OCEAN MODELING II

Parameterizations

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SPACE - TIME SCALES and OCEAN MODELS

WORKHORSE (CLIMATE)
$O(10-100 \text{ years/day})$

HI-RES
$O(<<10 \text{ years/day})$

FAST
$O(100-1000 \text{ years/day})$
PARAMETERIZATIONS:
- ACCOMPLISH PHYSICAL EFFECTS OF UNRESOLVED SUB-GRID-SCALE PROCESSES (ALL INVOLVING TURBULENCE),
- PHYSICALLY-BASED / JUSTIFIED,
- AS SIMPLE AS POSSIBLE,
- AS FEW PARAMETERS AS POSSIBLE.

• Development,
• Implementation,
• Verification,
• Impacts
PARAMETERIZATIONS IN CESM1 POP2

- Vertical mixing (momentum and tracers)
  - surface boundary layer,
  - interior
- Horizontal viscosity (momentum)
- Lateral mixing / mesoscale eddies (tracers)
- Overflows

- Submesoscale eddies (tracers)
- Diurnal cycle for short-wave heat flux
- Solar absorption
VERTICAL MIXING SCHEME: K-PROFILE PARAMETERIZATION (KPP)
Large, McWilliams, and Doney (1994, Rev. Geophys.)

A first-order turbulent closure scheme

\[ \partial_t X = -\partial_z w'X' \]

where \( X \) is a generalized variable (i.e., \( U, \theta \), etc.), \( w \) is the vertical velocity component, and the primes denote turbulent fluctuations.

\[ w'X' = -K_X \partial_z X \]

where \( K_X \) is an eddy diffusivity or viscosity.
KPP BOUNDARY LAYER DEPTH

The boundary layer depth, $h$, is determined based on a bulk Richardson number,

$$Ri_b(d) = \frac{[B_r - B(d)] d}{|V_r - V(d)|^2 + V_t^2(d)}$$

where $d$ is depth. Also

- $V_r$: near-surface reference horizontal velocity vector
- $V(d)$: boundary layer horizontal velocity profile
- $B_r$: near-surface reference buoyancy
- $B(d)$: boundary layer buoyancy profile
- $V_t$: velocity scale of turbulent velocity shear

$h$ is equated to the smallest value of $d$ at which the bulk $Ri$ equals $Ri_{cr} = 0.3$. 

INTERIOR MIXING

- Shear instability: $K_X^s$
- Internal wave breaking: $K_X^w$
- Double diffusion: $K_X^d$
- Local static instability (convection): $K_X^c$
- Tidal mixing: $K_X^t$

$$K_X(\text{interior}) = K_X^s + K_X^w + K_X^d + K_X^c + K_X^t$$
KPP BOUNDARY LAYER MIXING

\[ K_X(l) = h \ w_X(l) \ G(l) \]

with

\[ l = d / h, \]

\[ w_X(l): \text{turbulent velocity scale,} \]

\[ G(l): \text{cubic shape function.} \]

• \( K_X \) is non-local,

• Interior mixing at the base of the boundary layer influences the turbulence throughout the boundary layer,

• There is also a non-local counter-gradient term.
HORIZONTAL VISCOSITY

Spatially uniform, isotropic, Cartesian, $\Delta=250\text{km}$ grid for illustration

\[ D(U) = A U_{xx} + A U_{yy} \]
\[ D(V) = A V_{xx} + A V_{yy} \]

Grid Re (Diffuse Noise) \[ \rightarrow A > 0.5 \ V \ \Delta = 100,000 \text{ m}^2/\text{s} \]

Resolve WBC (Munk Layers) \[ \rightarrow A > \beta \ \Delta^3 = 80,000 \text{ m}^2/\text{s} \]

Diffusive CFL \[ \rightarrow A < 0.5 \ \Delta^2 / \Delta t = 8000,000 \text{ m}^2/\text{s} \]

Realism (EUC, WBC) \[ \rightarrow A \sim \text{physical} = 1,000 \text{ m}^2/\text{s} \]

Smagorinsky \[ \rightarrow A = C \ \Delta^2 \int (\partial_x U)^2 + (\partial_y V)^2 + (\partial_x V + \partial_y U)^2 \]
ANISOTROPIC HORIZONTAL VISCOSITY

\[
\begin{align*}
    \partial_t u + \ldots &= \partial_x (A \partial_x u) + \partial_y (B \partial_y u) \\
    \partial_t v + \ldots &= \partial_x (B \partial_x v) + \partial_y (A \partial_y v)
\end{align*}
\]

