SCIENCE OF OCEAN CLIMATE MODELS

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Introduction: Models as Essential Tools for Ocean Science

The rich textures and features of the global ocean circulation and its tracer distributions cannot be fully studied in a controlled laboratory setting using a direct physical analog. Consequently, ocean scientists have increasingly turned to numerical models as a rational experimental tool, along with theoretical methods and a growing array of observational measurements, for rendering a mechanistic understanding of the ocean. Indeed, during the 1990s and early 2000s, global and regional ocean models have become the experimental tool of choice for many oceanographers and climate scientists.

The scientific integrity of computer simulations of the ocean has steadily improved during the past decades due to deepened understanding of the ocean and ocean models, and from enhanced computer power facilitating more realistic representations of the huge range of scales relevant for ocean fluid dynamics. Additionally, ocean models are a key component in earth system models (ESMs), which are used to study interactions between physical, chemical, and biological aspects of the Earth's climate system. ESMs are also used to help anticipate modifications to the Earth's climate arising from humanity's uncontrolled greenhouse experiment. Quite simply, without computer models, our ability to develop a robust and testable scientific basis for ocean and climate dynamics of the past, present, and future would be absent.

There is a tremendous amount of science forming the foundations of ocean models. This science spans a broad interdisciplinary spectrum, including physics, mathematics, chemistry, biology, computer science, climate science, and all aspects of oceanography. This article describes the fundamental principles forming the basis for physical ocean models. We are particularly interested here in global ocean climate models used to study large-scale and long-term phenomena of direct importance to climate.

Scales of Motion and the Subgrid-scale Problem

To appreciate the magnitude of the task required to simulate the ocean circulation, consider the scales of motion involved. We start at the large scale, where a typical ocean basin is on the order of 10^3–10^4 km in horizontal extent, with depths reaching on average to *c.* 5 km. The ocean's massive horizontal gyre and overturning circulations occupy nearly the full extent of these basins, with typical recirculation times for the horizontal gyres decadal, and overturning timescales millennial.

At the opposite end of the spectrum, the ocean microscale is on the order of 10^{-3} m, which is the scale where molecular viscosity can act on velocity gradients to dissipate mechanical energy into heat. This length scale is known as the Kolmogorov length, and is given by $(\nu^3/\varepsilon)^{1/4}$, where $\nu = 10^{-6}$ $\mathrm{m^2\,s^{-1}}$ is the molecular kinematic viscosity for water, and ε is the energy dissipation rate. In turn, molecular viscosity and the Kolmogorov length imply a timescale $T = L^2/\nu \approx 1$ s.

When formulating a computational physics problem, it is useful to estimate the number of discrete degrees of freedom required to represent the physical system. Consider a brute force approach, where all the space and timescales described above are explicitly resolved by the ocean simulation. One-second temporal resolution over a millennial timescale climate problem requires more than 3×10^{10} time steps of the model equations. Resolving space into regions of dimension 10^{-3} m for an ocean with volume roughly $1.3 \times 10^{18}\,\mathrm{m}^3$ requires 1.3×10^{27} discrete grid cells (roughly 10^4 times larger than Avogadro's number!). These numbers far exceed the capacity of any computer in use today, or for the forseeable future. Consequently, a truncated description of the ocean state is required.

Truncation Methods

Three general approaches to truncation are employed in fluid dynamics. One approach is to coarsen the space time resolution. Doing so introduces a loss of information due to the unresolved small scales. Determining how the resolved scales are affected by the unresolved scales is fundamental to computational fluid dynamics, as well as to a statistical description of fluid turbulence. This is a nontrivial problem in subgrid-scale (SGS) parametrization, a problem intimately related to the turbulence closure problem of fluid dynamics.

The second truncation method filters the continuum equations by truncating the fundamental modes of motion admitted by the equations.

