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 <u>much</u> larger than the atmosphere. (3m of ocean ≈ entire atmospheric column above). Important reservoir for heat, CO_2, & other constituents of the Earth system.
- There is extremely small diapycnal mixing (across density surfaces) once water masses are subducted below the mixed layer [$K_v = O(10^{-5}\ m^2/s)$]. 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/cm^3. 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

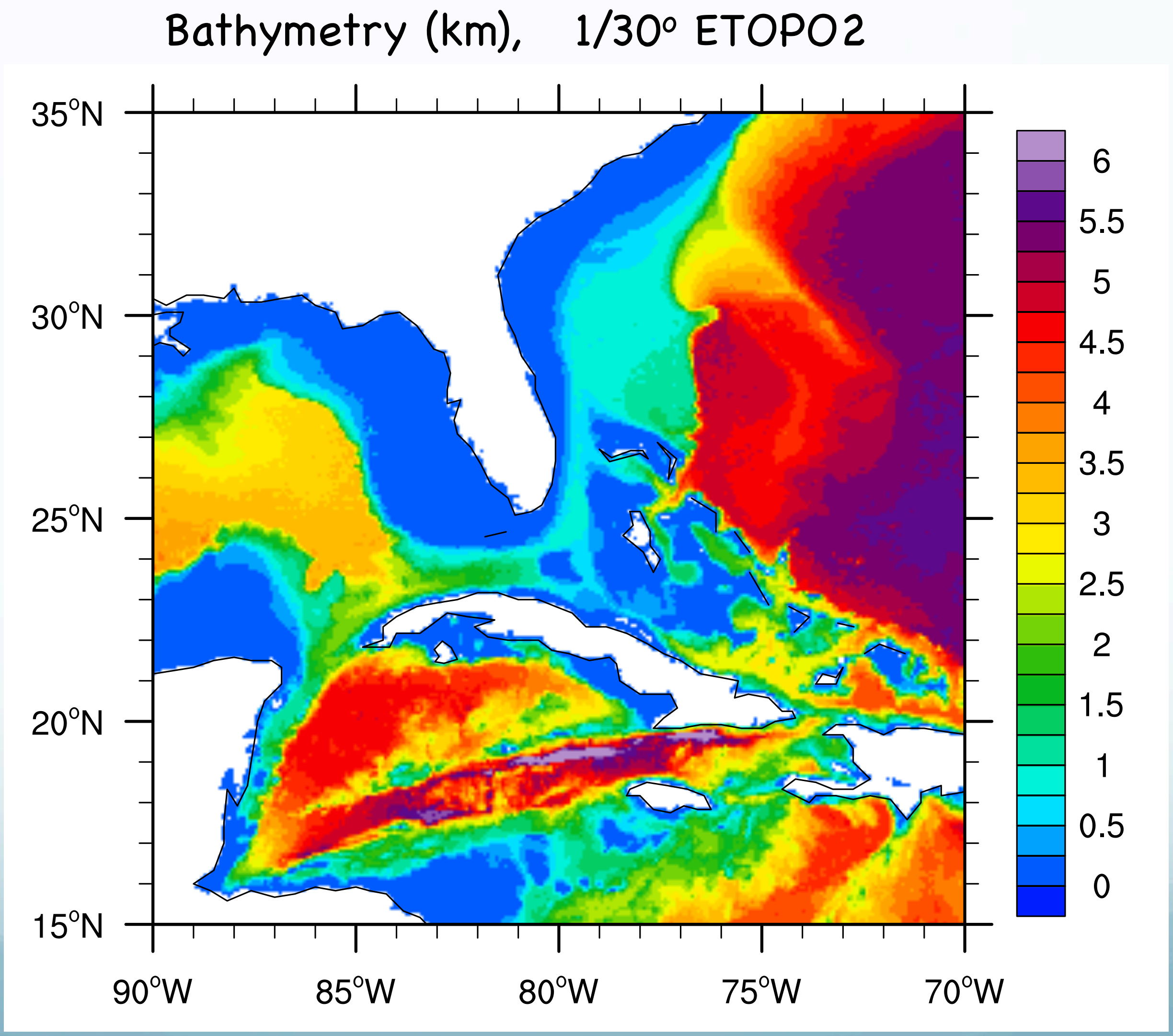
Ocean Modelling Challenges

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

Ocean Modelling Challenges

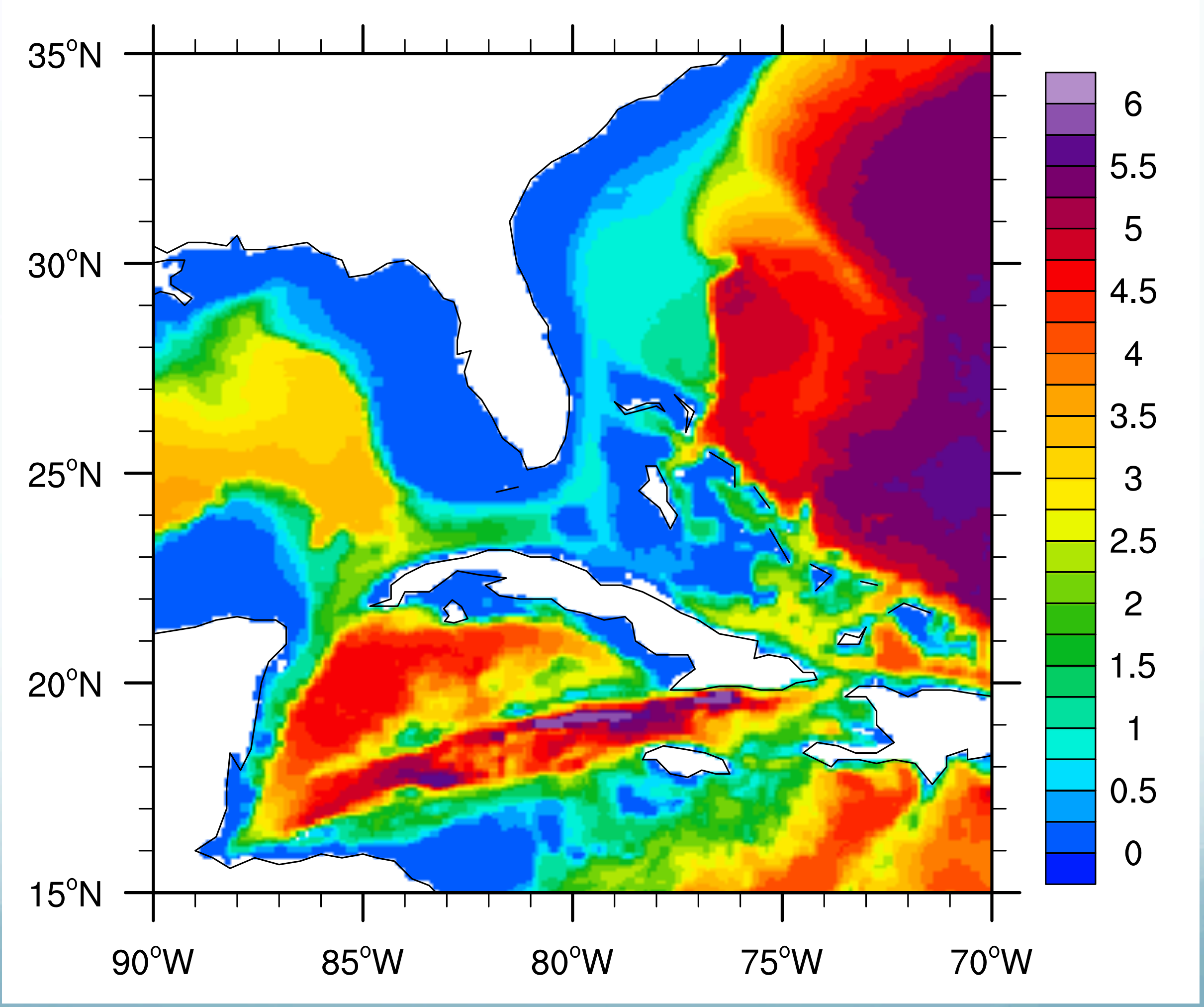
Highly irregular domain; land boundary exerts strong control on ocean dynamics

Ocean Modelling Challenges



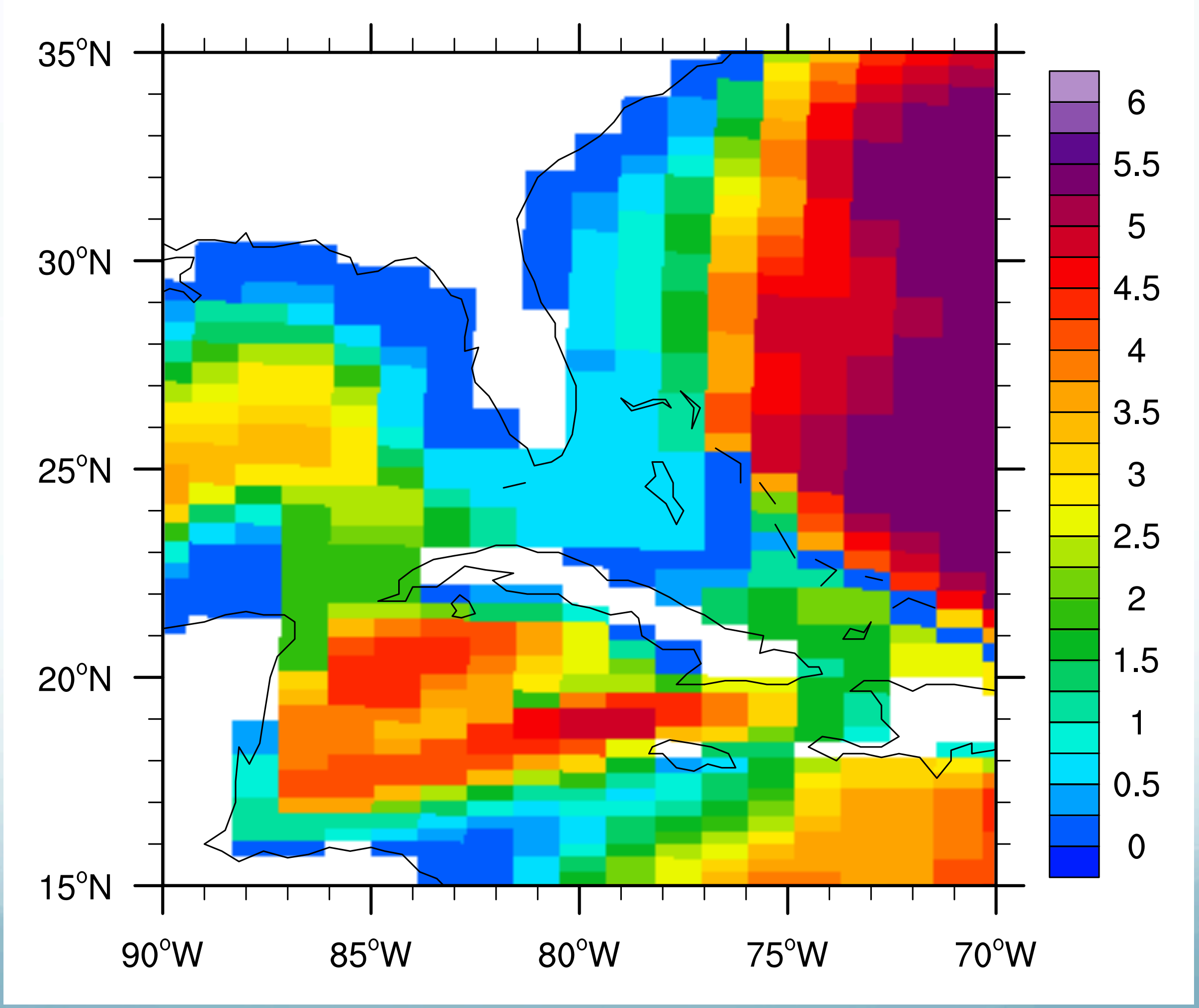
Ocean Modelling Challenges

Bathymetry (km), 1/10° POP ("tx0.1")



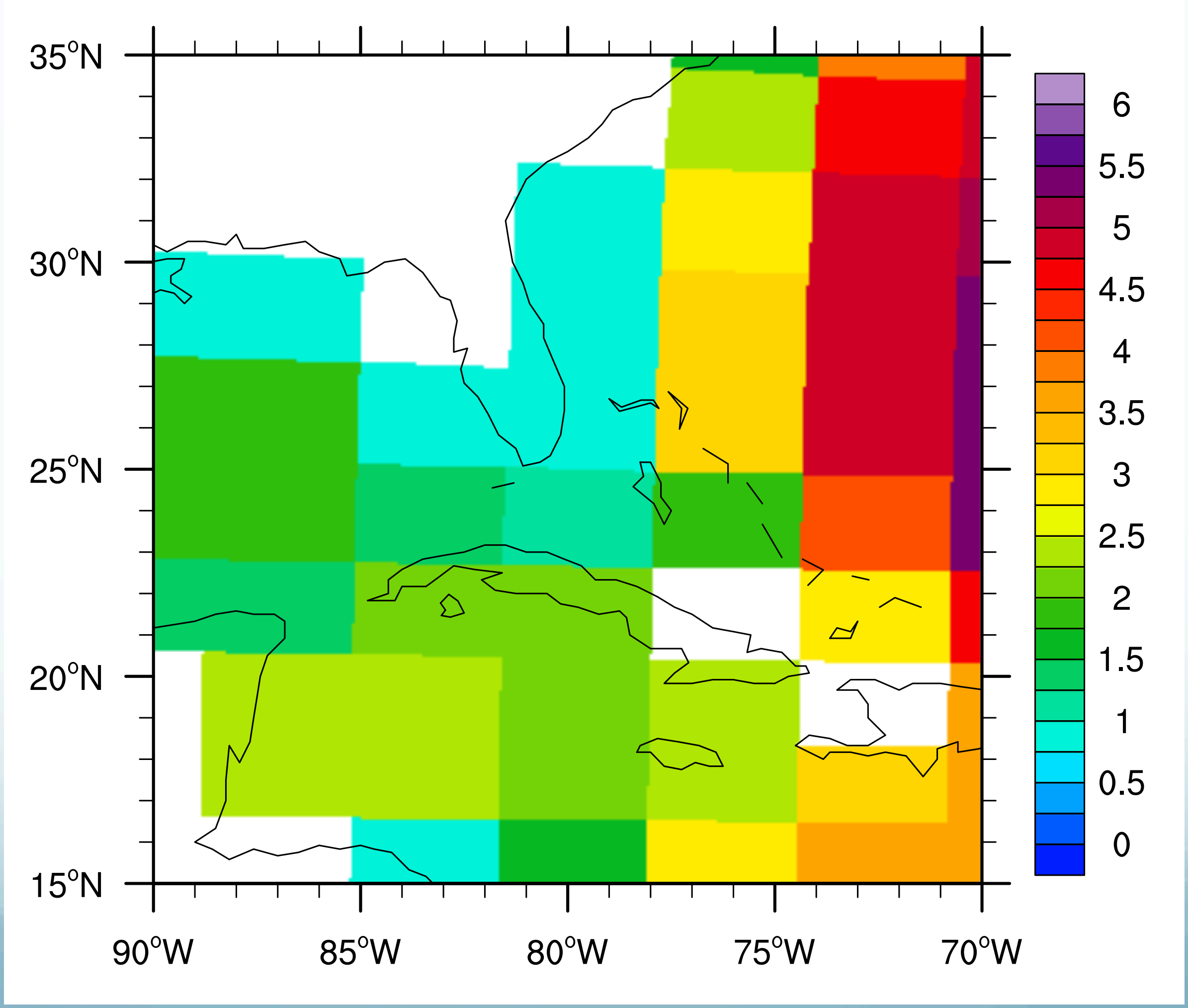
Ocean Modelling Challenges

Bathymetry (km), 1° POP ("gx1v6")



