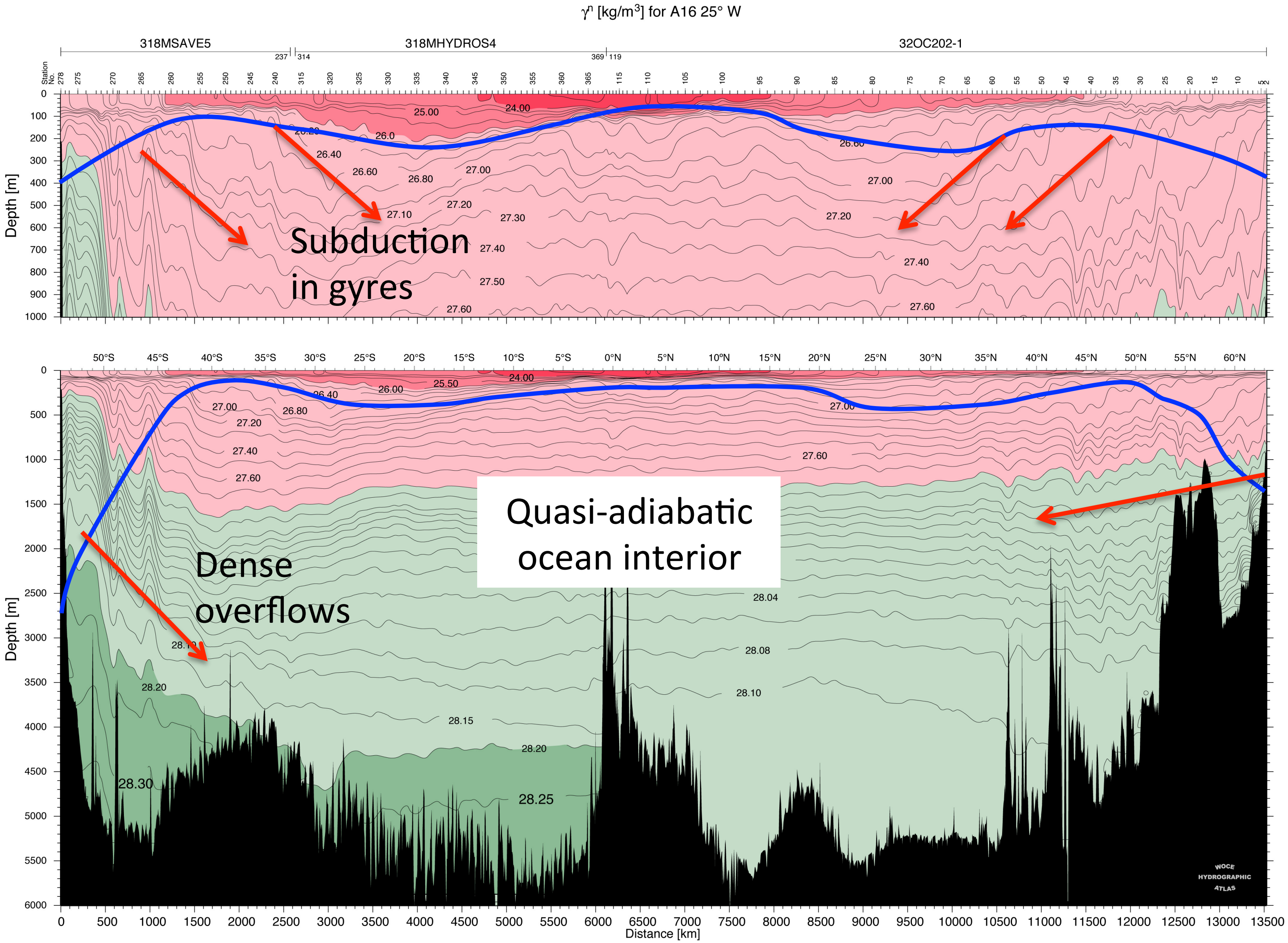
The oceanic boundary layers in models: our (unfortunate) reliance on parameterizations

Bulk of ocean properties are set in the boundary layers



Subduction in gyres

Quasi-adiabatic ocean interior

Dense overflows

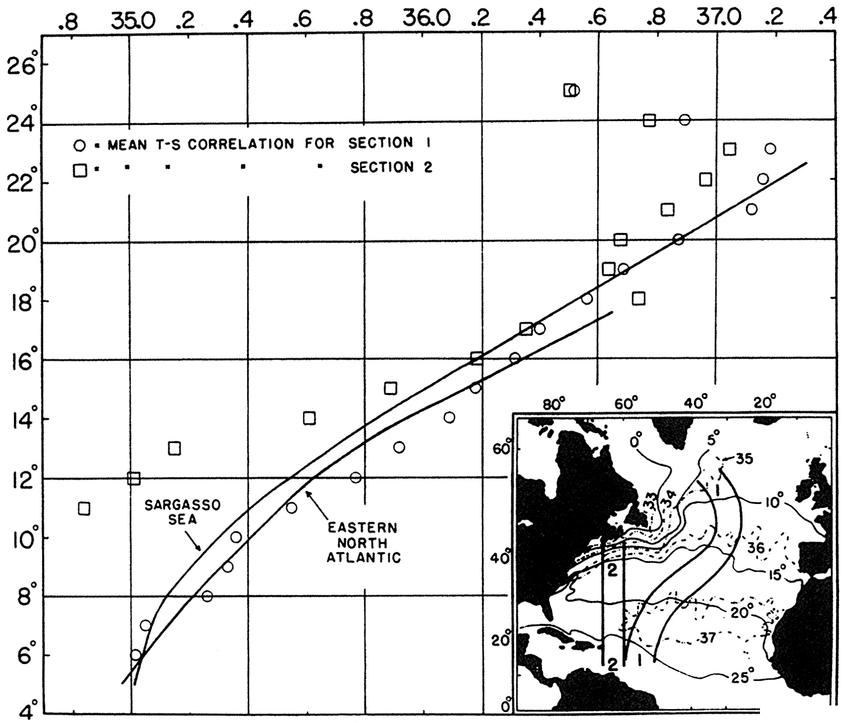
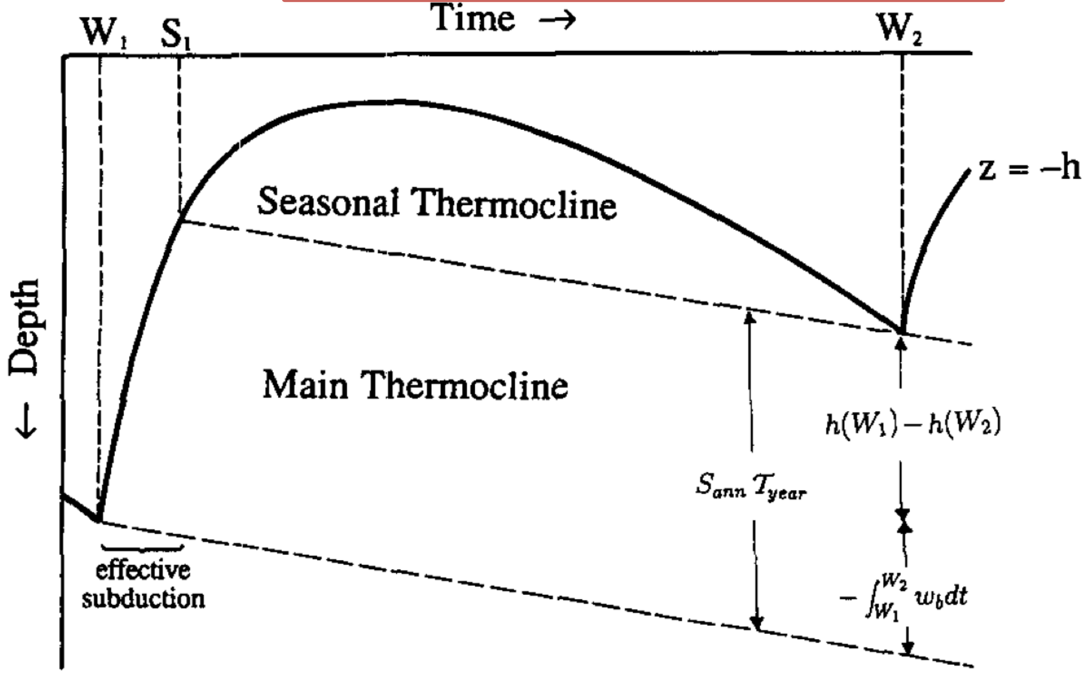