The result is an approximation to the original physical system. The advantage is that reducing the admitted motions also reduces the space timescales required to simulate the system, and/or it simplifies the governing equations thus facilitating computationally cheaper simulations. The hydrostatic approximation is a prime example of filtering used in all present-day global ocean climate models. Here, vertical motions are assumed to possess far less energy than horizontal, thus rendering a simplified vertical momentum balance where the weight of fluid above a point in the ocean determines the pressure at that point. The Boussinesq approximation provides another filtering of the fundamental equations. In this case, the near incompressibility of seawater is exploited to eliminate all acoustic motions by assuming that fluid parcels conserve volume rather than mass. The Boussinesq approximation is commonly used in ocean climate modeling, but it is becoming less so due to its inability to provide a prognostic budget for sea level that includes steric effects. Steric effects are associated with water expansion or contraction arising from density changes, and such changes are a key aspect of the ocean's response to anthropogenic climate change, with a warmer ocean occupying a larger volume which raises the sea level.

A final truncation method considers a much smaller space and time domain, yet maintains the very fine space and time resolution set by either molecular viscosity (direct numerical simulation (DNS)), or somewhat larger viscosity (large eddy simulation (LES)). Both DNS and LES are important for process studies aimed at understanding the mechanisms active in fine-scale features of the ocean. Insights gained via DNS and LES have direct application to the development of rational SGS parametrizations of use for ocean climate models.

Ocean Mesoscale Eddies

The Reynolds number UL/ν provides a dimensionless measure of the importance of advective effects relative to viscous or frictional effects. Consider a large-scale ocean current, such as the Gulf Stream, with a velocity scale $U = 1\,\mathrm{m\,s^{-1}}$, length scale $L = 100\,\mathrm{km}$, and molecular viscosity $\nu = 10^{-6}\,\mathrm{m^2\,s^{-1}}$. In this case, the Reynolds number is $Re \approx 10^{11}$, which is very large and so means that the ocean fluid contains extremely turbulent regimes. At these space and timescales, the effects of rotation and stratification are both critical. The turbulence relevant at this scale is termed geostrophic turbulence. A geostrophically turbulent fluid contains numerous mesoscale eddy features, which result from a conversion of potential energy (imparted to the ocean fluid by the atmospheric forcing) to kinetic energy through the process of baroclinic instability. Such eddy features are the norm rather than the exception in most of the ocean. They provide a chaotic or turbulent element to the ocean general circulation. Hence, quite generally the ocean flow is not smooth and laminar, unless averaging over many years. Rather, it is turbulent and full of chaotic fine-scale motions which make for an extremely challenging simulation task.

Mesoscale eddies have scales on the order of 100 km in the middle latitudes, 10 km in the higher latitudes, and their timescale for recirculation is on the order of months. The length scale is related to the first baroclinic Rossby radius, which is a scale that arises in the baroclinic instability process. Furthermore, mesoscale eddies are the ocean's analog of atmospheric weather patterns, with atmospheric weather occurring on scales roughly 10 times larger than the ocean eddies. Differences in vertical density stratification account for differences in length scales.

Ocean tracer properties are strongly affected by the mixing and stirring of mesoscale eddies. For example, biological activity is strongly influenced by eddies, especially with the strong upwelling in the cyclonic eddies bringing high nutrient water to the surface. Additionally, eddies are a leading order feature transporting properties such as heat and freshwater poleward across the Antarctic Circumpolar Current. They also mix properties across the time-mean position of the Gulf Stream and Kuroshio Currents, whose jet-like currents continually fluctuate due to baroclinic instability.

Given the smaller scales of mesoscale eddies in the ocean than in the atmosphere, the problem of simulating these ocean features is $10 \times 10 \times 10$ more costly than in the atmosphere. Here, two factors of 10 arise from the horizontal scales, and another from the associated refinement in temporal resolution. It is not rigorously known what grid resolution is required to resolve the ocean's mesoscale eddy spectrum. However, preliminary indications point to 10 km or finer globally, with roughly 50–100 vertical degrees of freedom needed to capture the vertical structure of the eddies, as well as the tight vertical gradients in tracer properties. This resolution is far coarser than that required to resolve the Kolmogorov scale (as required for DNS). Nonetheless, globally resolving the ocean mesoscale eddy spectrum over climate timescales remains beyond our means, with the most powerful computers only just beginning to be applied to such massive simulations.

