Ocean Modelling I

Ocean Modelling Basics and the CESM Ocean Model

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Outline

1. Ocean properties and (unique) modelling challenges
2. The CESM ocean model  
   a. Governing equations  
   b. Grids  
   c. Finite difference numerics  
   d. Surface boundary conditions  
3. Some model results

(Parameterization of unresolved processes will be covered in following lecture)
Important Ocean Properties

• The heat capacity per volume of the ocean is much larger than the atmosphere. (3m of ocean ≈ entire atmospheric column above). Important reservoir for heat, CO2, & other constituents of the Earth system.

• There is extremely small diapycnal mixing (across density surfaces) once water masses are subducted below the mixed layer \( [K_v = O(10^{-5} \text{ m}^2/\text{s})] \). This is why water masses can be named and followed around the ocean.

• The ocean is a 2 part density fluid (temperature and salinity). Form ice when temperature <-1.8°C & resulting brine rejection increases salinity of adjacent water parcels.

• Once formed, ocean density (heat/salt) anomalies persist ➔ The ocean contains the memory of the climate system... Important implications for climate variability & predictability.

• The density change from top to bottom is much smaller than the atmosphere – 1.02 to 1.04 gr/cm³. This makes the Rossby radius (NH/f) (turbulence scale) much smaller – 10s->100s km.

• Top to bottom “lateral” boundaries ➔ leading order influence of topography on dynamics ➔ ocean gyres & associated heat transport
Ocean Modelling Challenges

Highly irregular domain; land boundary exerts strong control on ocean dynamics
Ocean Modelling Challenges

Highly irregular domain; land boundary exerts strong control on ocean dynamics
Ocean Modelling Challenges

Bathymetry (km), 1/30° ETOPO2
Ocean Modelling Challenges

Bathymetry (km), 1/10° POP (“tx0.1”)
Ocean Modelling Challenges

Bathymetry (km), 1° POP ("gx1v6")

The image shows a map of bathymetry (depth) in kilometers, with color coding representing different depth ranges. The map covers a region from 15°N to 35°N and from 90°W to 70°W.
Ocean Modelling Challenges

Bathymetry (km), 3° POP ("gx3v7")
Ocean Modelling Challenges

Paleoclimate modelling can entail significant changes in ocean domain...
Ocean Modelling Challenges

**LO-RES (3°) O(100+ years/day)**

**WORKHORSE (1°) O(10-100 years/day)**

**HI-RES (0.1°) O(1 year/day)**

Circumference of Earth

\(~4 \times 10^5 \text{ km}\)
Ocean Modelling Challenges

Oceanic deformation radius $O(10-200)$ km $\ll$ Atmospheric $O(1000s)$ km, $\Rightarrow$ significantly higher resolution is needed $O(0.1^\circ)$ to resolve ocean “weather”

1$^{st}$ baroclinic Rossby radius (km) ( $<$ Eddy length scale )

Fig. 6. Global contour map of the $1^\circ \times 1^\circ$ first baroclinic Rossby radius of deformation $\lambda_1$ in kilometers computed by Eq. (2.3) from the first baroclinic gravity-wave phase speed shown in Fig. 2. Water depths shallower than 3500 m are shaded.

Chelton et al., JPO, (1998)
Figure 2. From Smith et al. [2000], showing the first baroclinic Rossby radius, temporally and zonally averaged from their 0.1° North Atlantic model, along with grid spacings of the 0.1° model and the 0.28° model of Maltrud et al. [1998].

