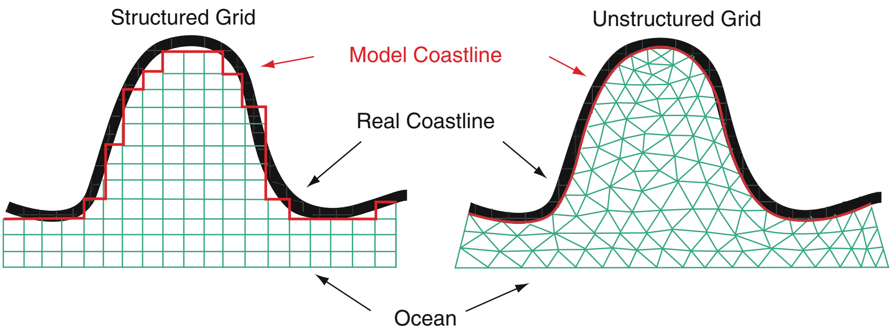

Fig. 4.3 An example of fitting a structured grid (*left*) and an unstructured grid (*right*) to a simple coastal embayment. The true coastline is shown in *black*, the model coastline in *red*. Note how the unstructured triangular grid can be adjusted so that the model coastline follows the true coastline, while the structured grid coastline is jagged – which can result in unrealistic flow disturbance close to the coast (Figure taken from Fig. 4.1 in Chen et al. [13] in *Oceanography*, reproduced with permission)

increased at the same time. An irregular gridding system that can increase resolution near the coast while keeping lower resolution over the open ocean coarse provides better computational efficiency more easily. In addition, because the unstructured grid can better represent the coastal area, it is advantageous over structured grid in simulating coastal currents and eddies.

To avoid a singular point at the North Pole, where grid lines converge, the ocean models either rotate the global gridding system, or create a tripolar grid with poles located over Canada, Russia, and Antarctica, such that the singular points of the grids are positioned over land instead of ocean.

An unresolved issue is the resolution–dependent physics in ocean models. Parameterizations of sub-grid scale processes inevitably produce some "free" (undefined) parameters (hence the name) and need to be in accordance with observations. These parameters are at varying horizontal grid sizes to capture the effects of processes from eddies with different sizes. The diffusion and viscosity terms will need to be adjusted accordingly.

Vertical Coordinate

Ocean models have developed several different vertical coordinate systems: (1) geopotential coordinate (or *z*-coordinate), (2) terrain-following coordinate, and (3) isopycnal coordinate, as shown in Fig. 4.4. Also, there is hybrid coordinate which combines various coordinate systems. Each one has its own advantages and disadvantages.

The most popular coordinate system is the geopotential coordinate [42]. For example, the first coupled climate model also utilized this vertical coordinate

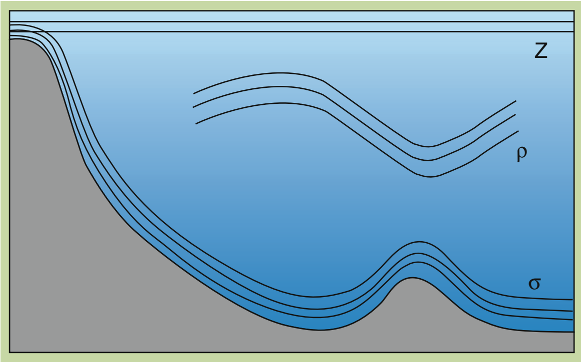

Fig. 4.4 A schematic diagram of an ocean basin illustrating the three regimes of the ocean germane to the considerations of an appropriate vertical coordinate. The surface mixed layer is naturally represented using fixed-depth z (or pressure p) coordinates, the interior is naturally represented using isopycnic ρ (density tracking) coordinates, and the bottom boundary is naturally represented using terrain-following σ coordinates (after [25]) (Figure taken from Box 1 in Chassignet et al. [11] in *Oceanography*, reproduced with permission)

system with the hydrostatic and Boussinesq approximations [26, 38]. The geopotential coordinate adopts a straightforward concept that the ocean water body is divided into regular boxes which are invariant with time. The vertical coordinate is based on the geopotential or depth of the ocean. Governing equations using lower-order finite difference methods are frequently used. In ocean models that adopt the geopotential coordinate (e.g., [35, 56]), the vertical resolution decreases as a function of depth (i.e., higher vertical resolution in the upper ocean and lower vertical resolution in the deep ocean) in order to better resolve processes around the thermocline. However, some studies have shown that the geopotential coordinate models have problems simulating the topographical effects on the ocean circulation because of their “stepwise” representation of the topography. They also have mesoscale eddy-induced vertical diffusion much greater than the observations.

The discontinued representation of the bathymetry and sidewall geometry in the geopotential coordinate system has been related to spurious topographic effects on the ocean circulation. Terrain-following (or sigma) coordinate models have been developed to handle complex lateral and bottom boundaries and avoid some of these problems. The sigma coordinate system is based on a fractional depth scaled from 0 to 1, with 0 being the surface and 1 being the bottom of the ocean. However, because the pressure gradient term is projected to be along the sigma surfaces, the horizontal pressure gradient produces considerable errors as a function of topographical slope and the stratification near the sea floor [2, 16, 32, 57]. Due to this problem, studies have shown that models that adopt the terrain-following coordinate are not suitable for global-scale simulations, unless the horizontal resolution is very high (the order of 10 km).

