Bergen Layered Ocean Model (BLOM)

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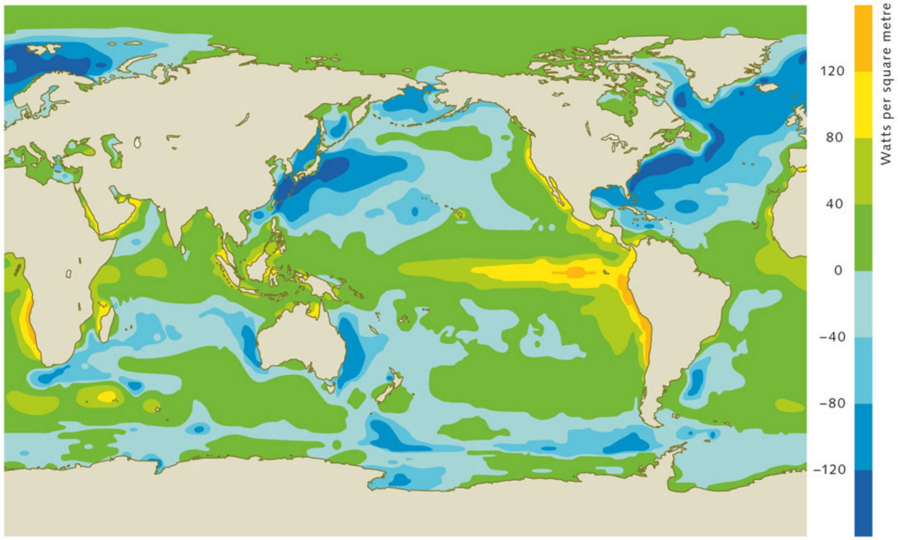
Outline

- Ocean's role in the climate system
- BLOM fundamentals
- BLOM dynamics and physics
- Horizontal grids
- Type of NorESM configurations involving BLOM
- New developments for NorESM3
- Well-known biases
- Namelist options to test out

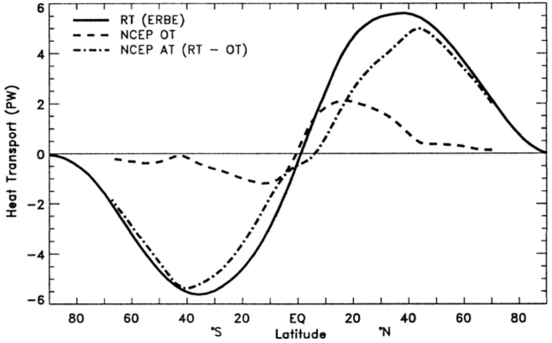
Ocean's role in the climate system

Large capacity heat reservoir

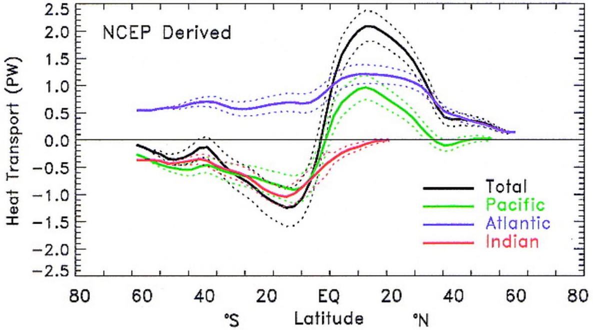


Mean heat exchange between the ocean and atmosphere

Meridional heat transport



Total, atmosphere and ocean

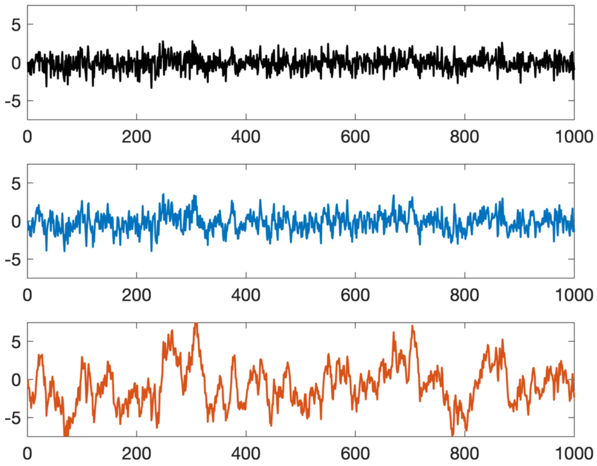


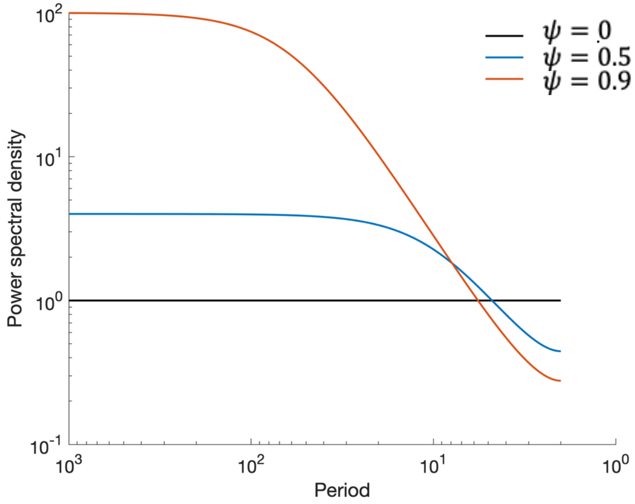
Different basin contributions

Courtesy: http://worldoceanreview.com

Ocean's role in the climate system

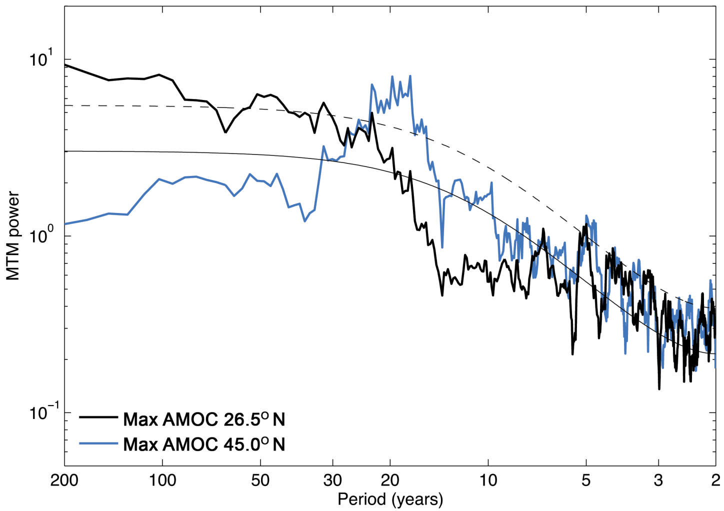
- The ocean is important for weather and climate variability from monthly time scales and upwards.
- Due to mass and thermal inertia, the ocean typically have a red noise response to atmospheric white noise forcing (Hasselmann, 1976).
- Consider a simple 1. order autoregressive model (AR(1)): $X_t = c + \psi X_{t-1} + \varepsilon_t$