Fig. 6. Global contour map of the 1° × 1° first baroclinic Rossby radius of deformation \( \lambda_1 \) in kilometers computed by Eq. (2.3) from the first baroclinic gravity-wave phase speed shown in Fig. 2. Water depths shallower than 3500 m are shaded.
Ocean Modelling Challenges

Fig. 1. The horizontal resolution needed to resolve the first baroclinic deformation radius with two grid points, based on a 1/8° model on a Mercator grid (Adcroft et al., 2010) on Jan. 1 after one year of spinup from climatology. (In the deep ocean the seasonal cycle of the deformation radius is weak, but it can be strong on continental shelves.) This model uses a bipolar Arctic cap north of 65°N. The solid line shows the contour where the deformation radius is resolved with two grid points at 1° and 1/8° resolutions.
At all (present-day) resolutions, OGCMs resolve the mesoscale in some regions but not others.
Ocean Modelling Challenges

Workhorse (1° ≈ 100km) ocean models for climate research cannot reproduce the rich mesoscale eddy field observed in Nature...
Mixing associated with sub-gridscale turbulence must be parameterized
Ocean Modelling Challenges

“The choice of vertical coordinate system is the single most important aspect of an ocean model’s design... Currently, there are three main vertical coordinates in use, none of which provide universal utility.”*

![Diagram of vertical coordinate systems]

Fig. 1. Schematic 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 z-coordinates; the interior is naturally represented using isopycnal ρ-coordinates; and the bottom boundary is naturally represented using terrain following σ-coordinates.

Ocean Modelling Challenges

• Long equilibration timescale $\Rightarrow$ deep ocean will in general be characterized by drift.

\[
\frac{H^2}{K_v} = \frac{(4000 \text{ m})^2}{(10^{-4} \rightarrow 10^{-5} \text{ m}^2/\text{s})} = O (5,000-50,000 \text{ years})
\]

Fig. 2. Horizontal-mean potential temperature difference time series for 1850 CONTROL minus PHC2 observations: (a) global, (b) Pacific, (c) Indian, and (d) Atlantic Oceans. The contour intervals are 0.1°, 0.2°, 0.25°, and 0.25°C in (a),(b),(c),(d), respectively. The shaded regions indicate negative differences. The time series are based on annual-mean fields smoothed using a 10-yr running mean.

Danabasoglu et al., J Climate, (2012)
CESM Ocean Model
Parallel Ocean Program version 2 (POP2)

• POP2 is a level- (z-) coordinate model developed at the Los Alamos National Laboratory (Smith et al., 2010).

• Descendant of the Bryan-Cox-Semtner class of models.

• Solves the 3-D primitive equations in general orthogonal coordinates with the hydrostatic and Boussinesq approximations.

• A linearized, implicit free-surface formulation is used for the barotropic mode (Dukowicz & Smith, 1994).

• Surface freshwater fluxes are treated as virtual salt fluxes, using a constant reference salinity ➔ net ocean volume remains constant (but not ocean mass).
Useful Resources

CESM1.1: PARALLEL OCEAN PROGRAM (POP2)

INTRODUCTION
The ocean component of the CESM1.1 is the Parallel Ocean Program version 2 (POP2). This model is based on the POP version 2.1 of the Los Alamos National Laboratory; however, it includes many physical and software developments incorporated by the members of the Ocean Model Working Group (see the notable improvements page for these developments).

DOCUMENTATION

- The Parallel Ocean Program (POP) Reference Manual (Los Alamos National Laboratory, LAUR-10-01853)
- Ocean Ecosystem Model Scientific Reference
- CESM1.1 POP2 User Guide
- CESM1.1 Ocean Ecosystem Model User Guide
- CESM1.1 POP2 FAQ

POP2 PORT VALIDATION AND MODEL VERIFICATION
Before running any experiments with CESM1.1 on a local machine, the user should make sure the POP2 code has ported to their machine properly and subsequently verify the POP2 model output.
The CCSM4 Ocean Component

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Model Equations

compressible fluid dynamics (Navier-Stokes) ➔
Boussinesq equations

Hierarchy of dynamical approximation

 Primitive equations

Balance equations

Planetary geostrophic equations

Quasigeostrophic equations

Model Equations

compressible fluid dynamics (Navier-Stokes)
\[ \rho = \rho_o + \delta \rho \]
\[ \frac{\delta \rho}{\rho} \ll 1 \]
Boussinesq equations

\[ \nabla \cdot \mathbf{v} = 0 \]
(non-divergent flow)

\[ \delta \rho \text{ is very small in the ocean, so ignore } \delta \rho \text{ except in gravitational force and equation of state.} \]