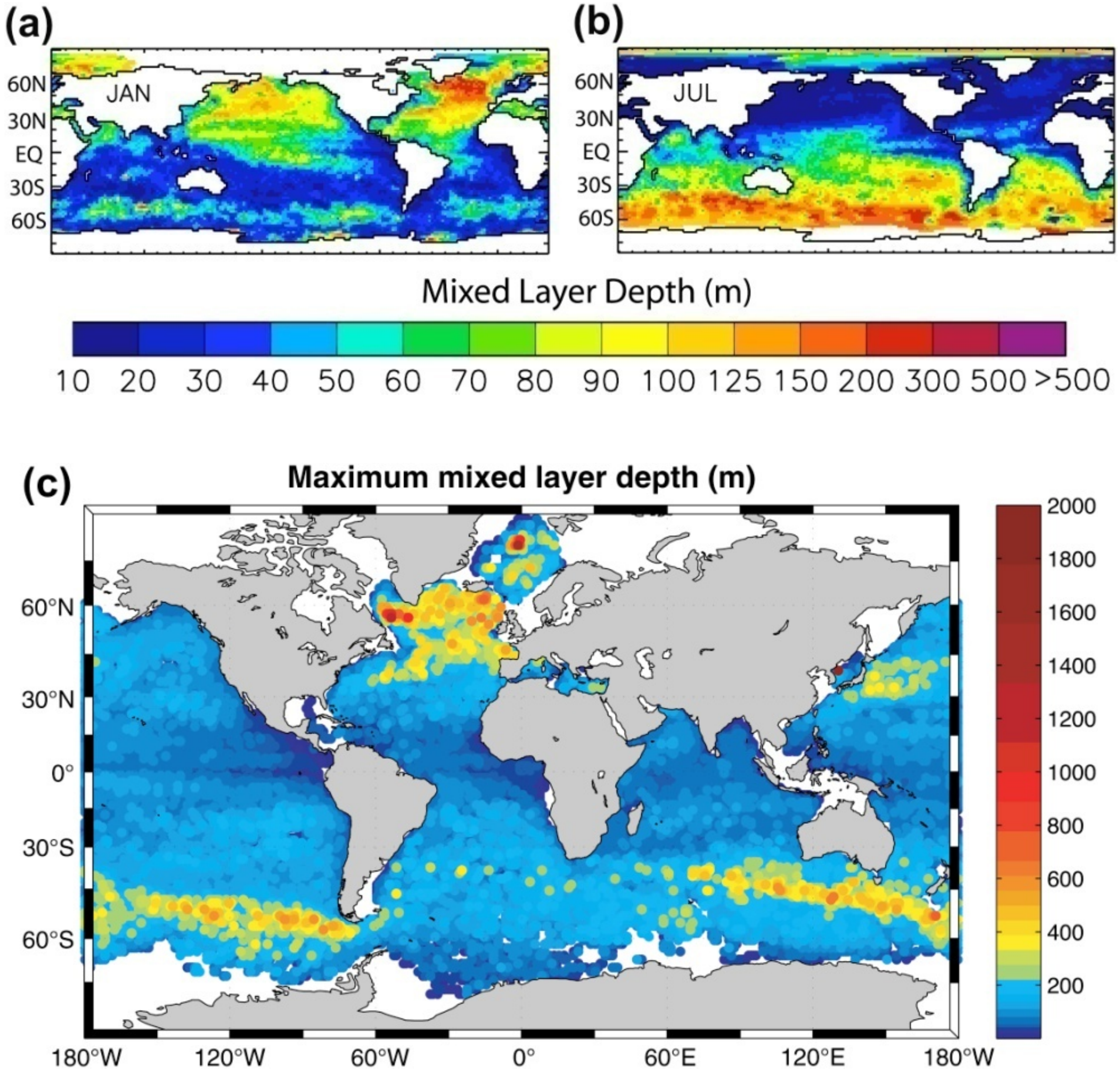
FIGURE 4.6

Temperature-salinity along surface swaths in the North Atlantic (dots and squares), and in the vertical (solid curves) at stations in the western North Atlantic (Sargasso Sea) and eastern North Atlantic. *Source: From Iselin (1939).*

Ocean interior properties are set by the late winter ML properties



Williams et al. (1994)

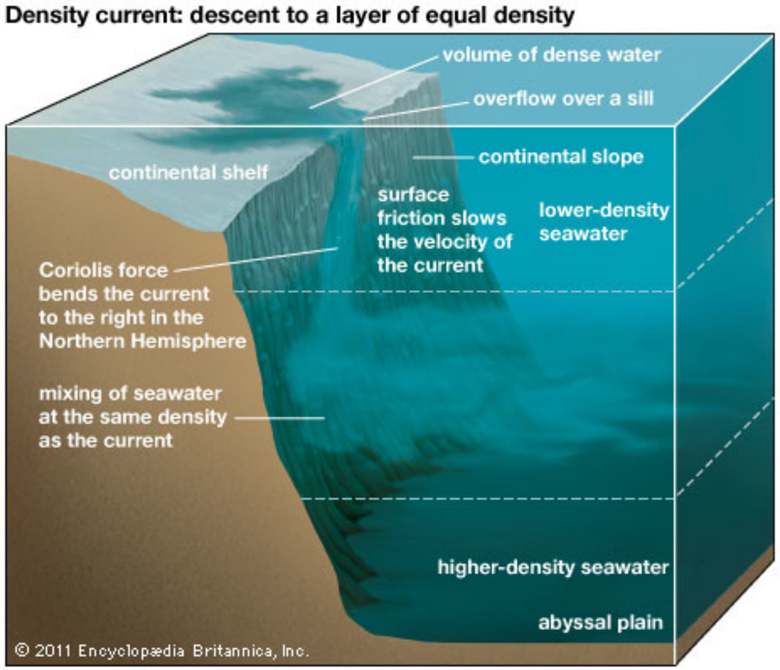


Mixed layer depth in (a) January and (b) July, based on a temperature difference of 0.2°C from the near-surface temperature. *Source: From deBoyer Montégut et al. (2004).* (c) Averaged maximum mixed layer depth, using the 5 deepest mixed layers in 1° × 1° bins from the Argo profiling float data set (2000-2009) and fitting the mixed layer structure as in Holte and Talley (2009). This figure can also be found in the color insert.

FIGURE 4.4

<u>At high latitudes</u>, final properties of water set in:

- Overflow from continental shelves or over oceanic ridges
- Exchanges between boundary currents and open ocean