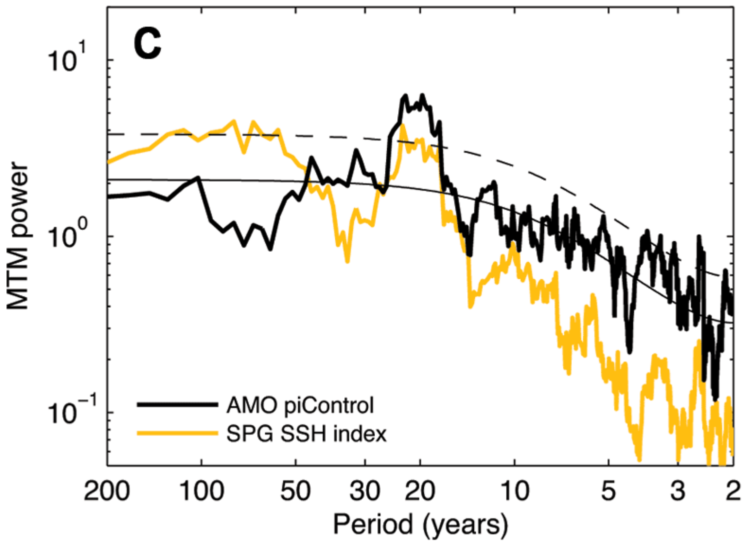




Ocean’s role in the climate system

- Sometimes one finds energy on certain time scales that exceeds what can be expected from a red noise process.
- Can one identify a mechanism that supports an oscillatory behavior?

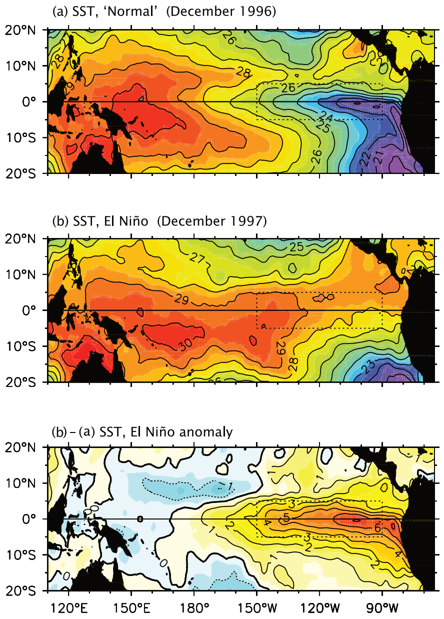


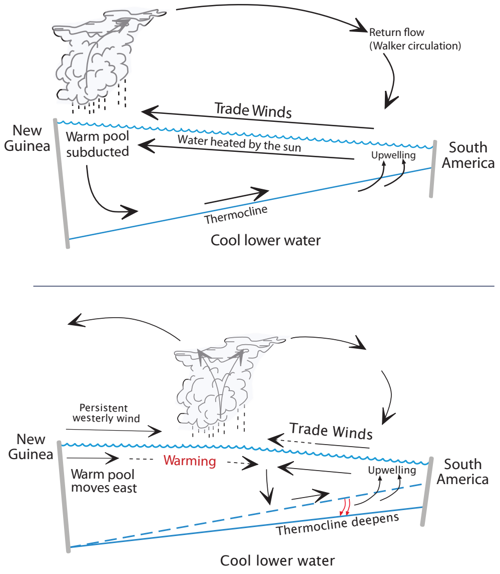


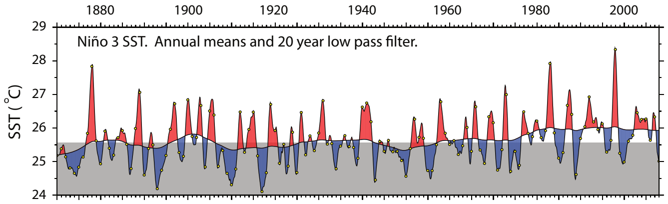
Examples from the NorESM1 pre-industrial control (Bentsen et al., 2013)

Ocean's role in the climate system

- El Niño, the largest and most important source of global interannual climate variability.
- A coupled atmosphere-ocean phenomena.



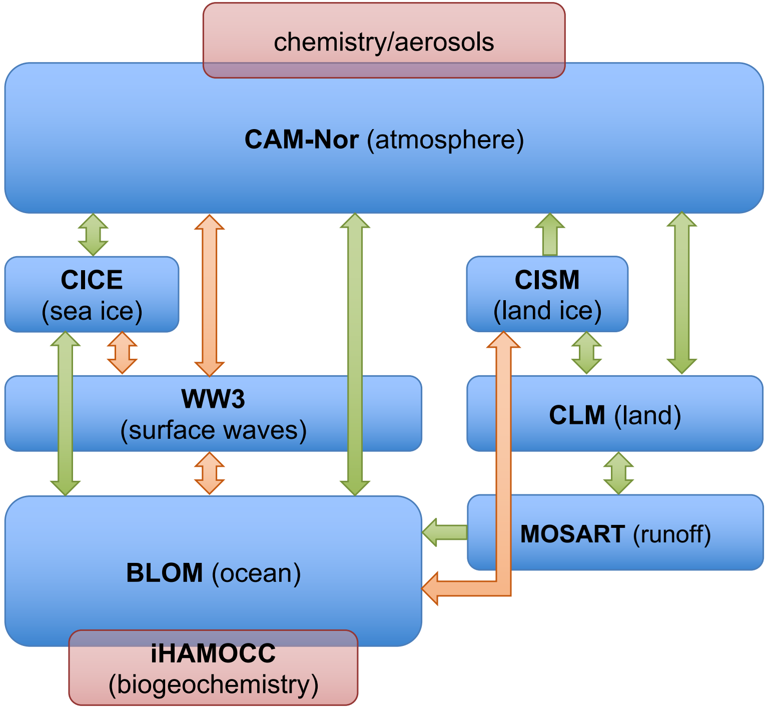


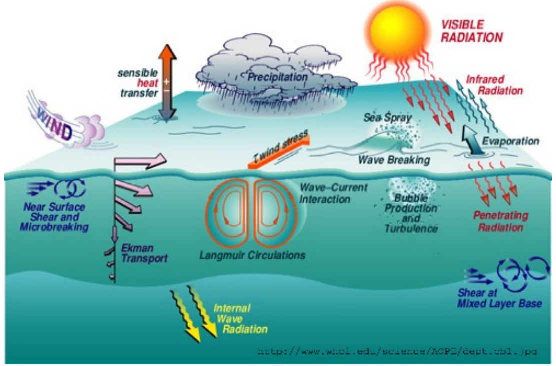


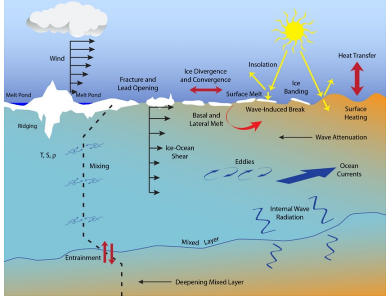
Vallis (2017)

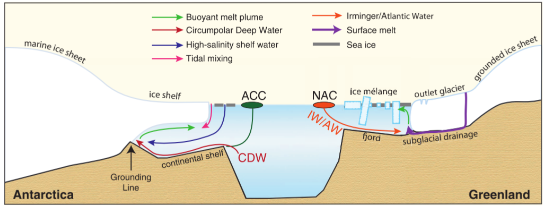
Ocean's role in the climate system

- Interface with other components.