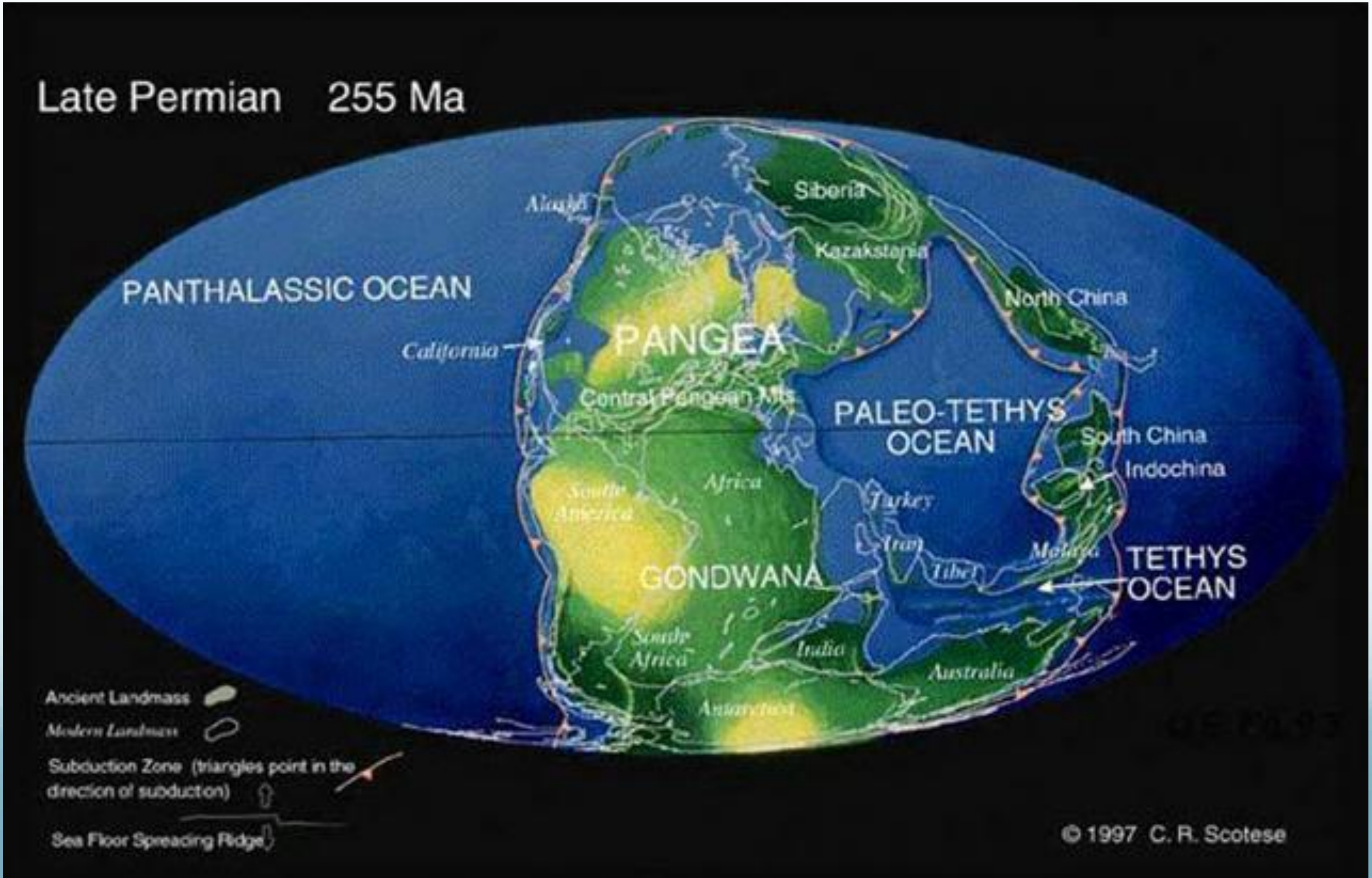
Ocean Modelling Challenges

Bathymetry (km), 3° POP ("gx3v7")



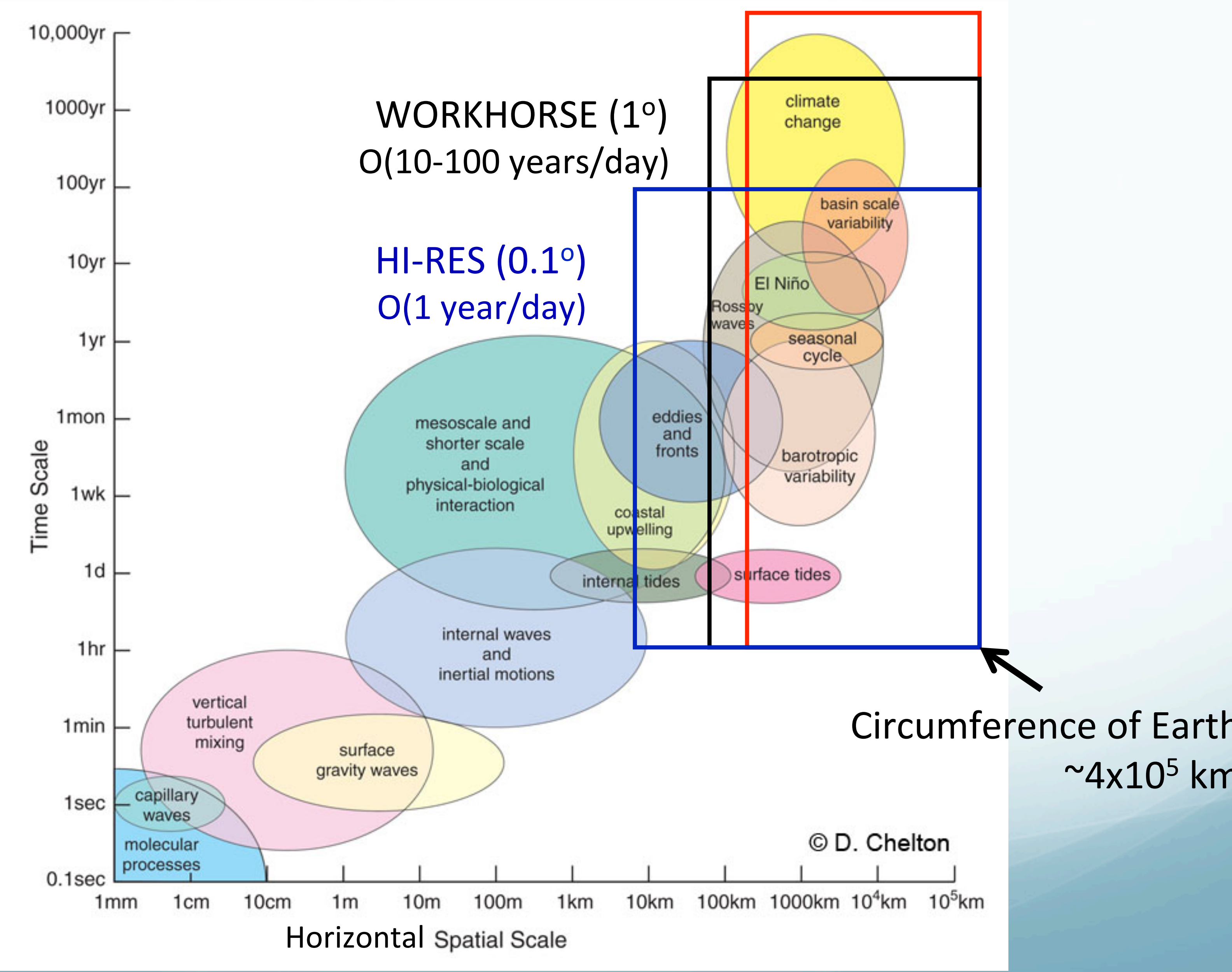
Ocean Modelling Challenges

Paleoclimate modelling can entail significant changes in ocean domain…



Ocean Modelling Challenges

LO-RES (3°)
O(100+ years/day)



Ocean Modelling Challenges

Oceanic deformation radius O(10-200) km << Atmospheric O(1000s) km,
➔ significantly higher resolution is needed O(0.1°) to resolve ocean "weather"

1st baroclinic Rossby radius (km) (< Eddy length scale)

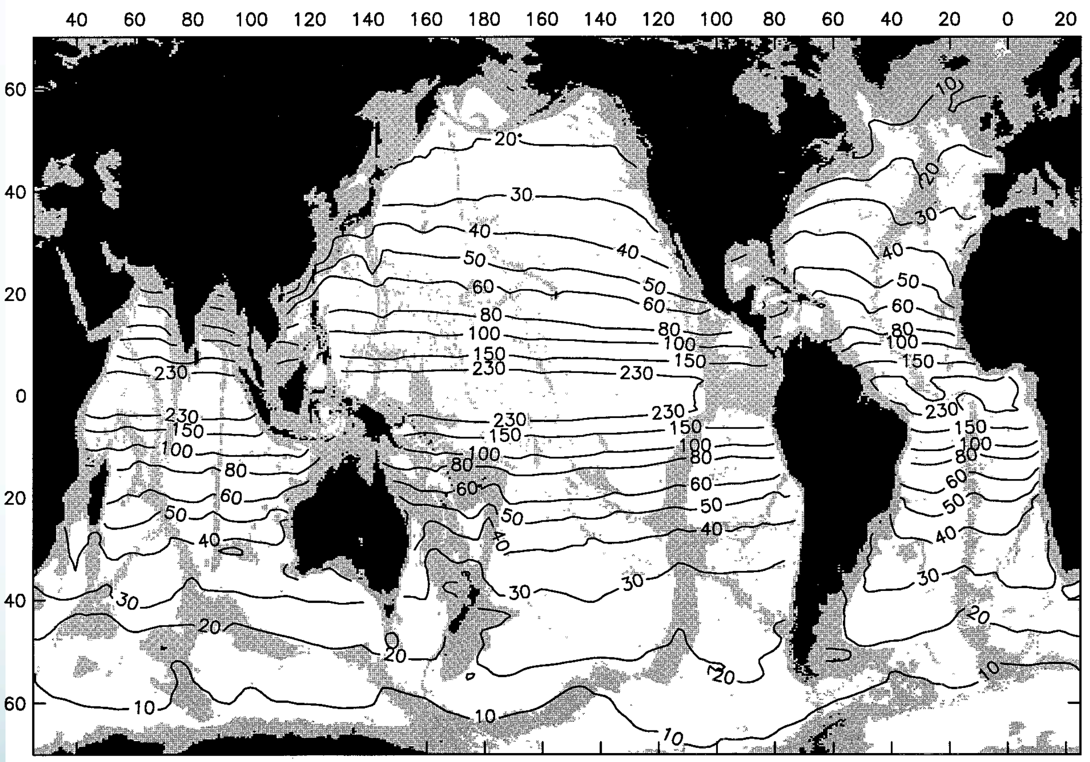


FIG. 6. Global contour map of the 1° × 1° first baroclinic Rossby radius of deformation λ_1 in kilometers computed by Eq. (2.3) from the first baroclinic gravity-wave phase speed shown in Fig. 2. Water depths shallower than 3500 m are shaded.

Chelton et al., JPO, (1998)

Ocean Modelling Challenges

spheric O(1000s) km,
ɔ resolve ocean "weather"

us (km) (< Eddy length scale)

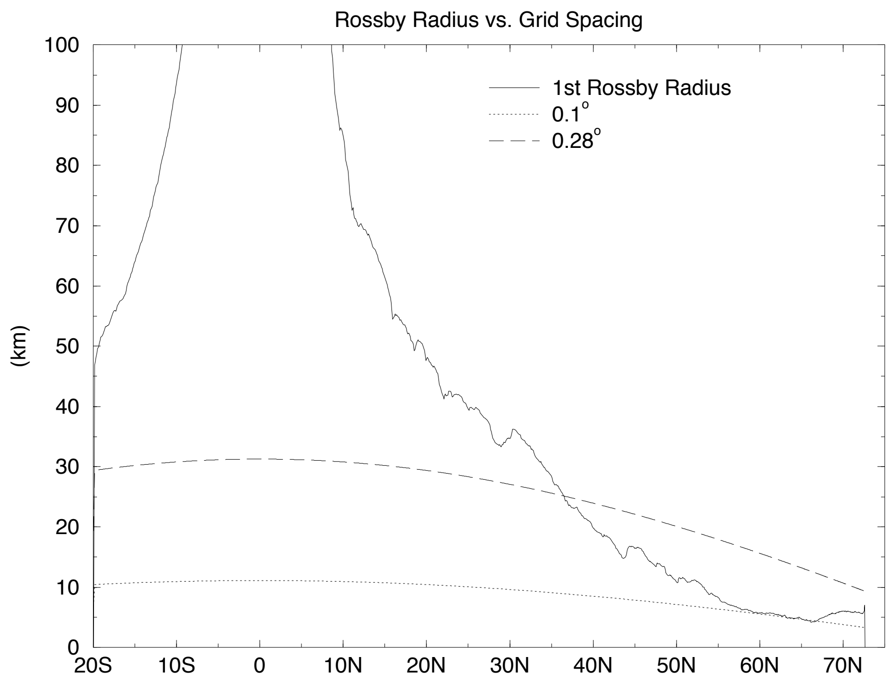
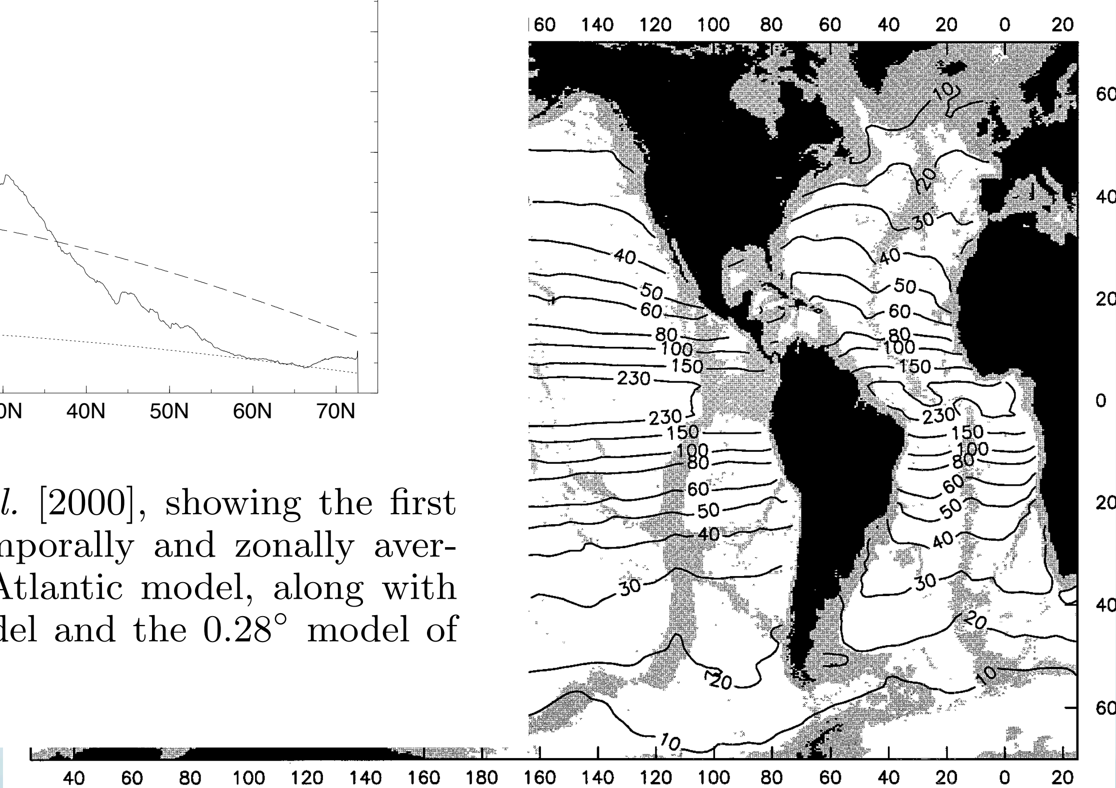


Figure 2. From *Smith et al.* [2000], showing the first baroclinic Rossby radius, temporally and zonally averaged from their 0.1° North Atlantic model, along with grid spacings of the 0.1° model and the 0.28° model of *Maltrud et al.* [1998].