Density current: descent to a layer of equal density

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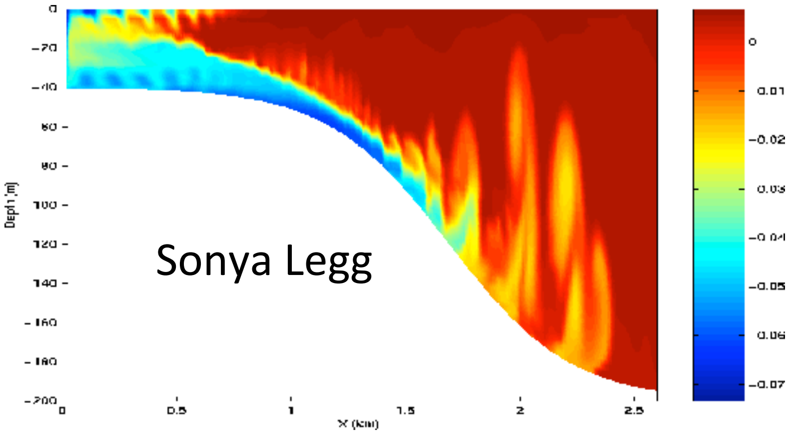
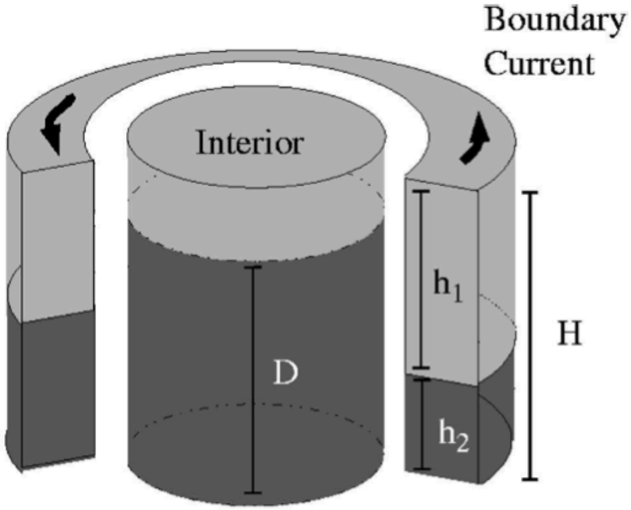
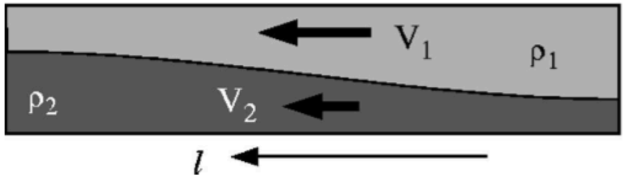


Sonya Legg



Boundary Current (unwrapped)



Straneo, 2008

How much boundary layer schemes influence the outcome of climate simulations?

→ Focus on the surface boundary layers

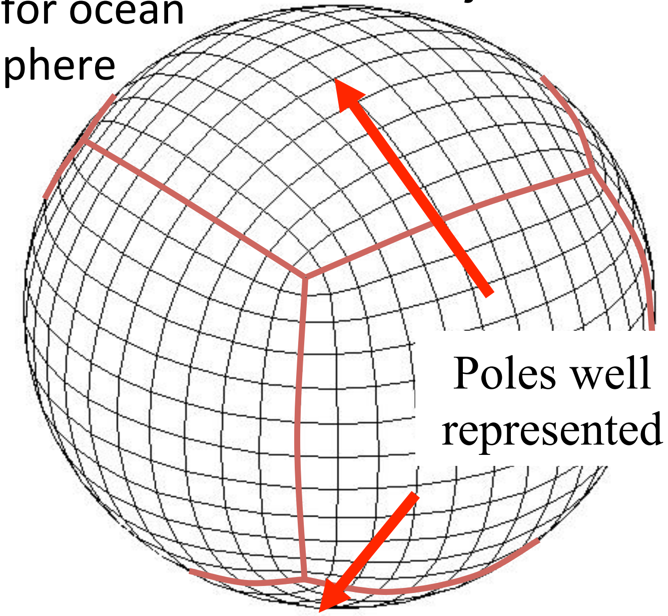
Compare:

- Simulation with/without a boundary layer
- Simulations with different boundary layers

No attempt to tune, take schemes as "out of box"

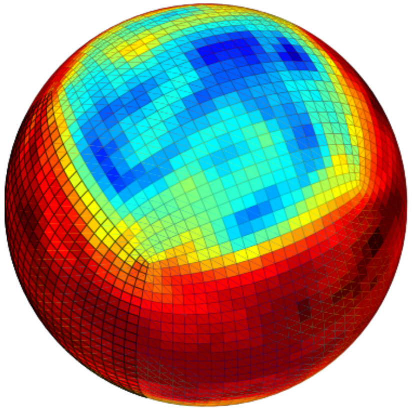
2 MIT GCM: Coupled Ocean-Atmosphere-Sea ice:

- Primitive equation models,
- Cube-sphere grid: ~3.75º,
- Synoptic scale eddies in the atmosphere,
- Gent and McWilliams eddy parameterization in the ocean,
- Simplified atmospheric physics (SPEEDY, Molteni 2003),
- Conservation to numerical precision (Campin et al. 2008)

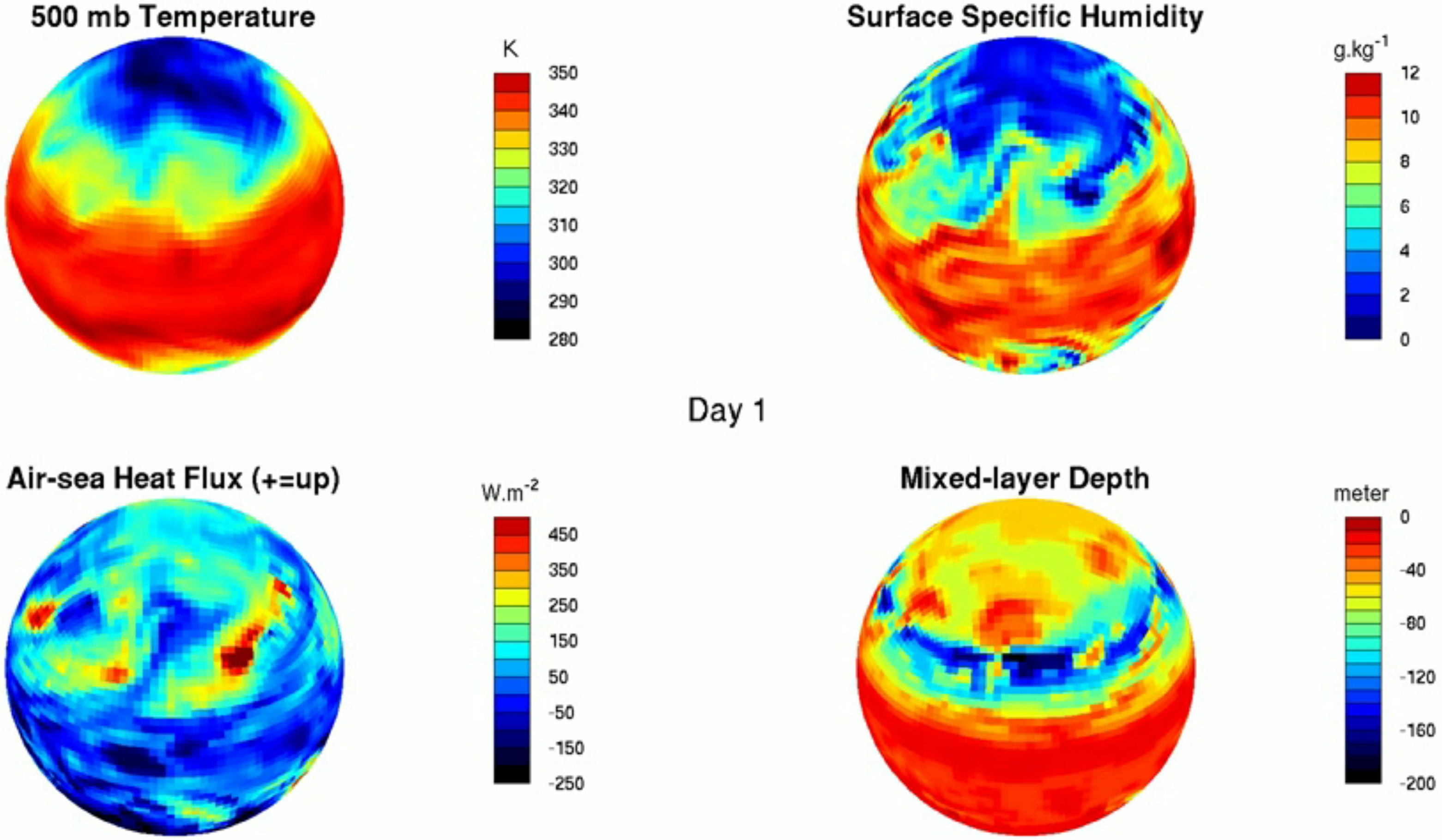


Same grid for ocean and atmosphere

Fully coupled: no adjustments



Temperature snap-shot at 500 mb.