The isopycnal coordinate system uses isopycnal surfaces to be the vertical levels. Its advantage is due to the fact that the ocean currents and the transport of tracers generally proceed along isopycnal surfaces, so that the model that adopts this coordinate can ease the treatment of advection. The diapycnal mixing due to eddies can be added to the model as a parameterization. Furthermore, since isopycnal surfaces evolve over time, the isopycnal coordinate is, by nature, an adaptive system. It can better resolve the frontal region and thermocline where sharp vertical gradients of density exist [4]. However, for areas with a low-density gradient such as the well-mixed shelf area or deepwater formation areas, the model has a lower resolution than it would with other vertical coordinate choices. Because it avoids some of the shortcomings of the two other coordinate systems, isopycnal models, or "hybrid" models using a combination of these coordinate systems have a great potential in becoming more and more viable for global climate simulations in the future. Examples of models that use hybrid coordinates include the Global Navy Coastal Ocean Model that uses sigma-z coordinate, and the experimental Hybrid Coordinate Ocean Model that uses the z- (or sigma) coordinate in the upper ocean and the isopycnal coordinate in the stratified ocean.

Sub-grid Scale Parameterization

With spatial discretization, one can solve the governing equation discussed in the section Equations of Motion. But, representing continuous fluid with limited number of grid points imposes some new challenges. For example, the global OGCM with horizontal grid scale of 200 km cannot explicitly simulate oceanic mesoscale eddies (Fig. 4.2). To overcome this, spatial resolution of the OGCM has been increasing greatly. Indeed, increasing the ocean resolution in the coupled climate model has resulted in many improvements in the simulation of ocean circulation features. However, the impact of higher ocean resolution on the atmospheric simulation in coupled models has been found to be relatively small and localized unless the resolution of the atmosphere is also changed [30]. Further it has been shown that increasing spatial resolution plays a secondary role in simulating ENSO properly [28]. To achieve realistic representation of the ocean circulation and physical status requires not only increasing resolution but also physical parameterization. In this section, these parameterizations are discussed.

There are many processes occurring in the ocean that are not expressed explicitly in the equations of motion, or that take place at time and space scales that are too small to be treated explicitly in a global ocean model. No OGCM can simulate process in the ocean explicitly due to computational constraints or lack of knowledge about these processes. However, these can be very important in influencing oceanic circulations. Sub-grid processes such as vertical mixing and eddies with a scale smaller than the grid that cannot be resolved are "parameterized" based on theoretical considerations, observational data, and/or results from finer-resolution models. In "non-eddy" resolving (or permitting) OGCMs, the effects of mesoscale eddies are represented by parameterizations called neutral diffusion [49, 59] and eddy-induced advection [19, 20]. Maltrud and McClean [37] have suggested that a horizontal grid-spacing of 50 km in mid-latitudes and 10 km in high latitudes is sufficient to resolve eddies. Higher resolution models are believed to be able to resolve these processes.

Turbulent mixing at the upper ocean surface must also be represented (e.g., [18, 34, 45]). The stirring and mixing at the ocean surface can penetrate to the interior of the ocean and trigger further vertical mixing. These waves can be reflected, scattered, or transformed by the seafloor at the bottom and lateral boundary of the ocean. The diapycnal mixing takes place as a result of baroclinic instability, and is brought to balance by geostrophic adjustment via radiation of gravity waves. These processes interact with the topography and the mean flow. Depending on the stability of the ocean, the diapycnal mixing can affect the general circulation or dissipate without having any effect.

Many sub-grid scale processes are recognized to be important but remain poorly understood. The lack of the four-dimensional observational data of ocean flows as well as the state variables of the ocean in the oceanic interior make validation of the parameterizations against observations difficult, so validations are usually performed by evaluating their bulk effect against a high-resolution model. Recently satellite measurements of sea surface height, sea surface temperature, winds, etc. that cover a large scale of domain with a reasonable resolution have proved very helpful for understanding the features of mesoscale eddies but more in situ measurements are necessary to validate the model and to calibrate satellite observations. With the coordinated efforts from both modeling and measurement communities, parameterizations in ocean models can be scientifically evaluated and improved accordingly.

Simple Conceptual Models

Continuing atmospheric-ocean coupling process, a brief review is provided in this section on two simple ocean models which describe two important oceanic phenomena that have a large impact on the entire climate system: (1) the El Niño Southern Oscillation and (2) the thermohaline circulation. The former feature produces the largest signal in interannual climate variability in the Earth system, and the latter is a feature that is very important in longer-term climate variability.

These features are most easily described using simple models based on the full sets of mathematical equations but with their complexity reduced to make it easier to describe the basic features of particular phenomena. For example, idealized geometries with simplified boundary conditions may be applied or several approximations can be made for some regions, which are not uniformly applicable to the global ocean. However, these simple models provide very useful insight on how a specific physical process is maintained and respond to individual forcing. For example, the ocean thermocline and wind-stress effects on El Niño Southern

Oscillation can be easily understood and modeled in a Cane–Zebiak model. In this subsection, a couple of simple models will be reviewed. One is this Cane–Zebiak model and the other describes the oceanic thermohaline circulation [15].