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Chelton et al., JPO, (1998)

Ocean Modelling Challenges

R. Hallberg / Ocean Modelling 72 (2013) 92–103 93

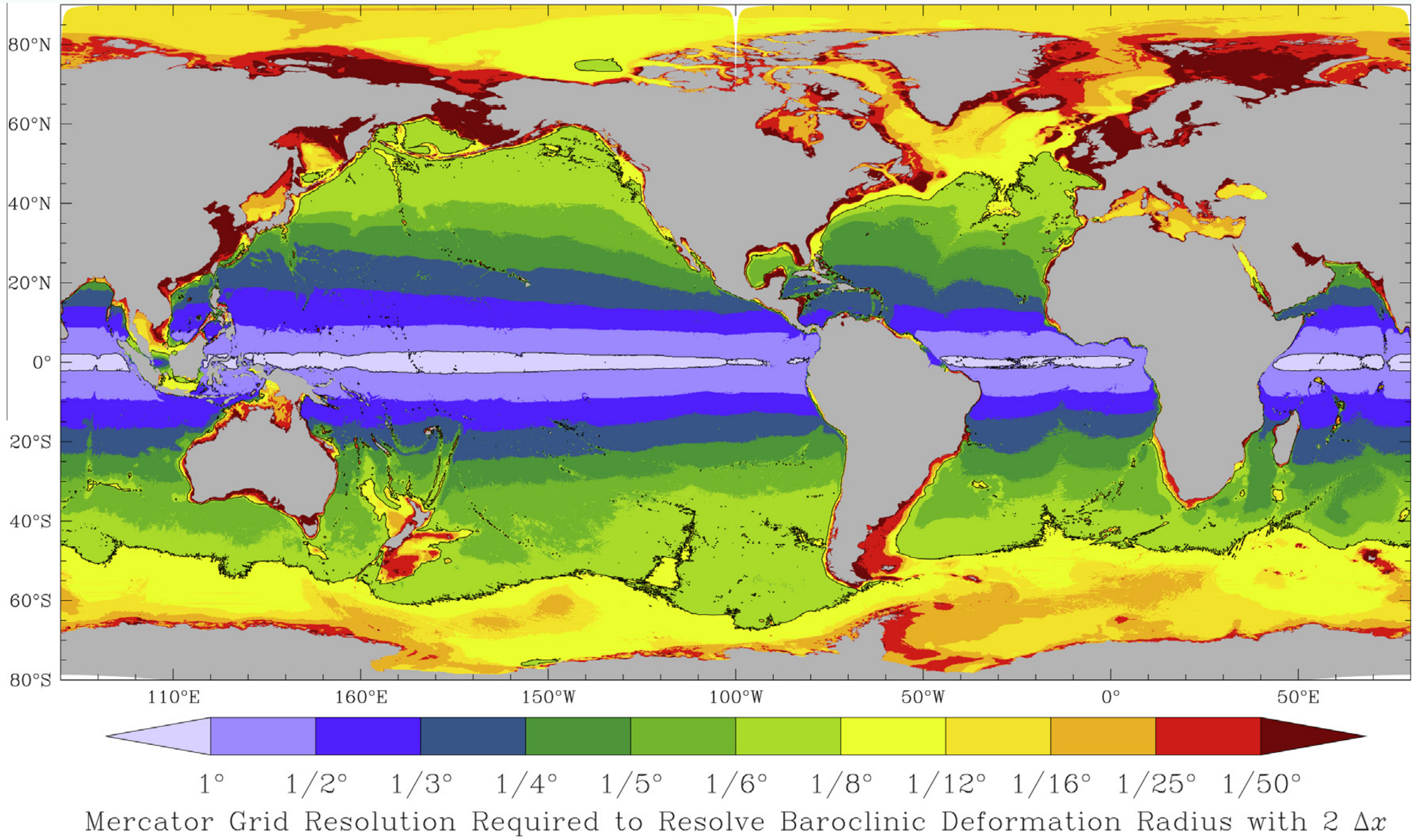
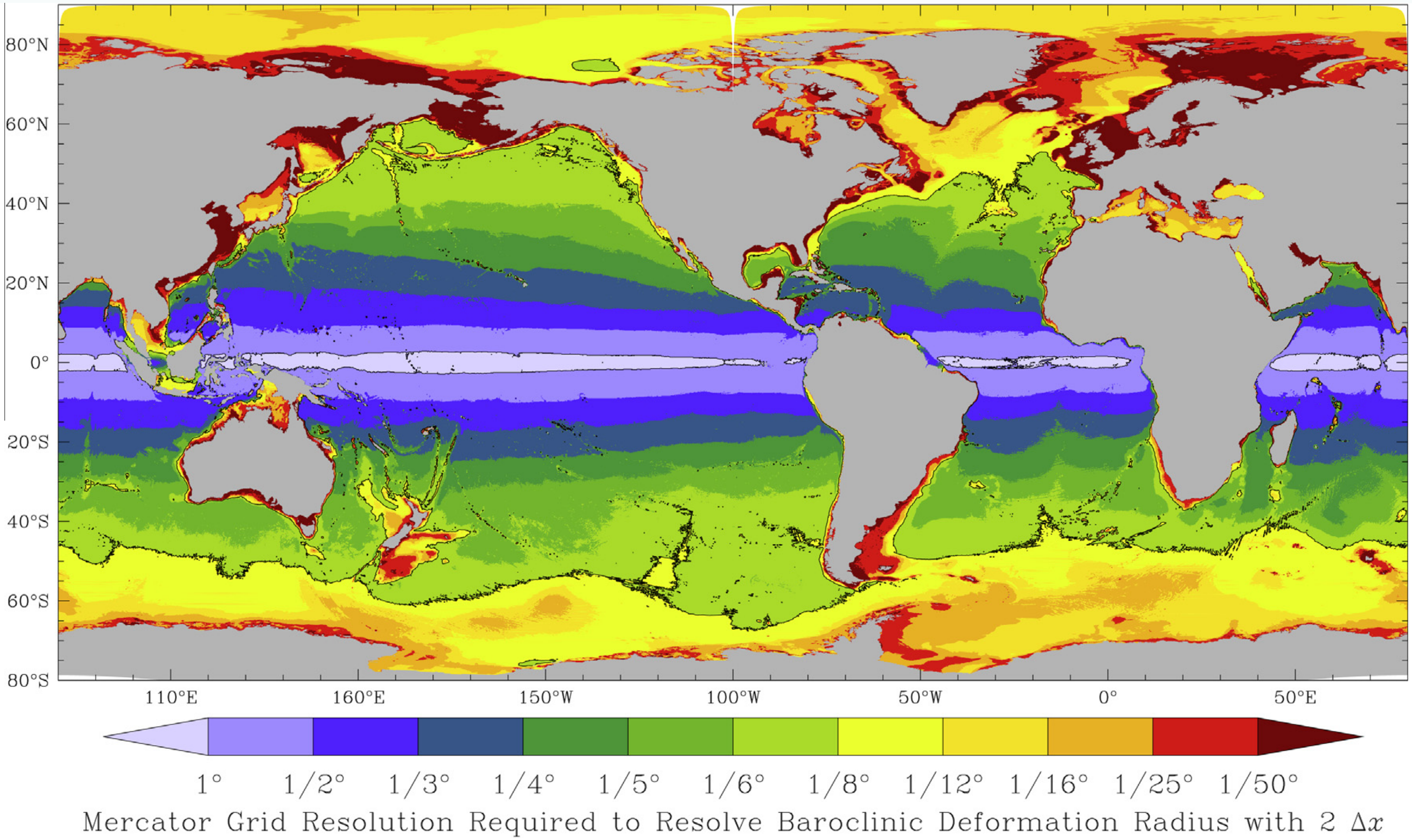


Fig. 1. The horizontal resolution needed to resolve the first baroclinic deformation radius with two grid points, based on a 1/8° model on a Mercator grid (Adcroft et al., 2010) on Jan. 1 after one year of spinup from climatology. (In the deep ocean the seasonal cycle of the deformation radius is weak, 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.

Hallberg, *Ocean Modelling*, (2013)

Ocean Modelling Challenges

R. Hallberg / Ocean Modelling 72 (2013) 92–103



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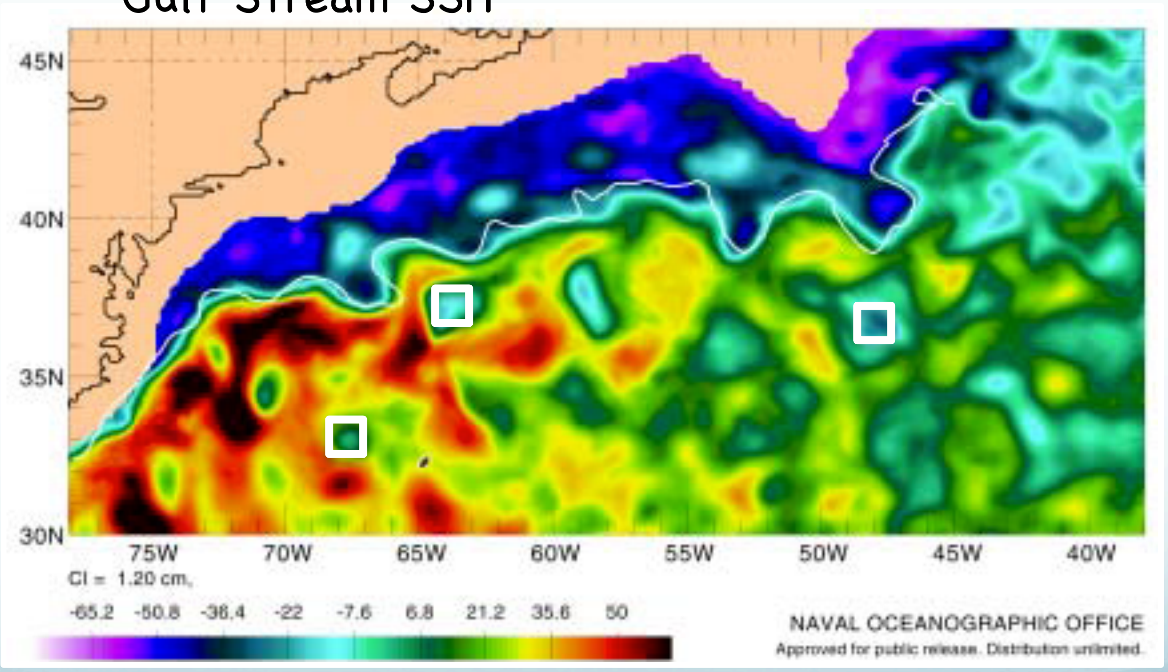
➔ At all (present-day) resolutions, OGCMs resolve the mesoscale in some regions but not others

Ocean Modelling Challenges

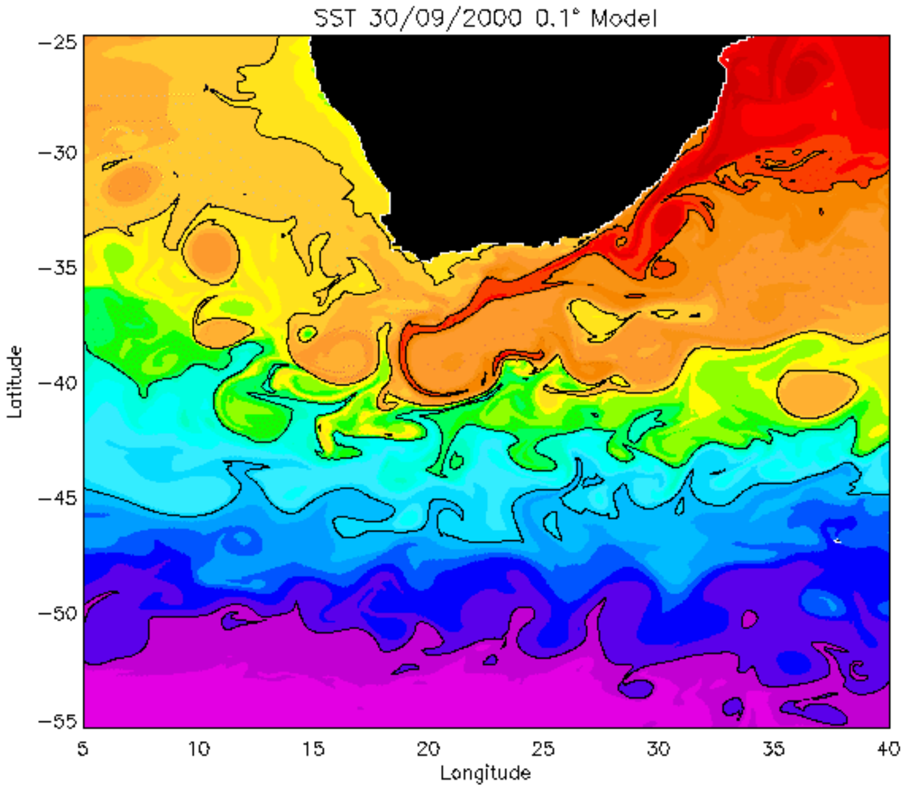
Workhorse (1° ≈ 100km) ocean models for climate research cannot reproduce the rich mesoscale eddy field observed in Nature...

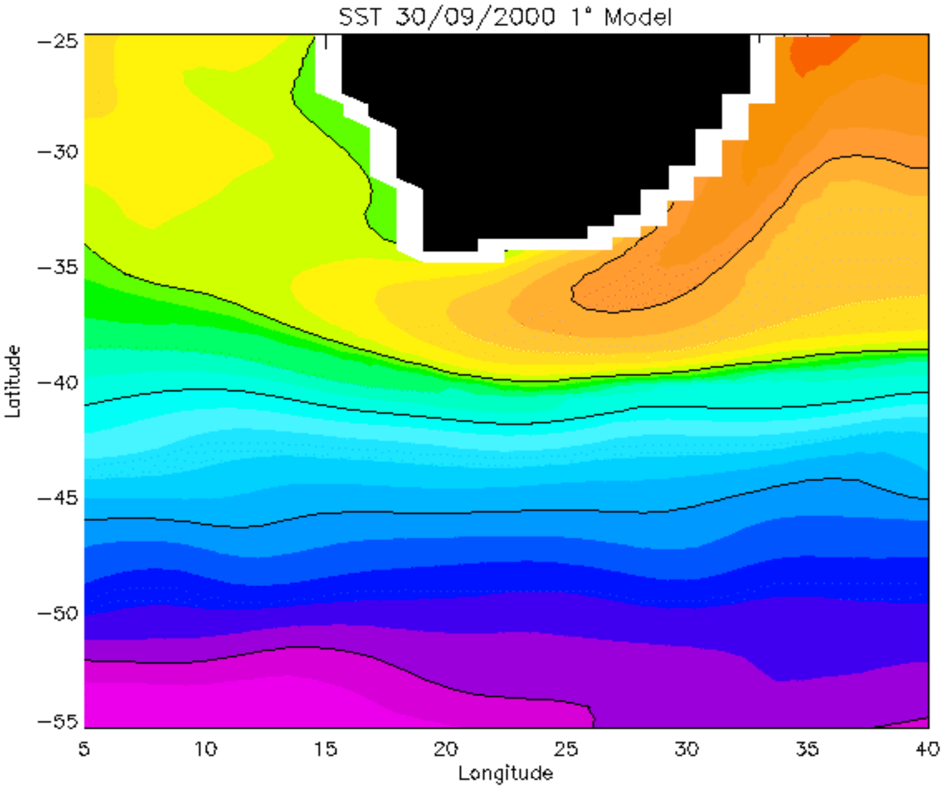
Gulf Stream SSH



Ocean Modelling Challenges

➔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."*

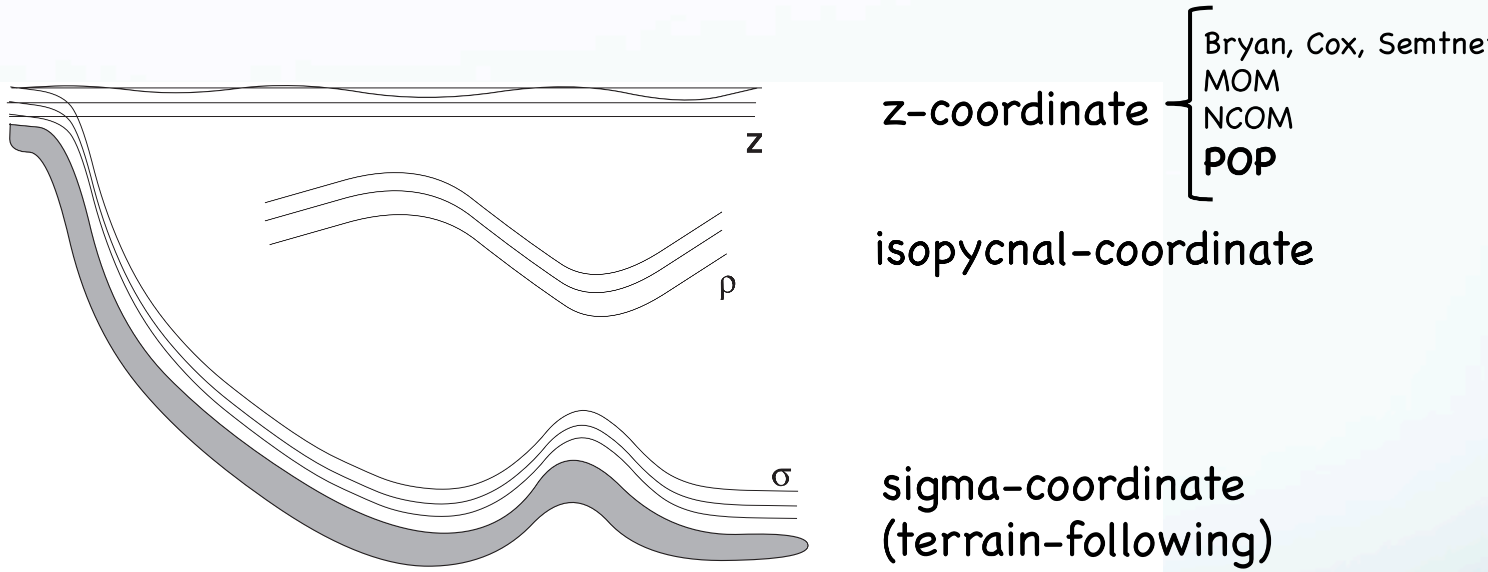


Fig. 1. Schematic of an ocean basin illustrating the three regimes of the ocean germane to the considerations of an appropriate vertical coordinate. The surface mixed layer is naturally represented using *z*-coordinates; the interior is naturally represented using isopycnal ρ-coordinates; and the bottom boundary is naturally represented using terrain following σ-coordinates.