500 mb Temperature

Surface Specific Humidity

Day 1

Air-sea Heat Flux (+=up)

Mixed-layer Depth

No boundary layer scheme (“Reference” run)

- First grid level: 30 m thick
  - → Mixed layer is always at least 30 m thick
- Convective adjustment: intense mixing for $N^2<0$
  - → Applies to T and S and other passive tracers
  - → Nothing of momentum

Non-local K profile parameterization (KPP), Large et al. (1994)

$$\overline{wx}(d) = -K_x(\partial_z X - \gamma_x)$$

Local term
(although non-local too)

Non-local term

depend on
forcing

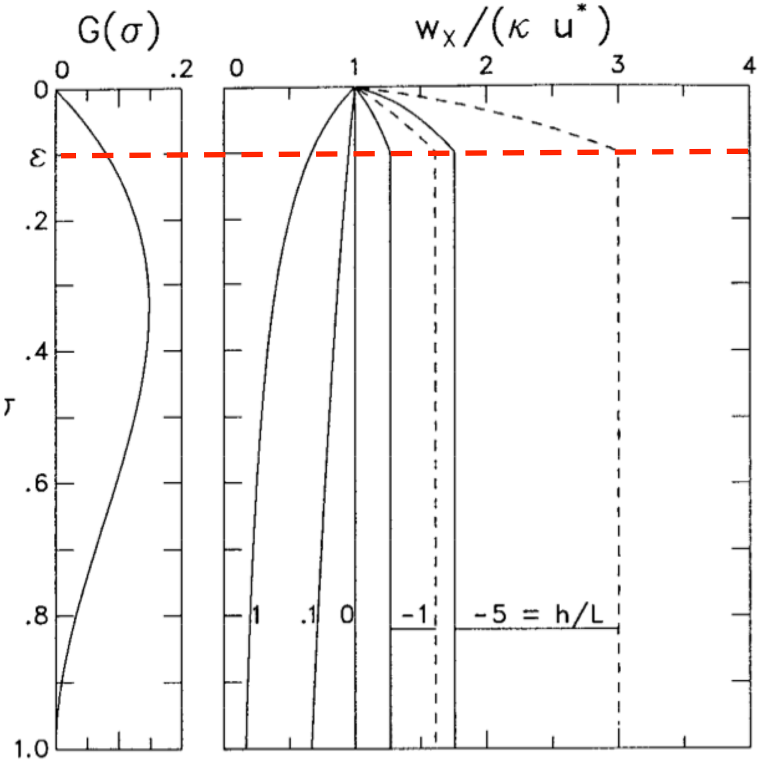
$$K_x(\sigma) = h w_x(\sigma) G(\sigma)$$

h boundary layer depth

$$\gamma_x = 0 \qquad \zeta \geq 0$$

$$\gamma_m = 0 \qquad \gamma_s = C_s \frac{\overline{ws_0}}{w_s(\sigma)h}$$

$$\gamma_\theta = C_s \frac{(\overline{w\theta_0} + \overline{w\theta_R})}{w_s(\sigma)h} \qquad \zeta < 0$$



TKE model Gaspard et al. (1990), Blanke and Delecluse (1993)

Second-order moment closure

TKE prognostic equation:

$$\partial_t \bar{e} = -\partial_z(\overline{ew'} + \rho_0^{-1}\overline{p'w'}) - \overline{\mathbf{u}'_h w'} \cdot \partial_z \mathbf{U}_h + \overline{b'w'} - \epsilon,$$

~ Transport

Source of TKE

Sink of TKE

dissipation

Closure

$$-\overline{\mathbf{u}'_h w'} = K_m \partial_z \mathbf{U}_h,$$

$$-\overline{b'w'} = -g\rho_0^{-1} K_\rho \partial_z \rho = K_\rho N^2$$

$$-(\overline{ew'} + \rho_0^{-1}\overline{p'w'}) = K_e \partial_z \bar{e}$$

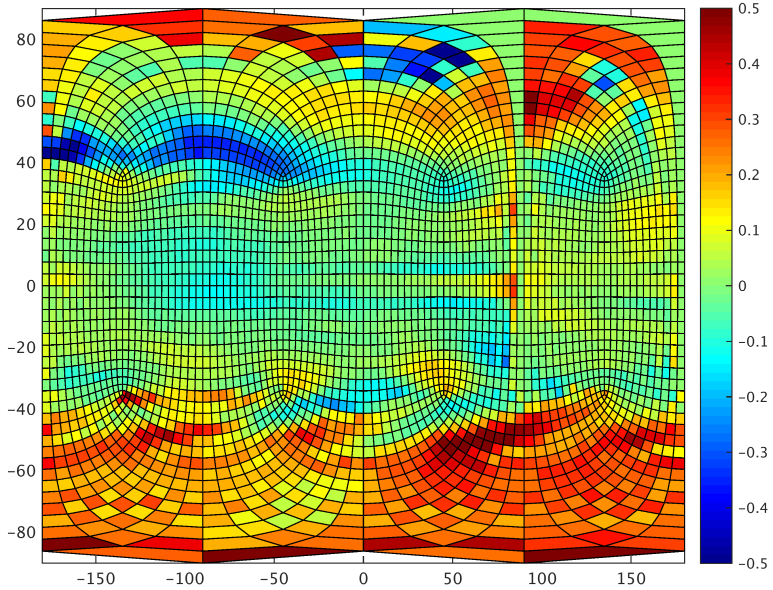
$$\epsilon = C_\epsilon \bar{e}^{3/2} l_\epsilon^{-1},$$

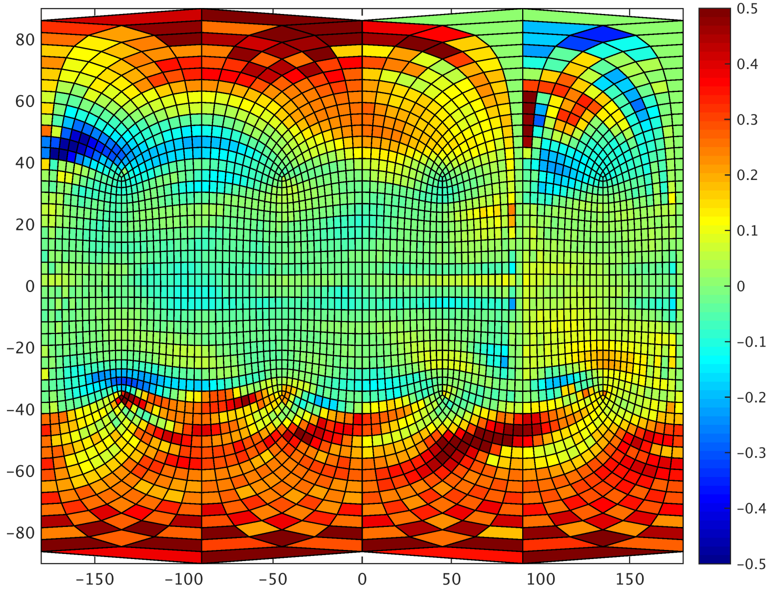
with

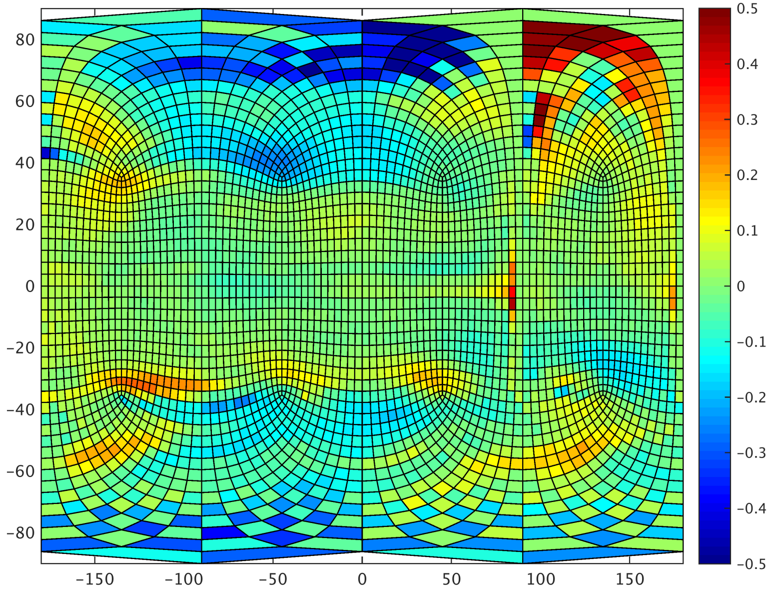
$$K_m = C_\kappa l_\kappa \bar{e}^{1/2},$$