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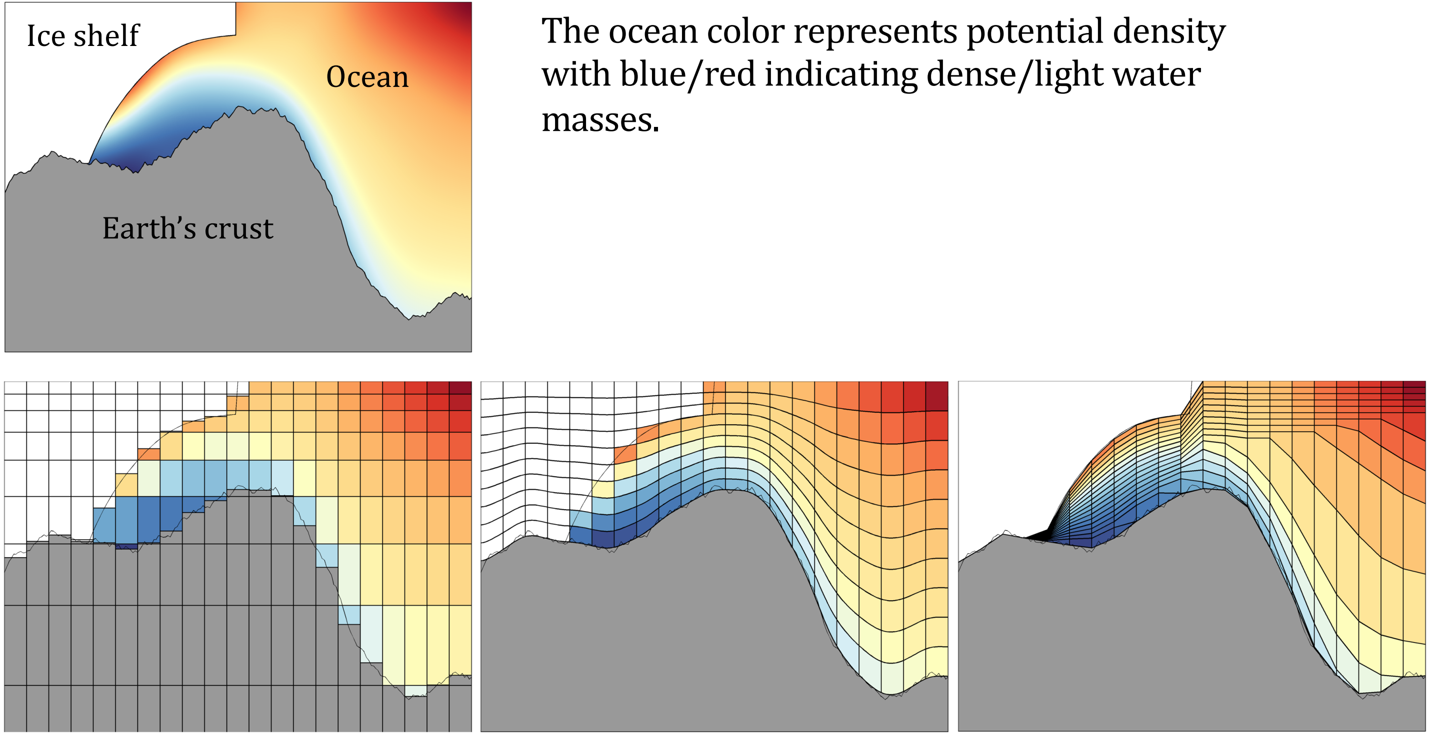
Bergen Layered Ocean Model (BLOM)

- Based on the Miami Isopycnic Coordinate Ocean Model (MICOM).
- Further developed at the Nansen Center and now primarily at NORCE.
- Earlier version used as the ocean component of the Bergen Climate Model (BCM).
- The ocean component of NorESM.

Ocean model vertical coordinate



The ocean color represents potential density with blue/red indicating dense/light water masses.

z-coordinate, partial cells

σ-coordinate, extended ocean

Hybrid coordinate

Primitive equations for ocean applications

A closed set of equations for an inviscid, isentropic and isohaline ocean in local Cartesian coordinates (x, y, z):

$$\frac{d\mathbf{u}}{dt} + f\mathbf{k} \times \mathbf{u} = -\frac{1}{\rho}\nabla_z p,$$

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

$$\frac{1}{\rho}\frac{d\rho}{dt} + \nabla \cdot \mathbf{v} = 0,$$

$$\frac{d\theta}{dt} = 0,$$

$$\frac{dS}{dt} = 0,$$

$$\rho = \rho(\theta, S, p).$$

Assumptions:

- Hydrostatic approximation.
- Shallow-fluid approximation.
- Spherical earth.
- Coriolis terms involving w and uw/r, vw/r terms are neglected.

Equations suitable for a layered model

A closed set of equations for an inviscid, isentropic and isohaline ocean with coordinates (x, y, s), where s is a generalized vertical coordinate:

$$
\begin{aligned}
&\left.\frac{\partial \mathbf{u}}{\partial t}\right|_s + \mathbf{u}\cdot\nabla_s\mathbf{u} + \dot{s}\frac{\partial \mathbf{u}}{\partial s} + f\mathbf{k}\times\mathbf{u} = -\frac{1}{\rho}\nabla_s p - \nabla_s\Phi,\\
&\frac{\partial p}{\partial s} = -\rho\frac{\partial \Phi}{\partial s},\\
&\left.\frac{\partial}{\partial t}\left(\frac{\partial p}{\partial s}\right)\right|_s + \nabla_s\cdot\left(\frac{\partial p}{\partial s}\mathbf{u}\right) + \frac{\partial}{\partial s}\left(\frac{\partial p}{\partial s}\dot{s}\right) = 0,\\
&\left.\frac{\partial}{\partial t}\left(\frac{\partial p}{\partial s}\theta\right)\right|_s + \nabla_s\cdot\left(\frac{\partial p}{\partial s}\theta\mathbf{u}\right) + \frac{\partial}{\partial s}\left(\frac{\partial p}{\partial s}\theta\dot{s}\right) = 0.\\
&\left.\frac{\partial}{\partial t}\left(\frac{\partial p}{\partial s}S\right)\right|_s + \nabla_s\cdot\left(\frac{\partial p}{\partial s}S\mathbf{u}\right) + \frac{\partial}{\partial s}\left(\frac{\partial p}{\partial s}S\dot{s}\right) = 0.\\
&\rho = \rho(\theta, S, p).
\end{aligned}
$$

Same assumptions as for the primitive equations and setting $s = z$, recovers the original equations.

The layer pressure thickness is $\partial p/\partial s$.