*Griffies et al, 2000: Developments in ocean climate modelling, *Ocean Modelling*, **2**, 123–192.

Ocean Modelling Challenges

- Long equilibration timescale ➔ deep ocean will in general be characterized by drift.

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

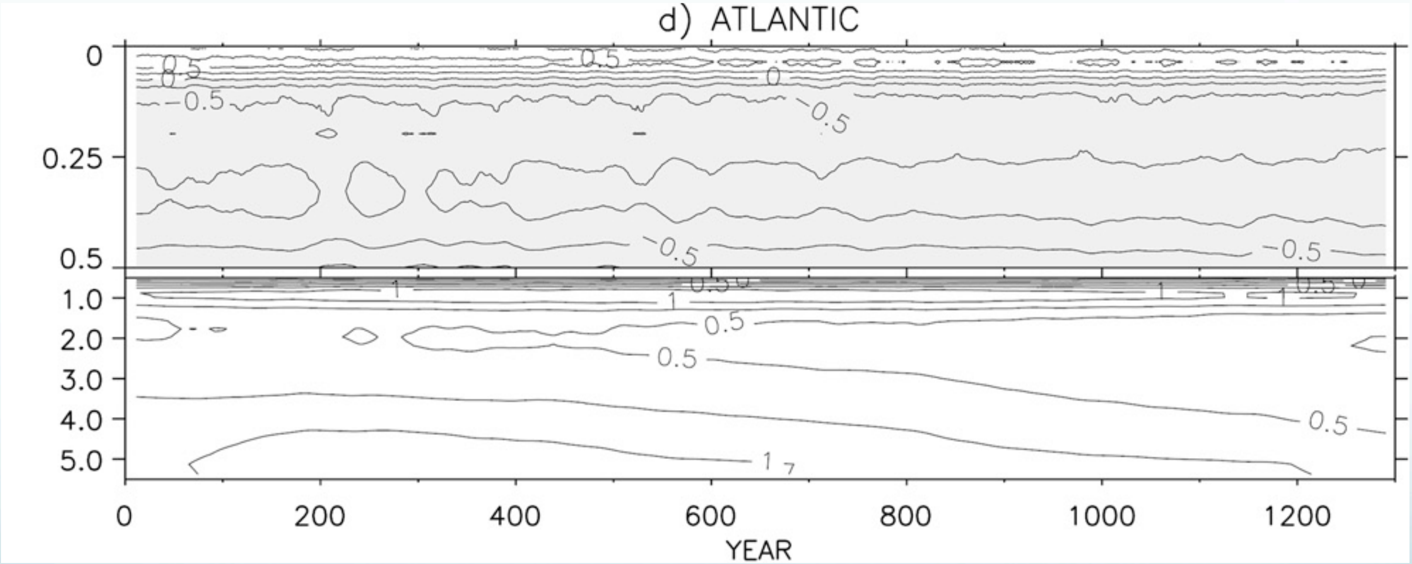


FIG. 2. Horizontal-mean potential temperature difference time series for 1850 CONTROL minus PHC2 observations: (a) global, (b) Pacific, (c) Indian, and (d) Atlantic Oceans. The contour intervals are 0.1°, 0.2°, 0.25°, and 0.25°C in (a),(b),(c),(d), respectively. The shaded regions indicate negative differences.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

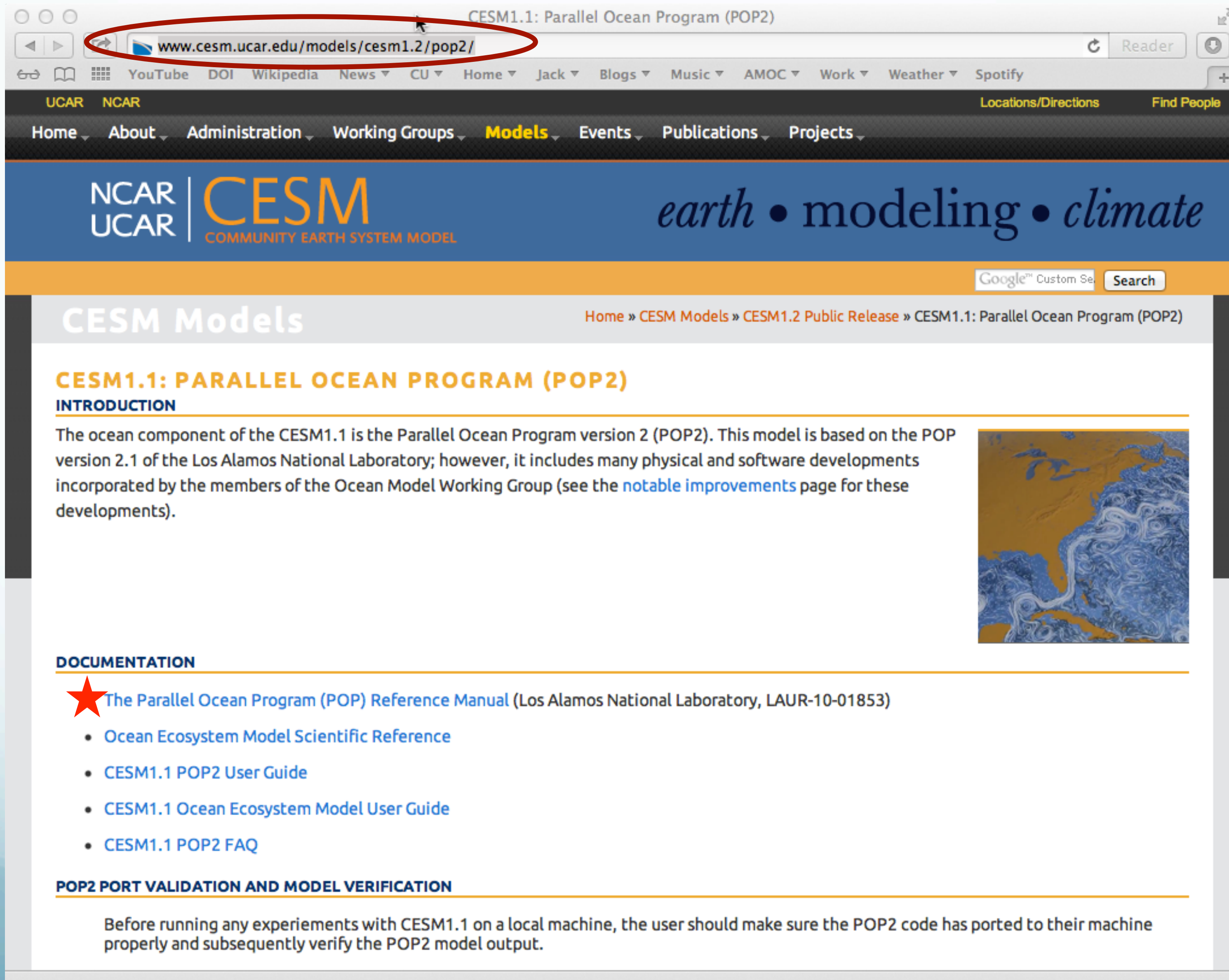
Danabasoglu et al., *J Climate,* (2012)

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



2010

Useful Resources

The Parallel Ocean Program (POP) Reference Manual

Ocean Component of the Community Climate System Model (CCSM) and Community Earth System Model (CESM) [1]

R. Smith[2], P. Jones[1], B. Briegleb[3], F. Bryan[2], G. Danabasoglu[2], J. Dennis[2], J. Dukowicz[1], C. Eden[4] B. Fox-Kemper[5], P. Gent[2], M. Hecht[1], S. Jayne[6], M. Jochum[2], W. Large[2], K. Lindsay[2], M. Maltrud[1], N. Norton[2], S. Peacock[2], M. Vertenstein[2], S. Yeager[2]

Contents

Useful Resources

VOLUME 25 JOURNAL OF CLIMATE 1 MARCH 2012

SPECIAL
Community Climate System Model CCSM4
COLLECTION

The CCSM4 Ocean Component

GOKHAN DANABASOGLU, SUSAN C. BATES, AND BRUCE P. BRIEGLEB
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STEVEN R. JAYNE
Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

MARKUS JOCHUM, WILLIAM G. LARGE, SYNTE PEACOCK, AND STEVE G. YEAGER
National Center for Atmospheric Research, Boulder, Colorado

Journal of Climate CCSM4 / CESM1 Special Collection Papers
(doi:10.1175/JCLI-D-11-00091.1)

Model Equations

Hierarchy of dynamical approximation

compressible fluid dynamics (Navier-Stokes)

↓

Boussinesq equations

↓

Primitive equations

↓

Balance equations

↙ ↘

Planetary geostrophic equations

Quasigeostrophic equations

*McWilliams, 1998, *Ocean General Circulation Models*, in Ocean Modeling and Parameterization, NATO Science Series.

Model Equations

compressible fluid dynamics (Navier-Stokes)

↓ $\rho = \rho_o + \delta\rho$, $\delta\rho / \rho \ll 1$

Boussinesq equations

↓

Primitive equations

↓

Balance equations

↙ ↘

Planetary geostrophic equations

Quasigeostrophic equations

δρ is very small in the ocean, so ignore δρ <u>except</u> in gravitational force and equation of state.

➔ Mass continuity equation becomes

$$\nabla_{3D} \cdot \mathbf{v} = 0$$

(non-divergent flow)

*McWilliams, 1998, *Ocean General Circulation Models*, in Ocean Modeling and Parameterization, NATO Science Series.

Model Equations

compressible fluid dynamics (Navier-Stokes)

↓ $\rho = \rho_o + \delta\rho$, $\delta\rho / \rho \ll 1$

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)

➔ 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).

*McWilliams, 1998, *Ocean General Circulation Models*, in Ocean Modeling and Parameterization, NATO Science Series.

Model Equations

compressible fluid dynamics (Navier-Stokes)

↓ $\rho = \rho_o + \delta\rho$, $\delta\rho / \rho \ll 1$

Boussinesq equations

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

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)

*McWilliams, 1998, *Ocean General Circulation Models*, in Ocean Modeling and Parameterization, NATO Science Series.

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):

Momentum equations:

1
$$\frac{\partial}{\partial t}u + \mathcal{L}(u) - (uv\tan\phi)/a - fv = -\frac{1}{\rho_0 a\cos\phi}\frac{\partial p}{\partial\lambda} + \mathcal{F}_{Hx}(u,v) + \mathcal{F}_V(u) \quad (2.1)$$

2
$$\frac{\partial}{\partial t}v + \mathcal{L}(v) + (u^2\tan\phi)/a + fu = -\frac{1}{\rho_0 a}\frac{\partial p}{\partial\phi} + \mathcal{F}_{Hy}(u,v) + \mathcal{F}_V(v) \quad (2.2)$$

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

$$\mathcal{F}_{Hx}(u,v) = A_M\left\{\nabla^2 u + u(1-\tan^2\phi)/a^2 - \frac{2\sin\phi}{a^2\cos^2\phi}\frac{\partial v}{\partial\lambda}\right\} \quad (2.4)$$

$$\mathcal{F}_{Hy}(u,v) = A_M\left\{\nabla^2 v + v(1-\tan^2\phi)/a^2 + \frac{2\sin\phi}{a^2\cos^2\phi}\frac{\partial u}{\partial\lambda}\right\} \quad (2.5)$$

$$\nabla^2\alpha = \frac{1}{a^2\cos^2\phi}\frac{\partial^2\alpha}{\partial\lambda^2} + \frac{1}{a^2\cos\phi}\frac{\partial}{\partial\phi}\left(\cos\phi\frac{\partial\alpha}{\partial\phi}\right) \quad (2.6)$$

$$\mathcal{F}_V(\alpha) = \frac{\partial}{\partial z}\mu\frac{\partial}{\partial z}\alpha \quad (2.7)$$

3 *Continuity equation*:

$$\mathcal{L}(1) = 0 \quad (2.8)$$

4 *Hydrostatic equation*:

$$\frac{\partial p}{\partial z} = -\rho g \quad (2.9)$$

5 *Equation of state*:

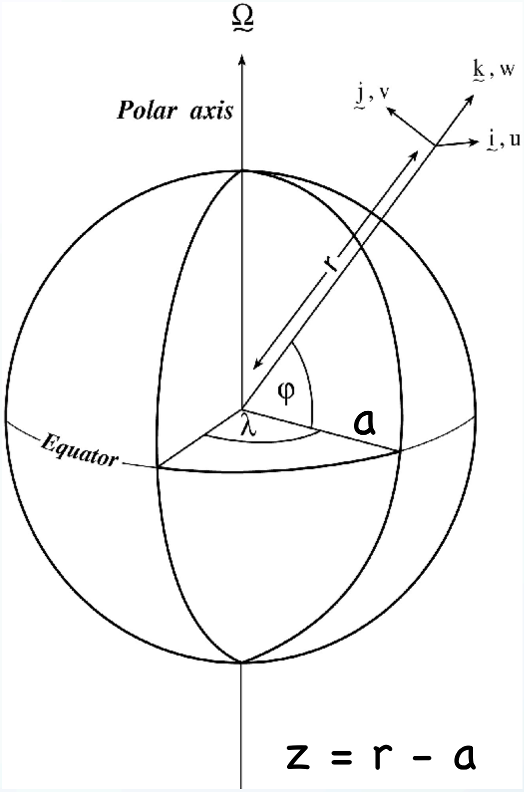
$$\rho = \rho(\Theta, S, p) \rightarrow \rho(\Theta, S, z) \quad (2.10)$$

6,7 *Tracer transport*:

$$\frac{\partial}{\partial t}\varphi + \mathcal{L}(\varphi) = \mathcal{D}_H(\varphi) + \mathcal{D}_V(\varphi) \quad (2.11)$$

$$\mathcal{D}_H(\varphi) = A_H\nabla^2\varphi \quad (2.12)$$

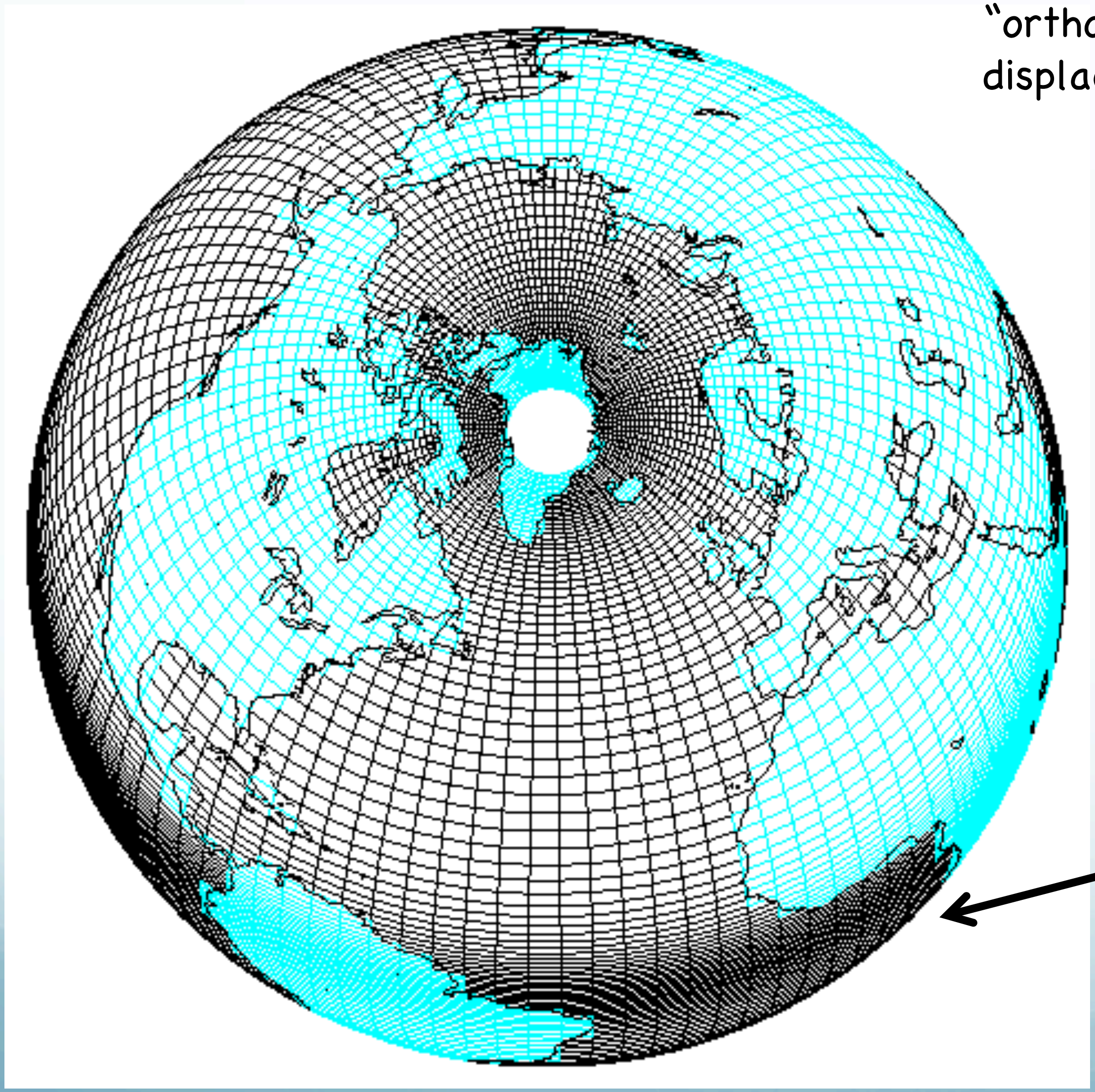
$$\mathcal{D}_V(\varphi) = \frac{\partial}{\partial z}\kappa\frac{\partial}{\partial z}\varphi, \quad (2.13)$$