$$K_\rho = K_m / P_{rt},$$

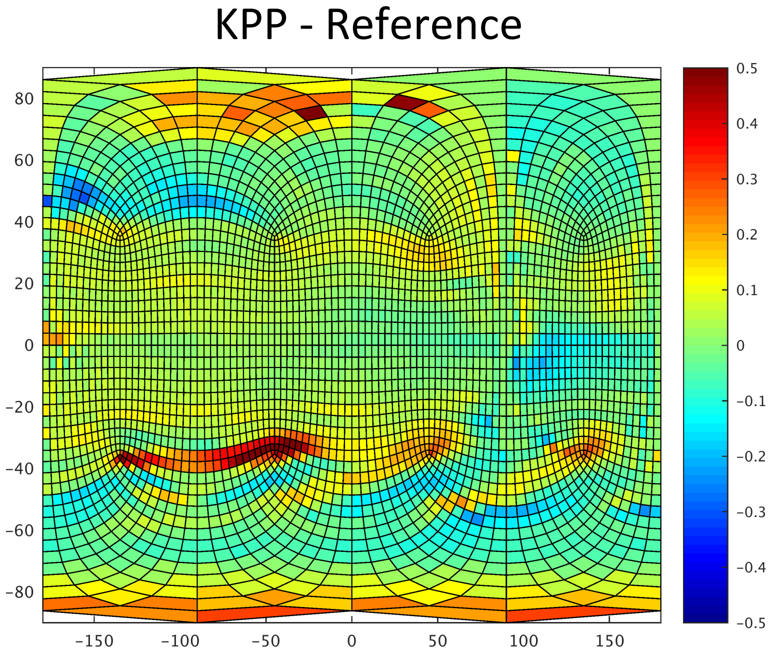
Temperature

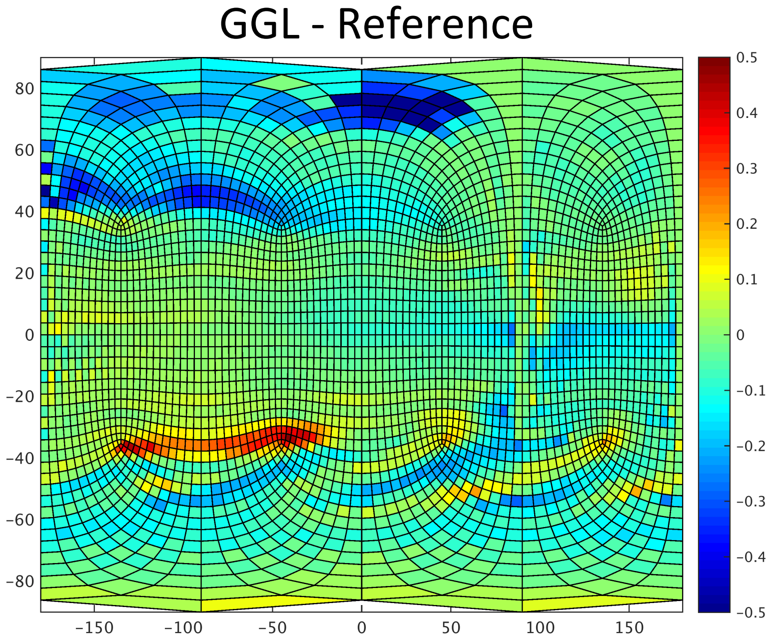

KPP minus Reference


GGL minus Reference


KPP minus GGL

Salinity


KPP - Reference
80
60
40
20
0
−20
−40
−60
−80
−150
−100
−50
0
50
100
150
0.5
0.4
0.3
0.2
0.1
0
−0.1
−0.2
−0.3
−0.4
−0.5


GGL - Reference
80
60
40
20
0
−20
−40
−60
−80
−150
−100
−50
0
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100
150
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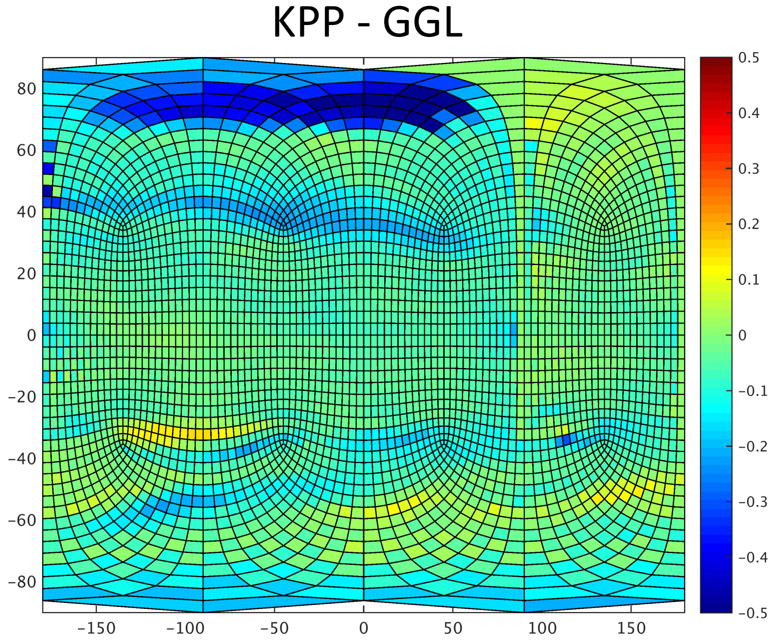
0
−0.1
−0.2
−0.3
−0.4
−0.5


KPP - GGL
80
60
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150
0.5
0.4
0.3
0.2
0.1
0
−0.1
−0.2
−0.3
−0.4
−0.5

Interaction between eddy parameterization and mixed layer

Gent and McWilliams scheme: $\Psi^* = K_{GM} S_y$ S_y: Isopycnal slope

- Near the surface $S_y \rightarrow$ infinity (or close)
- Need for a “tapering” scheme to control stability
- Mostly ad-hoc, but see Ferrari et al. (08, 10), Ferreira and Marshall (05)

