Mesoscale eddy dynamics in the Arctic Ocean: Requirements for tracer transport parametrizations

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The role of eddy transport and the resolution problem

- The oceanic heat transport into the Arctic is mediated by time-mean boundary currents
- But the bulk of the air-sea heat loss takes place away from the boundary currents.
- Boundary currents are baroclinically unstable. Lateral eddy heat transport crucial.

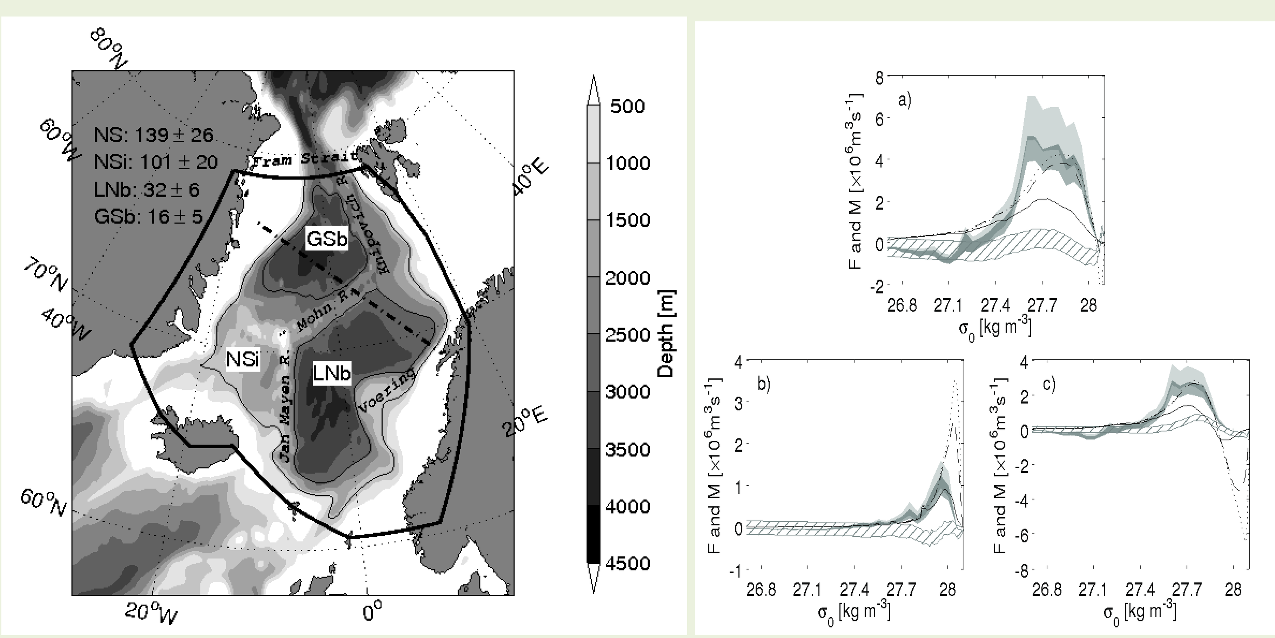


Fig. 1: (left) Surface heat loss and (right) diapycnal overturning transport in the Nordic Seas (from Isachsen and Nøst, 2013).

- The gravest internal deformation radius:
  - 5-25 km in deep basins
  - <5 km on shelves
- Topographic slopes will modify normal Eady scaling for EKE and eddy length scales.

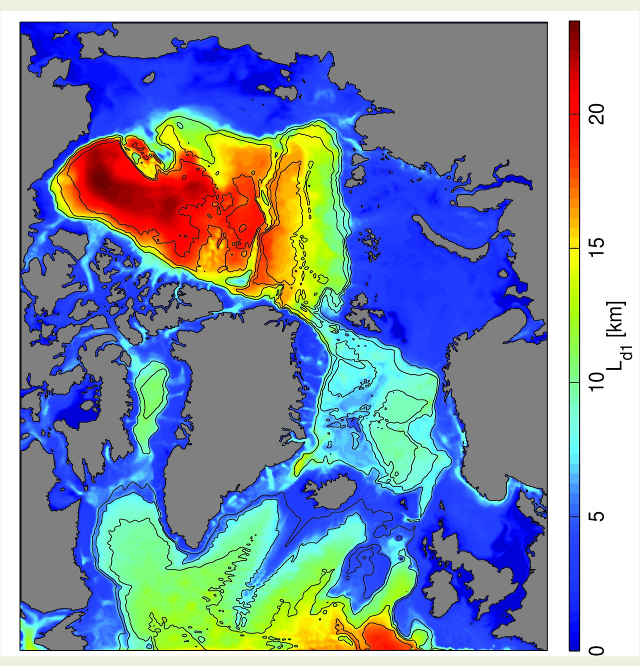


Fig. 2: Gravest internal deformation radius

- Eddy transport needs to be heavily parametrized in climate models.
- Models appear to have particular problems with topographic effects.