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°

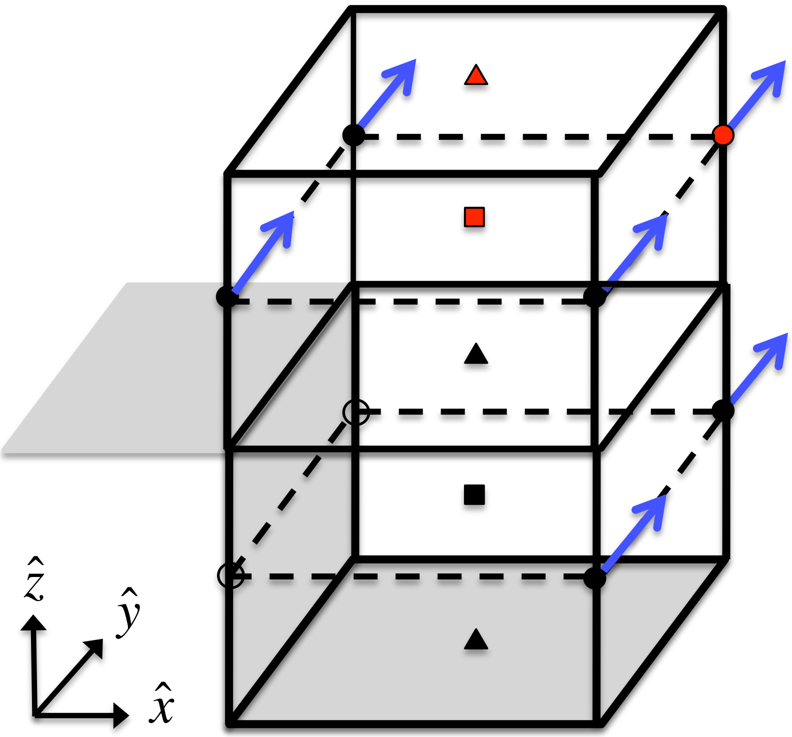
Equatorial refinement
(0.3° / 0.9°)

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)

$$\delta_x\psi = \left[\psi\left(x+\Delta_x/2\right)-\psi\left(x-\Delta_x/2\right)\right]/\Delta_x \qquad (3.4)$$

$$\overline{\psi}^x = \left[\psi\left(x+\Delta_x/2\right)+\psi\left(x-\Delta_x/2\right)\right]/2\,, \qquad (3.5)$$

● $$\nabla\psi = \hat{\mathbf{x}}\delta_x\overline{\psi}^y + \hat{\mathbf{y}}\delta_y\overline{\psi}^x \qquad (3.6)$$

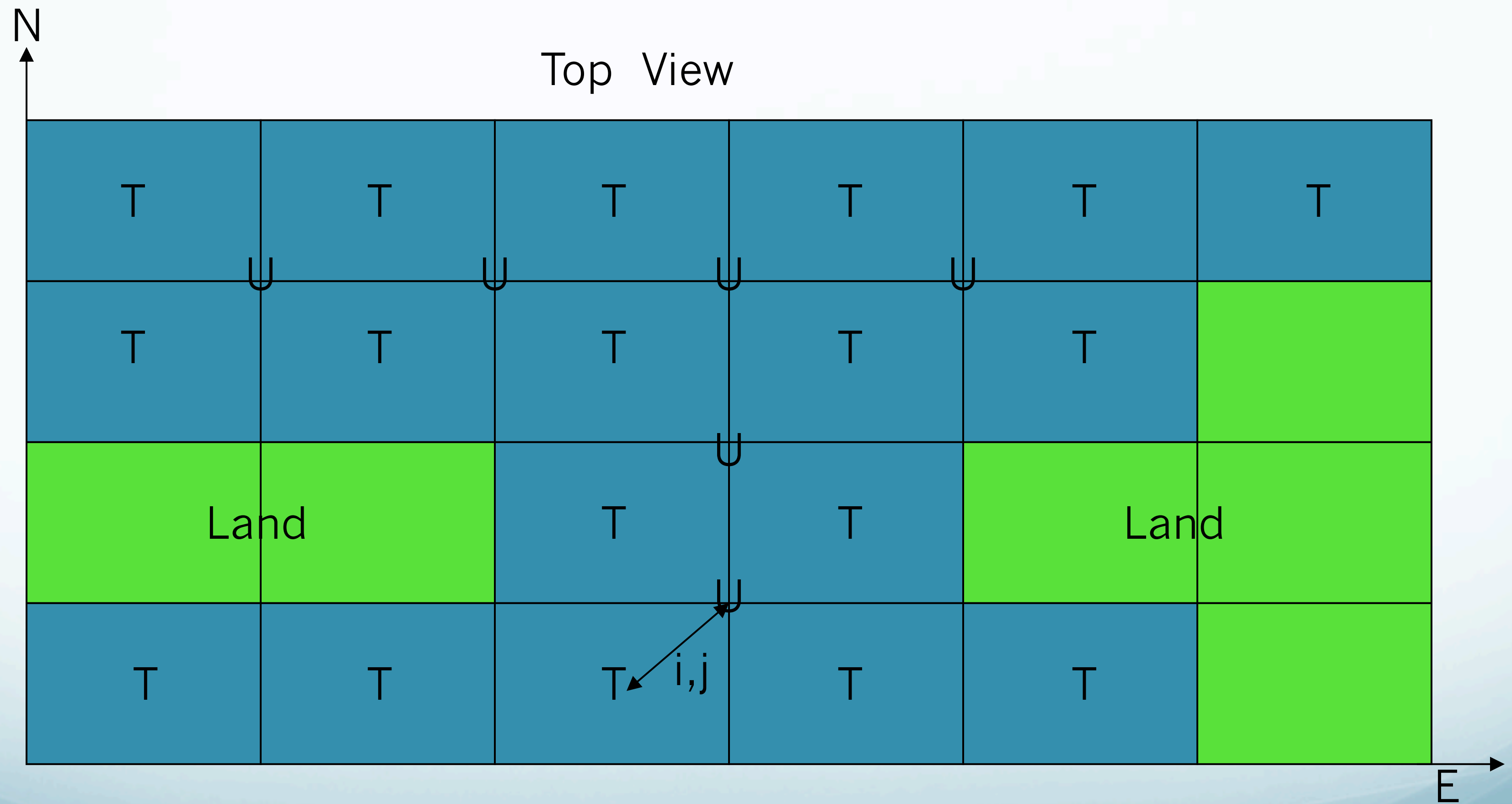
■ $$\nabla\cdot\mathbf{u} = \frac{1}{\Delta_y}\delta_x\overline{\Delta_y u_x}^y + \frac{1}{\Delta_x}\delta_y\overline{\Delta_x u_y}^x \qquad (3.7)$$

■ $$\hat{\mathbf{z}}\cdot\nabla\times\mathbf{u} = \frac{1}{\Delta_y}\delta_x\overline{\Delta_y u_y}^y - \frac{1}{\Delta_x}\delta_y\overline{\Delta_x u_x}^x \qquad (3.8)$$

■ $$\nabla\cdot G\nabla\psi = \frac{1}{\Delta_y}\delta_x\overline{\left[\Delta_y G\delta_x\overline{\psi}^y\right]}^y + \frac{1}{\Delta_x}\delta_y\overline{\left[\Delta_x G\delta_y\overline{\psi}^x\right]}^x\,. \qquad (3.9)$$

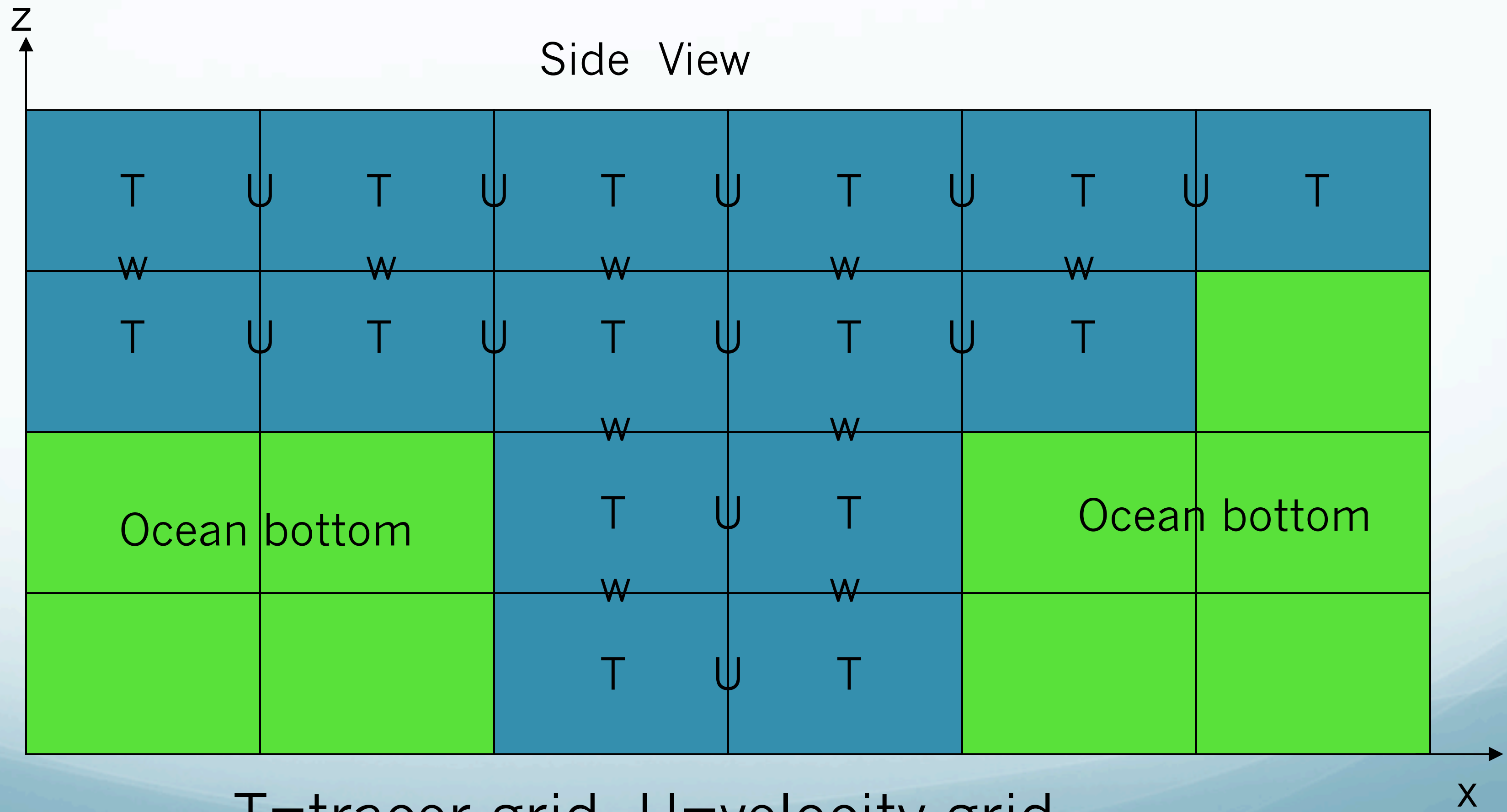
POP Spatial Discretization

- At least 2 adjacent active ocean T-cells are required for flow through channels



T=tracer grid, U=velocity grid

POP Spatial Discretization

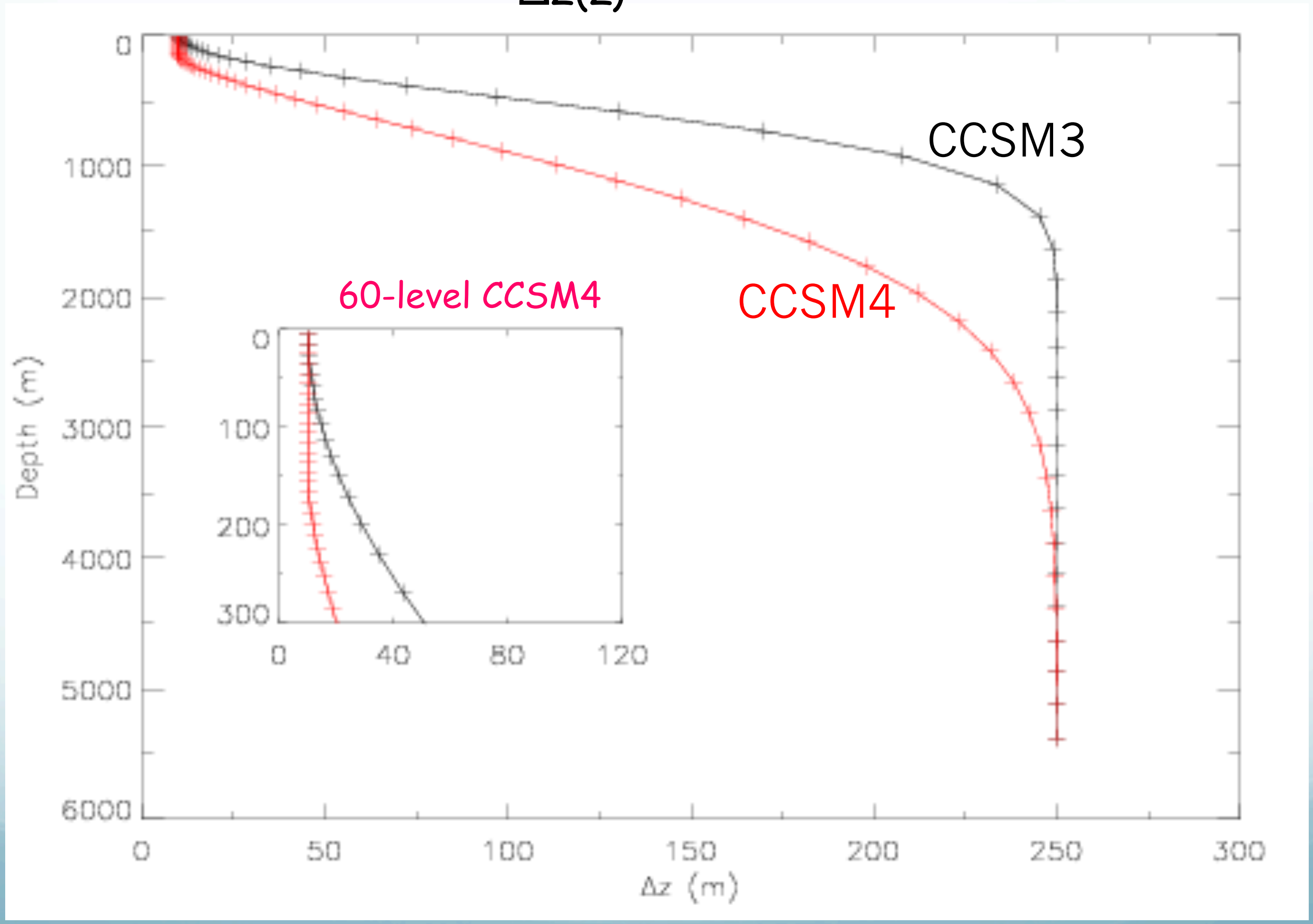


T=tracer grid, U=velocity grid

POP Vertical Discretization

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

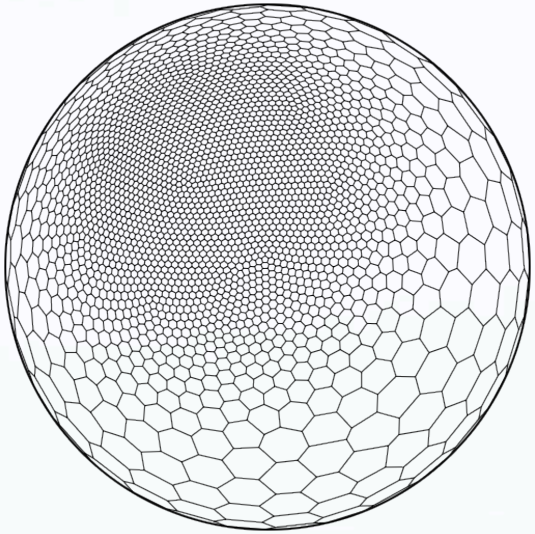
Δz(z)