Example: Gerdes et al. 1991

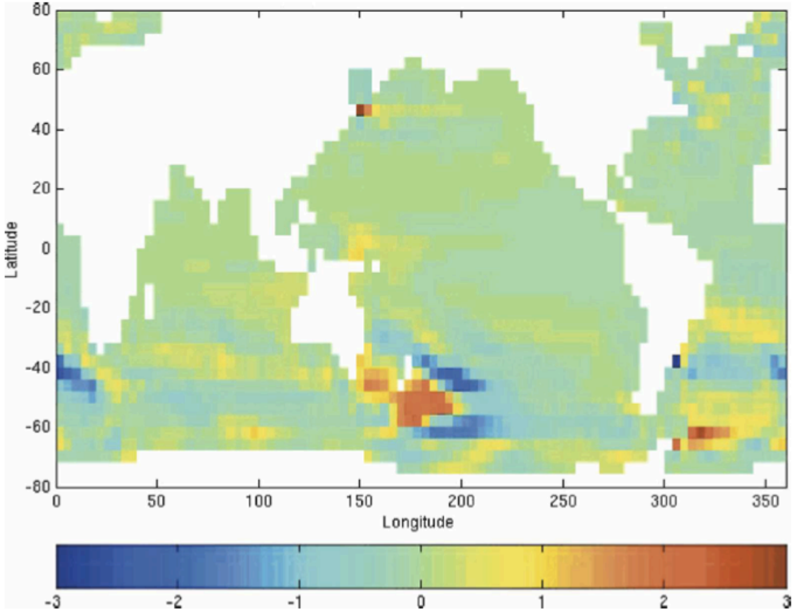
$$If\ \ S_{big} > |S_y| > S_{\max}$$

$$\Psi^* = \frac{S_{\max}^2}{S_y^2} K_{GM} S_y$$

$$If\ \ |S_y| > S_{big}$$

$$\Psi^* = 0$$

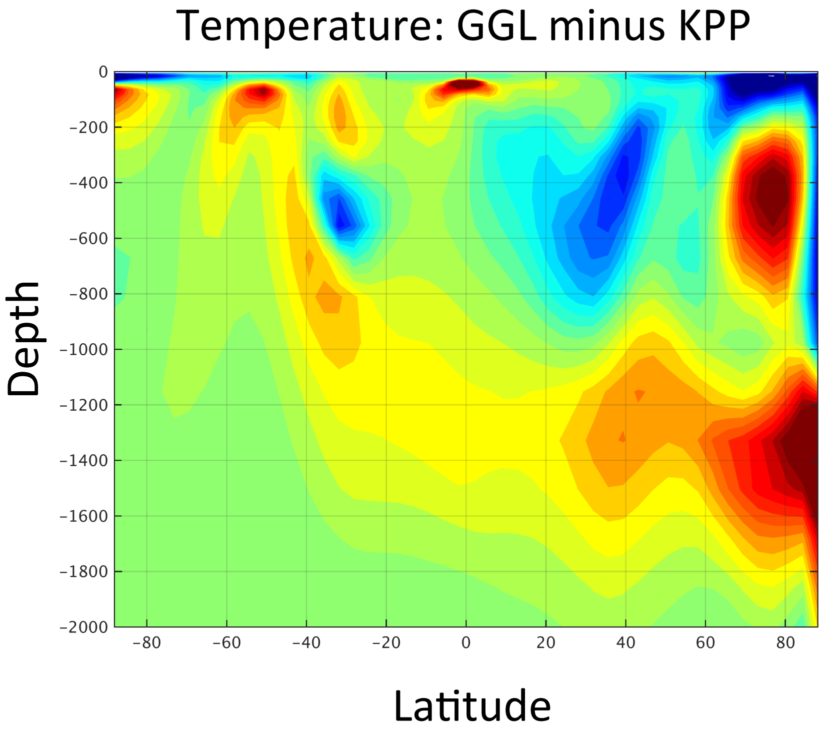
SST difference, 2 tapering schemes



°C

Differences:

- Without/with BL scheme :
  - Locally: +/- 0.3-0.5 ^{o}C
  - Globally: + 0.2 ^{o}C
- Between BL schemes:
  - Locally: +/- 0.3-0.5 ^{o}C
  - Globally ~ 0 ^{o}C



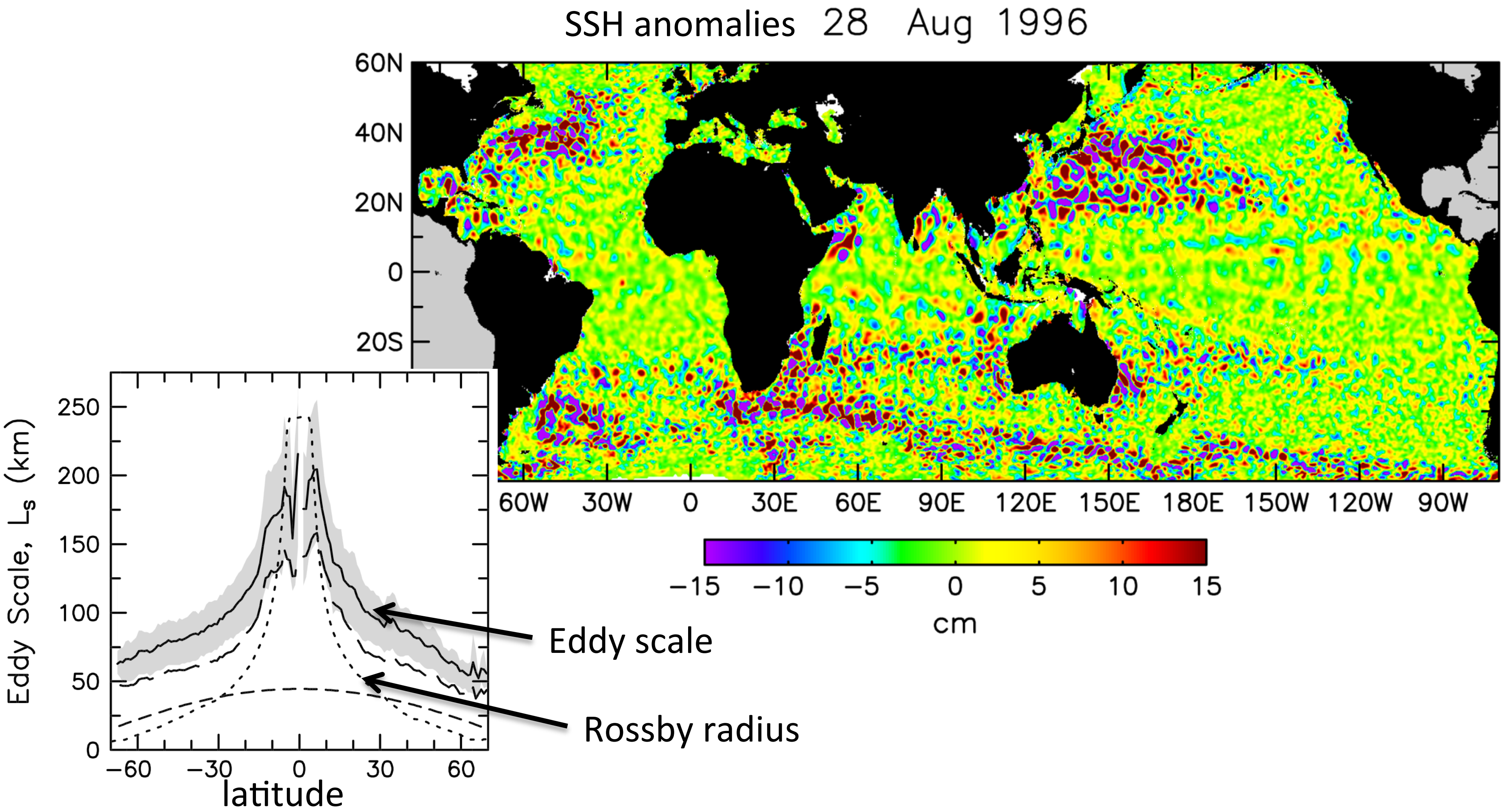
- Can one tune GGL and KPP so they do the same thing?
  → Probably yes

- Can we always do it?

- If yes, it is good?
  → Possibly no.
  → i.e. there are too many "unknown" parameters
  → Not enough O(1) constants

Ocean are full of mesoscale eddies

- Coherent vortices, radius of about 50-100 km
- Generated by baroclinic and barotropic instabilities
- ~215,000 eddies with 4 weeks or longer lifetime over 20 years (Chelton et al., 2011)
- They are everywhere

