Ocean General Circulation Models

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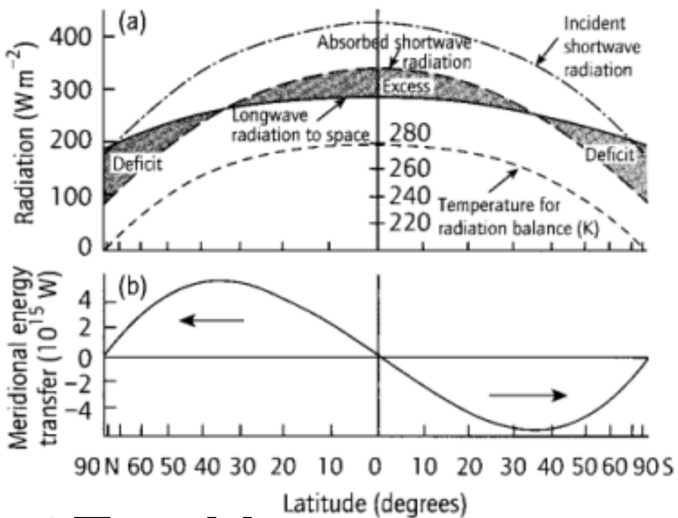
Outline

- The ocean in the Earth system
- Physics and Numerics
  - the equations of motion
  - approximations
  - discretization and model grids
- Parameterization of physical processes
  - Horizontal/vertical mixing
  - convection
- Grid configurations
- Coupling to the atmosphere
- Ocean models in E2SCMS3: MPIOM, NEMO, HADOM

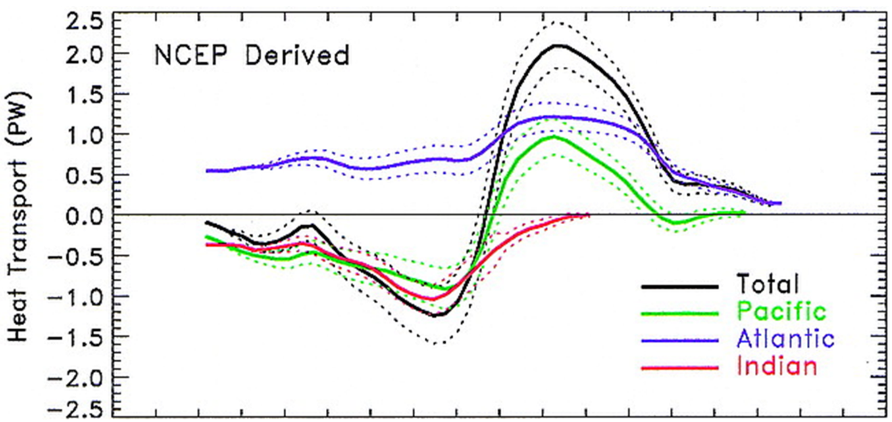
The ocean in the Earth system

- 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, 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 density change from top to bottom is much smaller than the atmosphere. This makes the Rossby radius (NH/f) ("scale of eddies") much smaller: 10s->100s km.
- leading order influence of topography on dynamics.

The ocean in the Earth system



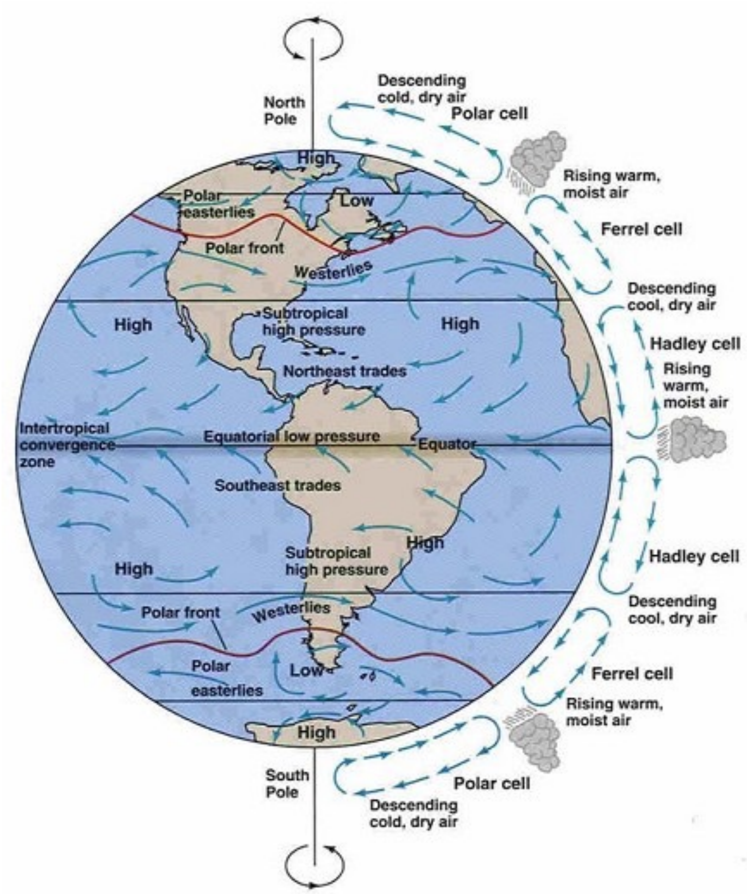
Total heat transport

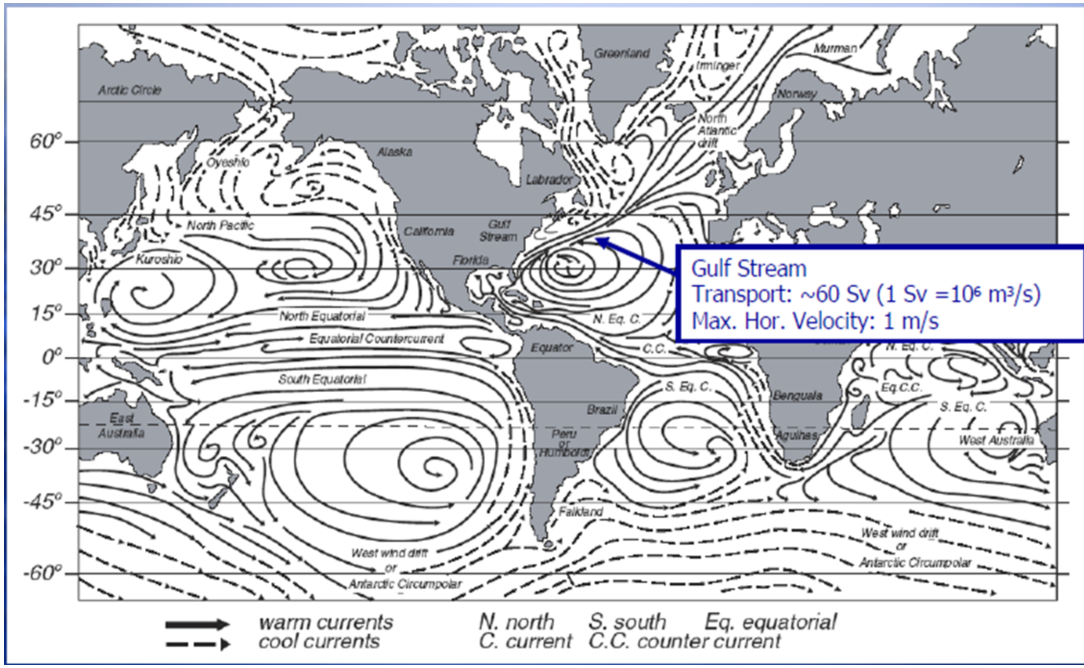


Ocean heat transport

- Atmospheric heat transport mainly due to transient eddies
- Ocean heat transport associated both to large-scale wind driven circulation (Gyres) and thermohaline circulation, driven by density differences
- Due to its large buffering capacity and "slow" time scales the ocean provides the memory of the climate system.

The wind-driven circulation





The wind-driven ocean: Sverdrup transport

Integrate the eq. of motion over the Ekman depth

$$\frac{\partial p}{\partial x} = f\,\rho\,v + \frac{\partial T_{xz}}{\partial z}$$

$$\frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial T_{yz}}{\partial z}$$

$$\frac{\partial P}{\partial x} = f\,M_y + T_x$$

$$\frac{\partial P}{\partial y} = -f\,M_x + T_y$$

$$\frac{\partial M_x}{\partial x} + \frac{\partial M_y}{\partial y} = 0$$

Mx, My: integrated mass transport

Note that the coriolis parameter is a function of latitude (ß = δf/δy)

$$\beta\,M_y = \frac{\partial T_y}{\partial x} - \frac{\partial T_x}{\partial y}$$

$$\beta\,M_y = \mathrm{curl}_z(T)$$

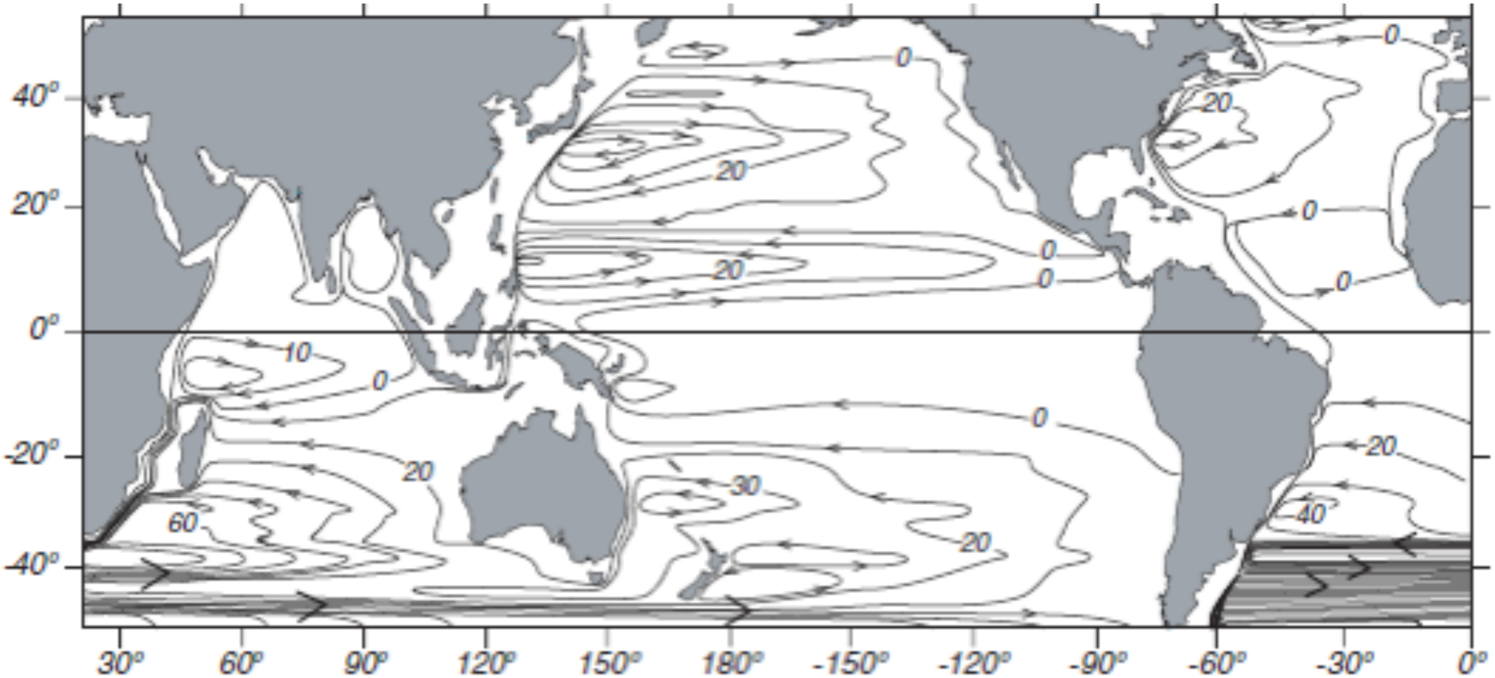
i.e. wind-driven circulation can, to a first order, be calculated just from wind stress

H.U. Sverdrup, 1947

The wind-driven ocean: Sverdrup transport

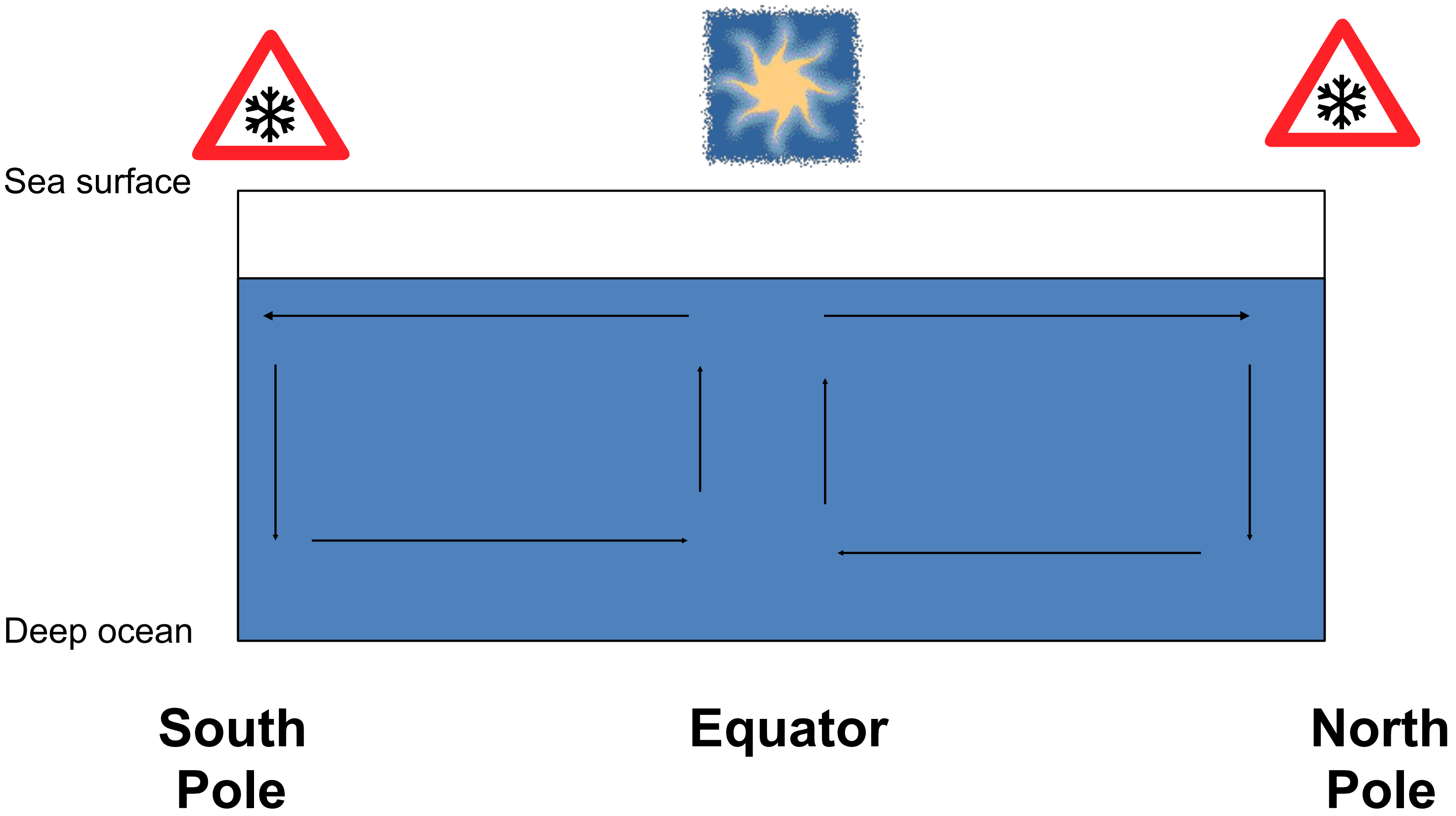
H.U. Sverdrup (1888-1957)