Mass continuity equation becomes

\[ \nabla_{3D} \cdot \mathbf{v} = 0 \]

Planetary geostrophic equations
Quasigeostrophic equations

Model Equations

compressible fluid dynamics (Navier-Stokes)

\[ \rho = \rho_o + \delta \rho \]
\[ \frac{\delta \rho}{\rho} \ll 1 \]

Boussinesq equations

\[ \frac{\partial p}{\partial z} = -g \rho \]

Primitive equations

Balance equations

Planetary geostrophic equations

Quasigeostrophic equations

Invoke hydrostatic approximation to simplify the vertical momentum equation (also, shallow-fluid approx)

\[ \Rightarrow \text{vertical velocity (w) is computed diagnostically from continuity eqn., rather than prognostically} \]

NOTE: There should be vertical acceleration when ocean becomes statically unstable \((\rho_z > 0)\), but w tendency has been excluded by the hydrostatic assumption. Therefore, vertical mixing must be parameterized by prognostic computation of vertical diffusivity (very large for an unstable column).

Model Equations

compressible fluid dynamics (Navier-Stokes)

\[ \rho = \rho_o + \delta \rho \]
\[ \delta \rho / \rho \ll 1 \]

Boussinesq equations

\[ \frac{\partial \rho}{\partial z} = -g \rho \]

Primitive equations

Balance equations

Planetary geostrophic equations

Quasigeostrophic equations

7 equations in 7 unknowns:

- 3 velocity components
- potential temperature
- salinity
- density
- pressure

Plus: 1 equation for each additional passive tracer (e.g. CFCs, Ideal Age)

Model Equations

3-D primitive equations in spherical polar coordinates with vertical z-coordinate for a thin, stratified fluid using hydrostatic & Boussinesq approx (Smith et al. 2010):

1. **Momentum equations:**
   \[
   \frac{\partial}{\partial t} \left( u + L(u) \right) - (uv \tan \phi)/a - fv = -\frac{1}{\rho_0 a \cos \phi} \frac{\partial p}{\partial \lambda} + F_{Hx}(u,v) + F_V(u)
   \]
   \[
   \frac{\partial}{\partial t} v + L(v) + (u^2 \tan \phi)/a + fu = -\frac{1}{\rho_0 a \tan \phi} \frac{\partial p}{\partial \phi} + F_{Hy}(u,v) + F_V(v)
   \]

2. **Continuity equation:**
   \[
   L(1) = 0
   \]

3. **Hydrostatic equation:**
   \[
   \frac{\partial p}{\partial z} = -\rho g
   \]

4. **Equation of state:**
   \[
   \rho = \rho(\Theta, S, p) \rightarrow \rho(\Theta, S, z)
   \]

5. **Tracer transport:**
   \[
   \frac{\partial}{\partial t} \phi + L(\phi) = D_H(\phi) + D_V(\phi)
   \]
   \[
   D_H(\phi) = A_H \nabla^2 \phi
   \]
   \[
   D_V(\phi) = \frac{\partial}{\partial z} \kappa \frac{\partial}{\partial z} \phi
   \]
CESM Ocean Model Grids

Horizontal discretization is done in generalized spherical coordinates to avoid N. Pole singularity:

“orthogonal curvilinear grid with displaced pole”

gx1v6: climate workhorse nominal 1°
gx3v7: testing, paleo apps nominal 3°

Equatorial refinement (0.3° / 0.9°)
CESM Ocean Model Grids

tripole mesh

tx0.1: “eddy-resolving” nominal 0.1°
POP Spatial Discretization

- Tracers (T, S, ρ, ψ) @ “T-points”
- Horizontal velocity (u,v) @ “U-points”
- Vertical velocity (w)
- No-slip, no normal flow b.c.’s

- Quadrilateral horizontal mesh (“Arakawa B-grid”)
- Note relative positions of T(i,j,k); u,v(i,j,k); w(i,j,k)