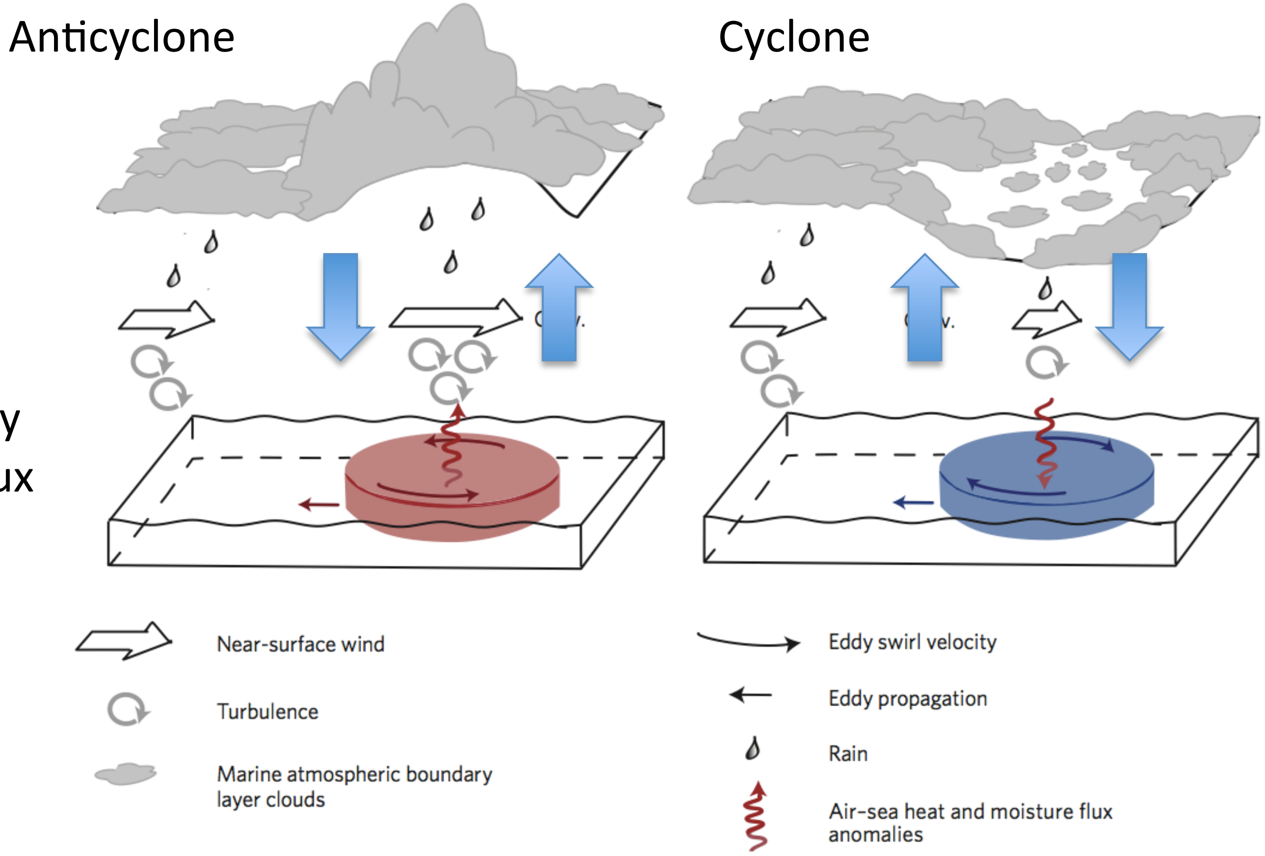

SSH anomalies 28 Aug 1996

Mechanism

- Dominant mechanism: Modulation of the BL vertical mixing

- Warmer SST

→ Decreased vertical stability
→ Downward momentum flux
→ Decreased shear
→ Larger surface winds

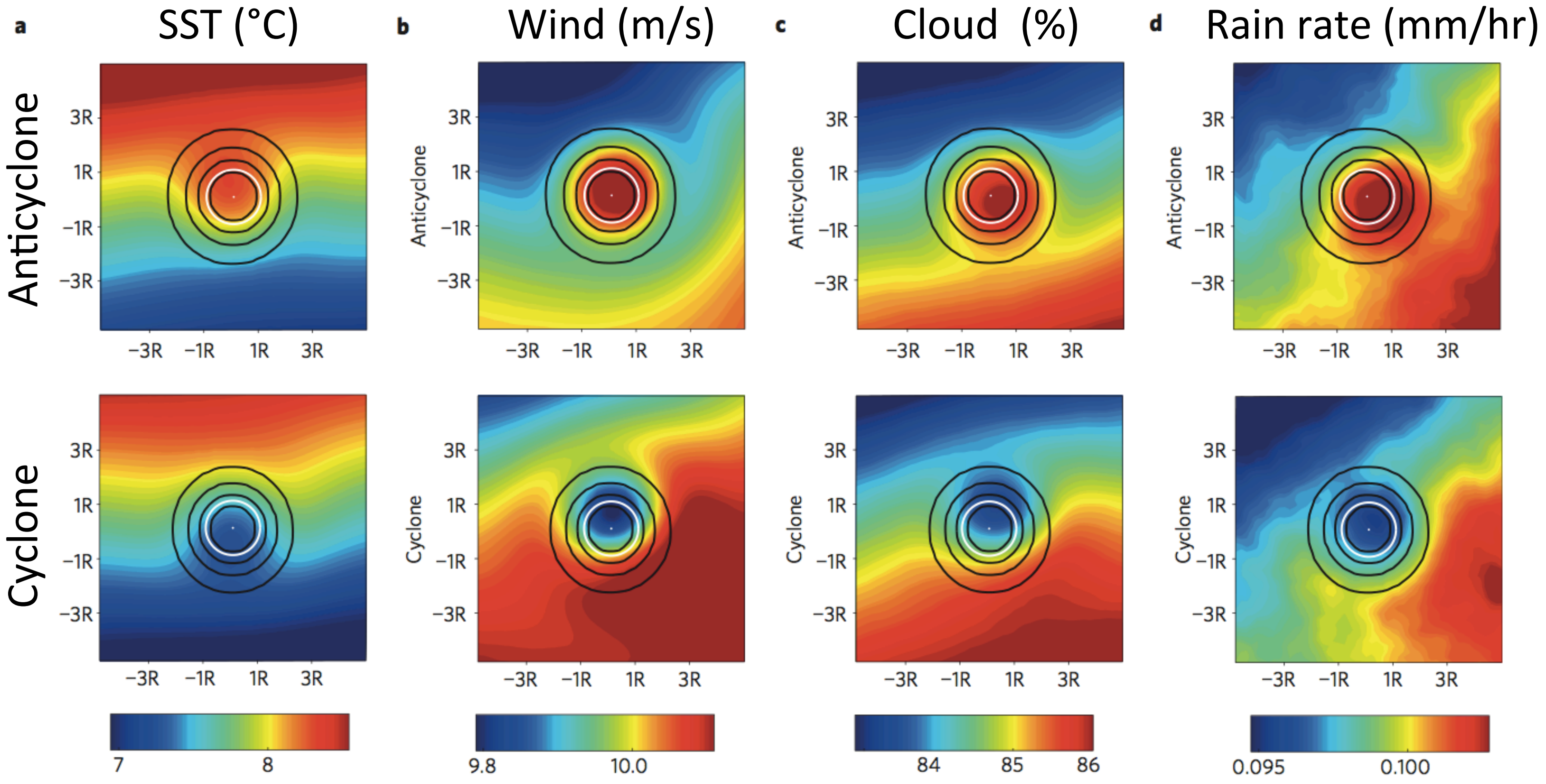


Pressure gradient adjustment does not appear important but it may depend on the scale.