Ocean climate modelers thus continually seek enhanced computer power in an aim to reduce the level of space time resolution coarsening. The belief, largely reflected in experience with simulations of refined resolution, is that reducing dependence on SGS parametrizations, such as those parametrizing mesoscale eddies, improves the simulation integrity. One reason for this improvement is that many SGS parametrizations are far from relevant for all of the multiple flow regimes of the ocean. Additionally, no SGS parametrization commonly used in ocean climate models includes the stochastic effects of turbulent eddies, and resolving these effects may be important for climate variability and predictability. Finally, refined resolution allows for a better representation of the very complex land–sea boundaries that strongly influence ocean currents and water mass properties. Hence, deducing the impact on ocean climate of explicitly resolved mesoscale eddies, and other fine-scale currents such as occur near boundaries, is the grand challenge of ocean climate modeling in the early twenty-first century.

As grid resolution is refined so that mesoscale eddy features are admitted, the resulting eddy permitting simulations become strongly dependent on the integrity of the numerical methods used for the transport of fluid properties. Although the representation of transport is important at coarse resolution, the fluid is also very viscous due to the huge viscosity required at the noneddying resolutions. Hence, a more energetic simulation, with enhanced energy in the resolved scale and much stronger fronts between ocean properties, generally stresses the ability of transport algorithms to maintain physically appropriate behavior. The integrity of the discrete transport operators is thus arguably the most critical feature of a numerical algorithm that sets the fidelity of eddying simulations.

Although mesoscale eddy permitting simulations are becoming more feasible as computer power increases, there remain important problems where SGS parametrizations must be used to partially capture features of the unresolved mesoscale spectrum, as well as smaller scales. The longer the timescale of the phenomena studied, the greater the need for SGS parametrization. In particular, no mesoscale eddy simulation has yet been run for more than a few decades. Climate modelers must therefore carefully employ truncation methods to make progress. Depending on particulars of the unresolved motions, the resulting SGS parametrizations can, and do, affect in nontrivial manners the simulation's physical integrity. Consequently, a great deal of intellectual energy in ocean model algorithm development relates to establishing numerically robust and physically rational SGS parametrizations. This is an extremely difficult problem. Nonetheless, some progress has been made during the past decades, at least enough to know what not to do in certain cases.

Posing the Problem of Ocean Modeling

The ocean is a forced and dissipative system. Forcing occurs at the upper boundary from interactions with the atmosphere, rivers, and sea ice, and at its lower boundary from the solid Earth. Forcing also occurs from astronomical effects of the Sun and Moon which produce tidal motions.

Important atmospheric forcing occurs over basin scales, with timescales set by the diurnal cycle, synoptic weather variability (days), the seasonal cycle, and interannual fluctuations such as the North Atlantic Oscillation and El Niño. Atmospheric momentum and buoyancy fluxes are predominantly responsible for driving the ocean's large scale horizontal and overturning circulations. Additional influences include forcing at continental boundaries from river inflow and calving glaciers, as well as in polar regions where sea ice dynamics greatly affect the surface buoyancy fluxes. Dissipation in the ocean generally occurs at the microscale, with turbulent processes converting mechanical energy to heat. Energy moves from the large to the small scales through nonlinear advective transport. In short, the ocean circulation emerges from an enormous number of processes spanning a huge space and time range. Furthermore, modes of ocean dynamical motions span space and timescales even greater than the forcing scales, with ocean circulation patterns, especially those in the abyssal regions, maintaining coherence over global scales for millennia.

From a mathematical physics perspective, the problem of modeling the ocean circulation is a problem of establishing and maintaining the kinematic, dynamic, and material properties of the ocean fluid: a fluid which is driven by a multitude of boundary interactions and possesses numerous internal transport and mixing processes. Kinematic balances are set by the geometry of the ocean domain, and by assuming that the fundamental fluid parcel constituents conserve mass. Dynamical balances result from applying Newton's law of motion to the continuum fluid so that the acceleration of a fluid parcel is set by forces acting on the parcel, with the dominant forces being pressure, Coriolis, gravity, and friction. And material balances of tracers, such as salt, heat, and biogeochemical species, are affected by circulation, mixing from turbulence, boundary fluxes, and internal sources and sinks. The ocean modeling problem also involves, at a fundamental level, a rational parametrization of SGS processes, such as fine-scale convective mixing in the upper ocean mixed layer due to strong mechanical and buoyant interactions with the atmosphere, breaking internal waves induced from tides rubbing against the ocean bottom, and mesoscale eddies spawned by baroclinically unstable currents. From these balances emerges the wonderful richness of the ocean circulation and its variety of tracer distributions. Faithfully emulating this system using computer simulations requires a massive effort in scientific and engineering ingenuity and collaboration.