Cane–Zebiak Model

El Niño Southern Oscillation (ENSO) is a climate pattern found in nature. The pattern is centered in the tropical pacific, but its signal can be detected globally. It is frequently described through an index measuring surface pressure differences between Tahiti and Darwin, or ocean surface temperature over the eastern Pacific but its signatures can be found in precipitation changes over tropics as well as mid-latitudes and in many other components of the Earth climate system. The feature has a very large interannual variability. On average there is an event every 3–6 years but it occurs irregularly. Various degrees of simple conceptual models have been proposed to explain and model ENSO. All these efforts can be summarized by the Bjerknes–Wyrtki model [3, 65], in which it was suggested that ENSO is an internal mode of oscillation of the coupled atmosphere-ocean system, driven by a continuous imbalance between the tightly coupled surface winds and temperatures on the one hand, and the more sluggish subsurface heat reservoir on the other. In the early 1980s, several highly idealized coupled models were developed based on the Bjerknes–Wyrtki model. The first one was that of Cane and Zebiak [9] and Zebiak and Cane [67]. Both the atmosphere and the ocean are highly simplified. These components are outlined in the next subsection.

The Atmosphere Model

Tropical atmosphere exhibits a couple of unique characteristics. One is that the wind is driven by latent heat release due to convection, and the other is a reversed polarity in the vertical structure, for example, low-level convergence with upper-level divergence in the divergent circulation. Therefore, linear dynamics with a single degree of freedom in the vertical coordinate has proven surprisingly accurate in reproducing the atmospheric circulation, classified as a Gill-type model [22, 44]. In ZC model, steady-state, linear shallow-water equation on an equatorial beta plane is used, just like a Gill-type model as follows:

$$\frac{\partial u}{\partial t} - \beta y v = -\frac{\partial \Phi}{\partial x} \tag{4.13}$$

$$\frac{\partial v}{\partial t} - \beta y u = -\frac{\partial \Phi}{\partial y} \tag{4.14}$$

$$\frac{\partial \phi}{\partial t} + c_a^2 \nabla \cdot \vec{V} = \frac{J}{C_p} \tag{4.15}$$

where $\vec{V} = (u, v)$ are the surface wind vector, ϕ is the geopotential, $\beta = df/dy$, with f being the Coriolis parameter, and c_a is a constant related to momentum and mass in the form of Newtonian cooling. Earlier, Matsuno [8] solved this equation for free tropics diabatic heating, and later Gill [22] found the steady solution for both the forced and damped versions. Lindzen and Nigam [36] introduced an alternative view on how surface wind was forced. They proposed the surface temperature gradient causes overlaying atmospheric pressure gradient and in turn surface wind anomalies.

The atmospheric heating anomaly Q is contributed by local evaporation (Q_S) and low-level moisture convergence (Q_L). It has been shown that the atmospheric moisture convergence is important for tropical heat budget and to maintain divergent circulation (e.g., [14]).

The Ocean Model

The ZC model represents the ocean as a single basin, the tropical Pacific, consisting of two vertical layers with a simplification of the ocean dynamics called a linear reduced gravity model. Vertically, the change of the thermocline needs to be considered, which is assumed motionless and linear. The set of equations is as follows:

$$\frac{\partial u}{\partial t} - \beta y v = -\frac{\Delta \rho}{\rho} \frac{\partial h}{\partial x} + \frac{\tau_{S_x}}{\rho \bar{h}} - ru \,, \tag{4.16}$$

$$\beta y u = -\frac{\Delta \rho}{\rho} \frac{\partial h}{\partial y} + \frac{\tau_{S_y}}{\rho \bar{h}} - rv \,, \tag{4.17}$$

$$\frac{\partial h}{\partial t} + h \nabla \cdot \vec{V} = -rh \,, \tag{4.18}$$

$$\frac{\partial T}{\partial t} = -\vec{V} \cdot \nabla T - M w_s \frac{\partial T}{\partial z} - \alpha_S T \,, \tag{4.19}$$

where (u,v) is the horizontal velocity in the layer, h is the derivation of the layer thickness from its mean value H, r is a Rayleigh friction parameter, τ_S is the surface wind stress, and $\Delta\rho$ a characteristic density difference between the layers. The upwelling is represented as follows:

$$w_S = H_1 \nabla \cdot \vec{V}_1 . \tag{4.20}$$

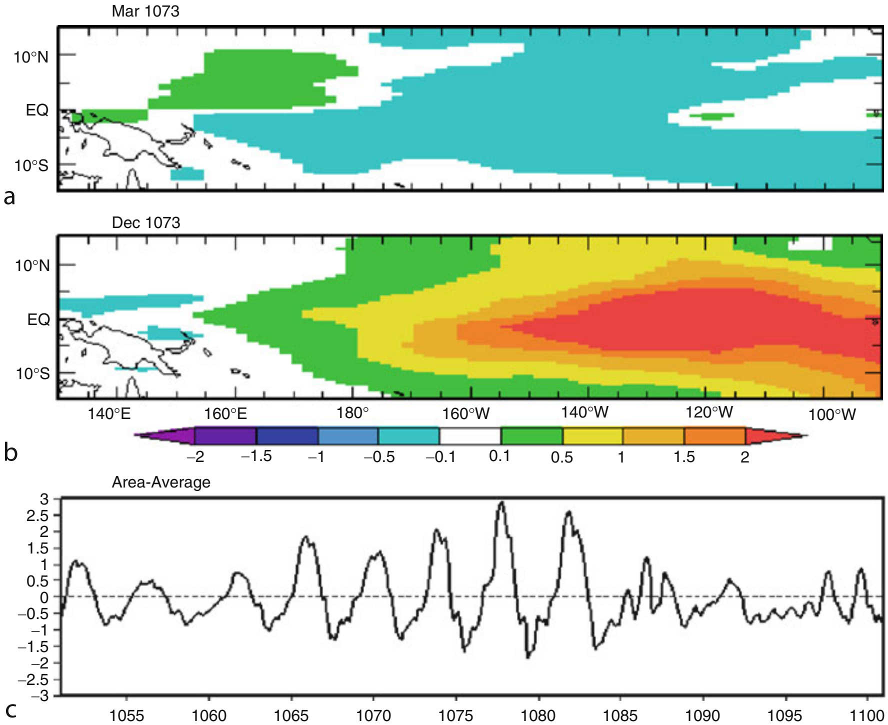

Fig. 4.5 An example of El Nino simulated by ZC model. (**a**) SST anomalies at March 1073, (**b**) same as (**a**) except December 1073, and (**c**) area averaged over the eastern equatorial Pacific (90° W to 150° W, 5° S to 5° N). The entire simulation was for the period of 1001–1200

The function M accounts for the fact that surface temperature is affected by upwelling. The anomalous temperature gradient, $\partial T/\partial z$, is defined by

$$\frac{\partial T}{\partial x} = \frac{(T - T_{\mathrm{e}})}{H_1}, \tag{4.21}$$

$$T_{\mathrm{e}} = (1 - \gamma)T + \gamma T_{\mathrm{d}}, \tag{4.22}$$

where H_1 is the surface layer depth, T_{e} measures temperature entrained into the surface layer, with γ as a mixing parameter (= 0.5), and T_{d} relates the subsurface temperature anomaly to the mean and anomalous thermocline depths.

In summary, the ocean model is essentially a reduced gravity model with one and half layer of ocean which includes the surface mixed layer and the layer below as one layer. Thus, the surface temperature is affected by both the atmospheric wind stress at the surface and the temperature entrained into the surface layer from the deep layer.

With reasonable choice of parameters, this coupled model has produced anomalous SST pattern and wind that resembles the observed ENSO anomalies (Fig. 4.5).

El Niño occurs every 3–4 years, followed by the La Niña events. A well-known deficiency of this model is that ENSO occurs too regularly with the fixed period around 2 years.

Although it is an idealized and theoretical model, this model included both the atmosphere and the ocean dynamics and was shown to have a reasonable evolution and life cycle of the ENSO compared to what was observed. Essentially, a Gill-type atmosphere with heating mainly tied to SST anomaly and convergence feedback term is coupled with a reduced gravity ocean model. In the mid-1980s, slightly different models were examined: one by Philander et al. [46], in which the SST anomaly was assumed proportional to both thermocline displacement and its anomaly; the other by Gill [68] in which the tendency of SST was assumed to be a function of advection of mean temperature by anomalous zonal current and atmospheric heating was determined by the SST anomaly. The former resulted a Kelvin wave response, while the latter a westward growing coupled mode.

These model variants, although far simpler than an OGCM, were constructed with careful thought on the processes influencing SST changes that are essential in producing the coupled ENSO phenomena. Simulating realistic ENSO remains a challenge for developers of climate models. Because ENSO is a dominant feature in interannual variability, and El Niño (La Niña) induces anomalous wet (dry) conditions all over the worlds, it is critical to simulate right. Understanding ENSO, and how it might change in an era of the global warming is an important research topic. For example, a recent study using the IPCC's AR4 model projections showed the El Niño with its center over the central Pacific may dominate in the future [66].

A Box Model for Thermohaline Circulation

The ZC model is primarily forced by surface wind stress and buoyancy plays a minor role. However, in climate application, ocean circulation can be driven by salinity and density gradients, which contributes much longer-term variability than ENSO which is predominately in interannual timescales. Although state-of-the-art OGCMs can now simulate the thermohaline circulation relatively well, it has been one of the long-standing challenges for modelers. Therefore, different types of simple conceptual models have been developed and used. Difference between the ZC model and the thermohaline models are not limited to the physical forcing mechanisms, but also through the introduction of nonlinearity in the thermohaline circulation and different mathematical treatment associated with the vertical stratification of buoyancy.

Therefore, an alternative to the formal mathematical derivation of simplified models from the full equations of motion is used to pose a conceptual model or simple physical analog for the ocean system. In this entry, an idealized model by Marotzke [40] will be briefly described. Two boxes which represent tropical and polar regions of the ocean are considered. Several assumptions are made: (1) ocean temperature is very close to atmospheric temperature in each box, and (2) salinity is forced by