- Finite difference numerics:
  (see POP Ref Manual for details)

\[
\begin{align*}
\delta_x \psi &= \left[ \psi \left( x + \Delta_x / 2 \right) - \psi \left( x - \Delta_x / 2 \right) \right] / \Delta_x \quad (3.4) \\
\psi^x &= \left[ \psi \left( x + \Delta_x / 2 \right) + \psi \left( x - \Delta_x / 2 \right) \right] / 2 , \quad (3.5) \\
\nabla \psi &= \hat{x} \delta_x \psi^y + \hat{y} \delta_y \psi^x \\
(3.6) \\
\nabla \cdot \mathbf{u} &= \frac{1}{\Delta_y} \delta_x \Delta_y u_x^y + \frac{1}{\Delta_x} \delta_y \Delta_x u_x^y \\
(3.7) \\
\hat{z} \cdot \nabla \times \mathbf{u} &= \frac{1}{\Delta_y} \delta_x \Delta_y u_y^y - \frac{1}{\Delta_x} \delta_y \Delta_x u_x^x \\
(3.8) \\
\nabla \cdot G \nabla \psi &= \frac{1}{\Delta_y} \delta_x \left[ \Delta_y G \delta_x \psi^y \right]^y + \frac{1}{\Delta_x} \delta_y \left[ \Delta_x G \delta_y \psi^x \right]^x . \quad (3.9)
\end{align*}
\]
• At least 2 adjacent active ocean T-cells are required for flow through channels

T=tracer grid, U=velocity grid
POP Spatial Discretization

T = tracer grid, U = velocity grid

Ocean bottom
POP Vertical Discretization

- Fixed z-levels, with non-uniform $\Delta z$
- Enhanced vertical resolution in surface diabatic layer ($\Delta z=10m$ at sfc)
- 60-lvl for gx1v6/gx3v7; 62-lvl for tx0.1

![Diagram showing vertical discretization for CCSM3 and CCSM4, with 60-level CCSM4 and 62-lvl for tx0.1]
MPAS-Ocean model grids (CESM3.0?)

**Horizontal:**
- unstructured
- quasi-uniform or variable resolution
- Voronoi Tesselations
- 4, 5, or 6-sided cells

**Vertical:** Arbitrary Lagrangian-Eulerian (ALE): z-level, z-star, sigma, isopycnal

Figures courtesy of Mark Petersen (LANL)

POP numerics in a nutshell

\[
\mathcal{L}(\alpha) = \frac{1}{a \cos \phi} \left[ \frac{\partial}{\partial \lambda} (u \alpha) + \frac{\partial}{\partial \phi} (\cos \phi v \alpha) \right] + \frac{\partial}{\partial z} (w \alpha) \quad (2.3)
\]

advection operator in analytic form

Finite difference advection

- Momentum: centered differencing (2\textsuperscript{nd} order)
  \[
  \mathcal{L}_U(\alpha) = \frac{1}{\Delta y} \delta_x \left[ (\Delta y u_x^y)^{xy} \alpha^x \right] + \frac{1}{\Delta x} \delta_y \left[ (\Delta x u_y^x)^{xy} \alpha^y \right] + \delta_z (w U \alpha^z) \quad (3.23)
  \]

- Tracers: upwind3 scheme (3\textsuperscript{rd} order)
  - Operator stencil is a function of \(v=(u,v,w)\)
  - Complex form (see POP Ref Manual)
  - Stronger conservation & monotonicity requirements
  - Other alternatives for tracers (e.g., flux-limited Lax-Wendroff scheme), but more expensive
Time Discretization

- 3-time-level modified leapfrog scheme (2nd order)
- Occasional averaging timestep to suppress the computational mode associated with decoupled even/odd timestep solutions
- For tracer X:

\[
\frac{X_{t+1} - X_{t-1}}{2\Delta t} = L^t(X^t) + D_H(X_{t-1}^t) + D_v^t(X_{t+1}^t)
\]