Frenger et al. 2013

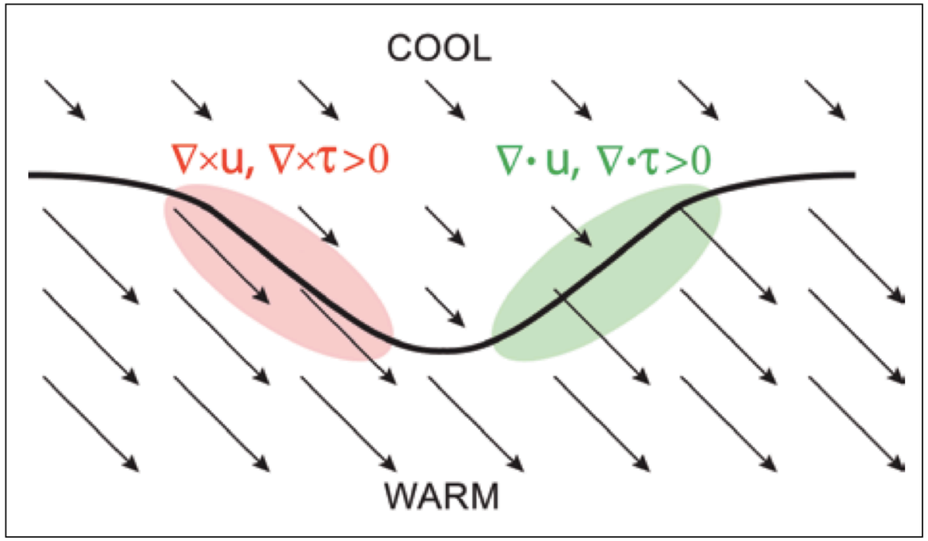
Observations: Mid-latitudes

Southern Ocean case (Frenger et al. 2013):

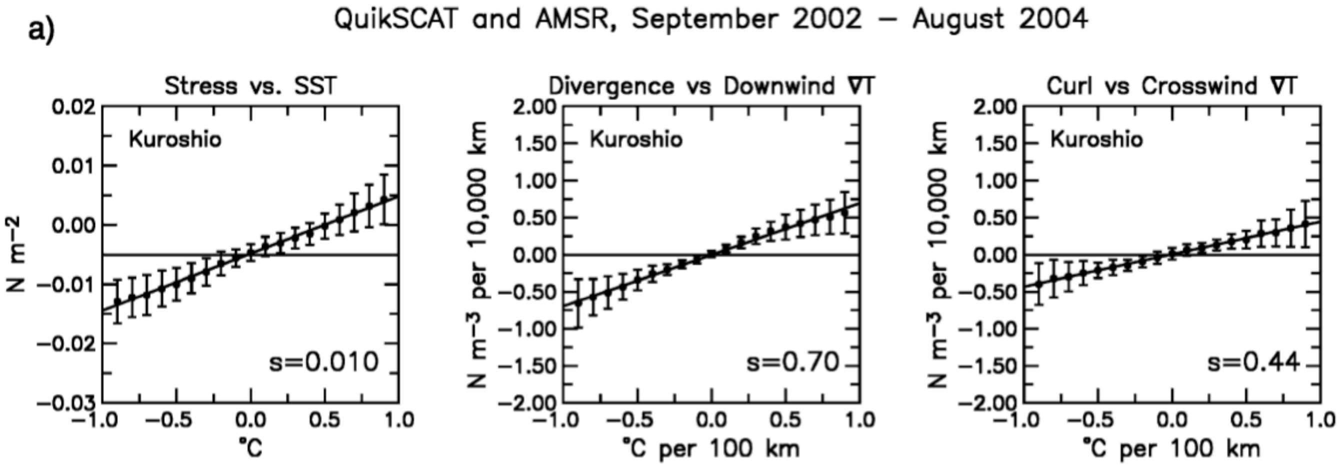


Wind direction →

See Bourras et al. (2004) for a North Atlantic example.



Maloney and Chelton (2006)



On the oceanic mesoscale:

$$\vec{u}' \propto T'$$

$$\nabla \times \vec{u} \propto \text{ crosswind } \nabla T$$

$$\nabla \cdot \vec{u} \propto \text{ downwind } \nabla T$$

Effects:

- $\vec{\tau}_a = \rho_a C_D |\vec{u}_a - \vec{u}_o| (\vec{u}_a - \vec{u}_o)$ → eddies modify the wind stress they experience
- Surface fluxes modified by presence of eddies (latent, sensible, radiative)

Abyssal circulation, diapycnal mixing and Bottom boundary layer

de Lavergne et al. 2016

1D advection-diffusion:

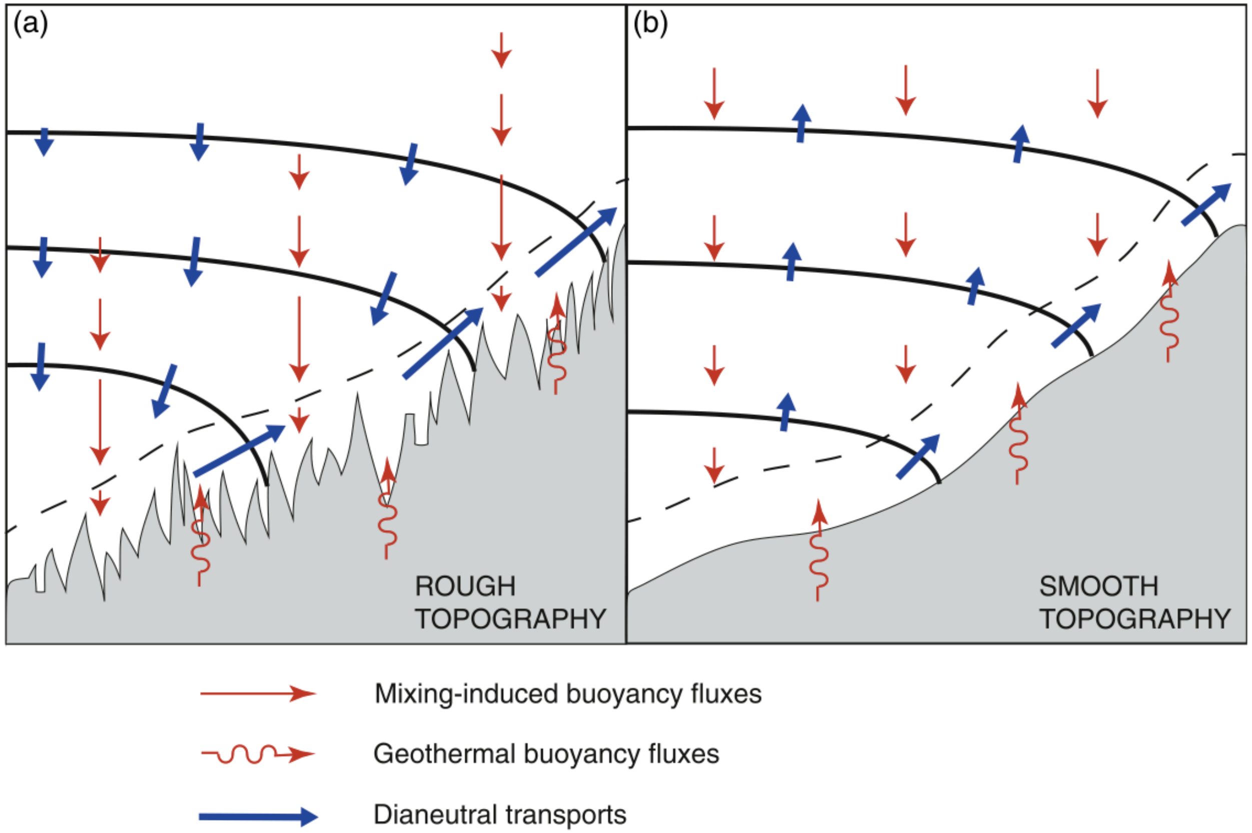
$$w^* N^2 = \frac{\partial}{\partial z}\left(k_V N^2\right)$$

Available energy for mixing

$$k_V N^2 = R_f \varepsilon_T$$

<u>Observations</u>: ε_T increases upward

→ Global overturning is closed within the BBL



$\varepsilon_T \downarrow$ with height

downwelling

$\varepsilon_T \uparrow$ with height

upwelling

<u>But</u> $k_v N^2 = 0$ @ bottom

- What do a ML scheme should do?
  - → KPP does ocean interior mixing: Double-diffusion, shear mixing, etc…
  - → GGL90 predicts TKE/Ks for the whole water column although in practice TKE is surface intensified
  - → What about convective adjustment?

- Interaction between eddy parameterization and surface boundary layer
  - → More generally interactions between subgriscale processes (e.g. Eden & Olbers' work)
  - → Horizontal v. vertical mixing in BL?
  - → When do we introduce new parameterization or shut off a parameterization?

- Under sea ice/ice shelf?
  - Models use mixed layer schemes which have been developed/tested for open ocean conditions

- Bottom/benthic boundary layer? Downslope/overflows mixing scheme?
  - These are relatively unexplored

- Eddy-ABL interactions:
  - Does it matter ? Should it be parameterized in ocean-only and coarse climate models?