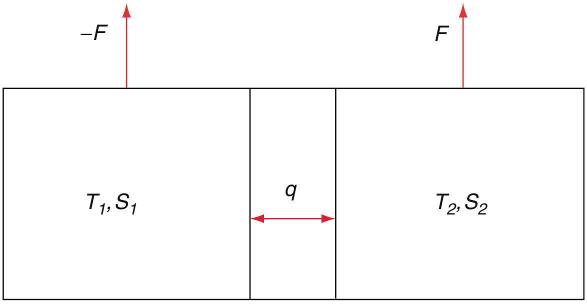

Fig. 4.6 A schematic diagram of the two-box model of the thermohaline circulation (modified from [29])

a flux of moisture through the atmosphere from the low-latitude to the high-latitude box. Salinity forcing is used to make the low-latitude box saltier and denser, and the high-latitude box fresher and less dense (Fig. 4.6). The governing equation of salt can be described as follows following Haidvogel and Bryan [29]:

$$V\frac{dS_1}{dt} = -FS_0 + |q|(S_2 - S_1), \tag{4.23}$$

$$V\frac{dS_2}{dt} = -FS_0 + |q|(S_1 - S_2), \tag{4.24}$$

where V is the volume of the boxes which is assumed to be equal, S_0 is a constant reference salinity, and q is the rate of volume exchanged in the pipes which connect the low-latitude and polar boxes. Pressure differences between the two boxes drive the system, which can be modeled using resistivity k^{-1}:

$$q = -\frac{k}{\rho_0}(\rho_2 - \rho_1)\,. \tag{4.25}$$

These are completed using a linear equation of state:

$$\rho = \rho_0\,(1 - \alpha T + \beta S)\,, \tag{4.26}$$

where α is the thermal expansion coefficient and β is the haline contraction coefficient.

As mentioned above, the state-of-the-art climate models now can simulate decent thermohaline circulation. However, it is still difficult to observe this circulation and bias exists in its location and intensity simulated by OGCM.

Biogeochemical Cycle Modeling

The oceans play a role in not only through physical quantities, such as wind stress and surface energy exchange, but also through chemical species, especially

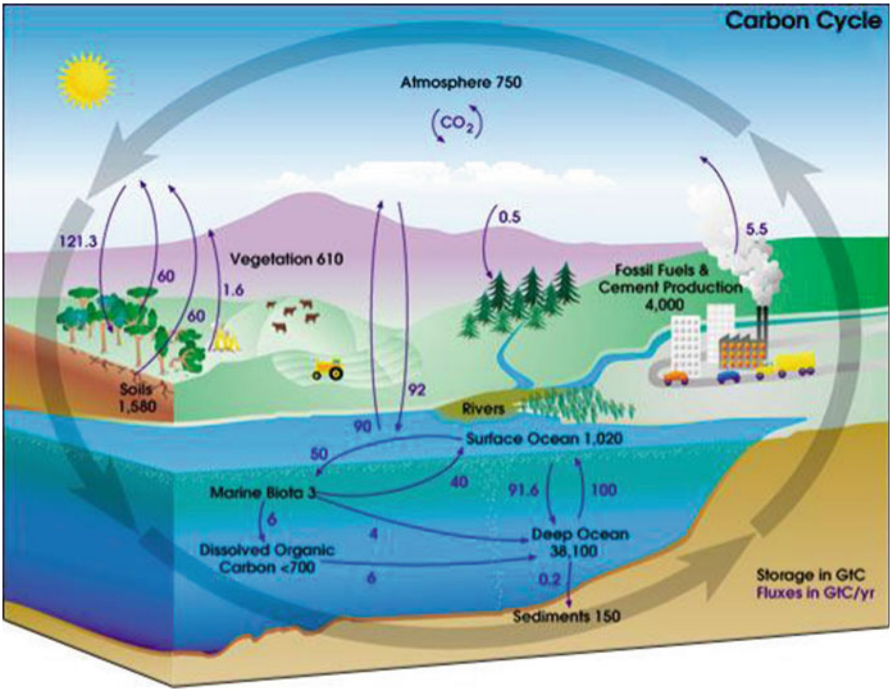

Fig. 4.7 A schematic diagram of the global carbon cycle, obtained from http://earthobservatory.nasa.gov/Features/CarbonCycle/carbon_cycle4.php

greenhouse gases, such as CO_2 and N_2O. Also, sea salt from the ocean surface is an important source of natural aerosols which have impacts on climate through aerosol–climate interaction. In this section, a few aspects on modeling of the biogeochemical cycle in the OGCM are briefly reviewed.

Currently, approximately 6 Gt/year of carbon is released in the atmosphere due to anthropogenic activity. Only about half of carbon remains in the atmosphere and the rest is stored either in the vegetation and soil, or in biota and dissolved trace species in the ocean. This stored carbon is typically identified as terrestrial and ocean carbon pools. These fluxes and reservoirs for carbon are summarized in Fig. 4.7. The size of the terrestrial and atmospheric carbon pools are about 2,200 and 750 Gt, while the Ocean carbon pool is estimated to hold about 40,000 Gt, the largest carbon pool except sediments and rocks.

The first generation of the Ocean Carbon model to treat the exchange process between atmosphere and ocean was developed more than 50 years ago by Revelle and Suess [50], which used two boxes of oceanic carbon pools with exchange processes between atmosphere and ocean. The rationale of this simple box model has been employed consistently through later generations of the ocean carbon model (e.g., [31]). Research questions about how much of the extra CO_2 released by human activity in the atmosphere can be stored in the ocean carbon pools resides at the core of climate change science, and how these large carbon pools will respond to global warming is also critical to understand.

The conservation equation for any trace constituent (like CO_2) in the ocean or atmosphere can be written as follows:

$$\frac{\partial C}{\partial t} = -\vec{V} \cdot \nabla C - C\nabla \cdot \vec{V} + C_{\text{Source}} - C_{\text{Sink}}, \tag{4.27}$$

where C is a mixing ratio of any trace constituent (units mass of trace constituent to mass of sea water).