(Implicit vertical mixing)
POP numerics in a nutshell

Time Discretization

Barotropic/Baroclinic split

• \( U = <U> + U', \) where \(<U>\) is depth-average (barotropic mode)
• Explicitly resolving fast barotropic gravity waves (\( \sqrt{gH} \approx 200 \text{ m/s} \)) would place severe limitations on model timestep due to Courant-Friedrichs-Lewy (CFL) stability condition: \( u(\Delta t/\Delta x) \leq 1 \)
• Therefore, barotropic gravity waves are filtered out by solving for \(<U>\) as a separate 2D system using implicit free-surface formulation with barotropic timestep = (much longer) baroclinic timestep.
• Explicitly solve for \( U' \) from momentum eqns without surface pressure gradient

\( \rightarrow \Delta t \approx 1 \text{ hour in } 1^\circ \text{ POP} \)

Refer to POP reference manual for further details on numerics!
POP surface forcing

- Ocean model forcing = fluxes of momentum, heat, and freshwater, (...) and other tracers) applied as surface boundary conditions to vertical mixing terms:

\[ F_v(\alpha) = \frac{\partial}{\partial z} \mu \frac{\partial}{\partial z} \alpha \]

momentum flux

\[ D_v(\varphi) = \frac{\partial}{\partial z} \kappa \frac{\partial}{\partial z} \varphi \]

tracer flux

- "Flux boundary conditions" at the surface (z=0):

\[
\begin{bmatrix}
\mu \frac{\partial}{\partial z} \tilde{\mathbf{u}} \\
\end{bmatrix}
|_{z=0} = \frac{\bar{\tau}}{\rho_o}
\]

\[
\begin{bmatrix}
\kappa \frac{\partial T}{\partial z} \\
\end{bmatrix}
|_{z=0} = \frac{Q_{net}}{\rho_o C_p}
\]

\[
\begin{bmatrix}
\kappa \frac{\partial S}{\partial z} \\
\end{bmatrix}
|_{z=0} = \frac{F_{net}}{\rho_o} S_o
\]

(see Barnier, 1998)

- Wind stress vector: \( \bar{\tau} \)

- Net heat flux: \( Q_{net} = Q_S + Q_L + Q_E + Q_H + Q_P + Q_{oi} \)

- Net freshwater flux: \( F_{net} = P + E + R + F_{oi} \)
POP surface forcing

- Bulk formulae parameterize the turbulent fluxes in terms of the near surface atmospheric state \((U, q, \theta)\) with a feedback of the surface ocean state \((U_o, SST)\) onto the fluxes:

\[
\tau_{as} = \rho C_D |\Delta \tilde{U}| \Delta \tilde{U} \tag{3a}
\]

\[
E = \rho C_E (q - q_{sat}(SST))|\Delta \tilde{U}| \tag{3b}
\]

\[
Q_E = \Lambda_v E \tag{3c}
\]

\[
Q_H = \rho c_p C_H (\theta - SST)|\Delta \tilde{U}|, \tag{3d}
\]

(see Large & Yeager, 2009)
POP surface forcing

- Fully coupled mode (B compset): active atmospheric model

- Forced ocean (C compset) or ocean_sea-ice (G compset): data atmosphere
  - Generally use CORE atmospheric state fields for surface b.c.’s
  - http://data1.gfdl.noaa.gov/nomads/forms/core.html
  - Interannual (1948-2009) as well as Normal Year Forcing (NYF) are available

- Default is for POP to “couple” to surface b.c.’s once per day

- Useful References:
  - Large & Yeager, 2004: Diurnal to decadal global forcing for ocean and sea-ice models: the data sets and climatologies, NCAR Tech Note TN-460.