Vertically discretized equations for a layered model

We assume s to be monotonically increasing with depth and discrete vertical layers can be formed by setting $\Delta s = 1$:

$$\left.\frac{\partial \mathbf{u}_k}{\partial t}\right|_s + \mathbf{u}_k \cdot \nabla_s \mathbf{u}_k + \dot{s}\frac{\partial \mathbf{u}_k}{\partial s} + f\mathbf{k} \times \mathbf{u}_k = -\nabla_p \Phi_k,$$

$$\frac{\partial p}{\partial s} = -\rho \frac{\partial \Phi}{\partial s},$$

$$\left.\frac{\partial\,(\Delta p_k)}{\partial t}\right|_s + \nabla_s \cdot (\Delta p_k \mathbf{u}_k) + \frac{\partial}{\partial s}\left(\frac{\partial p}{\partial s}\dot{s}\right) = 0,$$

$$\left.\frac{\partial\,(\Delta p_k \theta_k)}{\partial t}\right|_s + \nabla_s \cdot (\Delta p_k \theta_k \mathbf{u}_k) + \frac{\partial}{\partial s}\left(\frac{\partial p}{\partial s}\theta\dot{s}\right) = 0.$$

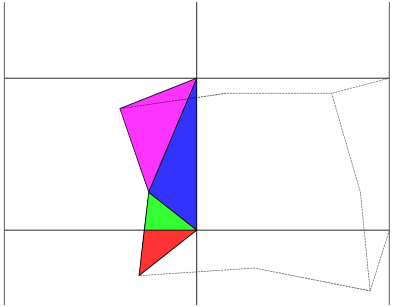
$$\left.\frac{\partial\,(\Delta p_k S_k)}{\partial t}\right|_s + \nabla_s \cdot (\Delta p_k S_k \mathbf{u}_k) + \frac{\partial}{\partial s}\left(\frac{\partial p}{\partial s}S\dot{s}\right) = 0.$$

$$\rho = \rho(\theta, S, p).$$

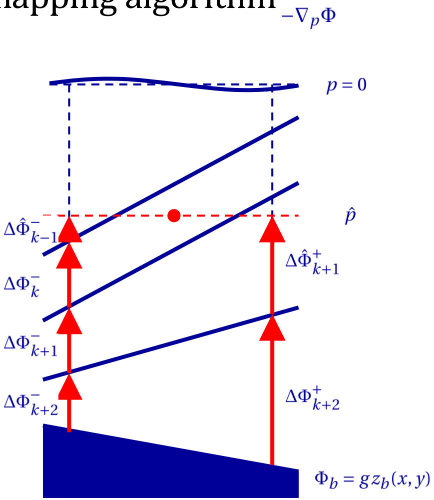
Specifying $\sigma_k = \sigma(s_k)$ and $\dot{s} = 0$ defines an adiabatic isopycnic model (ignoring a slight inconsistency that $\sigma(\theta_k, S_k)$ may drift away from σ_k).

BLOM dynamical core

- Mass conserving formulation (non-Boussinesq).
- Leap-frog and forward-backward time-stepping for the baroclinic and barotropic mode, respectively.
- Arakawa C-grid horizontal discretization.
- Momentum equations formulated in vector invariant form and solved with a potential vorticity/enstrophy conserving scheme (Sadourny, 1975).
- Layer thickness and tracer advection by incremental remapping (Dukowicz and Baumgardner, 2000).*
- Accurate vertical integration of the in-situ density in the evaluation of the pressure gradient force.*



Fluxing area in the incremental remapping algorithm



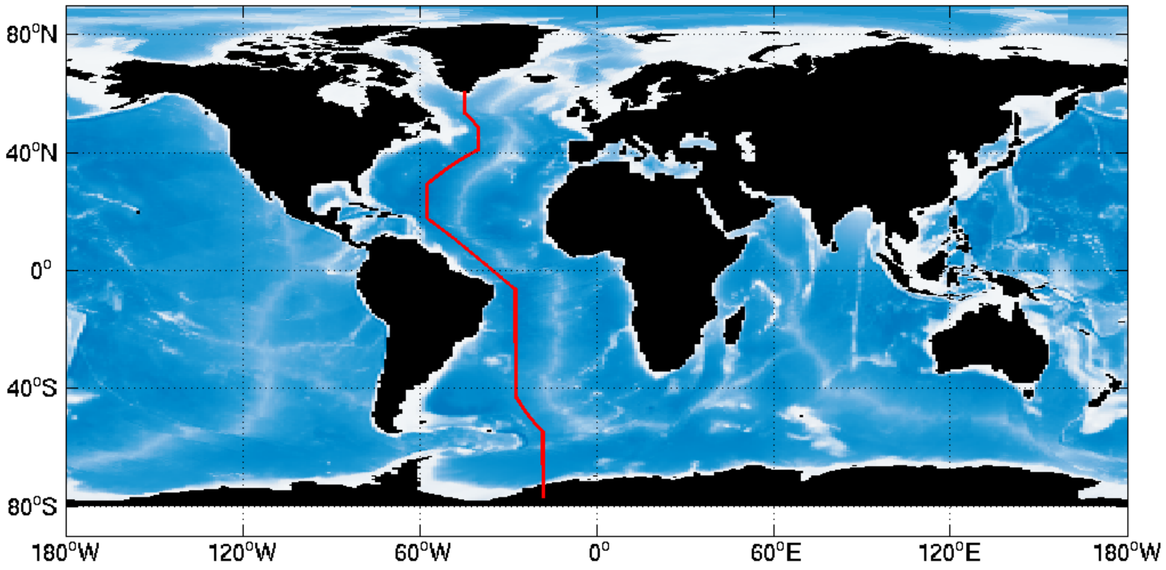
* Updated in NorESM3

BLOM physical parameterizations (NorESM2)

- Surface bulk mixed layer:
  - The depth is estimated with a Kraus-Turner type turbulent kinetic energy model (Oberhuber, 1993)
  - Extended with a parameterization of restratification by submesoscale eddies (Fox-Kemper et al., 2008).
- Eddy mixing:
  - The parameterization of thickness (Gent and McWilliams (GM), 1990) and isopycnal eddy diffusivities follows the diagnostic version of the eddy closure of Eden and Greatbatch (2008).
  - The eddy diffusivities are reduced when the grid resolved the first baroclinic Rossby radius.
- Diapycnal mixing:
  - Background diffusivity is vertically constant but with a latitude dependence following Gregg et al. (2003).
  - k-ε model for shear driven vertical mixing.
  - Mixing driven by energy extracted from the mean flow by bottom drag.
  - Tidally driven mixing according to Simmons et al. (2004).

Animation from a western Atlantic section

The frames are monthly mean variables from a coupled ice-ocean simulation driven with CORE-II forcing. The animations are from the last 20 years of the 5th cycle with forcing for the years 1948-2007.