There exist two important source and sink processes for carbon in the ocean, that are typically labeled as the biological and solubility pumps. The biological pump refers to the transport of carbon from the oceanic surface to its interior in the form of carbon bound to other elements and labeled organic or inorganic carbon, according to whether the carbon resides in chemical compound that arises primarily through biological activity (organic) or other chemical reactions (inorganic). Various routes for the transport of carbon from the atmosphere into different parts of the ocean have been discovered and are treated in modern climate models treating biogeochemistry. The strength of this biological pump is measured by the so-called f-ratio, a fraction of total primary production fueled by nitrate [17].

Second, the solubility pump is a transport process that takes place through physical-chemical interactions. Both temperature in the ocean layers and the aforementioned thermohaline circulation are two important players in determining the strength of the solubility pump. The most of the climate models in the IPCC 4th Assessment Report (AR4) does not have representation of carbon cycle. However, the next generation of OGCMs to be used in the IPCC 5th Assessment Report (AR5) has some degree of numerical representation of these two pumps.

Future Directions

The simulations of the oceans by contemporary OGCM in coupled climate model still exhibit various degrees of biases compared to observations despite improvements. Further reduction of these biases is required. Some dynamical, physical, and even biological processes are poorly represented, and others are entirely missing. Thus, much emphasis has been weighted on developing more comprehensive and computationally efficient ocean climate model at the same time. As a first step, increasing spatial resolution has been pursued as a way forward by many modeling groups in the world and reduces the need for some sub-grid scale processes which must currently be parameterized. Also, better numerical algorithm suitable for extreme computing environment has been sought at the same time.

In summary, several key aspects in the development of the next generation of the OGCM are as follows:

1. Multi-decadal natural variability-Atlantic Meridional Overturning Circulation (AMOC): This phenomenon is important for climate not only in the pan-Atlantic regions, but also for other regions in longer-term climate variability [60]. With limited availability of observational data, the OGCM has been playing an important role in understanding many processes involved with the AMOC and in simulating its historical evolution as well as current status. For example, most of climate simulations produce a weakening of the AMOC under the global warming due to increasing temperature [23, 55]. However, a role played by the fresh water budget on the AMOC has large uncertainty [61]. Many regional and international efforts have been made to collect more data and to improve model simulation of the AMOC [53].
2. Sea-level change in the warming world: Sea-level change in the global ocean is not only a function of temperature increase, but also many regional processes. As reviewed earlier, the OGCM with the Boussinesq approximation has its limitation in suitably simulating this sea-level change. Also, the ice sheet and sea ice variability and change have to be simulated and incorporated into the OGCM.
3. Simulation of marine ecosystem in the OGCM: The OGCM was typically for description of ocean circulation and physical properties. However, as our planet changes due to the anthropogenic activities, there are increasing demands on what would be the impact of climate change on marine ecosystem in a very regional scale (such as the coastal region of the Gulf of Mexico). On the other hand, proper simulation of marine ecosystem benefits better representation of the biological pump in the ocean biogeochemical cycle.

This entry introduces some of the basic concepts and examples of the OGCM. More details on each subject can be found in the following articles or textbooks. For example, general introductory and review on the contemporary OGCM can be found in Randall et al. [48], Griffies et al. [27], and Haidvogel and Bryan [29]; numerical formulation can be found in Griffies [24], Griffies et al. [26], and Bryan [5]; and biogeochemistry in the OGCM can be described in much more detail in Sarmiento [54].

Bibliography

1. Adcroft A, Campin J-M, Hill C, Marshall J (2004) Implementation of an atmosphere–ocean general circulation model on the expanded spherical cube. Mon Weather Rev 132:2845–2863
2. Beckmann A, Haidvogel D (1993) Numerical simulation of flow around a tall isolated seamount. Part I: Problem formulation and model accuracy. J Phys Oceanogr 23:1736–1753
3. Bjerknes J (1969) Atmospheric teleconnections from the equatorial Pacific. Mon Weather Rev 97:163–172
4. Bleck R (1998) Ocean modeling in isopycnic coordinates. In: Chassignet EP, Verron J (eds) Ocean modeling and parameterization. Kluwer, Dordrecht, pp 423–448