Fundamental Budgets and Methods

We now aim to place the previous discussions onto a more rigorous mathematical basis. Doing so is a necessary first step toward developing a suite of equations amenable to numerical methods for use on computers. Due to limitations in space, the material here will be quite terse and selective, allowing many results to be presented without full derivation and many topics to be omitted.

The basic equations of ocean fluid dynamics can be readily formulated by focusing on the dynamics of a mass conserving parcel of seawater. A fluid parcel is a region that is macroscopically small yet microscopically large. That is, from a macroscopic perspective, the parcel's thermodynamic properties may be assumed uniform, and the methods of continuum mechanics are applicable to describing the mechanics of an infinite number of these effectively infinitesimal parcels. However, from a microscopic perspective, these fluid parcels contain many molecules (on the order of Avogadro's number), and so it is safe to ignore the details of molecular interactions. Regions of a fluid with length scales on the order of 10^{-5} m satisfy these properties of a fluid parcel.

Mass Conservation

The mass of a parcel is written as $\mathrm{d}M = \rho \mathrm{d}V$, where ρ is the mass per volume (i.e., the density), and $\mathrm{d}V$ is the parcel's volume. Assuming the parcel mass to be conserved as the parcel moves through the fluid leads to the differential statement of mass conservation

$$\frac{\mathrm{d}}{\mathrm{d}t}\ln(\mathrm{d}M) = 0 \qquad [1]$$

where we assume there to be no internal sources of mass. The time derivative $\mathrm{d}/\mathrm{d}t$ measures changes occuring with respect to the moving parcel. This Lagrangian or material perspective complements the Eulerian perspective rendered by viewing the fluid from a fixed point in space. Transforming to the Eulerian perspective leads to the relation

$$\frac{\mathrm{d}}{\mathrm{d}t} = \partial_t + \mathbf{v}\cdot\nabla \qquad [2]$$

where ∂_t measures time changes at a fixed space point, and $\mathbf{v}$ is the parcel's velocity. The transport or advective operator $\mathbf{v}\cdot\nabla$ reveals the fundamentally nonlinear character of fluid dynamics arising from motions of fluid parcels with velocity $\mathbf{v}$. In the Eulerian perspective, mass conservation takes the form

$$\partial_t \rho + \nabla\cdot(\rho\mathbf{v}) = 0 \qquad [3]$$

Tracer Budgets

In addition to freshwater, a seawater parcel contains numerous constituents known as tracers. Heat (a thermodynamical tracer) and salt (a material tracer) are seawater's two active tracers. Heat and salt, along with pressure, determine the mass density through a complex empirical relation known as the equation of state. Density in turn impacts the hydrostatic pressure which then affects currents. Other ocean tracers include radioactive species, such as those introduced in the 1950s and 1960s from nuclear bomb tests; biological tracers such as phytoplankton and zooplankton fundamental to ocean ecosystems; and chemical elements such as carbon, iron, and nitrogen prominent in ocean biogeochemical cycles.

In describing the evolution of tracer within a seawater parcel, it is convenient to consider the tracer concentration C, which represents a mass of tracer per mass of seawater for material tracers. In addition to material tracers, we are concerned with a thermodynamical tracer that measures the heat within a fluid parcel. In this case, C is typically taken to be the potential temperature or potential enthalpy.

The evolution of tracer mass within a Lagrangian parcel of mass conserving fluid is given by

$$\rho\frac{\mathrm{d}C}{\mathrm{d}t} = -\nabla\cdot\mathbf{J} \qquad [4]$$

where for simplicity we ignore tracer sources. The tracer flux $\mathbf{J}$ arises from SGS transport of tracer in the absence of mass transport. Such transport consists of diffusion and/or unresolved advection. As this flux is not associated with mass transport, it vanishes when the tracer concentration is uniform, in which case the tracer budget reduces to the mass budget of eqn [1]. Use of the material time derivative relation [2] and mass conservation [3] renders the Eulerian conservation law for tracer

$$\partial_t(\rho C) + \nabla\cdot(\rho C\mathbf{v}) = -\nabla\cdot\mathbf{J} \qquad [5]$$

The convergence $-\nabla\cdot(\rho C\mathbf{v})$ is the flux form of advective tracer transport. As stated in the section titled

'Ocean mesoscale eddies', advective transport is critical, especially in eddying simulations where both the tracer and velocity distributions possess nontrivial structure. The flux form of advective tracer transport is employed in ocean climate models rather than the alternative advective form $\rho\mathbf{v} \cdot \nabla C$. The reason is that the flux form is amenable to conservative numerical schemes, as well as to a finite volume interpretation described in the section titled 'Basics of the finite volume method'.