1 Sverdrup (1 Sv) = $1 \times 10^6 m^3 s^{-1}$



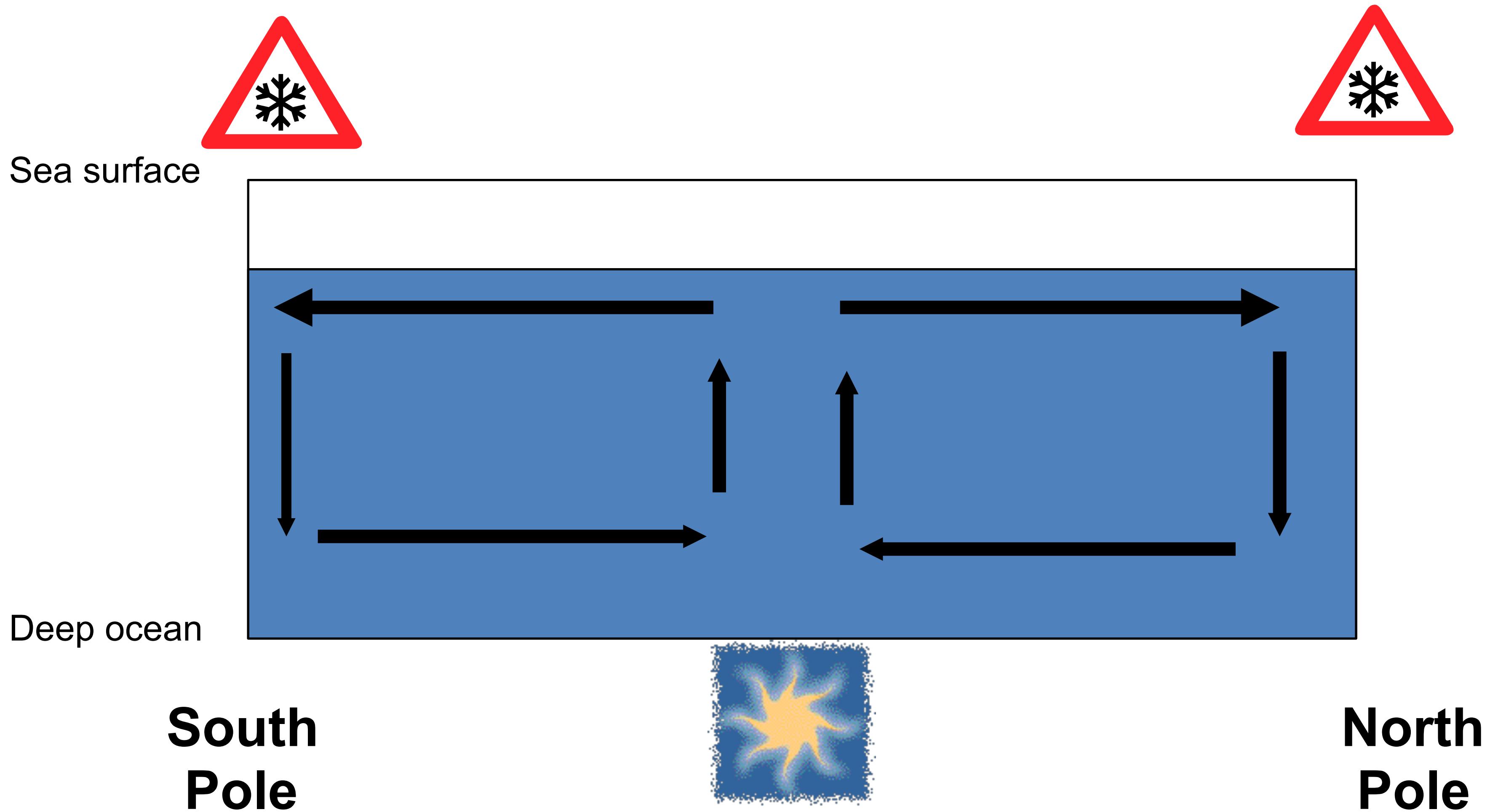
Sverdrup relation explains some elements of the large-scale ocean circulation, but not boundary currents and thermohaline circulation

A thermally-driven System ?



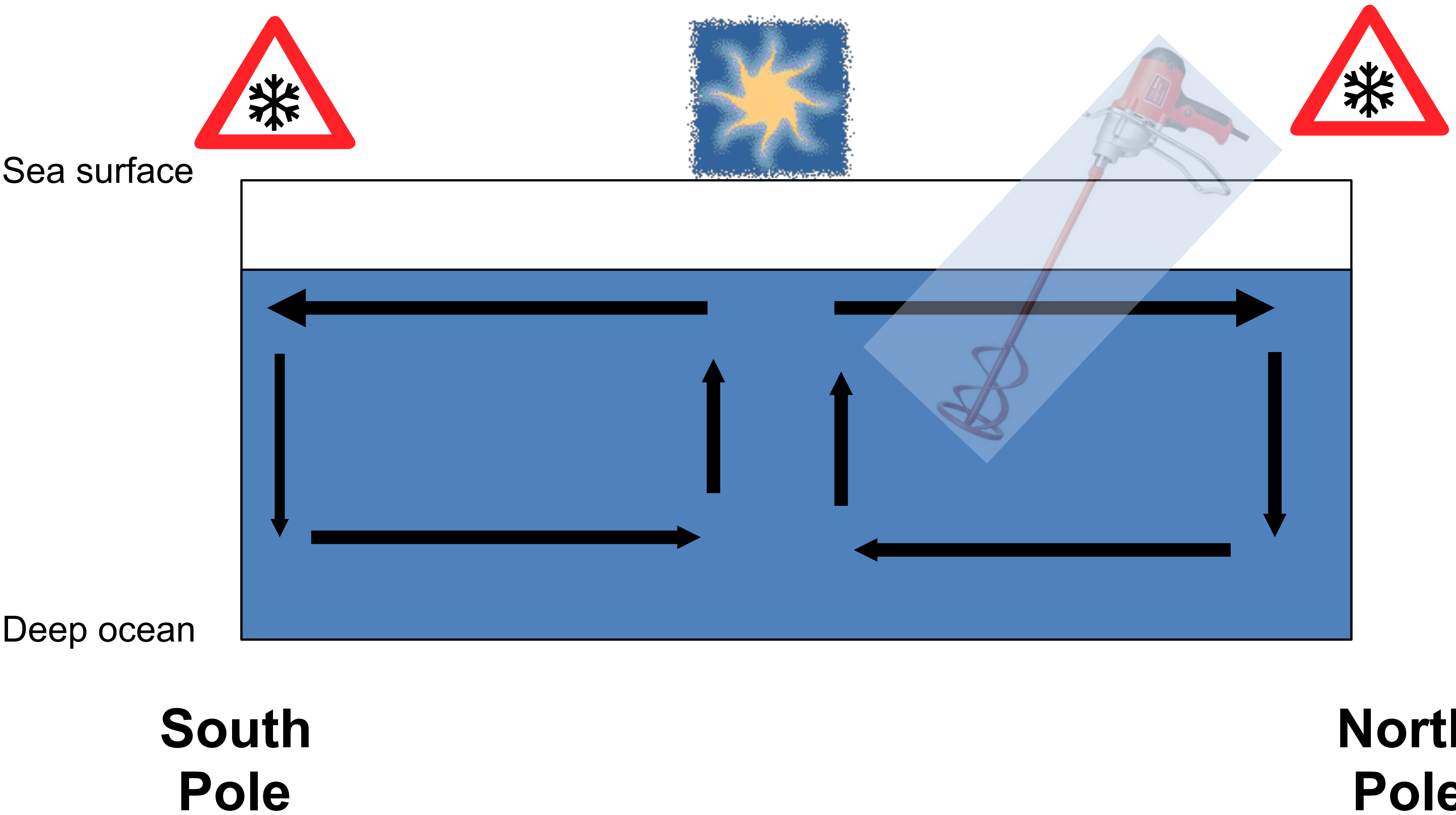
Consideration of energetics (Sandstroem, 1908) indicate a rather weak overturning !

A thermally-driven System ?



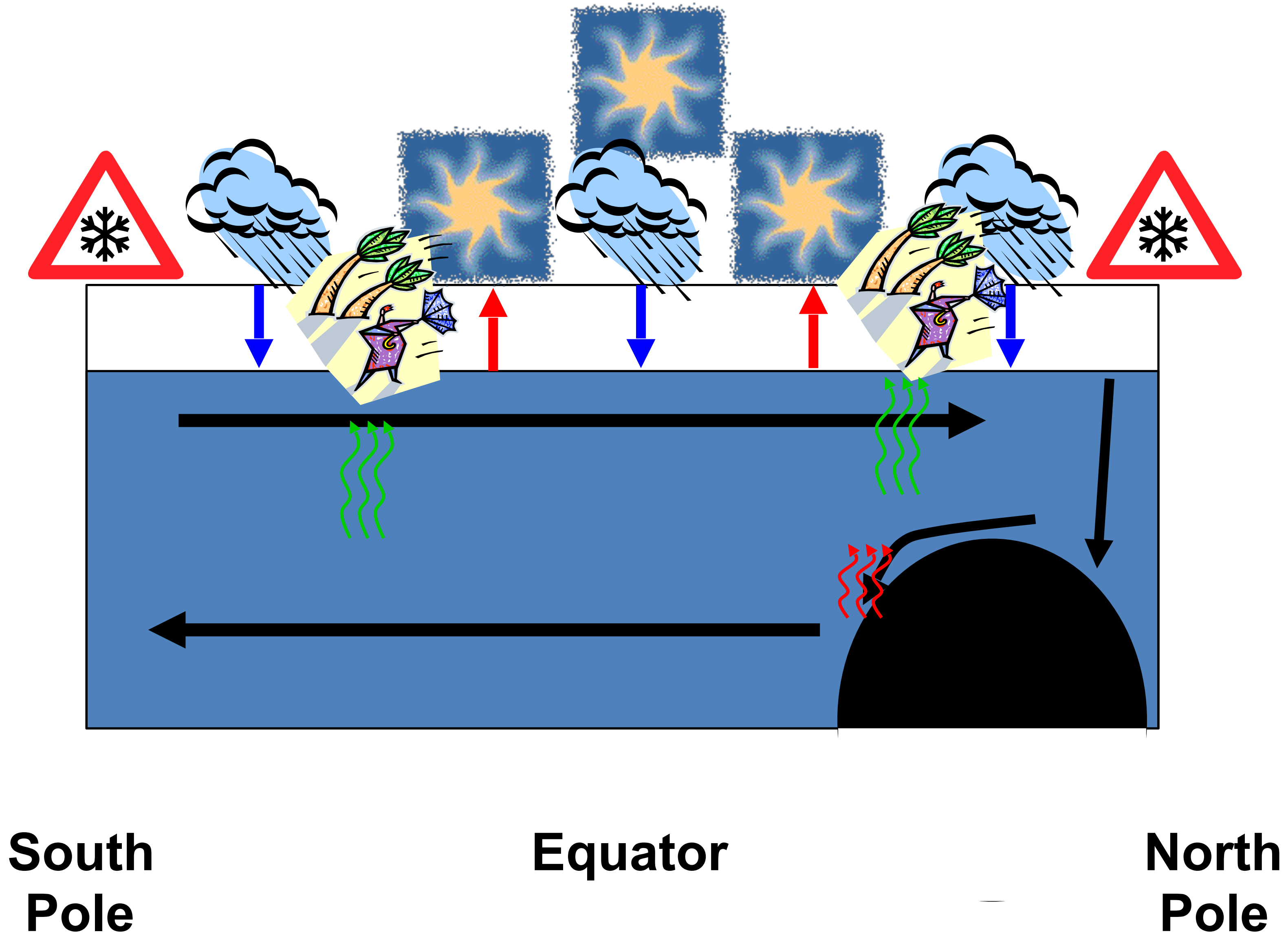
Substantial transport can only be achieved by a source of potential energy

A thermally-driven System ?