5. Bryan K (1989) The design of numerical model of the ocean circulation. In: Anderson DLT, Willebrand J (eds) Oceanic circulation models: combining data and dynamics. Kluwer, Dordrecht, pp 465–500
6. Breugem W.-P, Chang P, Jang CJ, Mignot J, Hazeleger W (2008) Barrier layers and tropical Atlantic SST biases in coupled GCMs. Tellus A 60:885–897. doi: 10.1111/j.1600-0870.2008.00343.x
7. Gill AE (1982) Atmosphere-ocean dynamics. Academic Press, 662 pp
8. Matsuno T (1966) Quasi-geostrophic motions in the equatorial area. J Meteorol Soc Jpn 44:25–43
9. Cane MA, Zebiak S (1985) A theory for El Niño and the southern oscillation. Science 228:1085–1087
10. Casulli V, Walters RA (2000) An unstructured grid, three-dimensional model based on the shallow water equations. Int J Numer Methods Fluids 32:331–346
11. Chassignet EP, Hurlburt HE, Smedstad OM, Halliwell GR, Wallcraft AJ, Metzger EJ, Blanton BO, Lozano C, Rao DB, Hogan PJ, Srinivasan A (2006) Generalized vertical coordinates for eddy resolving global and coastal ocean forecasts. Oceanography 19(1):118–129
12. Chen C, Liu H, Beardsley RC (2003) An unstructured grid, finite-volume, three-dimensional, primitive equations ocean model: applications to coastal ocean and estuaries. J Atmos Ocean Technol 20:159–186
13. Chen C, Beardsley RC, Cowles G (2006) An unstructured-grid finite-volume coastal ocean model (FVCOM) system. Oceanography 19(1):78–89
14. Cornejo-Garrido AG, Stone PH (1977) On the heat balance of the Walker circulation. J Atmos Sci 34:1155–1162
15. Cox MD (1989) An idealized model of he world ocean. Part I: The global-scale water masses. J Phys Oceanogr 19:1730–1752
16. Deleersnijder E, Beckers J-M (1992) On the use of the σ-coordinate system in regions of large bathymetric variations. J Mar Syst 3:381–390
17. Eppley RW, Peterson BJ (1979) Particulate organic matter flux and planktonic new production in the deep ocean. Nature 282:677–680
18. Gaspar P, Gregoris Y, Lefevre J (1990) A simple eddy kinetic energy model for simulations of the oceanic vertical mixing: Tests at station Papa and long-term upper ocean study site. J Geophys Res 95:16179–16193
19. Gent PR, McWilliams JC (1990) Isopycnal mixing in ocean circulation models. J Phys Oceanogr 20:150–155
20. Gent PR, Willebrand J, McDougall TJ, McWilliams JC (1995) Parameterizing eddy induced tracer transports in ocean circulation models. J Phys Oceanogr 25:463–474
21. Ghil M, Malanotte-Rizzoli P (1991) Data assimilation in meteorology and oceanography. Adv Geophys 33:141–266
22. Gill AE (1980) Some simple solutions for heat induced tropical circulation. Q J R Meteorol Soc 106:447–462
23. Gregory JM et al (2005) A model intercomparison of changes in the Atlantic thermohaline circulation in response to increasing atmospheric CO_2 concentration. Geophys Res Lett 32: L12703. doi:10.1029/2005GL023209
24. Griffies SM (2004) Fundamentals of ocean climate models. Princeton University Press, Princeton, 518 pp
25. Griffies SM et al (2000) Developments in ocean climate modeling. Ocean Model 2:123–192
26. Griffies SM, Ganadesikan A, Dixon KW, Dunne JP, Gerdes R, Harrison MJ, Rosati A, Russell JL, Samuels BL, Spelman MJ, Winton M, Zhang R (2005) Formulation of an ocean model for global climate simulations. Ocean Sci 1:45–79
27. Griffies SM, Adcroft AJ, Banks H, Böning CW, Chassignet EP, Danabasoglu G, Danilov S, Deleersnijder E, Drange H, England M, Fox-Kemper B, Gerdes R, Gnanadesikan A, Greatbatch RJ, Hallberg RW, Hanert E, Harrison MJ, Legg S, Little CM, Madec G, Marsland SJ, Nikurashin M, Pirani A, Simmons HL, Schröter J, Samuels BL, Treguier A-M, Toggweiler