Grid Re (Diffuse Noise) → Live with the “noise”

Resolve WBC (Munk Layers) → \( A = B = \beta \Delta^3 \), only near WBC

elsewhere:

Realism (EUC, WBC) → 
\[
\begin{align*}
    A &= 300 \text{ m}^2/\text{s} \\
    B &= 300 \text{ m}^2/\text{s} \text{ in the tropics} \\
    &= 600 \text{ m}^2/\text{s} \text{ polewards of } 30^\circ
\end{align*}
\]

Subject to diffusive CFL, but NO Smagorinsky
ANISOTROPIC HORIZONTAL VISCOSITY at 100-m DEPTH

CCSM4 Ocean:
- Minimally Numerically Viscous
- Maximally Physically Viscous

Large et al. (2001, JPO), Jochum et al. (2008, JGR)
IMPACTS ON LABRADOR SEA CIRCULATION AND SEA-ICE

w/ SMAGORINSKY

NO SMAGORINSKY

Jochum et al. (2008, JGR)

T (46.6m)
MESOSCALE EDDY PARAMETERIZATION FOR TRACERS

Ocean Observations suggest mixing along isopycnals is $\sim 10^7$ times larger than across isopycnals. Horizontal mixing causes spurious diapycnal mixing.

An example: The Veronis (1975) effect

For steady WBC ... $w \rho_z = \kappa \rho_{xx}$
Mimics effects of unresolved mesoscale eddies as a sum of
- diffusive mixing of tracers along isopycnals (Redi),
- an additional advection of tracers by an eddy-induced velocity ($u^*$, divergence-free),

Quasi-adiabatic and valid for the ocean interior,

Flattens isopycnals, thereby reducing PE,

Eliminates any need for horizontal diffusion, no Veronis effect.

Gent & McWilliams (1990, JPO)
Here, $T$ is a generic tracer, $s$ is the 2D isopycnal slope vector, and $K$ is the isopycnal diffusion tensor.

There are two diffusivities: $A_I$: isopycnal in $K$, $A_{ITD}$: thickness
IMPACTS ON DEEP WATER FORMATION / CONVECTION

HORIZONTAL MIXING

GM90 PARAMETERIZATION

Danabasoglu et al. (1994, Science)

4°x3°x20L ocean model
NEAR-SURFACE EDDY FLUX (NSEF) SCHEME

GM90 is valid only in the quasi-adiabatic ocean interior, therefore the usual practice has been to taper both $A_I$ and $A_{ITD}$ to zero as the surface is approached.

NSEF replaces the usual approach of applying near-surface taper functions for the diffusivities.

Ferrari et al. (2008, J. Climate), Danabasoglu et al. (2008, J. Climate)
EDDY-INDUCED MERIDIONAL OVERTURNING CIRCULATION

Vertical profiles of zonally-integrated total advective heat transport at 49°S

units: $Sv = 10^6 \text{ m}^3 \text{ s}^{-1}$

Danabasoglu et al. (2008, J. Climate)
SPATIAL VARIATIONS OF THE EDDY DIFFUSIVITIES

Following Ferreira et al. (2005), the diffusivities are specified as

\[ A = A_{REF} \left( \frac{N^2}{N^2_{REF}} \right) \]

\( N^2 \): Local buoyancy frequency,
\( N^2_{REF} \): Reference buoyancy frequency just below the transition layer,
\( A_{REF} \): Constant reference value of A within the surface diabatic region.

\( N_{min} \) is a lower limit, \( N_{min} = 0.1 \).

\( (N^2/N^2_{REF}) = 1 ; A_{REF} \)

\( N_{min} \leq (N^2/N^2_{REF}) \leq 1 \)

Ferreira et al. (2005, JPO)
THICKNESS DIFFUSIVITY

UPPER-OCEAN [0-945 m] MEAN

ZONAL-MEAN

Model Eddy-Induced Transport Comparisons with Roemmich and Gilson (2001) Observational Estimate

(Repeat hydrographic line in the North Pacific at an average latitude of 22°N)

The eddy-induced meridional velocity is given by

\[ v^* = -\frac{\partial \left( A_{ITD} S_y \right)}{\partial z} = -\left( \frac{\partial A_{ITD}}{\partial z} S_y + A_{ITD} \frac{\partial S_y}{\partial z} \right) \]

where \( S_y \) is the meridional slope of the isopycnal surfaces and \( z \) is the vertical coordinate (positive upwards).