The specification of SGS parametrizations appearing in the flux $\mathbf{J}$ is critical for the tracer equation, especially in models not admitting mesoscale eddies. The original approach, whereby the most common class of ocean climate models employed a diffusion operator oriented according to geopotential surfaces, greatly compromised the simulation's physical integrity for climate purposes. The problem with the horizontally oriented operators is that they introduce unphysically large mixing between simulated water masses. Given the highly ideal fluid dynamics of the ocean interior, most of the tracer transport in the interior arising from eddy effects occurs along a locally referenced potential density direction, otherwise known as a neutral direction. This mixing preserves water mass properties over basin scales for decades. Altering the orientation of the model's diffusion operator from horizontal–vertical to neutral–vertical brought the models more in line with the real ocean. In addition, mesoscale eddies stir tracers in a reversible manner. This stirring corresponds to an antisymmetric component to the SGS tracer transport tensor. The combination of neutral diffusion and skew diffusion have become ubiquitous in all ocean climate models that do not admit mesoscale eddies, even those models not based on the geopotential vertical coordinate.

Linear Momentum Budget

The linear momentum of a fluid parcel is given by $\mathbf{v}\rho \mathrm{d}V$. Through Newton's law of motion, momentum changes in time due to the influence of forces acting on the parcel. The forces acting on the ocean fluid parcel include internal stresses in the fluid arising from friction and pressure; the Coriolis force due to our choice of describing the ocean fluid from a rotating frame of reference; and gravity acting in the local vertical direction. The resulting equation for linear momentum takes the form

$$\rho \frac{\mathrm{d}\mathbf{v}}{\mathrm{d}t} = -\rho g\,\hat{\mathbf{z}} - f\hat{\mathbf{z}} \wedge \rho\mathbf{v} - \nabla p + \rho\mathbf{F} \qquad [6]$$

In this equation, $g \approx 9.8\,\mathrm{m\,s^{-2}}$ is the acceleration due to gravity, which is generally assumed constant for ocean climate modeling. The Coriolis force per mass is written $-f\hat{\mathbf{z}} \wedge \mathbf{v}$, with the Coriolis parameter $f = 2\Omega \sin\phi$, where $\Omega = 7.292 \times 10^{-5}\,\mathrm{s}^{-1}$ is the Earth's angular rotation rate, and ϕ is the latitude. Gradients in pressure p impart an acceleration on the fluid parcel, with the parcel accelerated from regions of high pressure to low.

The friction vector $\mathbf{F}$ arises from the divergence of internal friction stresses. In an ocean model, these stresses must be sufficient to maintain a near-unit-grid Reynolds number. Otherwise, the simulation may go unstable, or at best produce unphysical noise-like features. This constraint on the numerical simulation is unfortunate, since a unit grid Reynolds number requires an effective viscosity many orders of magnitude larger than molecular, since the grid sizes (order 10^4–10^5 m) are much larger than the Kolmogorov scale (10^{-3} m). This level of numerical friction is not based on physics, but arises due to the discrete lattice used for the numerical simulations. Various methods have been engineered to employ the minimal level of friction required to meet this, and other, numerical constraints, so as to reduce the effects of friction on the simulation. Such methods are *ad hoc* at best, and lead to some of the most unsatisfying elements in ocean model practice since the details of the methods can strongly influence the simulation.