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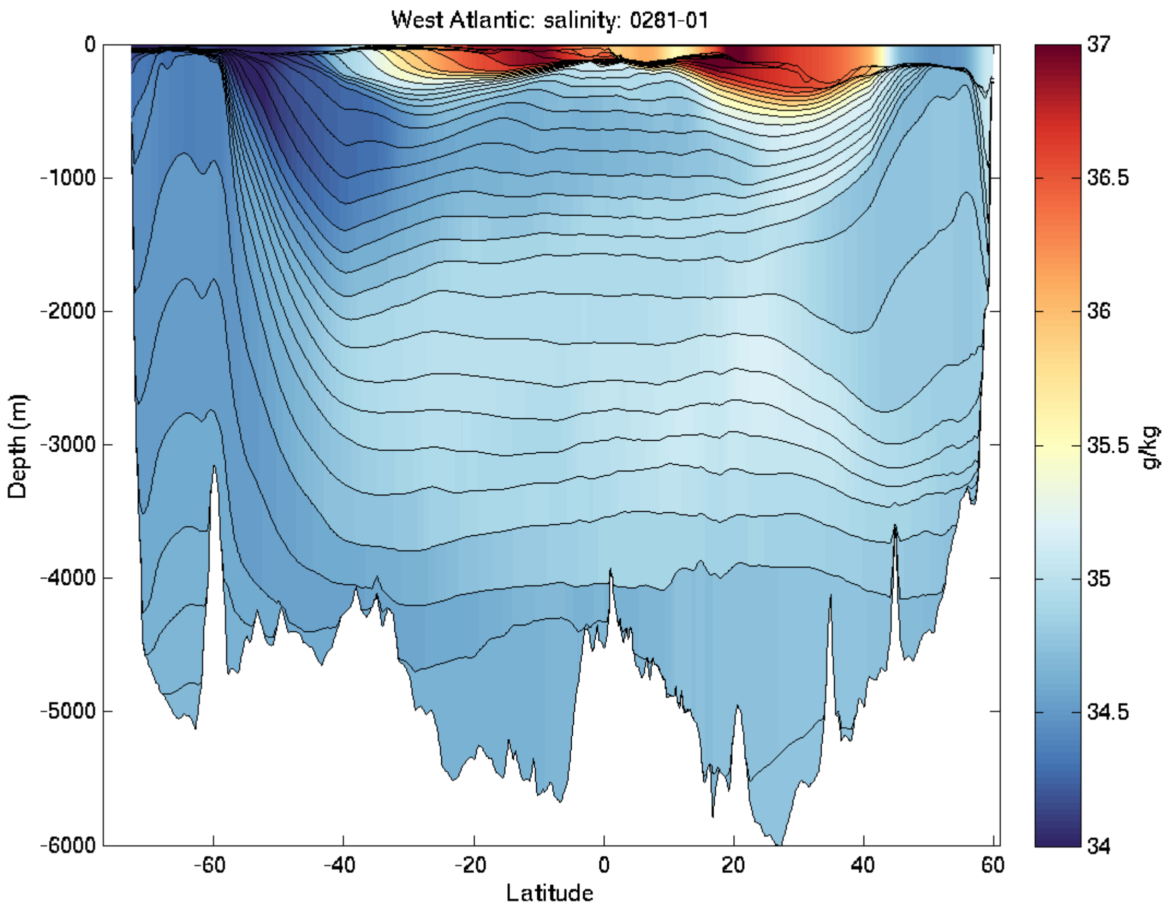
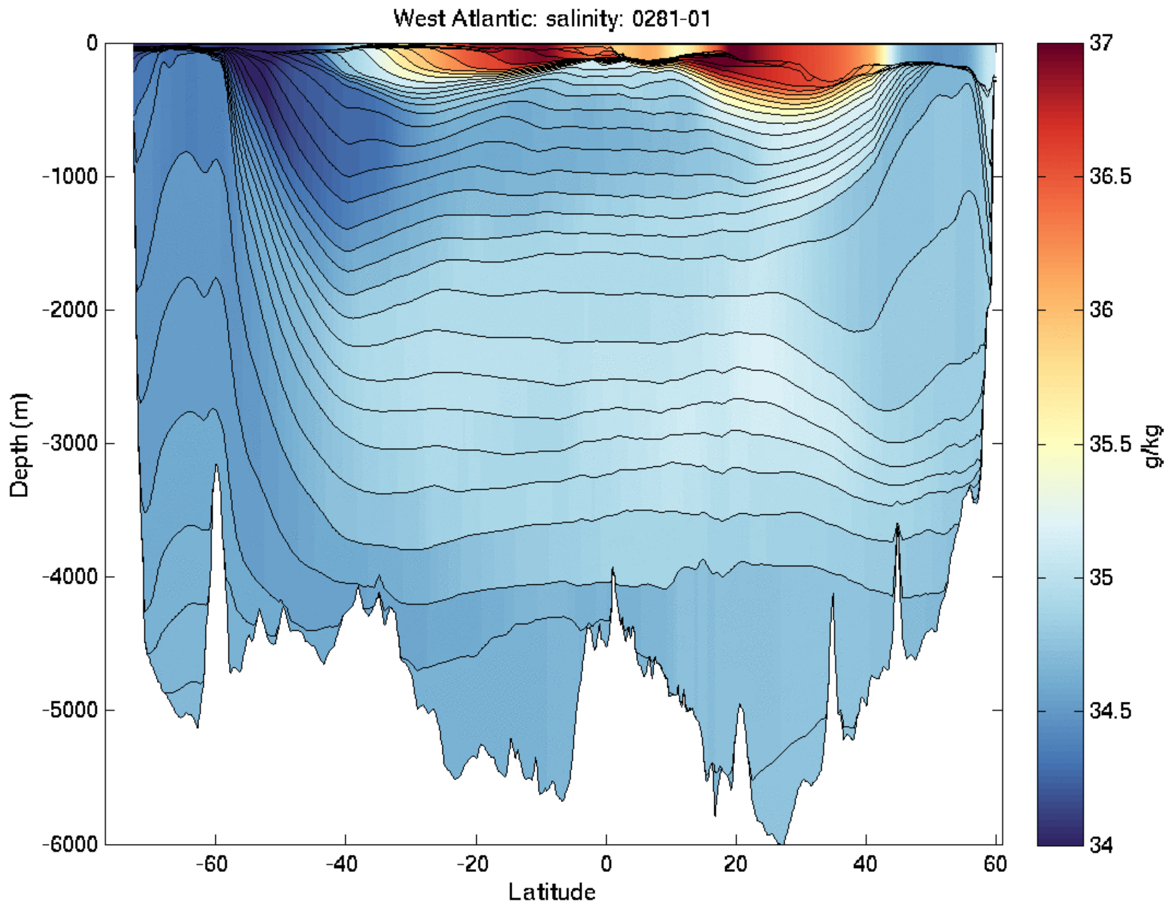

West Atlantic: salinity: 0281-01
Depth (m)
0
-1000
-2000
-3000
-4000
-5000
-6000
-60
-40
-20
0
20
40
60
Latitude
37
36.5
36
35.5
35
34.5
34
g/kg



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36.5
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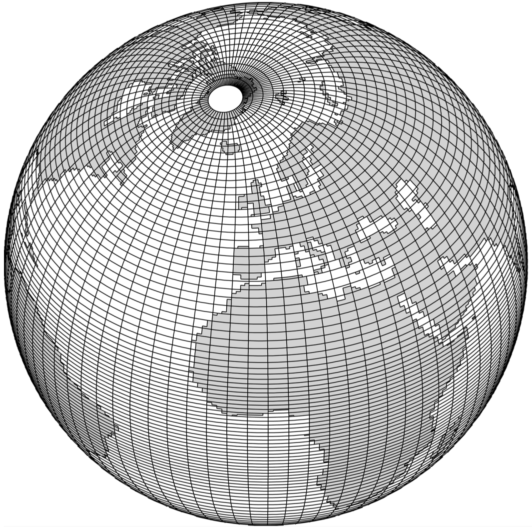
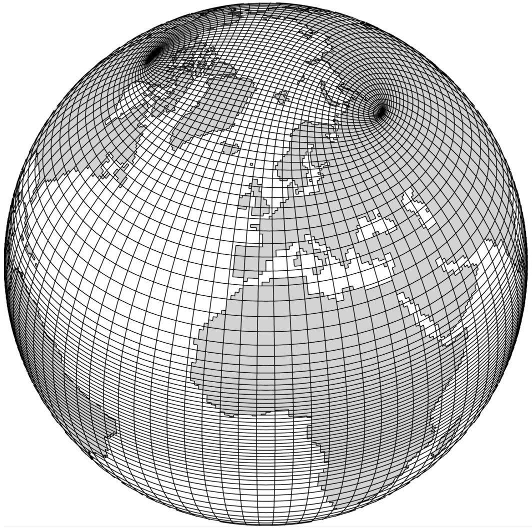
g/kg