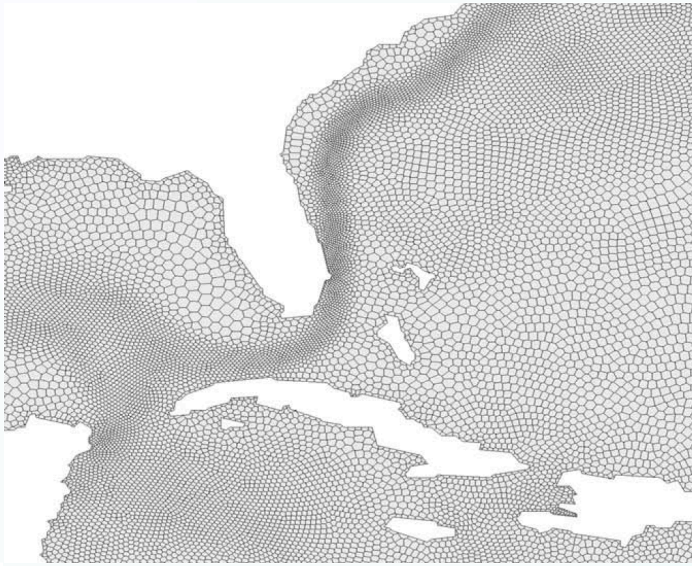


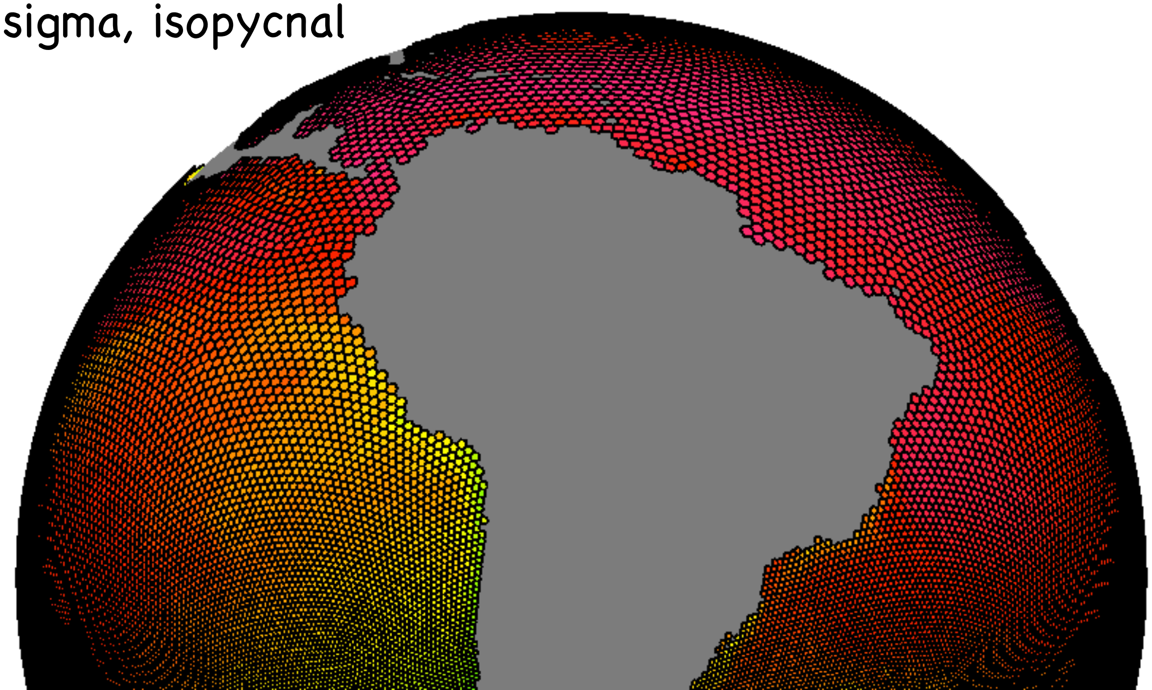
MPAS-Ocean model grids (CESM3.0?)

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)

Reference: Ringler et al, 2014, A multiresolution approach to global ocean modelling, *Ocean Modelling*, in revision.

POP numerics in a nutshell

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

advection operator in analytic form

Finite difference advection

- Momentum: centered differencing (2nd order)

$$\mathcal{L}_U(\alpha) = \frac{1}{\Delta_y}\delta_x\left[\overline{(\Delta_y \overline{u_x}^y)}^{xy}\overline{\alpha}^x\right] + \frac{1}{\Delta_x}\delta_y\left[\overline{(\Delta_x \overline{u_y}^x)}^{xy}\overline{\alpha}^y\right] + \delta_z(w^U\overline{\alpha}^z)\ . \quad (3.23)$$

- Tracers: upwind3 scheme (3rd 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

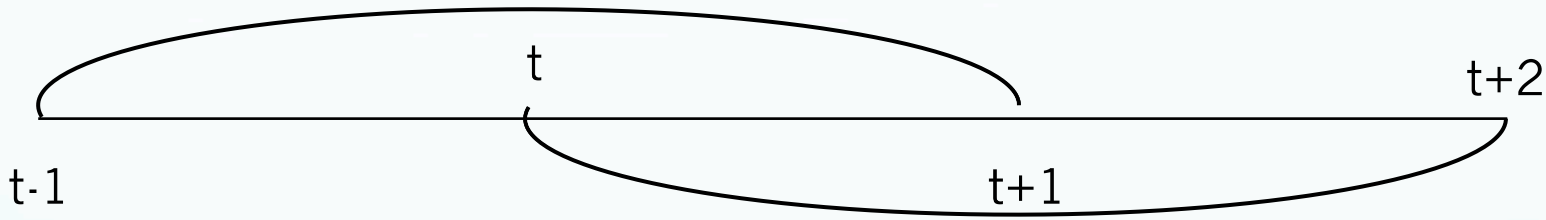
POP numerics in a nutshell

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:

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

(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 ($\sqrt{gH}$ ~200 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

➔ $\Delta t \approx 1$ hour in 1° 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:

u,v: $F_V(\alpha) = \frac{\partial}{\partial z} \underbrace{\mu \frac{\partial}{\partial z}\alpha}_{\text{momentum flux}}$ tracers: $D_V(\varphi) = \frac{\partial}{\partial z} \underbrace{\kappa \frac{\partial}{\partial z}\varphi}_{\text{tracer flux}}$

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

$$\left[\mu \frac{\partial}{\partial z}\vec{\mathbf{u}}\right]_{z=0} = \frac{\vec{\tau}}{\rho_o} \qquad \left[\kappa \frac{\partial}{\partial z}T\right]_{z=0} = \frac{Q_{net}}{\rho_o C_p}$$

(see Barnier, 1998)

$$\left[\kappa \frac{\partial}{\partial z}S\right]_{z=0} = \frac{F_{net}}{\rho_o} S_o$$

"virtual salinity flux"

➔ Primitive equation surface b.c.'s require specification of:

- Wind stress vector : $\vec{\tau}$
- 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, θ) with a feedback of the surface ocean state (U_o, SST) onto the fluxes:

$$\vec{\tau}_{as} = \rho C_{\mathrm{D}} |\Delta \vec{U}| \Delta \vec{U} \tag{3a}$$

$$E = \rho C_{\mathrm{E}} (q - q_{\mathrm{sat}}(\mathrm{SST})) |\Delta \vec{U}| \tag{3b}$$

$$Q_E = \Lambda_v E \tag{3c}$$

$$Q_H = \rho c_p C_{\mathrm{H}} (\theta - \mathrm{SST}) |\Delta \vec{U}|, \tag{3d}$$

(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:

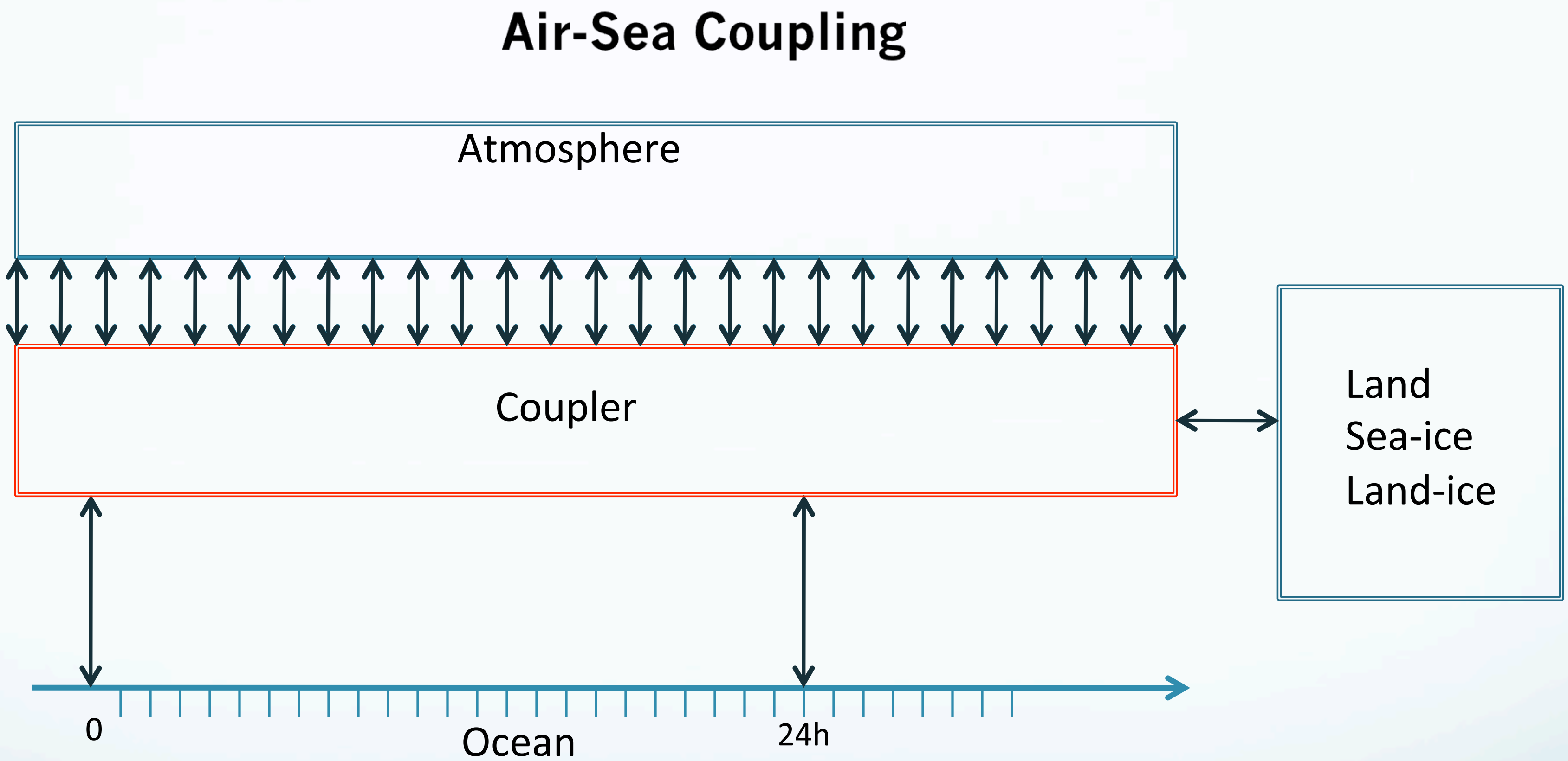
Barnier, 1998: Forcing the Ocean, in Ocean Modeling and Parameterization, NATO Science Series.

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

Large & Yeager, 2009: The global climatology of an interannually varying air-sea flux data set, *Clim Dyn*, **33**, 341-364, doi:10.1007/s00382-008-0441-3

Quality of POP model solution is strongly tied to quality of surface b.c.'s !

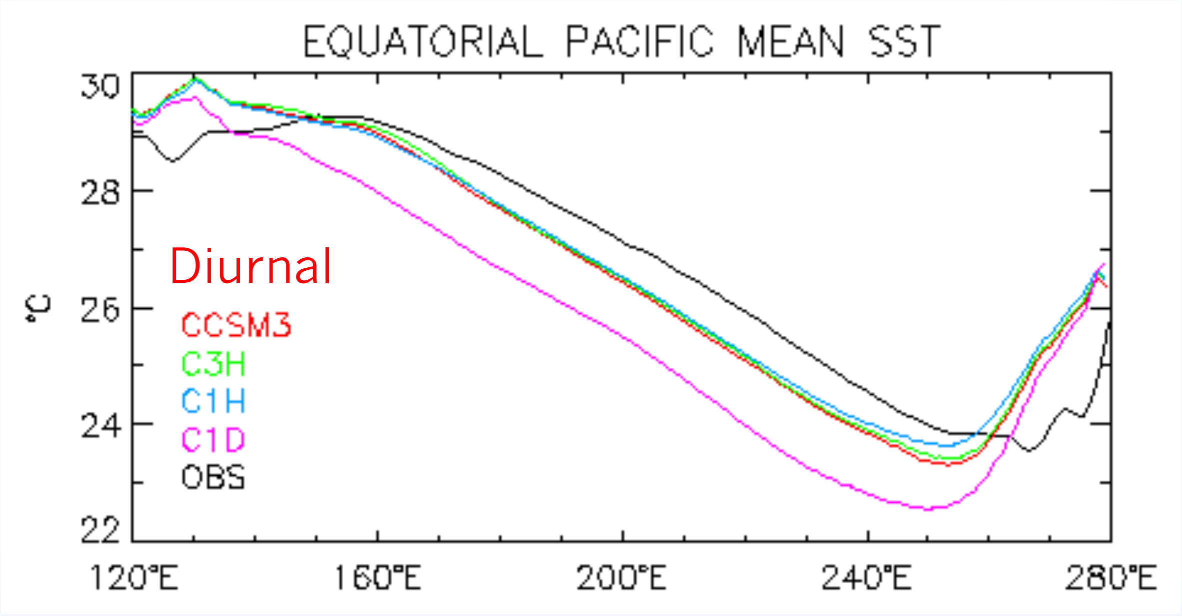
POP diurnal cycle



➔ Need to parameterize the diurnal cycle of (shortwave) radiative heat flux (ie., night & day). This is done with a zenith-angle dependent SW(lat, lon, hour, day of year) heat flux parameterization.