The hydrostatic approximation mentioned in the section titled 'Scales of motion and the subgrid-scale problem' exploits the large disparity between horizontal motions, occurring over scales of many tens to hundreds of kilometers, and vertical motions, occurring over scales of tens to hundreds of meters. In this case, it is quite accurate to assume that the moving fluid maintains the hydrostatic balance, whereby the vertical momentum equation takes the form

$$\partial_z p = -\rho g \qquad [7]$$

This approximation is a fundamental feature of the ocean's primitive equations, which are the equations solved by all global ocean climate models in use today. By truncating, or filtering, the vertical momentum budget to the inviscid hydrostatic balance, we are obliged to parametrize strong vertical motions occurring in convective regions, since the primitive equations cannot explicitly represent these motions. This has led to various convective parametrizations in use by ocean climate models. These parametrizations are essential for the models to accurately simulate various deep-water formation processes, especially those occurring in the open ocean due to strong buoyancy fluxes.

General Strategy for Time-Stepping Momentum

Acoustic waves are three-dimensional fluctuations in the pressure field. They travel at roughly $1500\,\mathrm{m\,s^{-1}}$. There is no evidence that resolving acoustic waves is essential for the physical integrity of ocean climate models. The hydrostatic approximation filters all acoustic modes except the Lamb wave (an acoustic mode that propogates only in the horizontal direction). As the Lamb wave is close in speed to the external gravity waves, it can generally be subsumed into an algorithm for gravity waves, and so is of little consequence to ocean model algorithms.

External or barotropic gravity waves are roughly 100 times the speed of the next fastest internal wave or advective signal. In their linear form, they travel at speed $\sqrt{gH}$, with H the ocean depth and g the acceleration of gravity. In the deep ocean, they are about 5–10 times slower than acoustic waves.

External gravity waves are nearly two-dimensional in structure, so they are largely represented by dynamics of the vertically integrated fluid column. This property motivates the formulation of primitive equation model algorithms which split the relatively fast vertically integrated dynamics from the slow and more complicated vertically dependent dynamics. Doing so allows for a more efficient time-stepping method to update the ocean's momentum field. Details of this split are often quite complex, and require care to ensure that the overlap between the fast and slow modes is trivial, or else suffer consequences of an unstable simulation. Nonetheless, these algorithms form a fundamental feature in all ocean climate models.

Basics of the Finite Volume Method

The previously derived budgets for infinitesimal fluid parcels represent a starting point for ocean climate model algorithm designs. The next stage is to pose these budgets on a discrete lattice, whereby the continuum fields take on a finite or averaged interpretation. There is actually more than one one way to interpret the relation between the continuum variables and those living on the numerical lattice. We introduce one method here, as it has achieved some recent popularity in the ocean model literature.

The finite volume method takes the flux form continuum equations and integrates them over the finite extent of a discrete model grid cell. The resulting control volume budgets are the basis for establishing an algebraic algorithm amenable to computational methods. For example, vector calculus allows for the parcel budget of a scalar field, such as a tracer concentration described by eqn [5], to be written over an arbitrary finite region as

$$\partial_t\left(\int\int\int \rho C\,\mathrm{d}V\right) = -\int\int \mathrm{d}A_{(\hat{\mathbf{n}})}\hat{\mathbf{n}}\cdot(\mathbf{v}^{\mathrm{rel}}C + \mathbf{J}) \qquad [8]$$

The volume integral is taken over an arbitrary fluid region, and the area integral is taken over the bounding surface to that volume, with outward normal $\hat{\mathbf{n}}$ and area element $\mathrm{d}A_{(\hat{\mathbf{n}})}$. The velocity $\mathbf{v}^{\mathrm{rel}}$ is determined by the relative velocity of the fluid parcel, $\mathbf{v}$, and the moving boundary. As the advective and diffusive fluxes penetrate the boundary, they alter the tracer mass in the region. The budget for the vector linear momentum can also be written in this form, with the addition of body forces from gravity and Coriolis which act over the extent of the volume. Once formulated in this manner, the discretization problem shifts from fundamentals to details, with details differing on how one represents the SGS behavior of the continuum fields. This then leads to the multitude of discretization methods available for such processes as transport, time-stepping, etc.

Elements of Vertical Coordinates

The choice of how to discretize the vertical direction is the most important choice in the design of a numerical ocean model. The reason is that much of the model's algorithms and SGS parametrizations are fundamentally influenced by this choice. We briefly outline here some physical considerations which may prejudice a choice for the vertical coordinate. For this purpose, we identify three regimes of the ocean germane to the considerations of a vertical coordinate.