Horizontal grids





Other notable grids used:

- 2° tripolar grid
- 0.5° tripolar grid
- 0.25° tripolar grid
- 0.125° tripolar grid

1.125° bipolar grid (every 4th grid line):

- 320 × 384 grid cells.
- Used for the NorESM1 CMIP5 experiments.
- Enhanced meridional resolution near the equator ($f_e = 1/4$).
- Typical grid alias: **gx1v6**

1° tripolar grid (every 4th grid line):

- 360 × 384 grid cells.
- Used for the NorESM2 CMIP6 experiments.
- Enhanced meridional resolution near the equator ($f_e = 1/4$).
- Typical grid alias: **tnx1v4**

Type of NorESM configurations involving BLOM

Coupled ocean, sea-ice simulations:

- Active ocean and sea-ice.
- Using data components for other components.
- Forcing protocols:
  - OMIP1: CORE-II, based on NCEP reanalysis)
  - OMIP2: based on JRA-55 reanalysis)
  - CPLHIST: forcing produced by a fully coupled NorESM simulation
- Long simulations can be carried out by repeating forcing cycles or repeating specific years (RYF).

Fully coupled simulations:

New developments for NorESM3

Dynamics

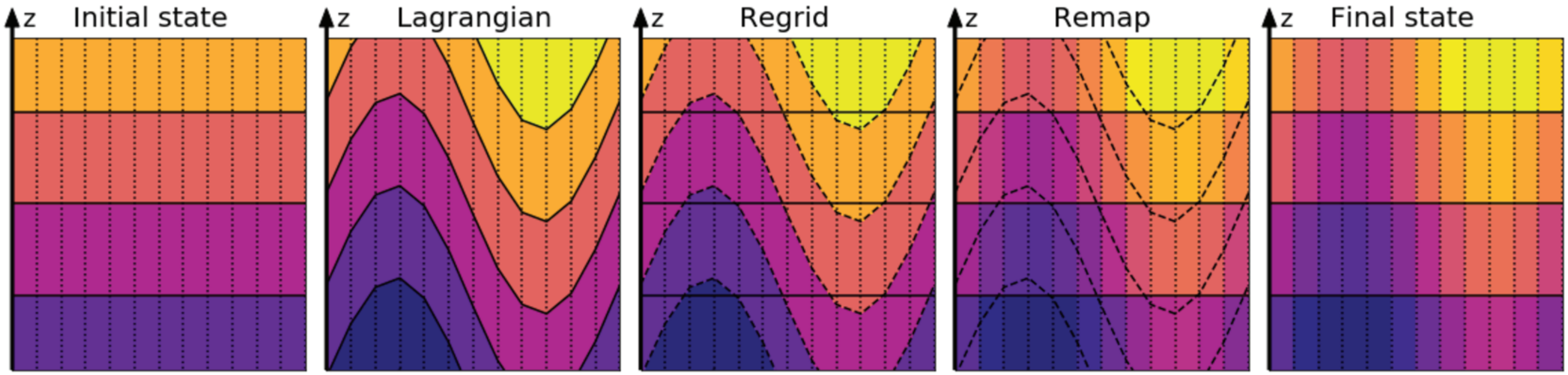
- Hybrid vertical coordinate with the ALE method
- More accurate advection
- Reformulated pressure gradient force
- Adapting reference density classes to model state

Physics

- Eddy parameterization
    - Topography-aware (Nummelin and Isachsen, 2024)
    - Neutral mixing
- Reformulated submesoscale eddy restratification
- KPP vertical mixing scheme
- Refined short-wave absorption options
- Wave effects

The ALE method

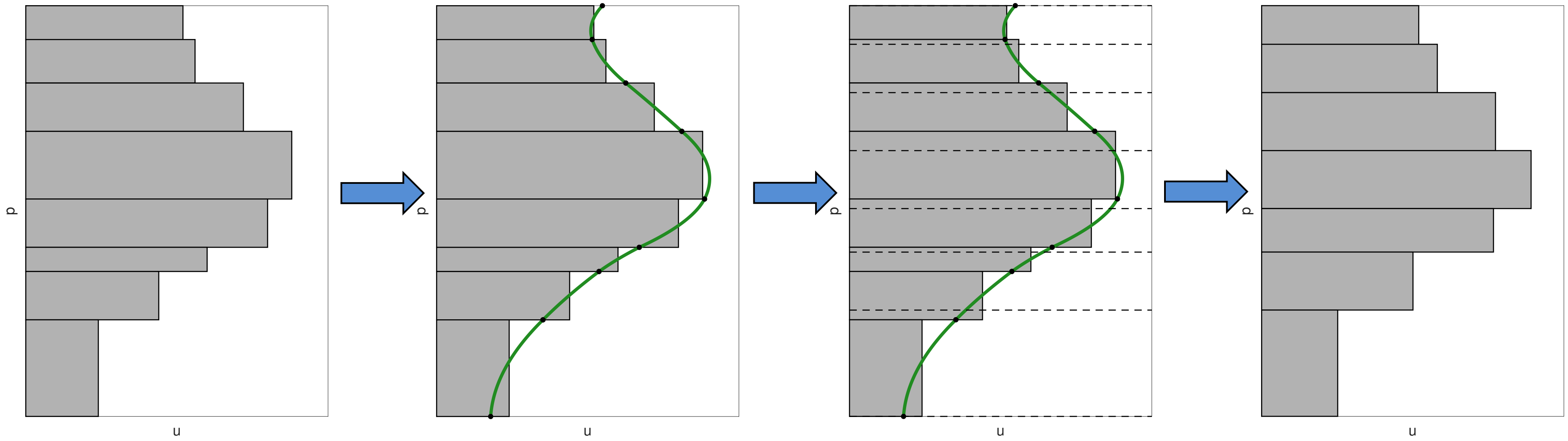
- The **arbitrary Lagrangian-Eulerian** (ALE) method first integrates the layer conservation equations forward in a truly Lagrangian phase, then remap variables to a desired vertical grid in a second phase.