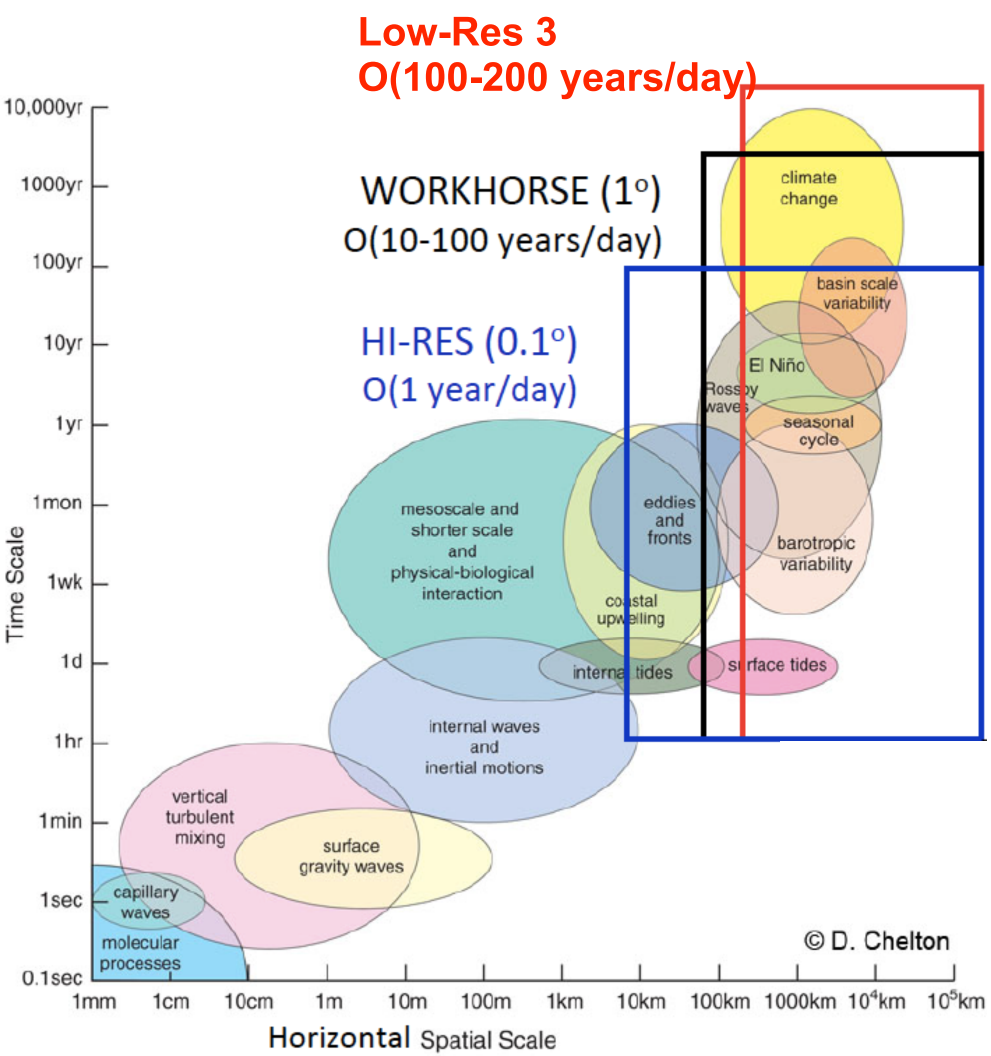


…or kinetic energy (stirring through tides and winds)

The Thermohaline System



Space and time scales



- Current global OGCMs resolve large-scale features
- some (barely) resolve "ocean weather" (eddies)
- many processes remain unresolved and need to be parameterized

Space and time scales

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

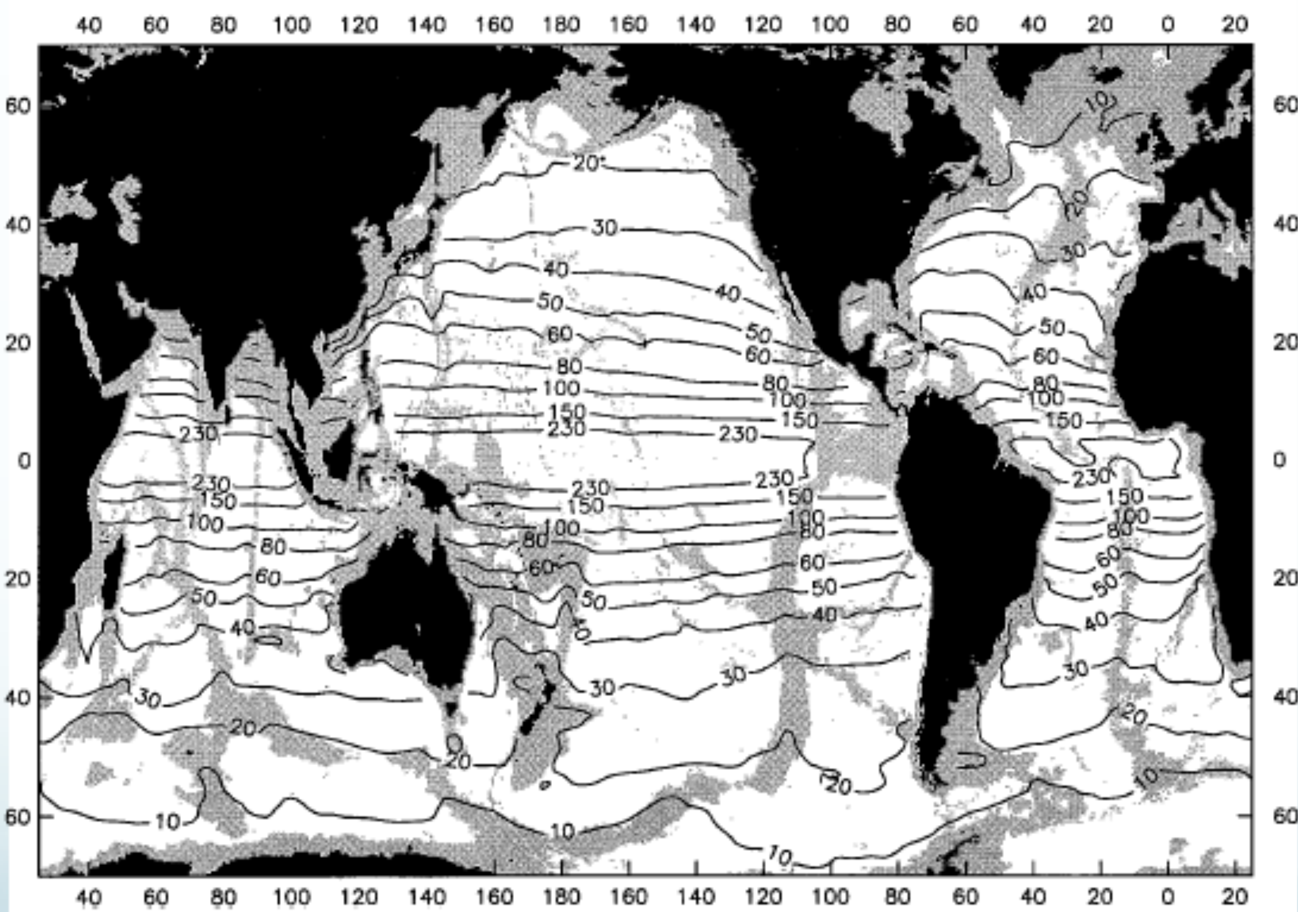
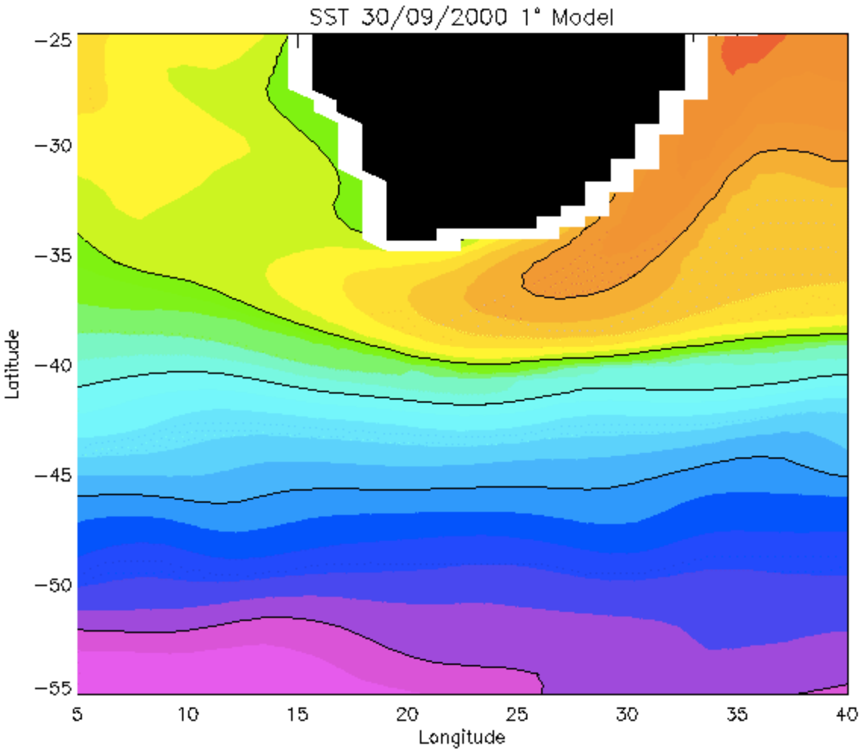


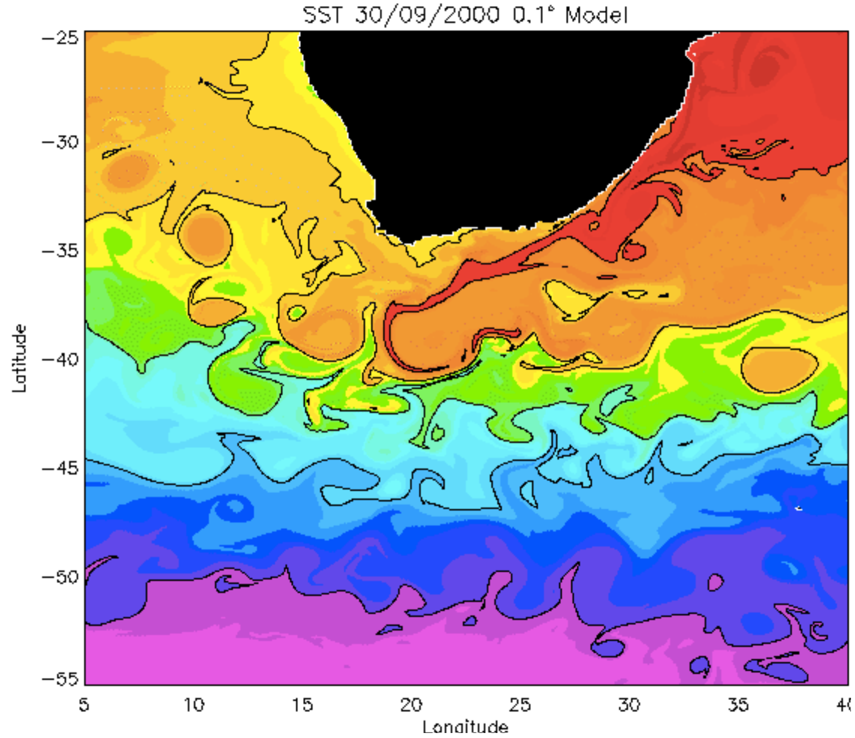
FIG. 6. Global contour map of the 1° × 1° first baroclinic Rossby radius of deformation λ_1 in kilometers computed by Eq. (2.3) from the first baroclinic gravity-wave phase speed shown in Fig. 2. Water depths shallower than 3500 m are shaded.

Chelton et al., JPO, (1998)

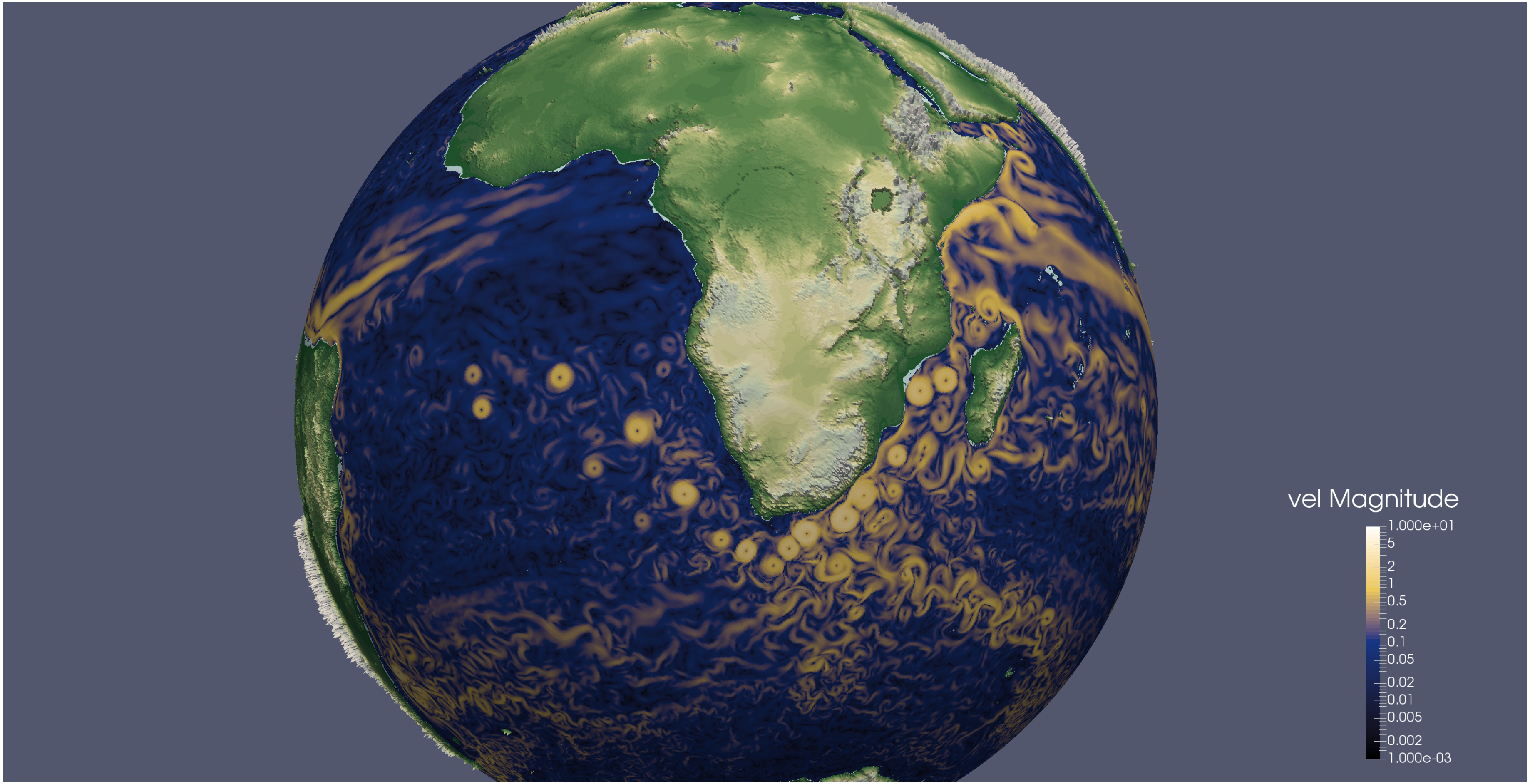
Space and time scales

resolution	lingo	meaning
≥ 1°	"coarse"	no eddies
~ 0.5°	"eddy-permitting"	some eddies
≤ 0.2°	"eddy-resolving"	eddies generate at realistic strength and rate