- *Upper ocean mixed layer.* This is a generally turbulent region dominated by transfers of momentum, heat, freshwater, and tracers with the overlying atmosphere, sea ice, rivers, etc. It is a primary region of importance for climate system modeling. It is typically very well mixed in the vertical through three-dimensional convective and turbulent processes. These processes involve nonhydrostatic physics which requires very high horizontal and vertical resolution (i.e., a vertical to horizontal grid aspect ratio near unity) to explicitly represent. A parametrization of these processes is therefore necessary in primitive equation ocean models which exploit the hydrostatic approximation. In this region, it is essential to employ a vertical coordinate that facilitates the representation and parametrization of these highly turbulent processes. Geopotential and pressure coordinates, or their derivatives, are the most commonly used coordinates as they facilitate the use of very refined vertical grid spacing, which can be essential to simulate the strong exchanges between the ocean and atmosphere, rivers, and ice.

- *Ocean interior.* Tracer transport processes in the ocean interior predominantly occur along neutral directions. The transport is largely dominated by large-scale currents and mesoscale eddy fluctuations. Water mass properties in the interior thus tend to be preserved over large space and timescales (e.g., basin and decade scales). This property of the ocean interior is critical to represent in a numerical simulation of ocean climate. A potential density, or isopycnal coordinate, framework is well suited to this task, whereas geopotential, pressure, and terrain-following models have problems associated with numerical truncation errors. The problem becomes more egregious as the model resolution is refined, due to the enhanced levels of eddy activity that pumps tracer variance to the grid scale. Quasi-adiabatic dissipation of this variance is difficult to maintain in nonisopycnal models.
- *Ocean bottom.* The ocean's bottom topography acts as a strong forcing on the overlying currents and so directly influences dynamical balances. In an unstratified ocean, the flow generally follows lines of constant f/H, where f is the Coriolis parameter and H the ocean depth. Additionally, there are several regions where density-driven currents (overflows) and turbulent bottom boundary layer (BBL) processes act as a strong determinant of water mass characteristics. Many such processes are crucial for the formation of deep-water properties in the World Ocean, and for representing coastal processes in regional models. It is for this reason that terrain-following models have been developed over the past few decades, with their dominant application focused on the coastal and estuarine problem.

The commonly used vertical coordinates introduced above can have difficulties accurately capturing all flow regimes. Unfortunately, each regime is important for accurate simulations of the ocean climate system. To resolve this problem, some researchers have proposed the use of hybrid vertical coordinates built from combinations of the traditional choices. The aim is to employ a particular vertical coordinate only in a regime where it is most suitable, with smooth and well-defined transitions to another coordinate when the regime changes. Research into the required hybrid vertical coordinate methods remains an active area in ocean model design.

Ocean Climate Modeling

The use of physical ocean models to simulate the ocean requires understanding and knowledge beyond the fundamentals discussed thus far in this article. Most notably, we require information about boundary fluxes. There are two basic manners that ocean models are generally used for simulations: as components of an ESM, whereby boundary fluxes are computed from atmosphere, sea ice, and river component models, based on interactions with the evolving ocean; or in a stand-alone mode where boundary fluxes are prescribed from a data set, in which case the uncertainties are huge due to sparse measurements over much of the ocean. This uncertainty in observed fluxes greatly handicaps our ability to unambiguously untangle model errors (i.e., errors in numerical methods, parametrizations, and formulations) from flux errors.

Correspondingly, the process of evaluating the fidelity of ocean simulations remains in its infancy relative to the situation in atmospheric modeling, where synoptic weather forecasts provide a stringent test of model fidelity. Nonetheless, ocean observations, including boundary fluxes, are steadily improving, with new observations providing critical benchmarks for evaluating the relevance of ocean simulations. Given the wide-ranging spatial and temporal scales of oceanic phenomena, it is essential that observations be maintained over a wide network in both space and time. In absence of this network, we are unable to provide a mechanistic understanding of the observed ocean climate system.

See also

Coupled Sea Ice–Ocean Models. Energetics of Ocean Mixing. Heat and Momentum Fluxes at the Sea Surface. Heat Transport and Climate. Mesoscale Eddies. Neutral Surfaces and the Equation of State. Ocean Carbon System, Modeling of. Ocean Circulation: Meridional Overturning Circulation. Wind Driven Circulation.

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