Courtesy: Griffies et al., (2020)

The ALE method

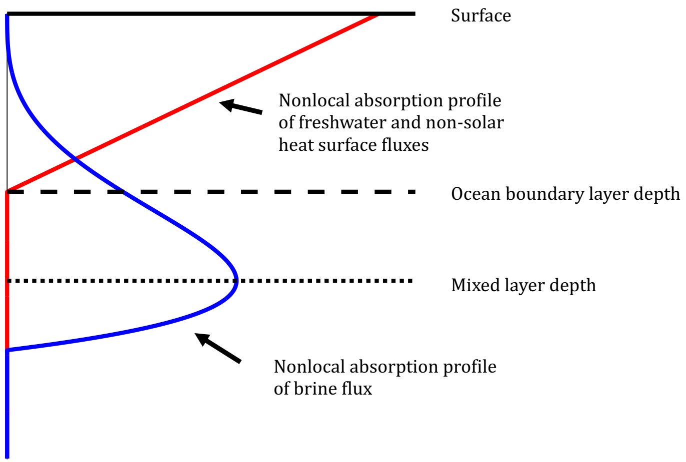
- The **arbitrary Lagrangian-Eulerian** (ALE) method first integrates the layer conservation equations forward in a truly Lagrangian phase, then remap variables to a desired vertical grid in a second phase.
- A suitable regridding approach and accurate remapping is crucial for the application of the ALE method for ocean climate modelling.



Cell-mean data on old grid | High-order reconstruction | Regridding | Cell-mean data after remapping

Physical parameterizations with the ALE method

- Vertical diffusivities are obtained using the nonlocal K Profile Parameterization (**KPP**; Large et al. 1994), as implemented in the **CVMix community package.**
- A deviation from the standard KPP is nonlocal distribution of surface momentum flux and nonlocal distribution of brine from sea-ice freezing.
- The KPP parameterization gives insufficient mixing in gravity currents and work remains to handle this in a satisfactory manner.
- Currently exploring surface wave effects and double diffusion.



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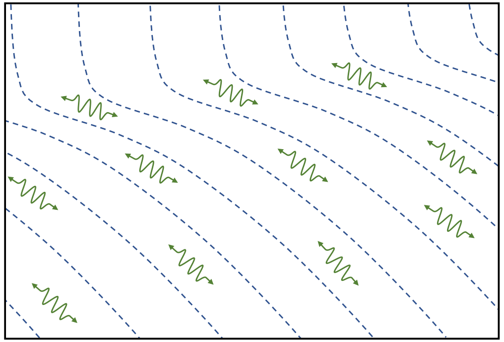
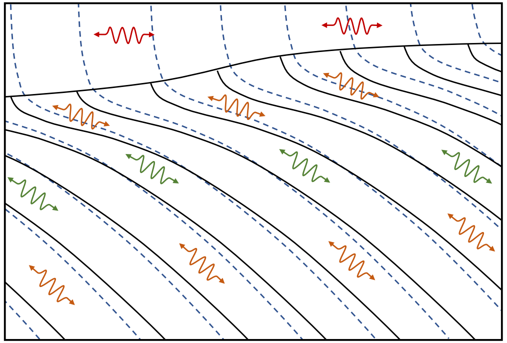
Physical parameterizations with the ALE method

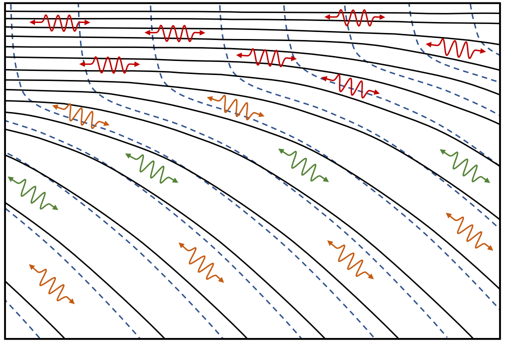
For eddy-induced diffusion with the ALE method, BLOM makes use of neutral diffusion along nonlocal neutral sublayers, as recently proposed by Shao et al. (2020).



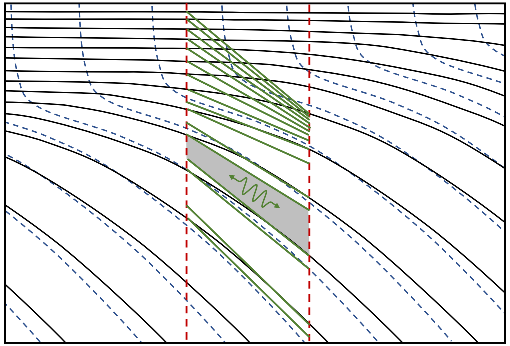
Neutral diffusion



Isopycnic layer diffusion



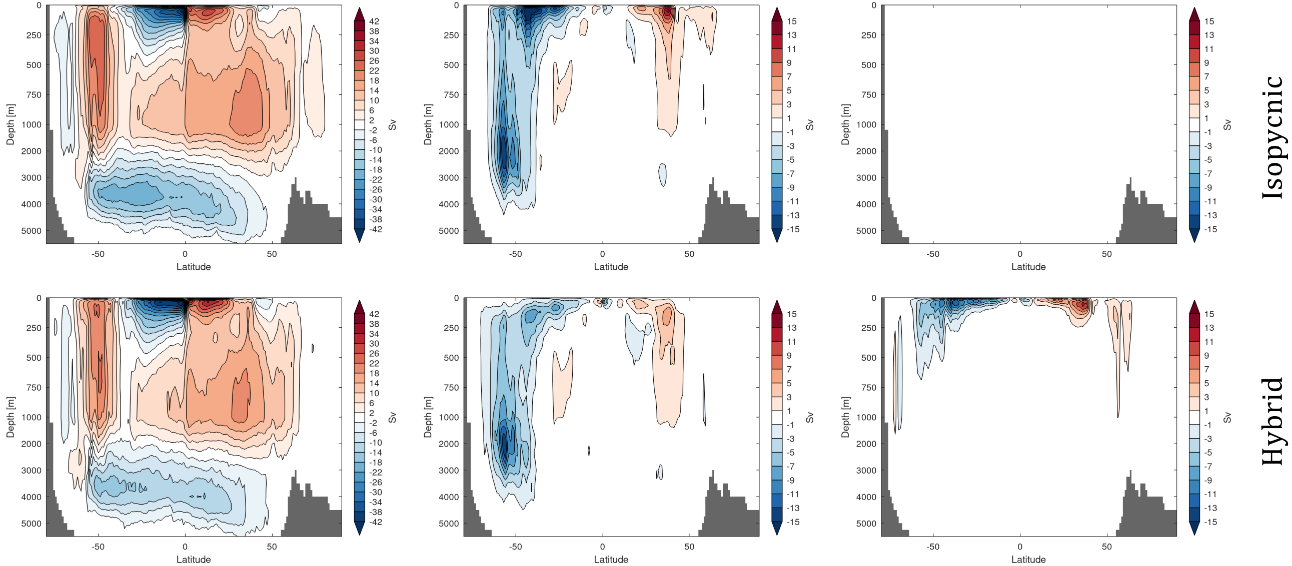
Hybrid layer diffusion



Neutral sublayers

Physical parameterizations with the ALE method

- The parameterization of Gent-McWilliams (**GM**) eddy-induced transport makes use of the neutral slope estimation of the neutral diffusion algorithm.
- With vertical resolution in the PBL with hybrid vertical coordinate, **submesoscale eddy-induced transport** (Fox-Kemper et al., 2008) has now been fully implemented.