Space and time scales



MPI ICON Ocean Model

Physics of ocean modelling: Eq. of Motion

Eq. of motion for a geophysical fluid on a rotating sphere

$$\frac{D\mathbf{v}}{Dt} = 2\Omega \sin\theta \times \mathbf{v} - \frac{1}{\rho}\nabla p + \boldsymbol{g} + \nabla \phi_t + \frac{\boldsymbol{F}_{fr}}{\rho}$$

- 1: acceleration term: local rate of change + advection
- 2: Coriolis term
- 3: acceleration due to pressure gradients
- 4: acceleration due to gravity
- 5: Tidal potential
- 6: Frictional forces

Boussinesq Approximation

$$\frac{\Delta\rho}{\rho_0} << 1$$

$$\rho_0 = 1025 kg/m3$$
$$\Delta\rho = 1 kg/m3$$

Density variations are small in the ocean

=> Density is replaced by ρ_o except in terms involving the gravitational acceleration g

- Approximation is made in the horizontal, but not the vertical momentum equations
- Effect on the mass of the fluid can be neglected, but not their effect on the weight (mass times gravity) of the fluid parcel
- Volume is conserved rather than mass
- Sound waves are filtered out

Hydrostatic Approximation

Vertical accelerations are small relative to the gravitational acceleration

- Vertical component of the Equation of Motion simplifies to:

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

- Ocean width is so much greater than its depth => Hydrostatic Approximation generally valid for large-scale ocean circulation
- Non-hydrostatic processes, e.g. convection, need to be parameterized

Equation of continuity

$$\nabla \cdot \mathbf{v} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$

- Conservation of volume (Incompressibility)
- Vertical velocity no longer prognostic variable, but determined from divergence of horizontal velocity components

Primitive Equations

Mathematical description of the ocean circulation by a coupled set of seven equations:

- Momentum Equations **u,v**
- Hydrostatic Equation **p**
- Continuity Equation **w**, $\delta\zeta/\delta t$
- Temperature Equation (representing change in temperature with time) **T**
- Salinity Equation (representing change in salinity with time) **S**
- Density Equation (calculating density from temperature and salinity) ρ

Barotropic and baroclinic sub systems

Horizontal velocities are decomposed into a vertically averaged part and the deviation from this.

$$\mathbf{V} = \int_{-H}^{0} \mathbf{v}\, dz$$

Barotropic sub system

$$\mathbf{v}' = \mathbf{v} - \frac{1}{H}\int_{-H}^{0} \mathbf{v}\, dz$$

Baroclinic sub system

- Decomposition yields one set of equations for each sub system
- Barotropic system is characterised by the phase speed of fast surface waves c=SQRT(gH)
- Internal dynamics are much slower (e.g. „Internal" wave speed of a two-layer fluid: c=SQRT(($g\,\Delta\rho/\rho_0$)H)

Barotropic sub-system

The vertically integrated momentum equation

$$\frac{\partial U}{\partial t} - fV + gH\frac{\partial \zeta}{\partial x} + \int_{-H}^{\zeta} \frac{\partial}{\partial x} p' dz = 0$$

$$\frac{\partial V}{\partial t} + fU + gH\frac{\partial \zeta}{\partial y} + \int_{-H}^{\zeta} \frac{\partial}{\partial y} p' dz = 0$$

The barotropic continuity equation

$$\frac{\partial \zeta}{\partial t} + \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = Q_\zeta$$

with $Q_\xi = E\ P + R$

OGCMs often solve the barotropic sub system semi-implicitly (i.e. the discretisation contains variables at the old and new time step), requiring the inversion of a matrix.

Baroclinic sub-system

The baroclinic momentum equation

$$\frac{\partial u'}{\partial t} - fv' = \frac{1}{H}\int_{-H}^{\zeta} \frac{\partial p'}{\partial x} dz - \frac{\partial p'}{\partial x}$$

$$\frac{\partial v'}{\partial t} + fu' = \frac{1}{H}\int_{-H}^{\zeta} \frac{\partial p'}{\partial y} dz - \frac{\partial p'}{\partial y}$$

a linearized pressure equation,

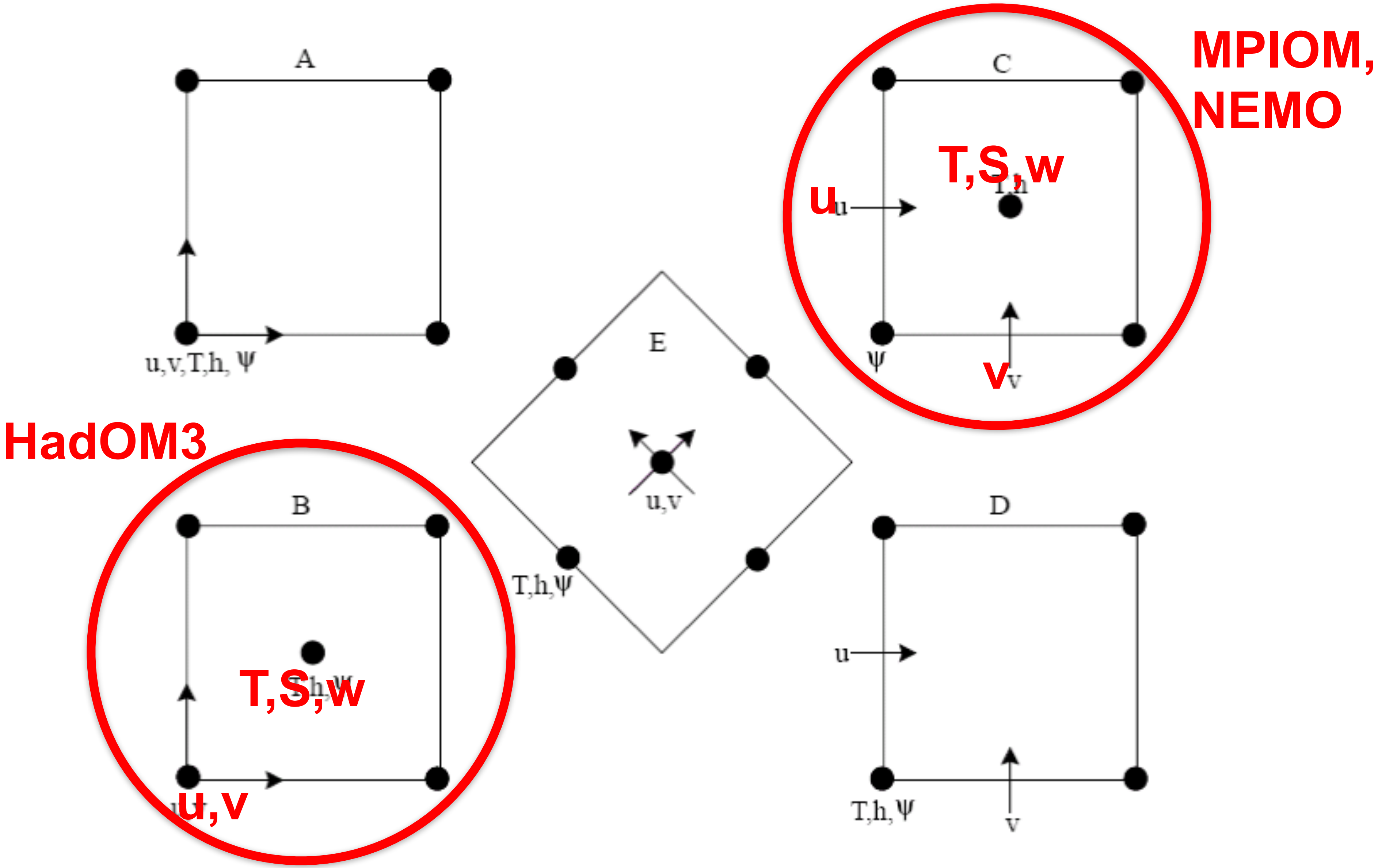
$$\frac{\partial^2 p'}{\partial z \partial t} = \frac{wg}{\rho_0}\frac{\partial \rho}{\partial z}.$$

ζ Sea surface elevation, f coriolis parameter, H total depth

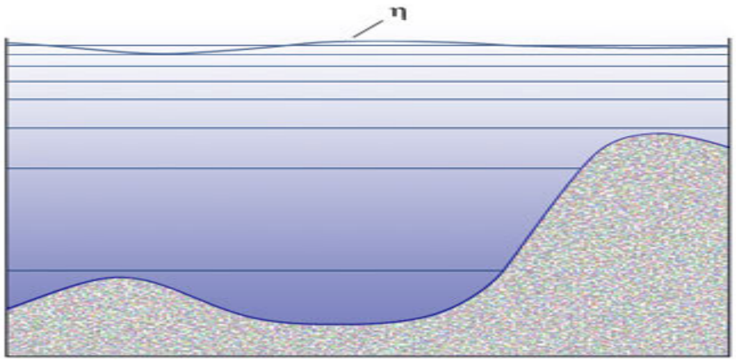
and the continuity equation

$$\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} = 0$$

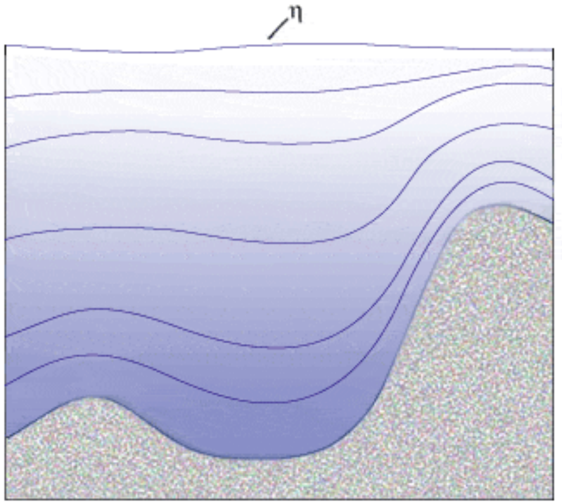
Discretization: Arakawa grid types



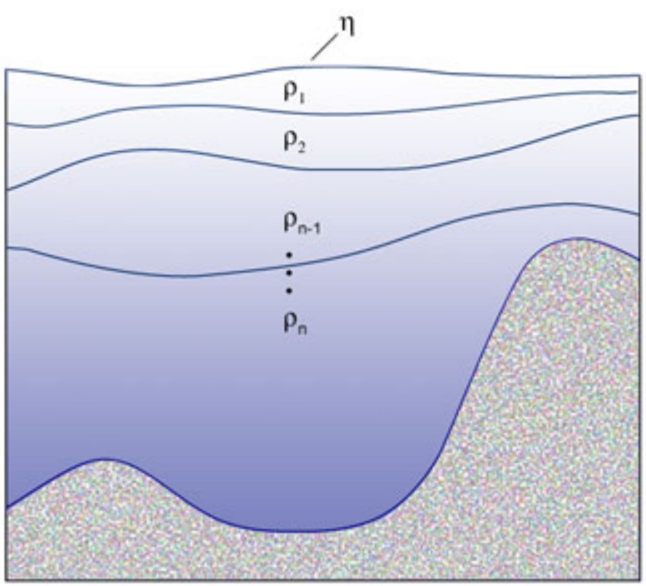
Choice of the vertical coordinate



Z (geopotential)



Terrain-following (sigma)



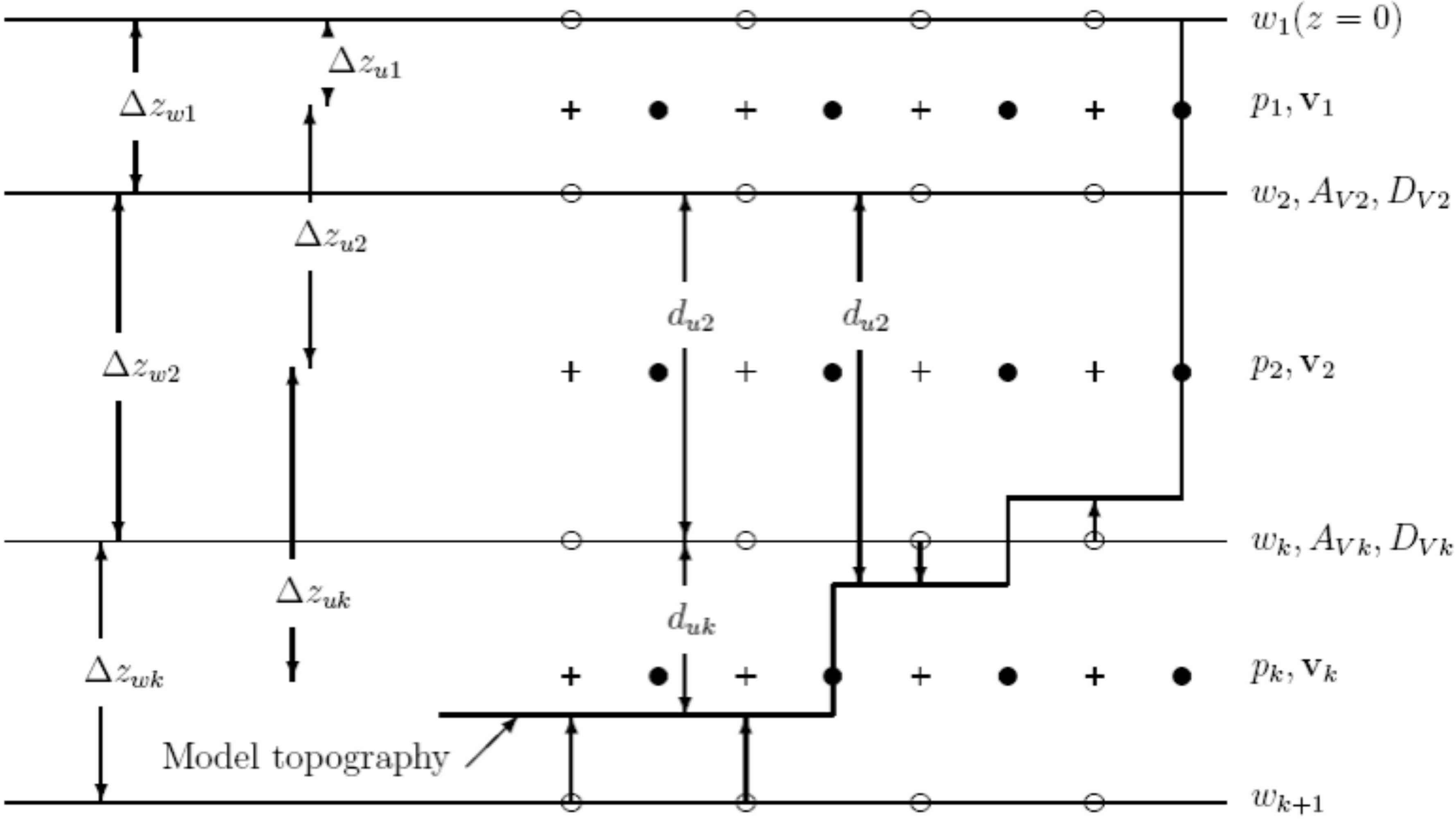
isopycnal

Each system has advantages and disadvantages (DYNAMO Project)