Quality of POP model solution is strongly tied to quality of surface b.c.’s!
Need to parameterize the diurnal cycle of (shortwave) radiative heat flux (i.e., night & day). This is done with a zenith-angle dependent SW(lat, lon, hour, day of year) heat flux parameterization.
The SW diurnal cycle results in dramatically improved equatorial SST
SST and Salinity Differences from Observations

**SST**

- **CCSM3**
  - Mean = -0.615, RMS = 1.355

- **CCSM4**
  - Mean = -0.304, RMS = 1.139

**SSS**

- **CCSM3**
  - Mean = -0.365, RMS = 0.879

- **CCSM4**
  - Mean = 0.036, RMS = 0.583

- **Forced ocean-ice**
  - Mean = 0.066, RMS = 0.406

**Coupled Models**

- **CCSM3**
- **CCSM4**
First, a $2^\circ$ atmosphere and land, $0.5^\circ$ ocean and sea ice. The 1870 Control run was integrated for 260 years, which had a good top of the atmosphere balance of $-0.12 \text{ W/m}^2$. The run was branched from year 123 of the Control run, and was integrated from 1870 to 2030. The greenhouse gas forcing was taken from observations between 1870 and 2000, and then followed the Special Report on Emissions Scenarios A1B future scenario. Additional forcings are the levels of dust, sea salt, and carbonaceous and sulphate aerosols. These aerosol levels are based on a historical reconstruction run using the CCSM chemistry component, and then projected forward for the period 2000–2030. The solar forcing was held constant at $1,365 \text{ W/m}^2$, and no volcanic forcing was applied to the run. The initial condition for the $0.5^\circ$ run was taken from 1 January 1980 of the $2^\circ$ run. The atmosphere and land fields were interpolated onto the $0.5^\circ$ grid, and the ocean and sea ice fields were used without modification. The $0.5^\circ$ run was integrated from 1980 to 2030, and was forced in exactly the same way as the $2^\circ$ run. The factor of 16 times more grid points in the atmosphere and land components, but the same number in the ocean and sea ice components, means the $0.5^\circ$ run takes about 12 times the computational resource of the $2^\circ$ run.

3 Comparison of mean climates
This section will compare several aspects of the $2^\circ$ and $0.5^\circ$ run climates averaged between the beginning of 1985 and 2000, and observations representing the end of the twentieth century. The rationale for averaging between 1985 and 2000 is the following. The $0.5^\circ$ run will obviously have a period of adjustment to its own climatology from that of the $2^\circ$ run. Figure 2 shows the globally averaged ocean temperature between the surface and 203 m depth between 1980 and 2000 from the two runs. It shows that most of the adjustment in the upper 200 m of the ocean occurs in the first 5 years, and the adjustment to the new $0.5^\circ$ climatology is almost complete after 10 years. Adjustment to the new Arctic Ocean sea ice thickness distribution in the $0.5^\circ$ run also takes 5–10 years, and the adjustment in the atmosphere and land components is faster than this. Thus, the choice was made to start the averaging in 1985, rather than in 1990, in order to have a longer averaging period, given that most of the upper ocean adjustment had occurred by then.

3.1 The upper ocean simulation
Figure 3 shows the difference between the SST in the two runs and a climatology from (Levitus et al. 1998) data and (a) (b) (c) (d) Gent et al., Clim Dyn, 2010
Interannual SST variability simulated by CORE-II POP

- CORE-forced ocean-ice hindcast simulation with 1° POP yields good reproduction of observed SST variability over late 20th century
subsurface observations, and only a weak restoring of model surface salinity to observed climatology is employed. The CORE forcing dataset, which has been adopted by the CLIVAR Working Group on Ocean Model Development for model intercomparisons, imparts realistic surface variability on a range of relevant time scales (Large and Yeager 2004; Large and Yeager 2009), and the resulting simulated upper-ocean variability shows good agreement with a variety of in situ observations (e.g., North Atlantic upper-ocean heat content in Fig. 1). The 240-yr CORE-IA integration is spun up through four consecutive 60-yr cycles of 1948–2007 forcing. The ocean and sea ice models in DP experiments are initialized with 1 January restart files for a particular year from the last (fourth) cycle of the CORE-IA simulation. No attempt is made to initialize the atmosphere and land models to historical states. Instead, the initial conditions for these component models are taken from corresponding years of a six-member ensemble of twentieth-century (20C) runs. Specifically, the 10-member DP ensembles are generated by randomly selecting atmosphere and land initial conditions from different 20C runs and/or from different days in the month of January. The reader is referred to Gent et al. (2011) and Meehl et al. (2012) for complete descriptions of the CCSM4 twentieth- and twenty-first-century control simulations and forcing details. We refer to all coupled experiments initialized from CORE-IA, whether of past or future time periods, as decadal prediction experiments.