Horizontal-mean \( v^* \) profiles computed between 20°N and 40°N in the North Pacific

Danabasoglu & Marshall (2007, Ocean Modelling)
GRAVITY CURRENT OVERFLOWS

Faroe-Bank Channel
Denmark Strait
Greenland
Iceland
Scotland

From J. Price
GRAVITY CURRENT OVERFLOW PARAMETERIZATION

\[ \text{Interior: } \rho_i T_i S_i \]
\[ \text{Inflow: } M_i \]
\[ \text{Outflow: } M_S \]
\[ \text{Source: } \rho_S T_S S_S \]

\[ \text{Entrainment: } \rho_E \]
\[ T_E S_E M_E \]

\[ \text{Product: } \rho_p T_p S_p \]

\[ M_P = M_S + M_E \]

\[ \text{Surface, } h_e \]

\[ \text{Slope, } \alpha \]

\[ \text{Sill Depth, } d_s \]
\[ \text{Width, } W_s \]

\[ \text{Sill Break, } X_{\text{ssb}} \]

Reduced Gravities:
\[ g_s' = \left(\frac{g}{\rho_o}\right) (\rho_s - \rho_i) \]
\[ g_e' = \left(\frac{g}{\rho_o}\right) (\rho_s - \rho_e) \]

Based on Price & Yang (1998); described in Briegleb et al. (2010, NCAR Tech. Note) and Danabasoglu et al. (2010, JGR)
OVERFLOW PARAMETERIZATION MODEL SCHEMATIC
BOTTOM TOPOGRAPHY OF THE x1 RESOLUTION OCEAN MODEL
VERIFICATION AND IMPACTS OF THE OVERFLOW PARAMETERIZATION

<table>
<thead>
<tr>
<th>UNCOUPLED</th>
<th>COUPLED</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control (no overflows)</td>
<td>OCN</td>
</tr>
<tr>
<td>With overflows</td>
<td>OCN*</td>
</tr>
</tbody>
</table>

Each experiment is run for 170 years.
# NORDIC SEA OVERFLOW TRANSPORTS

<table>
<thead>
<tr>
<th>All in Sv</th>
<th>Verification Steps</th>
<th></th>
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<tbody>
<tr>
<td></td>
<td>Observed</td>
<td>Diagnostic model (offline)</td>
<td>OCN*</td>
<td>CCSM*</td>
</tr>
<tr>
<td>$M_I$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$M_S$</td>
<td>4.1 - 7.5</td>
<td>5.2</td>
<td>4.7</td>
<td>4.4</td>
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<tr>
<td>$M_E$</td>
<td>1.5 – 3.7</td>
<td>1.2</td>
<td>0.9</td>
<td>1.1</td>
</tr>
<tr>
<td>$M_P$</td>
<td>6.4 – 9.4</td>
<td>6.4</td>
<td>5.6</td>
<td>5.5</td>
</tr>
</tbody>
</table>
ATLANTIC MERIDIONAL OVERTURNING CIRCULATION (AMOC)
AMOC TRANSPORT AT 26.5°N

RAPID is observational data
TEMPERATURE AND SALINITY DIFFERENCES FROM OBSERVATIONS AT 2649-m DEPTH

Obs: Levitus et al. (1998), Steele et al. (2001)
ZONAL VELOCITY ACROSS 69°W IN THE NORTH ATLANTIC

velocity: cm/s, density: kg/m³
# EQUATORWARD VOLUME TRANSPORTS

<table>
<thead>
<tr>
<th>$\sigma_o \geq$</th>
<th>44°W 27.80</th>
<th>49.3°W 27.80</th>
<th>49.3°W 27.74</th>
<th>69°W 27.80</th>
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</thead>
<tbody>
<tr>
<td>OCN</td>
<td>5.3</td>
<td>3.5</td>
<td>17.3</td>
<td>0.2</td>
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<tr>
<td>OCN*</td>
<td>10.7</td>
<td>9.3</td>
<td>26.7</td>
<td>2.0</td>
</tr>
<tr>
<td>OBS</td>
<td>13.3</td>
<td>14.7</td>
<td>26 ± 5</td>
<td>12.5</td>
</tr>
</tbody>
</table>

All in Sv
IMPACTS ON SEA-ICE CONCENTRATION

CCSM*

CCSM

% of grid area

MIN = -27.27 MAX = 3.31
THANK YOU