Regional models often apply sigma-coordinates (ROMS);

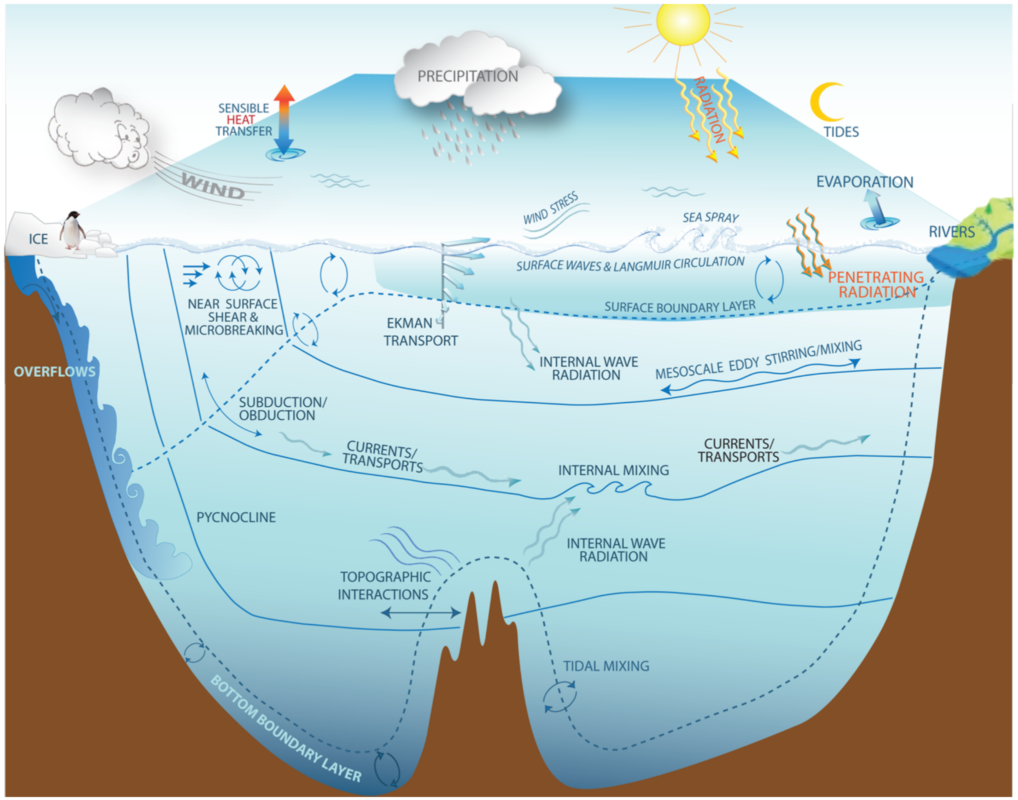
Global models prefer isopycnic (MICOM) or z – coordinates (MOM, MPIOM, NEMO, HADOM)

Vertical discretization in a z-coordinate model



„partial bottom cells“

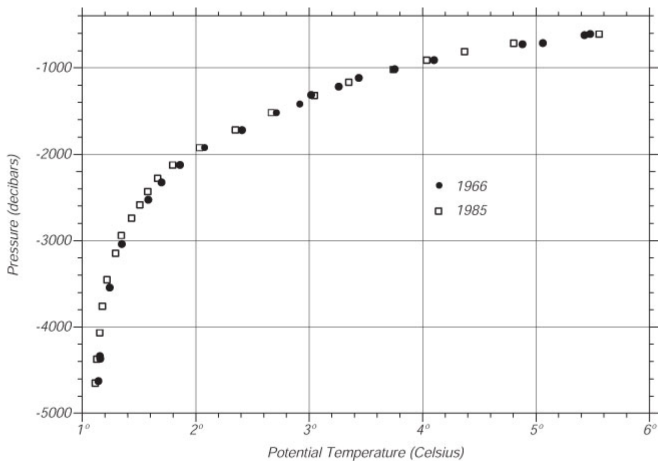
Mixing in the ocean



- Mixing processes in the ocean : turbulence, convection, instabilities, diffusion, friction, etc.
- These processes are usually not resolved by OGCMs, but need to be parameterized

Vertical mixing in the interior

The equation for vertical mixing of a tracer Θ such as salt or temperature is:



$$\frac{\partial \Theta}{\partial t} + W \frac{\partial \Theta}{\partial z} = \frac{\partial}{\partial z}\left(K_z \frac{\partial \Theta}{\partial z}\right) + S$$

Munk, 1966:

The equation has the solution: **$T \sim T_0 \exp(z / H)$,**

- $H = K_z / W$ is the scale depth of the thermocline, and T_0 is the temperature near the top of the thermocline.
- exponential function fit through the observations of T (z) to get H.
- W calculated from the observed vertical distribution of ^{14}C, a radioactive isotope of carbon.

The length and time scales gave :
W = 1.2 cm/day
K_z = 1.3 × 10^{-4} m²/s

In the real ocean, K_z is not uniform. Large parts of the ocean, K_z does not exceed 1×10^{-5} m²/s. Locally, at steep ridges etc, 1×10^{-3} m²/s are observed

Vertical mixing (MPIOM)

Philander and Pacanowski (1981) parameterize vertical mixing as a function of the Richardson Number $Ri = N^2/(du/dz)^2$, where N is the

Brunt-Väisäla frequency $N=(g/\rho\ d\rho/dz)^{-1/2}$

=> static stability vs. vertical current shear

$$\nu = \frac{\nu_0}{(1+aRi)**n} + \nu_b$$

ν = viscosity
ν_b = background viscosity

$$\kappa = \frac{\kappa_0}{(1+aRi)**(n+1)} + \kappa_b$$

κ = diffusivity
κ_b = background diffusivity

with a=5 n=2

In the ocean interior, v and k are close to the background values;

These are close to Munk's (1966) estimates but can be used as tuning parameters

Vertical mixing (MPIOM)

The PP scheme tends to underestimate wind induced mixing in the mixed layer

- MPIOM includes an additional wind mixing scheme
- turbulent energy input by the wind is proportional to the third power of the atmospheric 10m wind speed
- decays exponentially with depth

$$A_{\mathrm{w}}(1) = (1 - I) W_T V_{10\ \mathrm{m}}^3$$

$$A_{\mathrm{w}}(k) = A_{\mathrm{w}}(k-1) \frac{\frac{\lambda}{\Delta z}}{\frac{\lambda}{\Delta z} + \delta_z \rho} e^{\frac{\Delta z}{z_0}}$$

$A_w(k)$ = wind induced mixing at level k

W_T, λ, z_o are tuning coefficients

Vertical mixing

NEMO

Vertical mixing based on Turbulent Kinetic Energy (TKE) scheme, where mixing depends on local density profile (Gaspar et al., 1990)

Non-local K-Profile Parameterization (KPP) scheme (Large et al., 1994). Used here in NEMO3.3

Tidal mixing scheme

HadOM3:

Kraus-Turner mixed layer, with Richardson Number dependent vertical diffusion for momentum

Horizontal mixing

(Turbulent) Horizontal mixing is achieved by eddies of various size

$$K_x \sim UL$$

=> K_x scales with horizontal velocity and length scale

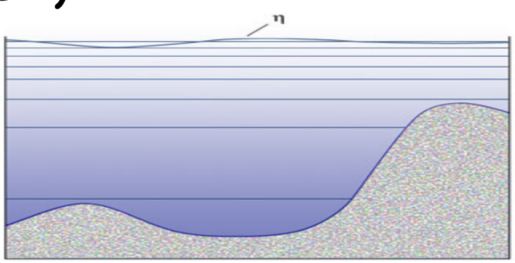
$K_x \approx 800\ m^2/s$ Geostrophic Horizontal Eddy Diffusivity

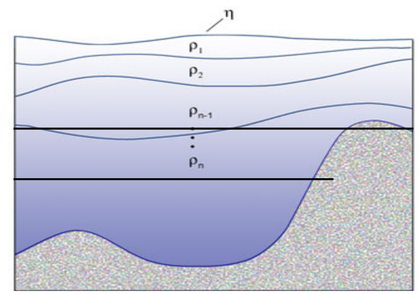
In MPIOM K_x is scaled linearly with the grid size (1000 m^2/s for a 400 km wide grid cell)

- Horizontal mixing is observed to occur **along isopycnals** rather than on horizontal surfaces
- Naturally achieved in isopycnal models

Horizontal mixing on isopycnals

In z-coordinate models, the diffusion tensor has to be rotated (Redi, 1982)





$$\mathbf{K} = D_\mathrm{H} \begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & \epsilon \end{bmatrix} \longrightarrow \mathbf{K} = D_\mathrm{H} \begin{bmatrix} 1 & 0 & S_x \\ 0 & 1 & S_y \\ S_x & S_y & \epsilon + S_\mathrm{dif}^2 \end{bmatrix}$$

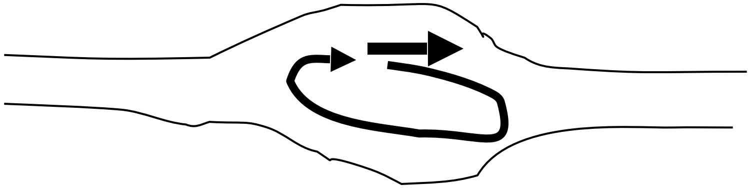
$$\epsilon = \frac{D_\mathrm{V}}{D_\mathrm{H}} \quad \text{and} \quad S_\mathrm{dif} = (S_x, S_y, 0) = \left(\frac{-\delta_x \rho}{\delta_z \rho}, \frac{-\delta_y \rho}{\delta_z \rho}, 0 \right)$$

D_V= vertical diffusion; D_H: horizontal diffusion

- Isopycnal diffusion in a z-coordinate model requires calculation of **isopycnal slopes** at every time step!
- D_H is scale-dependent as a function of the horizontal grid size

Gent-McWilliams eddy parameterization

- Meso-scale Eddy are typically unresolved by OGCMs
- Gent and McWilliams (1990) and Gent et al., 1995 proposed to include the advective transport by the moving eddies

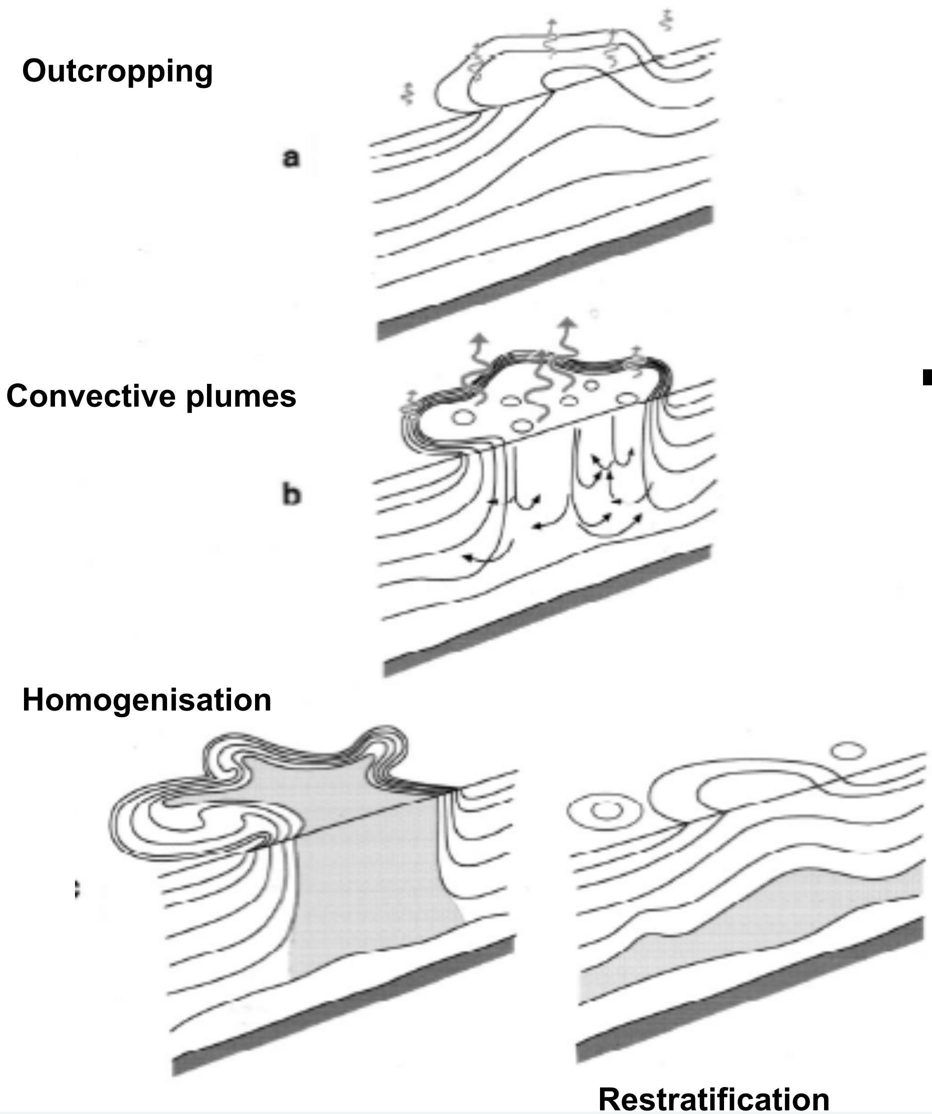


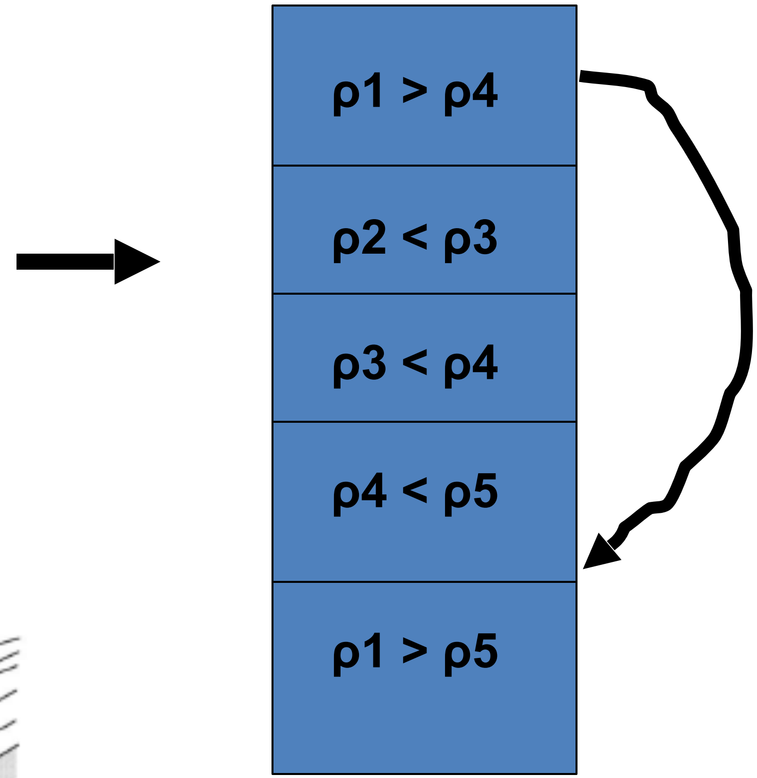
- Example: Lens with Mediterranean water in the Atlantic
- Divergence of this flux is added to the tendency equation for T and S
- „bolus velocity" is derived from isopycnal slopes (GM90), Griffies, 1997; or from stability consideration (Visbeck et al., 1996),