JR, Tsujino H, Valllis GK, White L (2009) Problems and prospects in large-scale ocean circulation models. Community White Paper for OceanObs09. https://abstracts.congrex.com/scripts/jmevent/abstracts/FCXNL-09A02a-1672431-1-cwp2a07.pdf
28. Guilyardi E et al (2004) Representing El Niño in coupled ocean-atmosphere GCMs: the dominant role of the atmospheric component. J Climate 17:4623–4629
29. Haidvogel DB, Bryan FO (1992) Ocean general circulation modeling. In: Trenberth KE (ed) Climate system modeling. Cambridge University Press, New York, pp 371–412
30. Roberts MBH, Gedney N, Gregory J, Hill R, Mullerworth S, Pardaens A, Rickard G, Thorpe R, Wood R (2004) Impact of an eddy-permitting ocean resolution on control and climate change simulations with a global coupled GCM. J Clim 17:3–20
31. Sarmiento JL, Gruber N (2002) Sinks for anthropogenic carbon. Physics Today 55(8):30–36
32. Haney RL (1991) On the pressure gradient force over steep topography in sigma-coordinate ocean models. J Phys Oceanogr 21:610–619
33. Henderson-Sellers B, Davies AM (1989) Thermal stratification modeling for oceans and lakes. Annu Rev Numer Fluid Mech Heat Transfer 2:86–156
34. Jackson L, Hallberg R, Legg S (2008) A parameterization of shear-driven turbulence for ocean climate models. J Phys Oceanogr 38:1033–1053
35. Lehmann A (1995) A three-dimensional baroclinic eddy-resolving model of the Baltic Sea. Tellus 47A:1013–1031
36. Lindzen RS, Nigam S (1987) On the role of sea surface temperature gradients in forcing low level winds and convergence in the tropics. J Atmos Sci 44:2418
37. Maltrud M, McClean J (2005) An eddy resolving global 1/10° ocean simulation. Ocean Model 8:31–54
38. Manabe S, Bryan K (1969) Climate calculations with a combined ocean-atmosphere model. J Atmos Sci 26:786–789
39. Manabe S, Bryan K, Spelman MJ (1975) A global ocean-atmosphere climate model. Part I. The atmospheric circulation. J Phys Oceanogr 5:3–29
40. Marotzke J (1989) Instabilities and multiple steady states of the thermohaline circulation. In: Anderson DLT, Willebrand J (eds) Oceanic circulation models: combining data and dynamics. Kluwer, Dordrecht, pp 501–511
41. Martin PJ (1985) Simulation of the mixed layer at OWS November and Papa with several models. J Geophys Res 90:903–916
42. Meehl G et al (2007) The WRCP CMIP3 Multimodel Dataset: A new era in climate change research. Bull Am Meteorol Soc 88:1383–1394
43. Mellor GL, Yamada T (1982) Development of a turbulence closure model for geophysical fluid problems. Rev Geophys Space Phys 20:851–875
44. Neelin JD (1989) On the interpretation of the Gill model. J Atmos Sci 46:2466–2468
45. Noh Y, Kim H-J (1999) Simulations of temperature and turbulence structure of the oceanic boundary layer with the improved near-surface process. J Geophys Res 104:15621–15634
46. Philander SGH, Yamagata T, Pacanowski RC (1984) Unstable air-sea interaction in the tropics. J Atmos Sci 41:604–613
47. Randall D, Ringler T, Heikes R, Jones P, Baumgardner J (2002) Climate modeling with spherical geodesic grids. Comput Sci Eng 4:32–41
48. Randall DA, Wood RA, Bony S, Colman R, Fichefet T, Fyfe J, Kattsov V, Pitman A, Shukla J, Srinivasan J, Stouffer RJ, Sumi A, Taylor KE (2007) Climate models and their evaluation. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, Miller HL (eds) Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, UK
49. Redi MH (1982) Oceanic isopycnal mixing by coordinate rotation. J Phys Oceanogr 12:1154–1158
50. Revelle R, Suess HE (1957) Carbon dioxide exchange between atmosphere and ocean and the question of increasing of atmospheric CO_2 during the past decades. Tellus 9:18–27

51. Ringler T, Ju L, Gunzburger M (2008) A multiresolution method for climate system modeling: application of spherical centroidal voronoi tessellations. Ocean Dyn 58:475–498
52. Sadourny R, Arakawa A, Mintz Y (1968) Integration of the non-divergent barotropic vorticity equation with an icosahedral-hexagonal grid for the sphere. Mon Weather Rev 96:351–356
53. Sanford TB, Kelly KA, Former DM (2011) Sensing the ocean. Phys Today 64(2):24–28
54. Sarmiento JL (1992) Biogeochemical ocean models. In: Trenberth KE (ed) Climate system modeling. Cambridge University Press, New York, pp 519–551
55. Schmittner A, Latif M, Schneider B (2005) Model projections of the North Atlantic thermohaline circulation for the 21st century assessed by observations. Geophys Res Lett 32:L32710. doi:10.1029/2005GL024368
56. Semtner AJ, Chervin RM (1992) Ocean general circulation from a global eddy-resolving model. J Geophys Res 97:5493–5550
57. Shchepetkin A, McWilliams J (2002) A method for computing horizontal pressure-gradient force in an ocean model with a non-aligned vertical coordinate. J Geophys Res 108:35.1–35.34
58. Smith R, Gent P (2004) Reference Manual for the Parallel Ocean Program (POP)—ocean component of the community climate system model. Los Alamos National Laboratory, LAUR-02-2484
59. Solomon H (1971) On the representation of isentropic mixing in ocean models. J Phys Oceanogr 1:233–234
60. Sutton RT, Hodson LR (2005) Atlantic forcing of North American and European summer climate. Science 309:115–118
61. Swingedouw D, Braconnot P, Marti O (2006) Sensitivity of the Atlantic Meridional overturning circulation to the melting from northern glaciers in climate change experiments. Geophys Res Lett 33:L07711. doi:10.1029/2006GL025765
62. Vernois G (1973) Large scale ocean circulation. In: Yih C-S (ed) Advances in applied mathematics, vol 13. Academic, New York, pp 1–92
63. Walton CC, Pichel WG, Sapper JF (1998) The development and operational application of nonlinear algorithms for the measurement of sea surface temperatures with the NOAA polar-orbiting environmental satellites. J Geophys Res 103(C12): 27999–28012
64. Weart SR (2008) The discovery of global warming: revised and expanded edition. Harvard University Press, Cambridge, MA. 240 p. ISBN: 978–0674031890
65. Wyrtki K (1975) El Niño – The dynamic response of the equatorial Pacific Ocean to atmospheric forcing. J Phys Oceanogr 4:91–103
66. Yeh S-W, Kug J-S, Dewtte B, Kwon M-H, Kirtman B, Jin F-F (2009) El Niño in a changing climate. Nature 461:511–514
67. Zebiak S, Cane MA (1987) A model El Niño-southern oscillation. Mon Weather Rev 115:2262–2278
68. Gill AE (1985) Elements of coupled ocean-atmosphere models for the tropics. In: Nihoul JCJ (ed) Coupled ocean-atmosphere models, vol 40. Elsevier Oceanogr. Ser., Amsterdam, pp 303–327