The DP experiments differ from 20C runs (and their future scenario extensions) in terms of initialization procedure and length of integration but are otherwise subject to the same external forcings of solar irradiance, greenhouse gases, aerosols, and volcanic activity. The forcings used are identical to those used in 20C experiments through 2005, and thereafter, the RCP 4.5 future scenario projection.

Fig. 1. Pentad-mean heat content anomalies expressed as the 275-m depth-averaged temperature anomaly relative to 1957–90 climatology from (a)–(d) Ishii and Kimoto (2009), (e)–(h) Levitus et al. (2009), and (i)–(l) CORE-IA. The boxes in each panel demarcate the SPG (50°–10°W, 50°–60°N) and STG (70°–30°W, 32°–42°N) regions.
Coupled decadal prediction of North Atlantic SST

Yeager et al., 2014, in prep.
For even more info...

Books:


Review Papers:


Questions?
Model Biases

Mixed Layer Depth

a) OBS

b) CCSM4 - OBS

c) OCN - OBS

d) CCSM3 - OBS
Model Biases

SST Differences from Observations

2° atmosphere

mean = 0.63°C
rms = 1.44°C

1° atmosphere

mean = -0.01°C
rms = 1.07°C

Obs: Levitus et al. (1998), Steele et al. (2001)
Helpful Guides

http://www.cesm.ucar.edu/models/cesm1.2/pop2/

CESM Webpage for POP

- POP2 User Guide
- Ocean Ecosystem Model User Guide
- POP Reference Manual
- Ocean Ecosystem Reference Manual
Friday’s breakout session

Sea-ice, Ocean, and Land-ice

- Create and run a low-resolution ice-ocean
- Change the namelists
  - turn off the overflow parameterization
  - change snow and sea ice albedo
- Advanced exercises: changing wind stress forcing within the source code
- Data Analysis using nco commands and ncview
Central Advection Discretization

\[
\text{ADV}_{i,j,k} = - \left( u_E T^*_{E} - u_W T^*_{W} \right)/DXT - \left( v_N T^*_{N} - v_S T^*_{S} \right)/DYT - \left( w_k T^*_{T} - w_{k+1} T^*_{B} \right)/dz
\]

\[
u_E(i) = (u_{i,j} DYU_{i,j} + u_{i,j-1} DYU_{i,j-1})/(2DXT_{i,j})
\]

\[
u_W(i) = u_E(i - 1)
\]

\[
v_N(j) = (v_{i,j} DXU_{i,j} + v_{i-1,j} DXU_{i-1,j})/(2DXT_{i,j})
\]

\[
v_S(j) = (v_{i,j-1} DXU_{i,j} + v_{i-1,j-1} DXU_{i-1,j-1})/(2DXT)
\]

\[
T^*_{E} = \frac{1}{2} \ast (T_{i+1,j} + T_{i,j})
\]
Baroclinic & Barotropic Flow

• Issue: Courant-Friedrichs-Lewy (CFL) stability condition associated with fast surface gravity waves.
  • $u(\Delta t/\Delta x) \leq 1$
  • Barotropic mode $\sqrt{gH} \sim 200$ m/s
• Split flow into depth averaged barotropic $\langle U \rangle$ plus vertically varying baroclinic ($U'$)
• Fast moving gravity waves are filtered out, but that’s okay because they don’t impact climate
Barotropic and Baroclinic Flow

\[ U = \langle U \rangle + U' \]

- \( \langle U \rangle \): Implicit, linearized free-surface formulation obtained by combining the vertically integrated momentum and continuity equations

- \( U' \): use a leapfrog time stepping to solve

\[ \frac{X^{t+1} - X^{t-1}}{2 \Delta t} = D^{t-1} + \text{ADV}^t + \text{SRC}^{t,t-1} \]

- Occasional time averaging to eliminate the split mode