Gent-McWilliams eddy parameterization

From Medicine: a **bolus** (from Latin bolus, ball) is a round mass of organic matter moving through the digestive tract

Open ocean convection

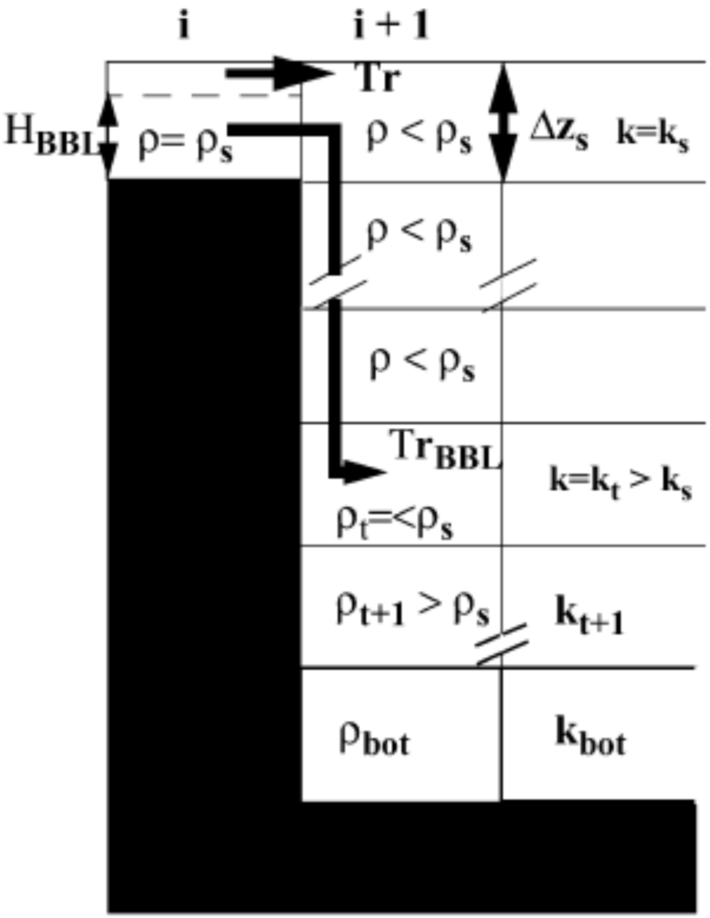




Open ocean convection

- **„Convective adjustment"** (Bryan, 1969): full mixing of adjacent vertical grid cells: used in NEMO (Madec et al., 1991)
- **„Mixing-only"** instabilities are mixied with the Richardson-Number dependent vertical mixing scheme (PP)
- **„enhanced convective mixing"**: the vertical mixing coefficients from the PP scheme are overwritten by much larger numbers: default for MPIOM
- **„plume mixing"**, a more physical, but CPU-time consuming parameterization of convective plumes (Paluskiewicz and Romea, 1997)

Slope Convection



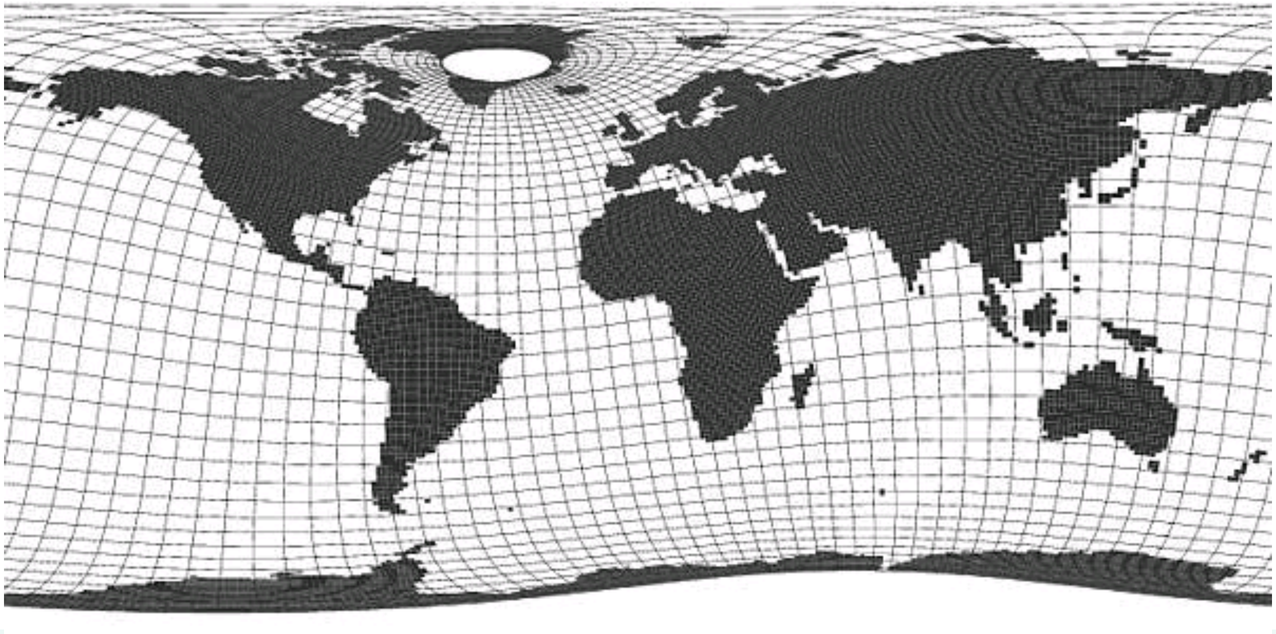
MPIOM and NEMO include Bottom Boundary Layer schemes

- BBL scheme identifies grid cells where the density of neighbouring grid cells indicates possible **slope convection**
- A certain amount of mass from the shallow source cell is transported into the target cell according to the density profile using the **horizontal advective velocity and/or increased diffusion**
- Scheme improves, for example, the **water mass properties** in the Atlantic where Mediterranean Water finds its equilibrium density at mid-depth (ca. 1200 m)

Horizontal grid configurations

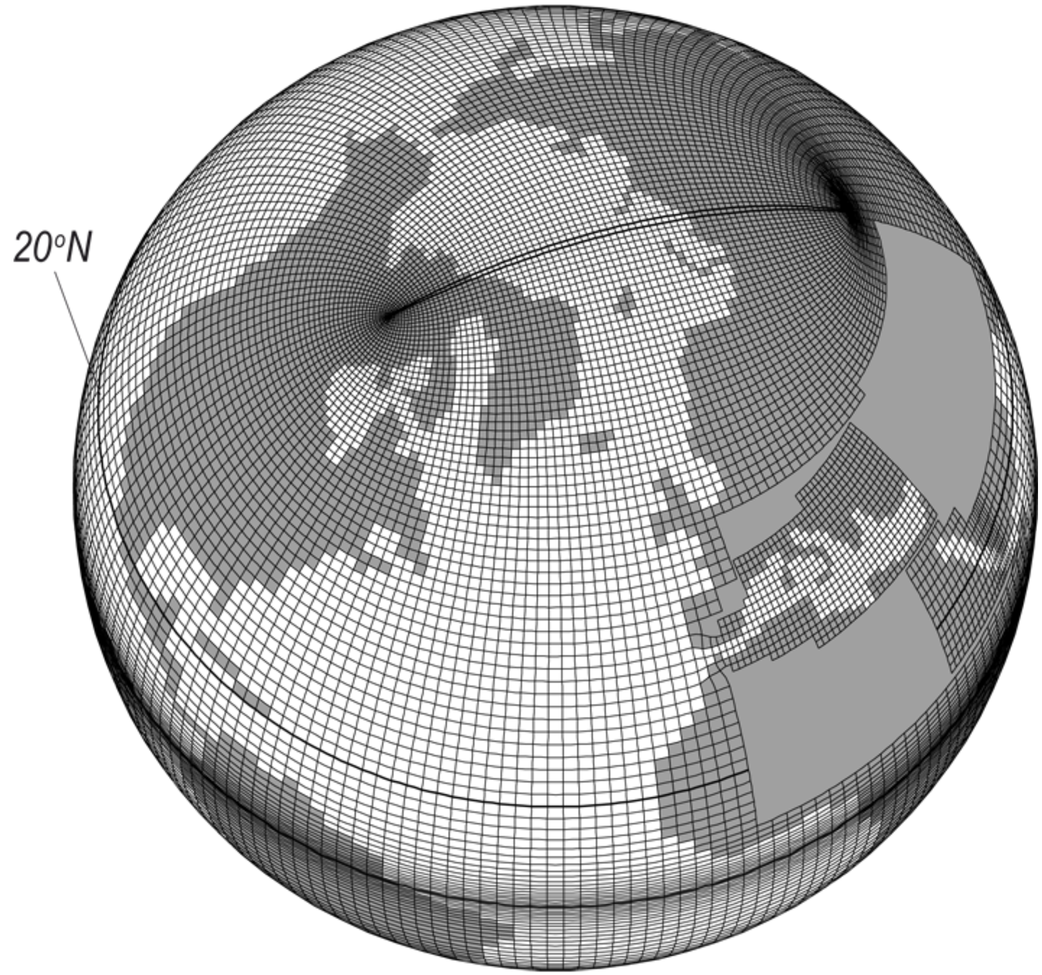
- Regular longitude-latitude grid

  HadOM3: 3.75º (lon) by 2.5º (lat)
- Orthogonal curvilinear grid: **MPIOM:** bipolar grid with northern pole in Greenland nominal resolution of 1.5º (GR1.5) or tri-polar grid with two northern poles (TP04, TP01)



Horizontal grid configurations

- **NEMO:** tripolar grid with two northern poles in Canada and Russia nominal resolution of 1º (ORCA1). Special treatment of marginal seas.



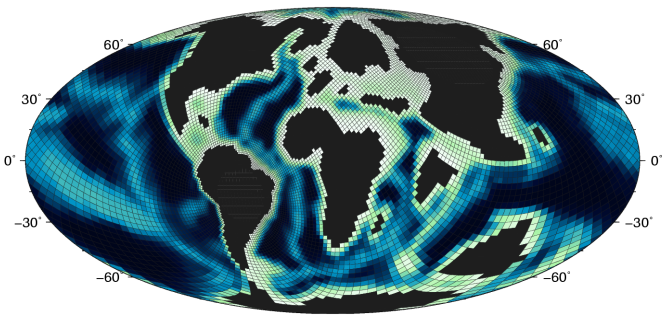
Horizontal grid configurations

PhD Thesis Malte Heinemann / Matthias Heinze

PETM: Paleocene/Eocene Thermal Maximum (55 Ma b.p.)

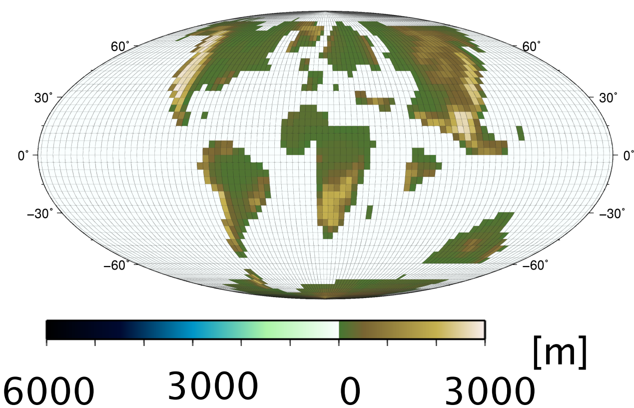
- tool: AO-GCM with P/E boundary conditions