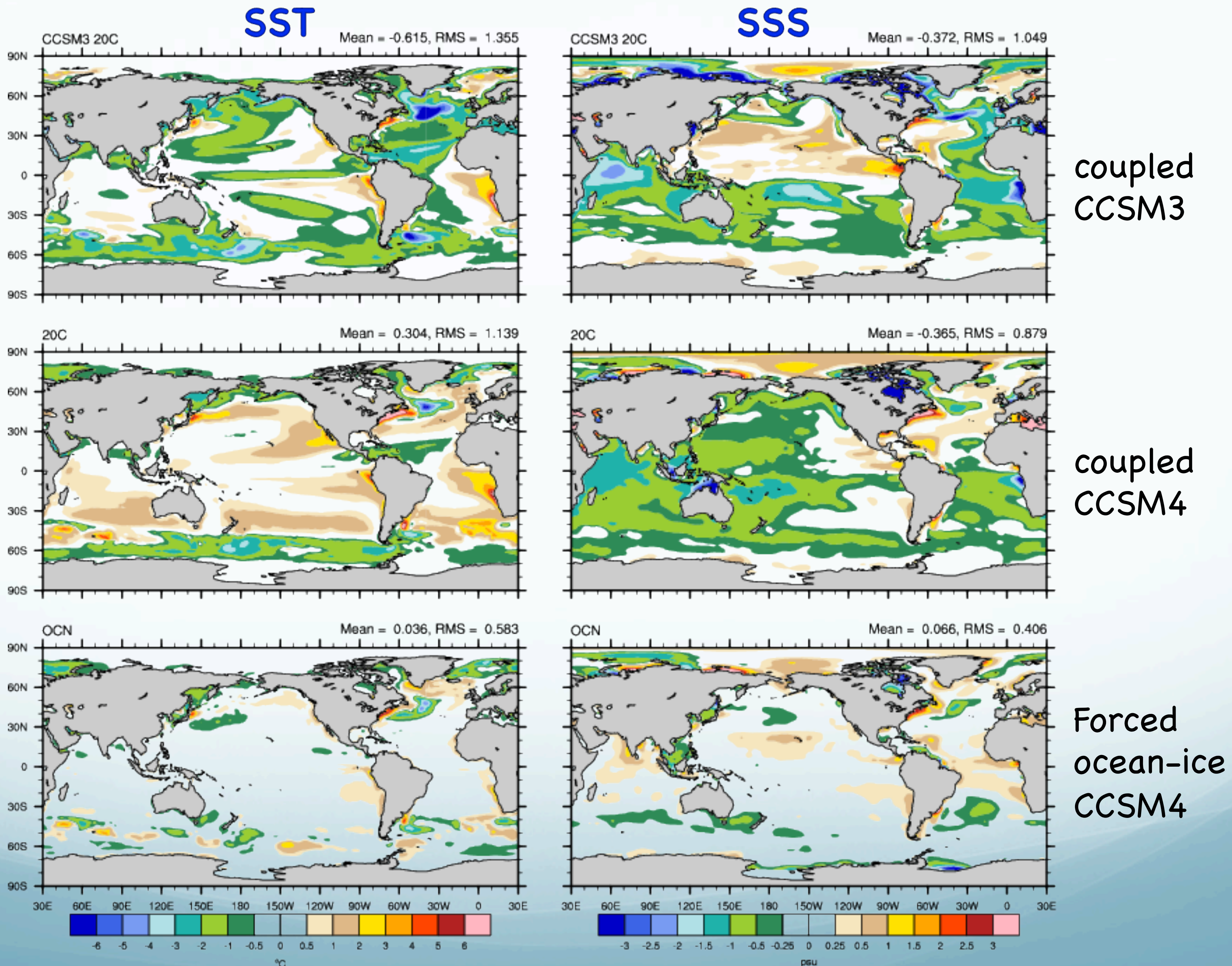
POP diurnal cycle

Air-Sea Coupling



➔ The SW diurnal cycle results in dramatically improved equatorial SST

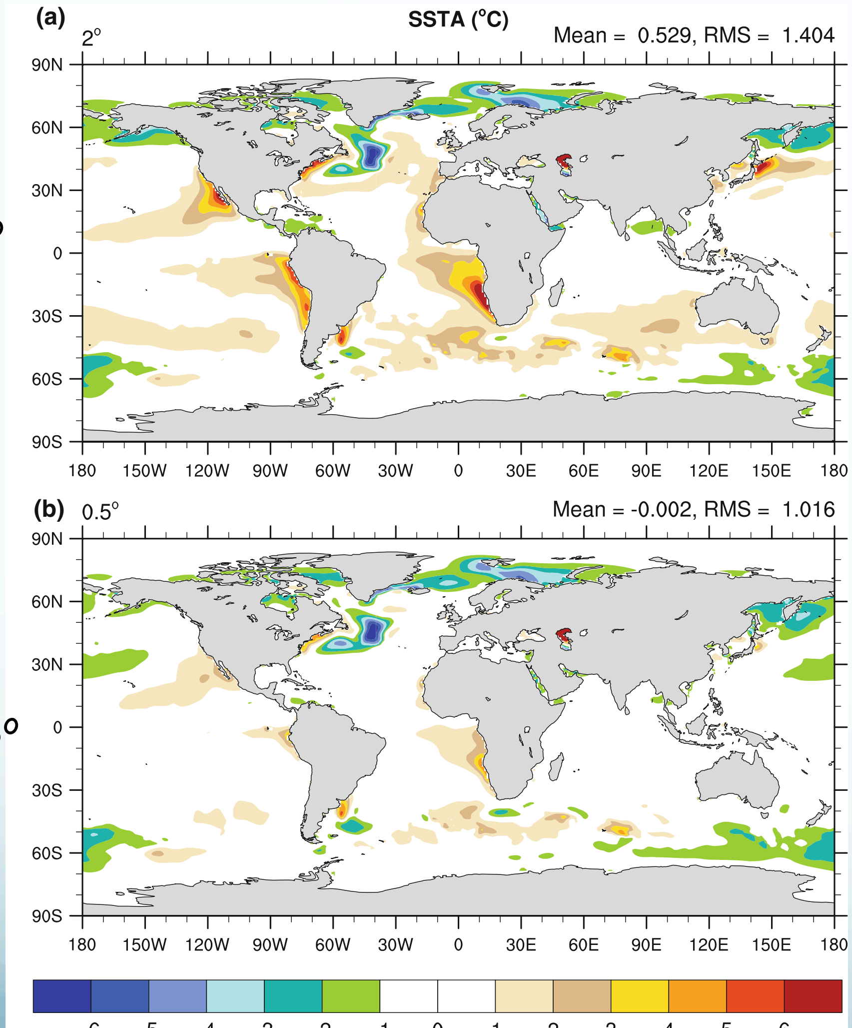
SST and Salinity Differences from Observations



Coupled CCSM4 SST bias as a function of atmospheric resolution

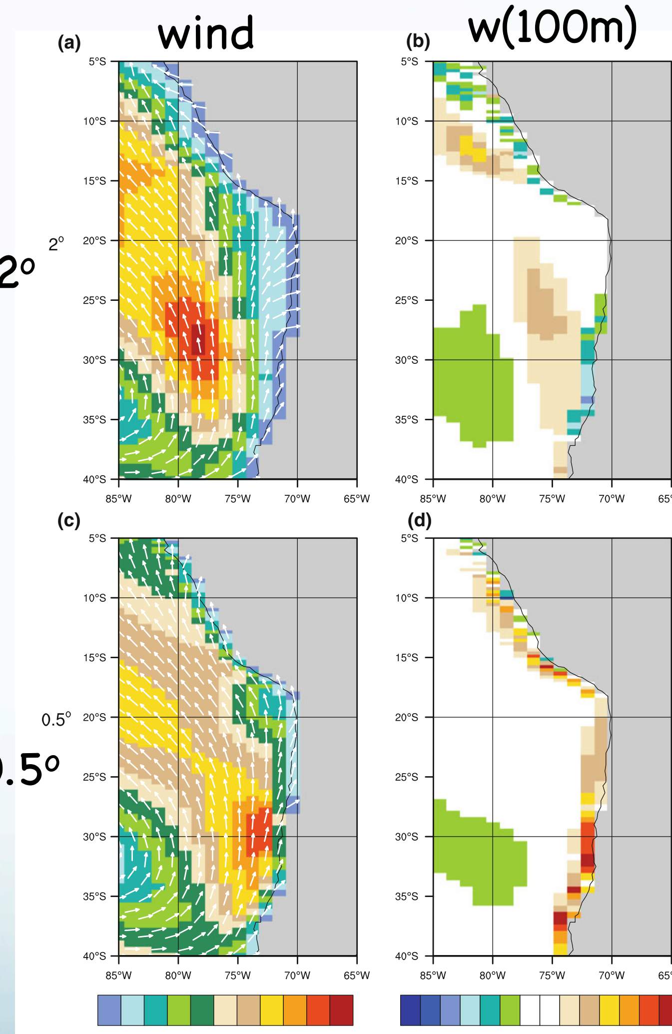
2°

0.5°



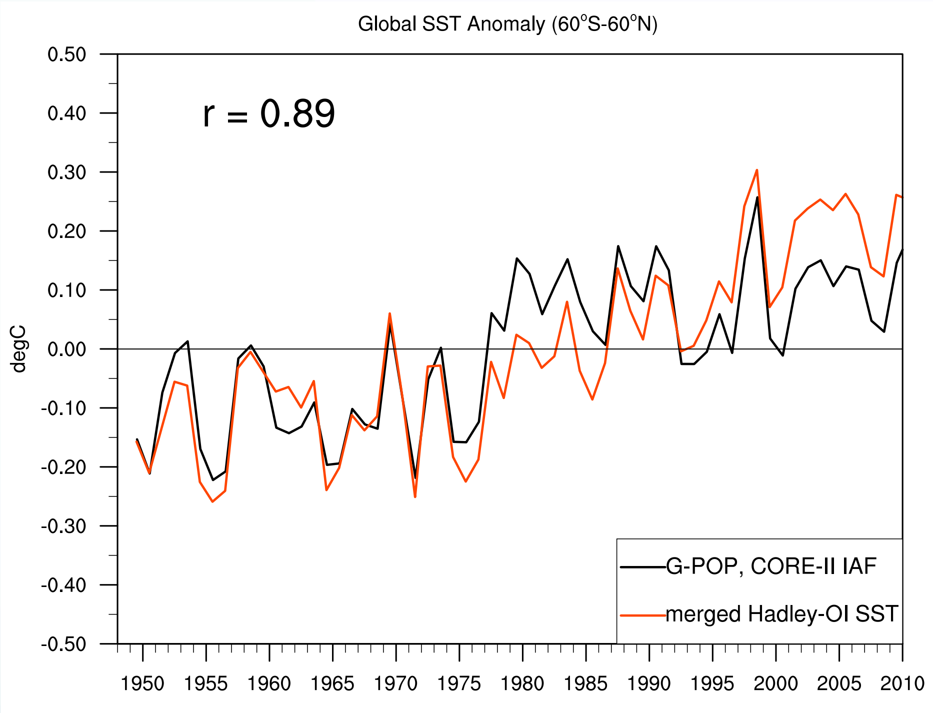
2°

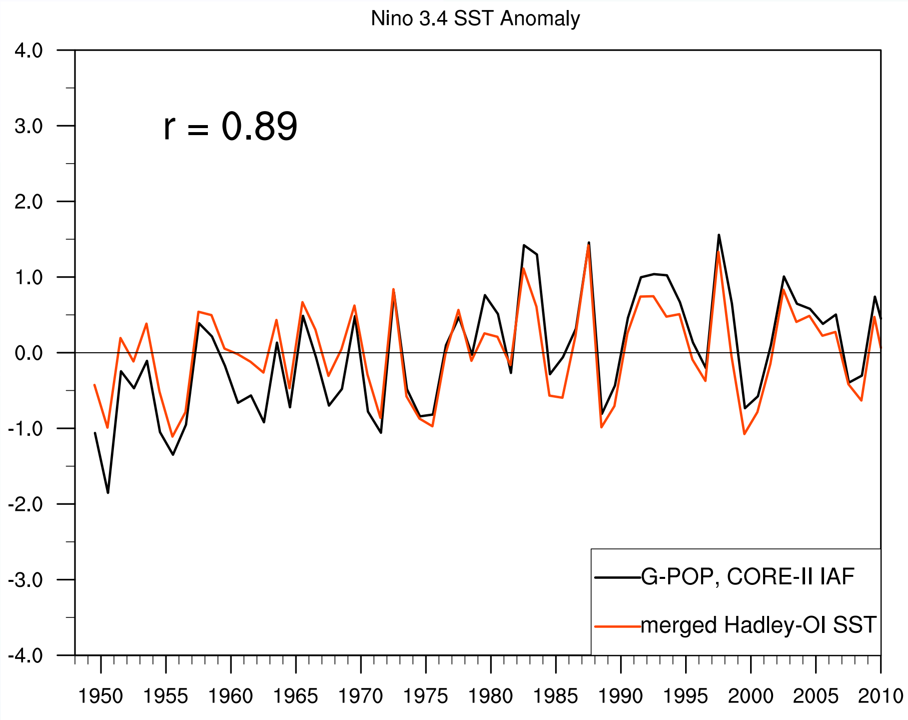
0.5°



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

Interannual variability of North Atlantic 275m Heat Content

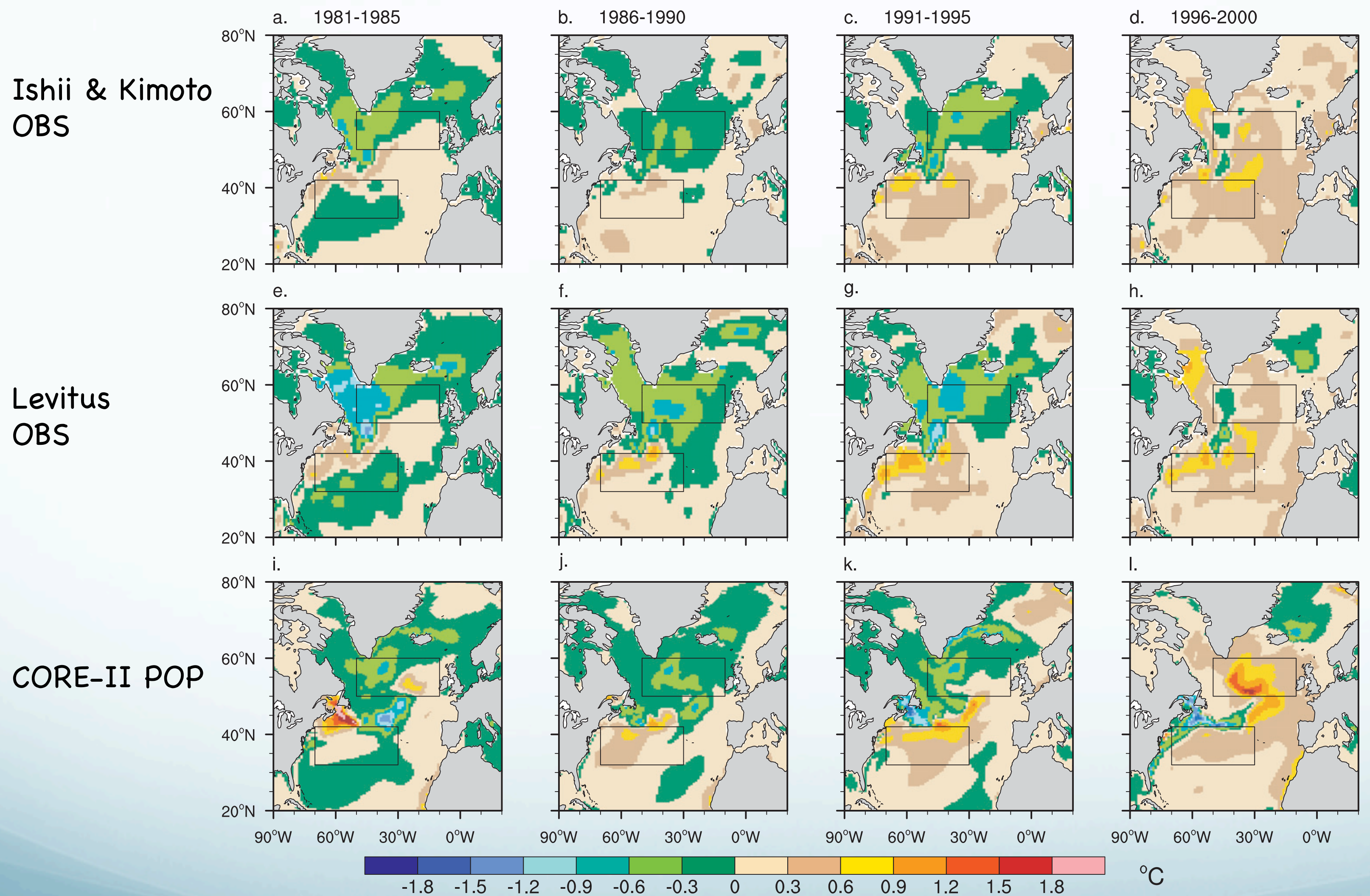
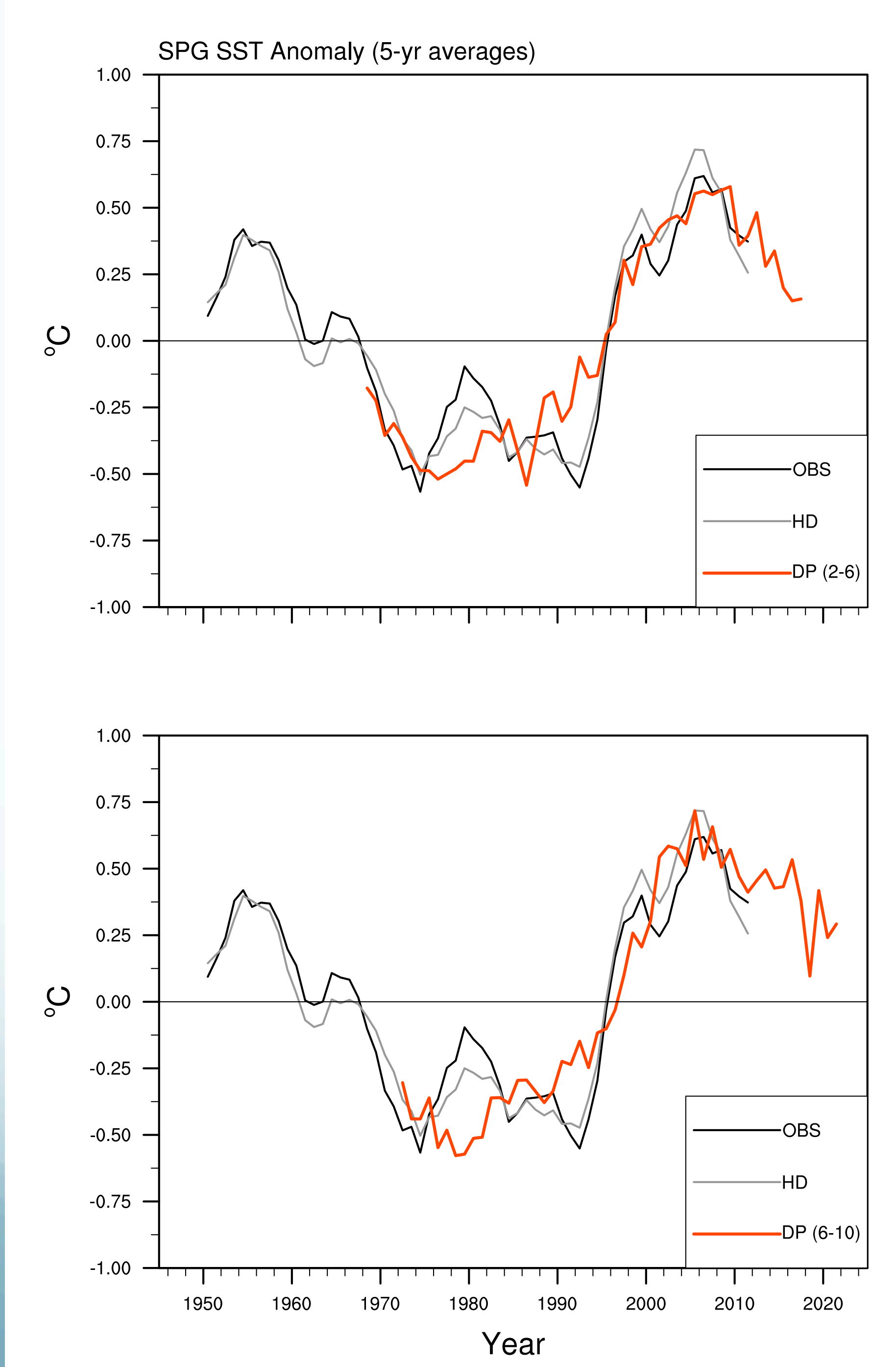