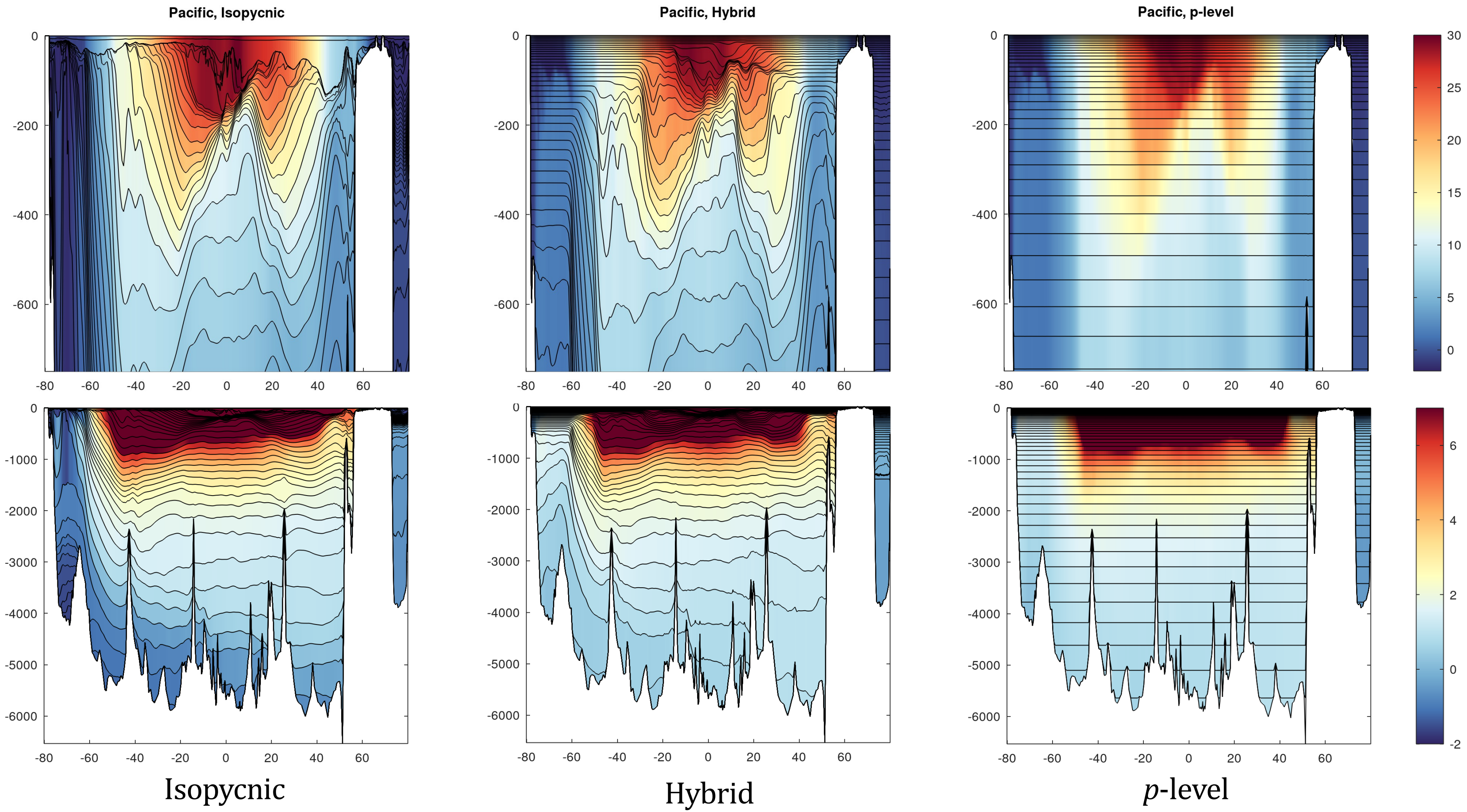


Total meridional overturning circulation (MOC)

Mesoscale eddy-induced MOC

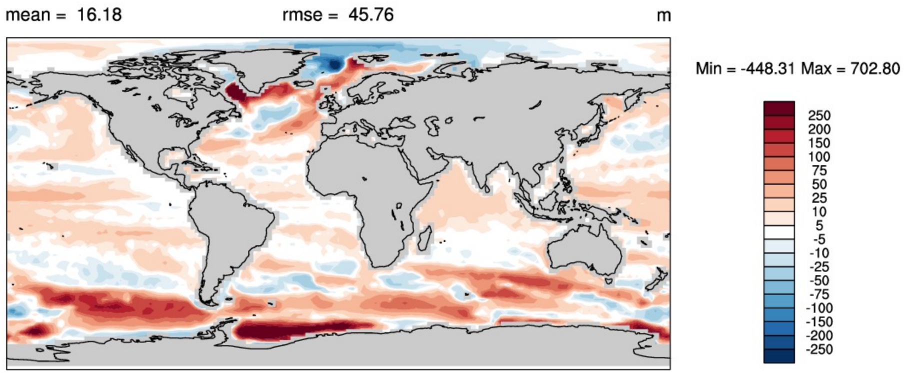
Submesoscale eddy-induced MOC

Potential temperature and layer structure in a Pacific section in the final month of a second OMIP2 forcing cycle.

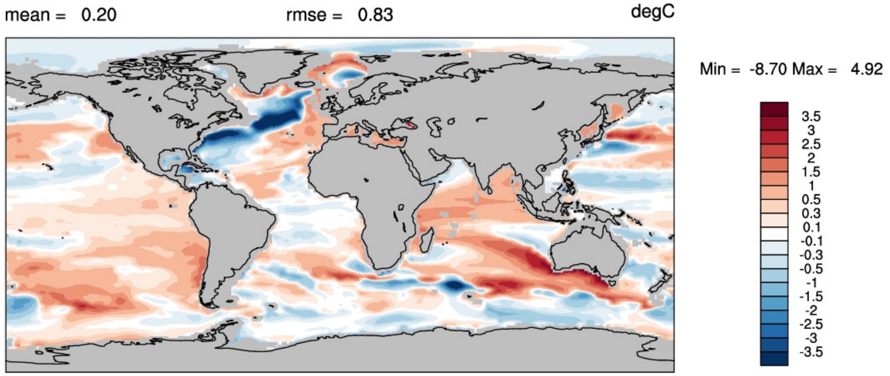


Well-known biases

- Too cold and well-mixed Southern Ocean (NorESM1, NorESM2).
- Deep mixed layer bias (NorESM1, NorESM2).
- Weak Atlantic inflow into the Arctic (NorESM1, NorESM2).
- Upper ocean temperature biases: warm Indian Ocean, cold of the coast of Newfoundland.
- Freshening global mean sea surface salinity in fully coupled simulations.



Mixed layer depth bias in NorESM2 OMIP2 simulation



Temperature bias at 500 m in NorESM3-dev OMIP2 simulation

Namelist options to test out

- Parameters controlling the strength of parameterized eddy transport and mixing (EGC, EGIDFQ) – *impact upper ocean stratification and meridional gradients/exchange*.
- Parameter controlling the strength of submesoscale restratification (CE) – *impacts mixed layer depth*.
- Background vertical diffusivity (BDMC2) – *impacts vertical gradients/exchanges*.