MPI-OM

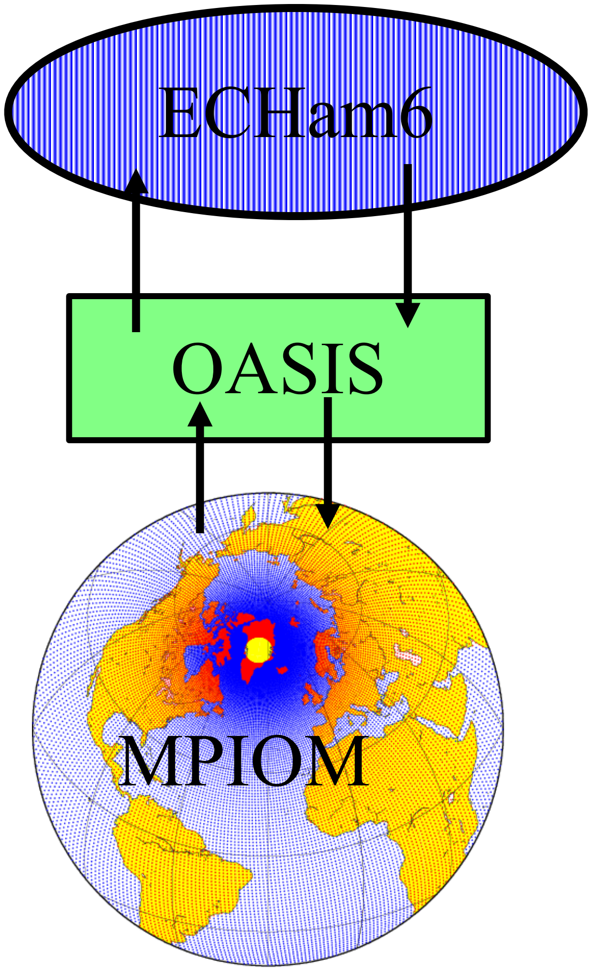


OASIS

ECHAM5



OGCM in the coupled model

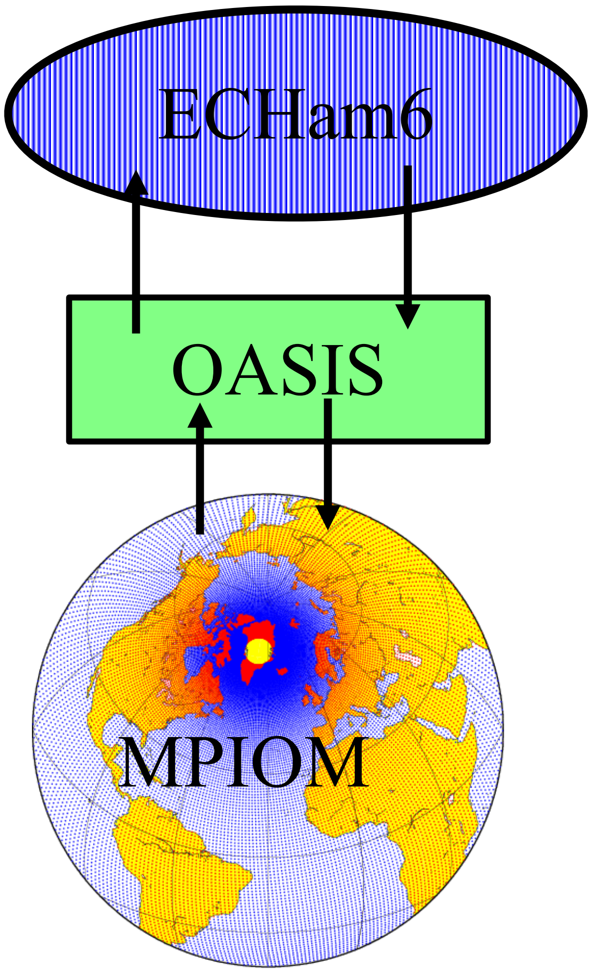


MPIOM is coupled to ECHAM via the OASIS coupler; coupling frequency is usually one day

MPIOM receives:

- **Total heat flux over water**
- **Downward short wave heat flux**
- **Conductive heat flux sea ice**
- **Residual heat flux sea ice**
- **Wind stress over water (τx, τy)**
- **Wind stress over ice (τx, τy)**
- **Wind speed**
- **Solid precipitation**
- **Liquid precipitation (incl. River runoff)**
- **CO_2 flux**
- **Atmospheric $C0_2$ concentration**

OGCM in the coupled model

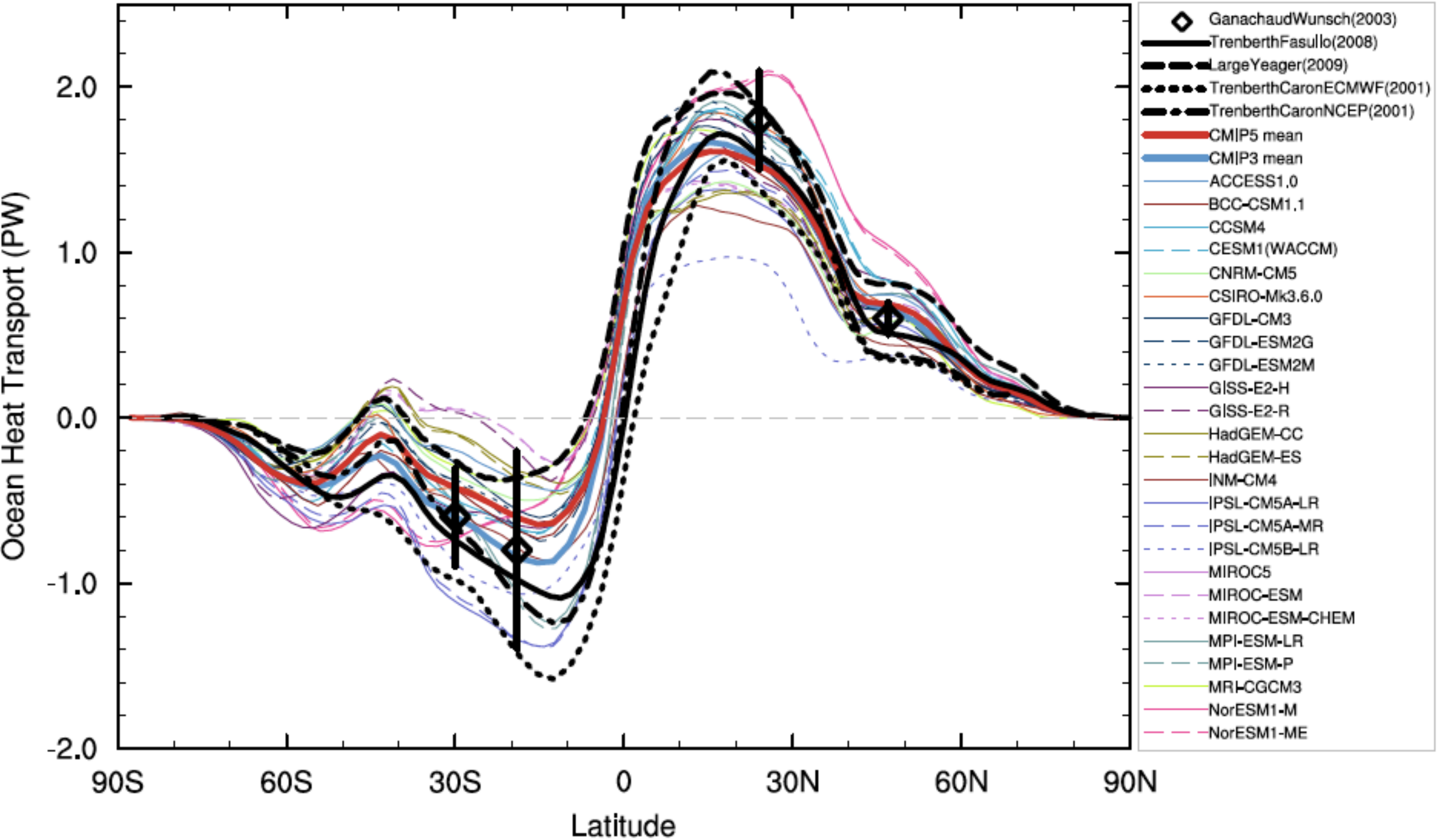


MPIOM sends:

- **Sea surface temperature**
- **Sea ice cover**
- **Sea ice thickness**
- **Snow thickness (on sea ice)**
- **Ocean surface velocities**

- **Partial pressure of CO_2 in the upper layer**
- **Chemical solubility of $C0_2$**

Ocean model evaluation



Flato et al., 2013

Ocean model evaluation

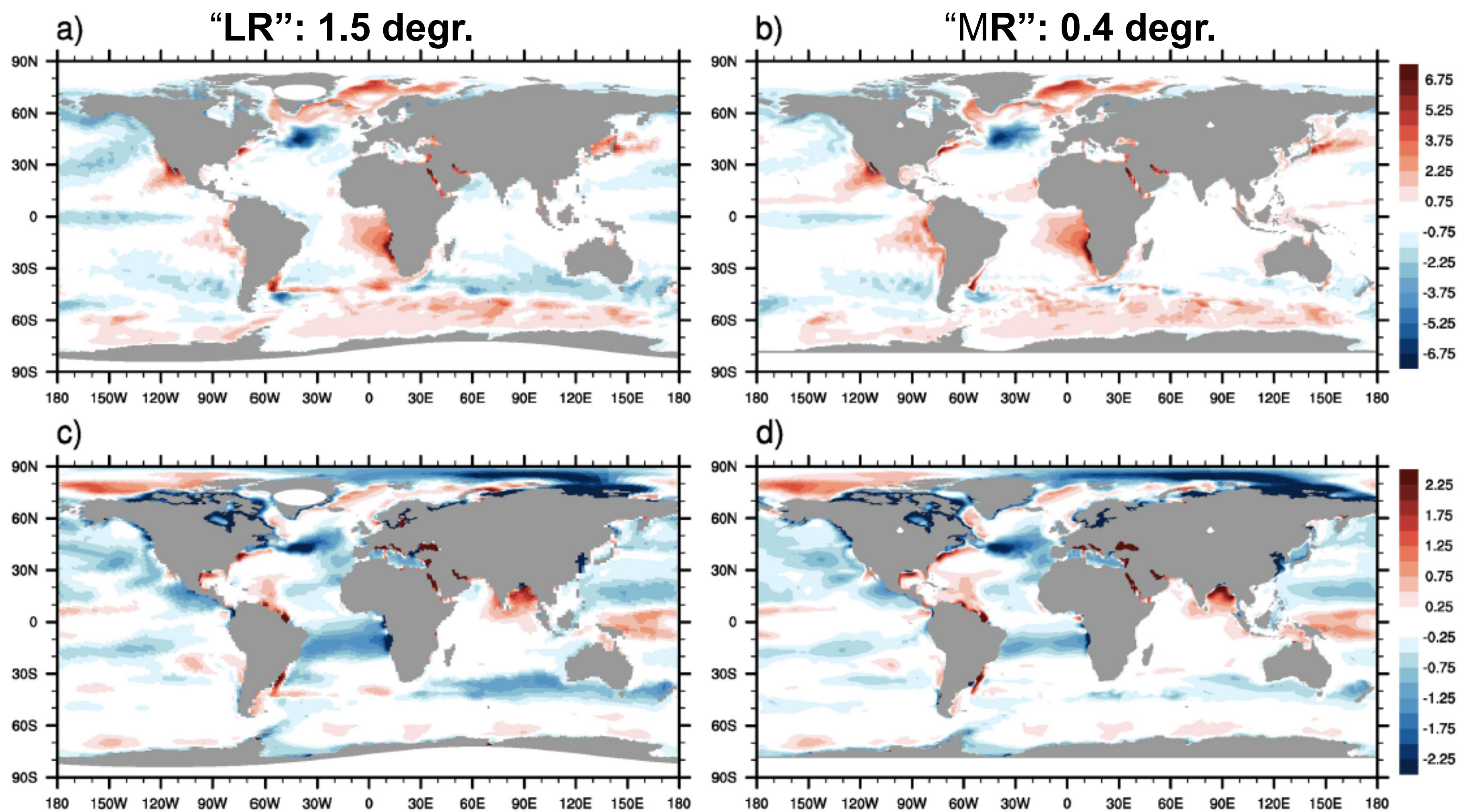


Figure 2. Biases in surface properties for (left column) MPI-ESM-LR, and (right column) MPI-ESM-MR: (a, b) sea surface temperature (°C), and (c, d) salinity (psu) differences between the ensemble means (1980–2005) from the historical experiments and the PHC3 climatology.

Jungclaus et al., 2013

MPIOM References

Wolff, J.O., E. Maier-Reimer, S. Legutke, 1997: The Hamburg Ocean Primitive Equation model HOPE, Technical Report 13, DKRZ, Hamburg.

Marsland, S.J., H. Haak, J.H. Jungclaus, M. Latif, and F. Röske (2003), The Max-Planck- Institute global ocean/sea ice model with orthogonal curvilinear coordinates. *Ocean Modelling*, 5, 91-127.

Jungclaus, J.H., M. Botzet, H. Haak, N. Keenlyside, J.-J. Luo, M. Latif, J. Marotzke, U. Mikolajewicz, and E. Roeckner, 2006: Ocean circulation and tropical variability in the coupled model ECHAM5/MPI-OM. *Journal of Climate*, 19, 3952-3972.

Jungclaus, J.H., N. Fischer, H. Haak, K. Lohmann, J. Marotzke, D. Matei, U. Mikolajewicz, D. Notz, and J.-S. von Storch, 2013: Characteristics of the ocean simulations in MPIOM, the ocean component of the MPI-Earth System Model. Journal of Advances in Modeling Earth Systems, 5, 1-25.

Documentation:

http://www.mpimet.mpg.de/en/science/models/mpiom/

HADOM3/FAMOUS References

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Jones, C., et al. (2005): Systematic optimisation and climate simulations of FAMOUS, a fast version of HADCM3. Clim. Dyn., 25, 189-204.