FIG. 1. Pentadal-mean heat content anomalies expressed as the 275-m depth-averaged temperature anomaly relative to 1957–90 climatology from (a)–(d) Ishii and Kimoto (2009), (e)–(h) Levitus et al. (2009), and (i)–(l) CORE-IA. The boxes in each panel demarcate the SPG (50°–10°W, 50°–60°N) and STG (70°–30°W,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

Yeager et al., *J Clim*, 2012

Coupled decadal prediction of North Atlantic SST



Yeager et al., *2014, in prep.*

For even more info…

Books:

- Chassignet & Verron (Eds.), 1998: *Ocean Modeling and Parameterization,* Proceedings of the NATO Advanced Study Institute, NATO Science Series C, vol. 516, 451pp.

- Haidvogel & Beckmann, 1999: *Numerical Ocean Circulation Modelling,* Imperial College Press, 318 pp.

- Griffies, 2004: *Fundamentals of Ocean Climate Models,* Princeton University Press, 518 pp.

Review Papers:

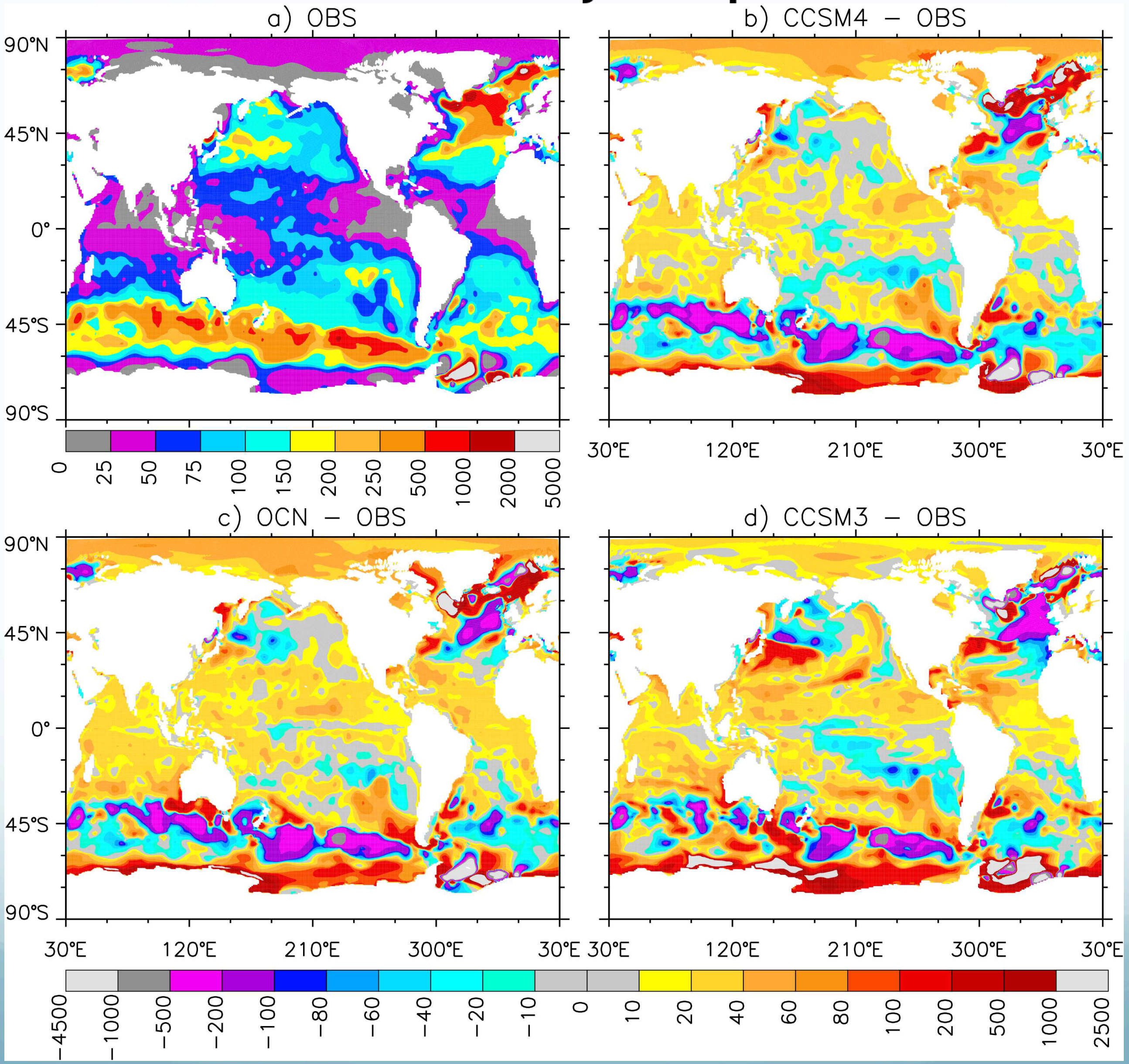
- Griffies et al, 2000: Developments in ocean climate modelling, *Ocean Modelling*, **2**, 123-192.

- Griffies et al., 2010, *Problems and prospects in large-scale ocean circulation models,* Ocean Obs '09 Community White Paper, doi:10.5270/OceanObs09.cwp.38.

Questions?

Model Biases

Mixed Layer Depth



Model Biases

SST Differences from Observations

2° atmosphere

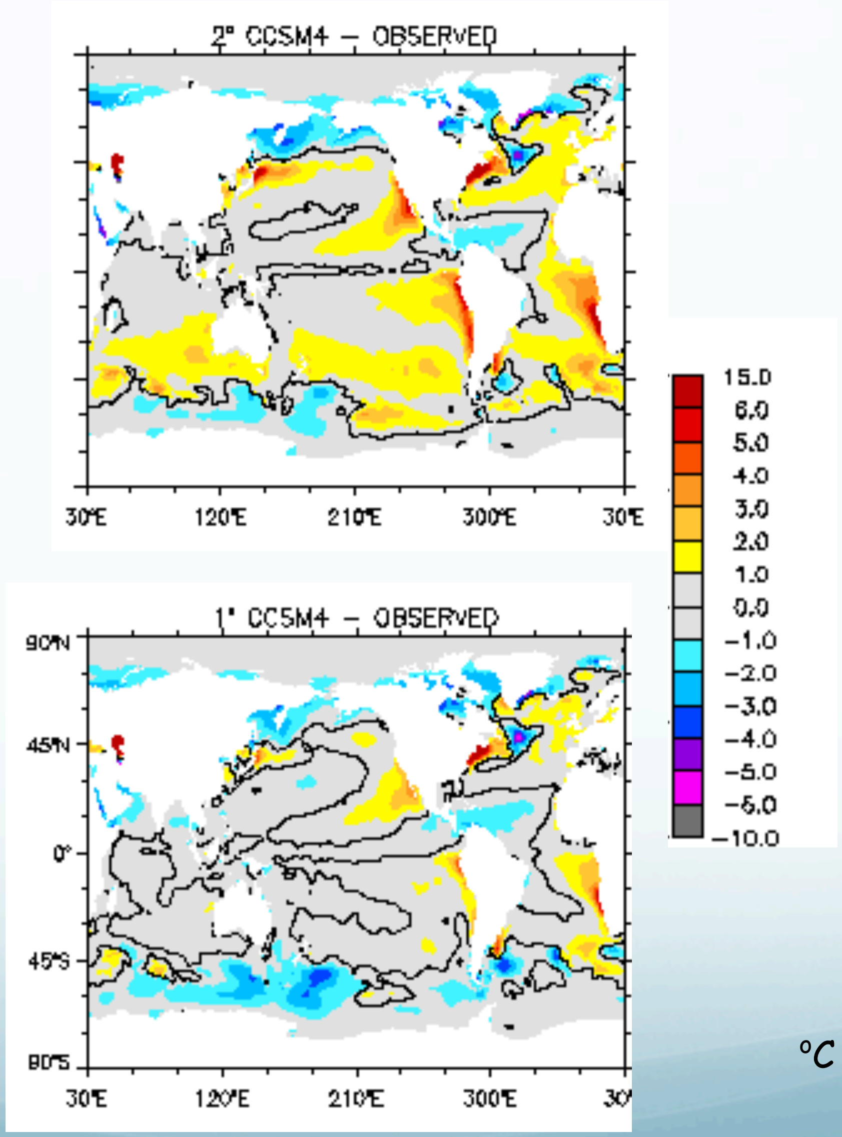
mean= 0.63°C

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

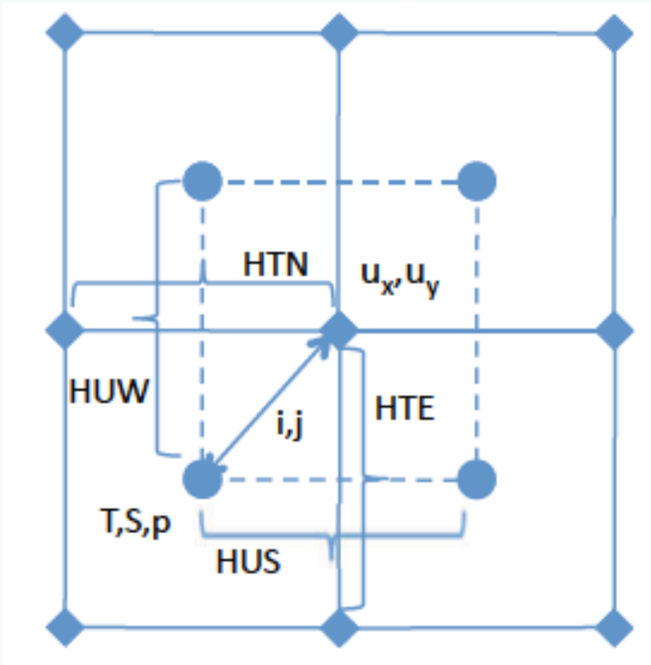
- POP2 User Guide
- Ocean Ecosystem Model User Guide
- POP Reference Manual
- Ocean Ecosystem Reference Manual

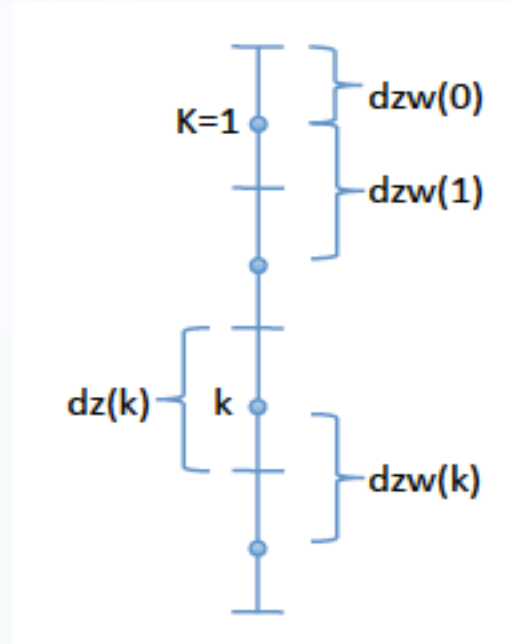
Friday's breakout session

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





$$ADV_{i,j,k} = -\,(u_E\ T^*_E - u_W\ T^*_W)/DXT - (v_N\ T^*_N - v_S\ T^*_S)/DYT - (w_k\ T^*_T - w_{k+1}\ T^*_B)/dz$$

$$u_E(i) = (u_{i,j}DYU_{i,j} + u_{i,j-1}DYU_{i,j-1})/(2DXT_{i,j})$$
$$u_W(i) = u_E(i-1)$$
$$v_N(j) = (v_{i,j}DXU_{i,j} + v_{i-1,j}DXU_{i-1,j})/(2DXT_{i,j})$$
$$v_S(j) = (v_{i,j\text{-}1}DXU_{i,j} + v_{i-1,j-1}DXU_{i-1,j})/(2DXT)$$

$$T^*_E = ½ * (T_{i+1,j} + T_{i,j})$$

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 $\sqrt{gH}$ ~200 m/s
- Split flow into depth averaged barotropic (<U>) plus vertically varying baroclinic (U')
- Fast moving gravity waves are filtered out, but that's okay because they don't impact climate

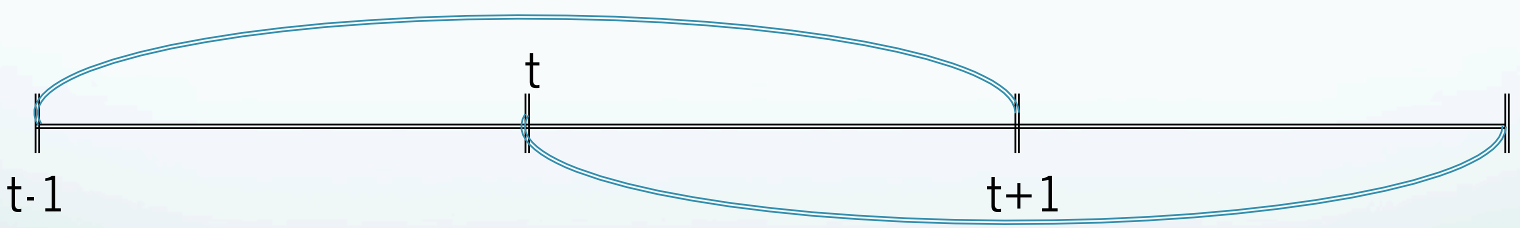
Barotropic and Baroclinic Flow

U = <U> + U’

- <U>: Implicit, linearized free-surface formulation obtained by combining the vertically integrated momentum and continuity equations

- U’: use a leapfrog time stepping to solve

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



- Occasional time averaging to eliminate the split mode