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Documentation:
http://cms.ncas.ac.uk/documents/vn4.5/p040.pdf (Ocean)

http://cms.ncas.ac.uk/documents/vn4.5/p045.pdf (Sea ice)

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Madec, G. (2008): NEMO ocean engine. Note du Pole de modelisation, Institut Pierre-Simon Laplace (IPSL), France, No 27

Fichefet T. and M.A. Morales Maqueda (1997): Sensitivity of a global sea ice model to the treatment of ice thermodynamics and dynamics. J Geophys Res, 102, 12609-12646

Gaspar, P., Y. Gregoris and J.M. Lefevre (1990): A simple eddy kinetic energy model for simulations of the oceanic vertical mixing. Tests at station Papa and long-term upper ocean study site. J Geophys Res, 95, 16179-16193

Documentation:

“The NEMO book”:

(http://www.nemo-ocean.eu/content/download/15482/73217/file/NEMO_book_v3_3.pdf)

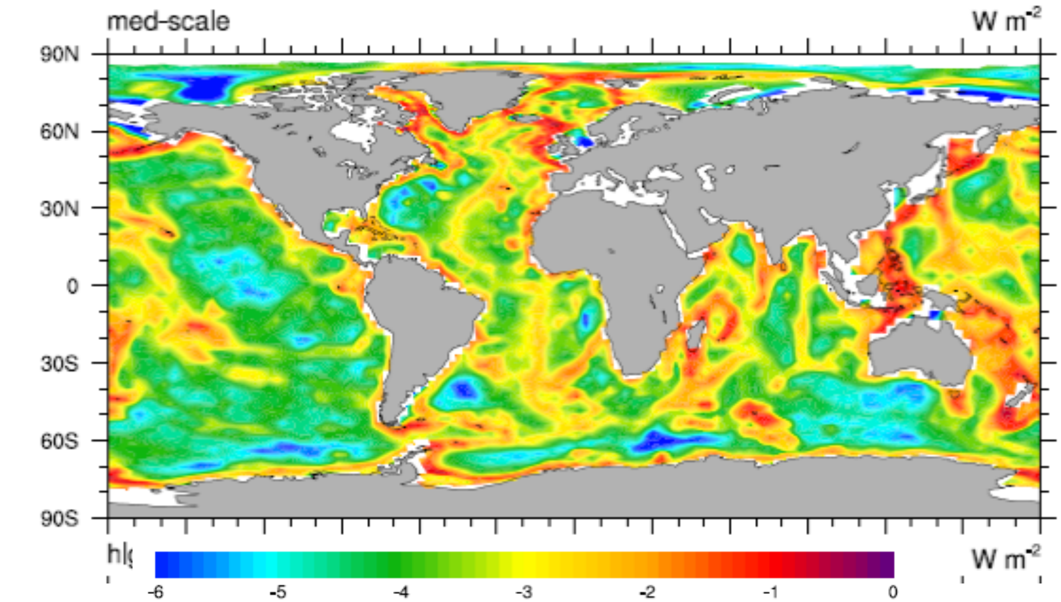
Conclusions

Through its buffering abilities and relatively slow time scales, the ocean represents the fly wheel of the Earth's climate system. It's most prominent role is the storage and redistribution of heat, (fresh) water, and matter (such as CO_2)

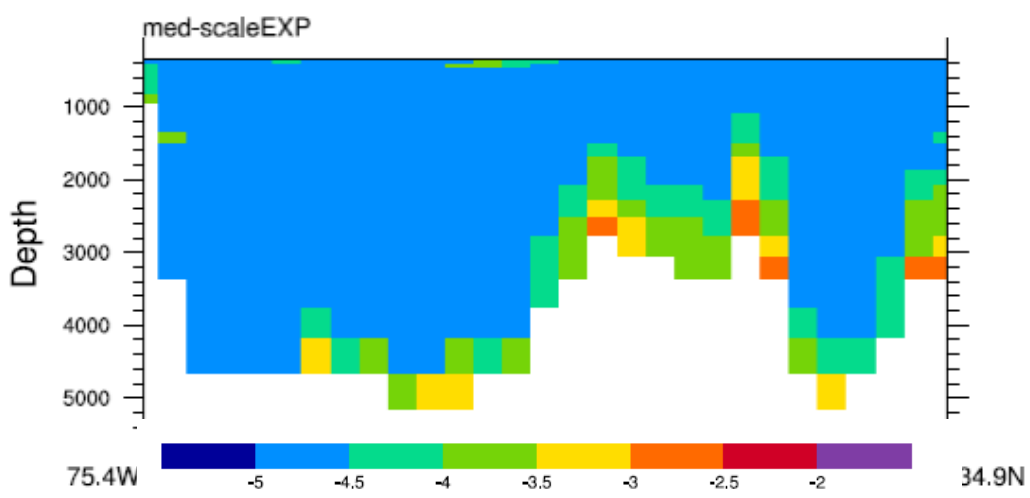
Ocean models, as part of current ESMs, resolve large-scale dynamics, and some (none in this summer school!) resolve the eddy-scale. Many process remain to be parameterised.

ESMs of different complexity serve different purposes (speed vs. complexity and accuracy), biases and limitations need to be considered

Vertical mixing in MPIOM



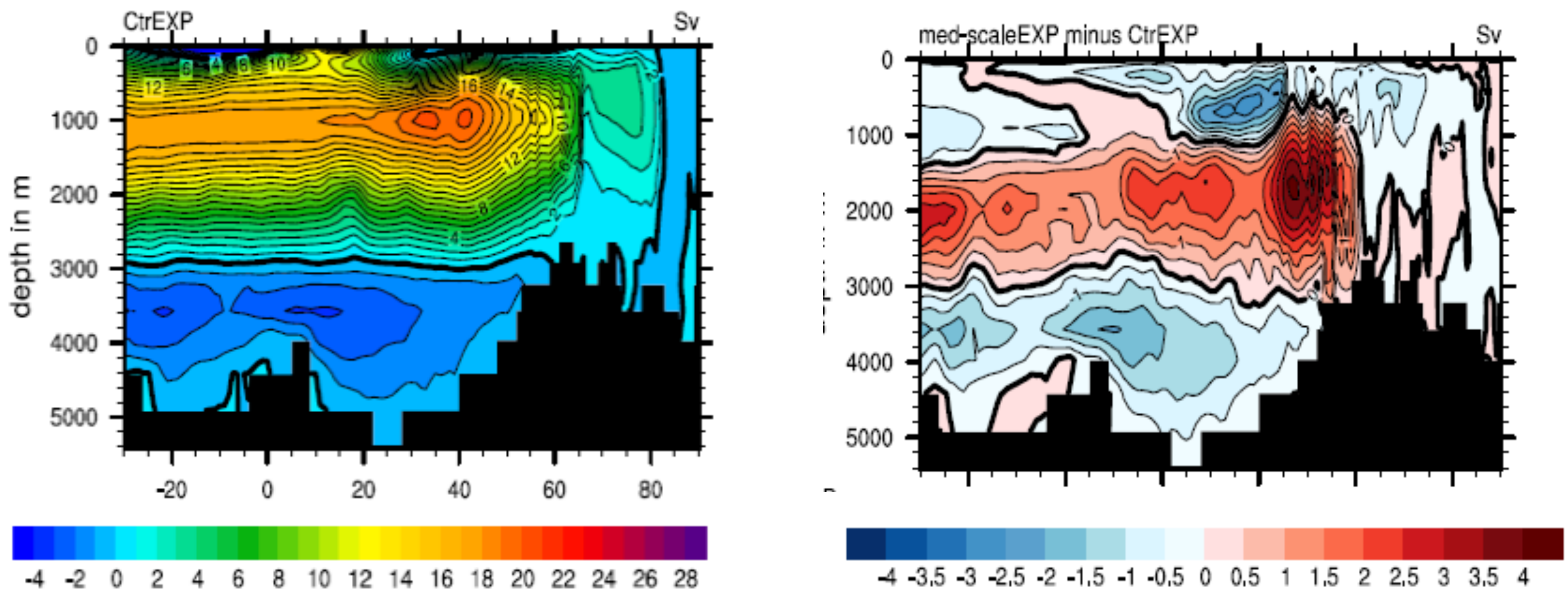
Tidal energy dissipation (log scale in W/m**2)



Vertical section of diffusity (log scale, units: m**2/s

Exarchou et al., 2012

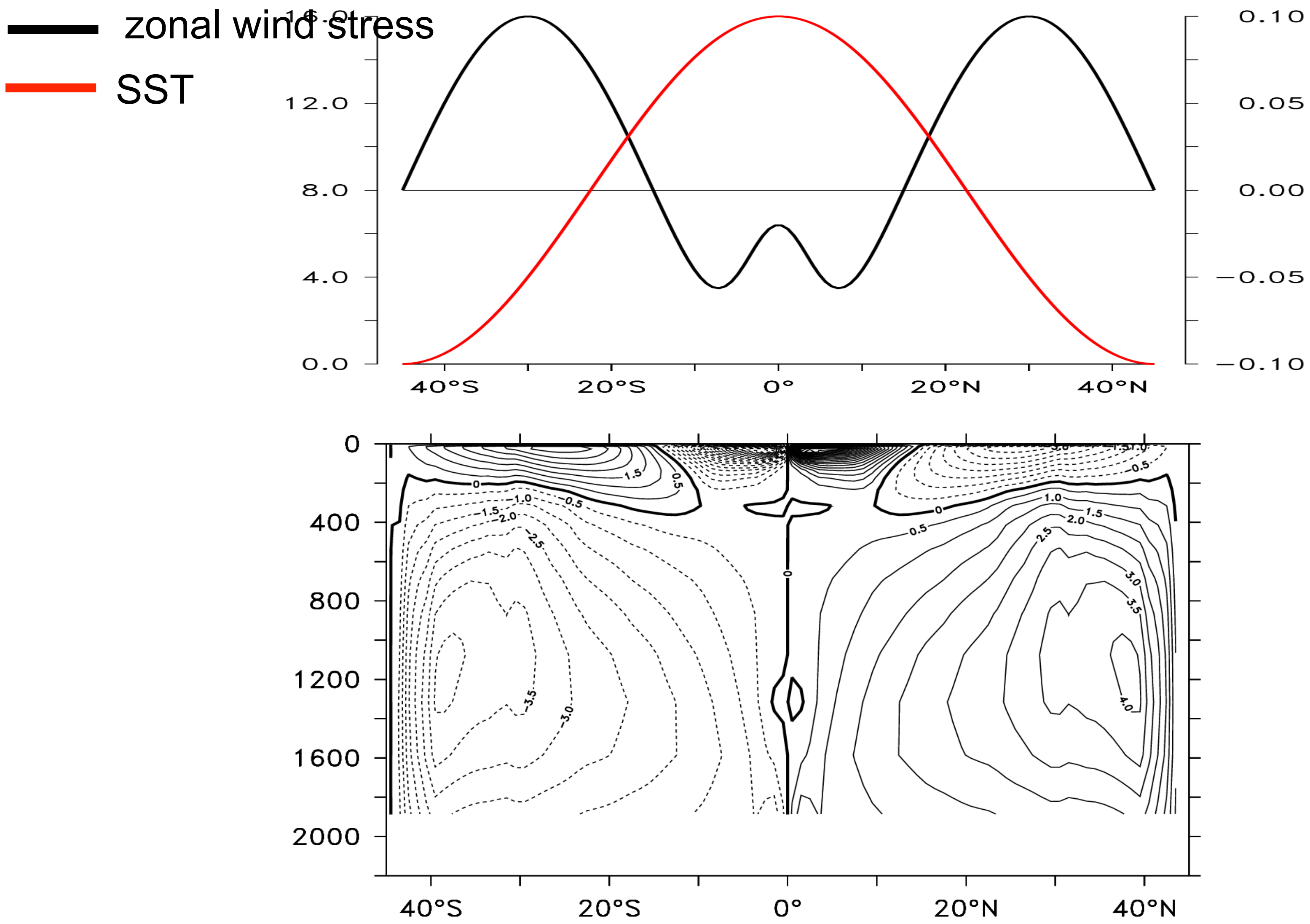
Vertical mixing in MPIOM



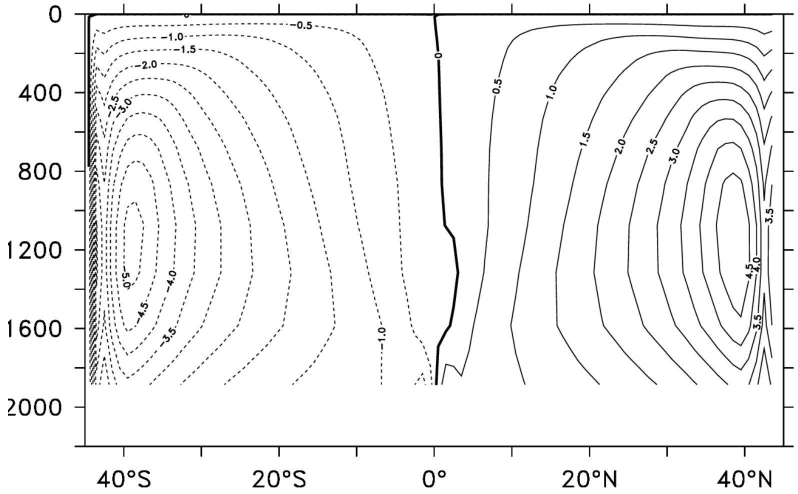
Introduction of tidal mixing scheme introduces considerable changes in water mass distribution, deep water formation, and overturning circulation.

Results depend on details of implementation (Exarchou et al., 2012)

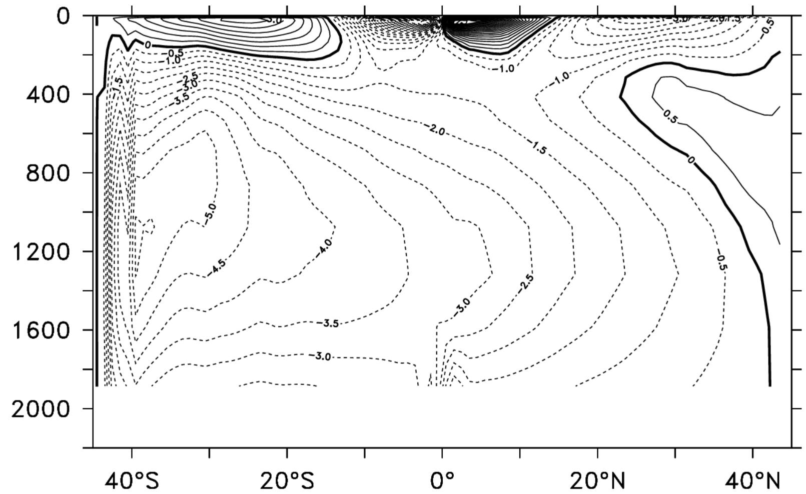
An idealised ocean set-up: rectangular box, rotation, idealised forcing



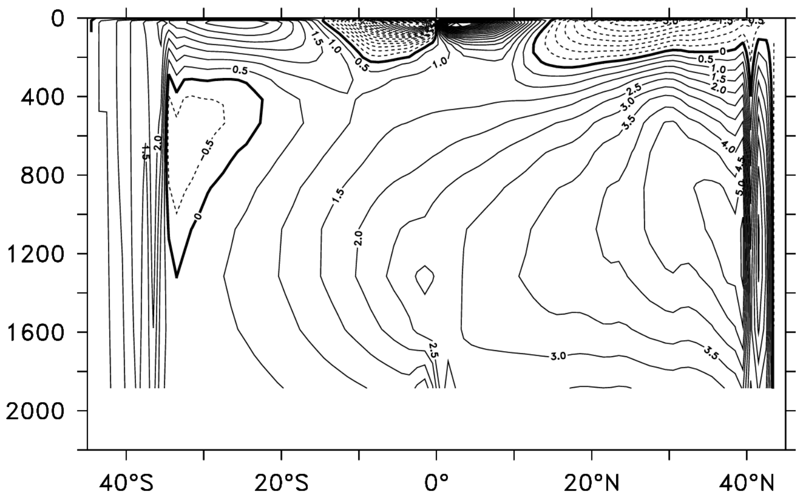
Overturning circulation in an ocean model



(no wind forcing)



N-Pole 2° warmer than S-Pole



open zonal channel (ACC)

Strength and shape of the overturning circulation is not only determined by the density gradients between high and low latitudes, but also by the wind systems and the basin geometry