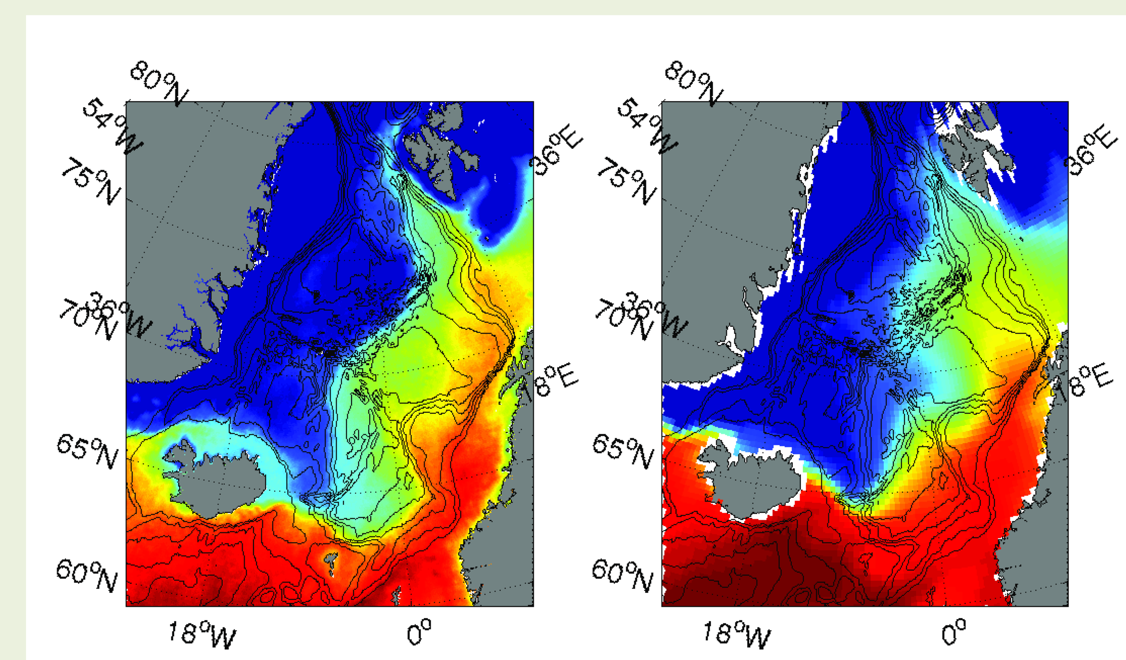


Fig. 3: SST from (left) observations (OSTIA analysis) and (right) NorESM climate model.

Useful *forms* of eddy parametrizations

- Fluxes should be adiabatic in interior → overturning streamfunction (GM)
- Observations and modeling from the Southern Ocean suggest that eddy overturning over continental slopes can lift dense water onto shelves.

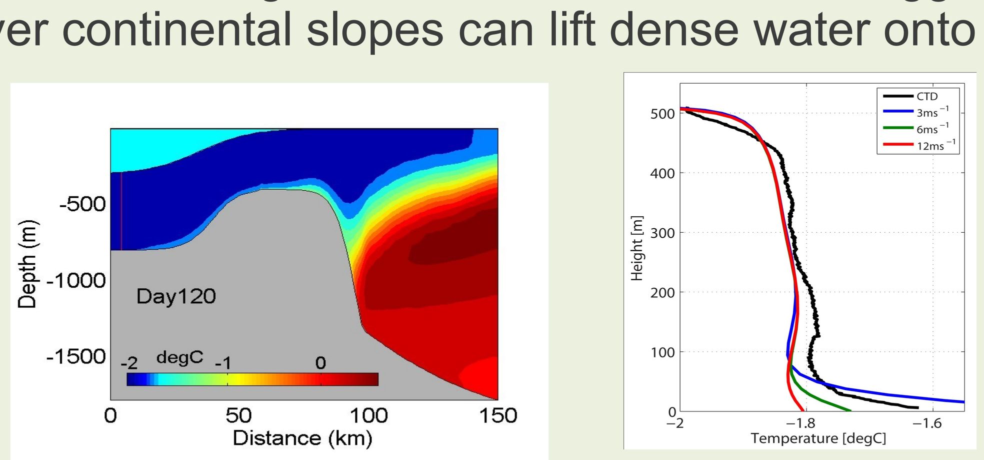


Fig. 4: (left) Modeled hydrography over Antarctic continental slope and (right) modeled and observed hydrographic profiles on the shelf (Nøst et al., 2011).

- Ferrari et al. (2010) proposed a GM-like streamfunction param. with top and bottom boundary layers:

$$\left(\rho_z - c^2\frac{\rho_0}{g}\frac{d^2}{dz^2}\right)\Psi^\# = -\kappa\nabla_h\rho, \qquad -H<z<0$$
$$\Psi^\# = 0, \qquad z\in[-H,0]$$

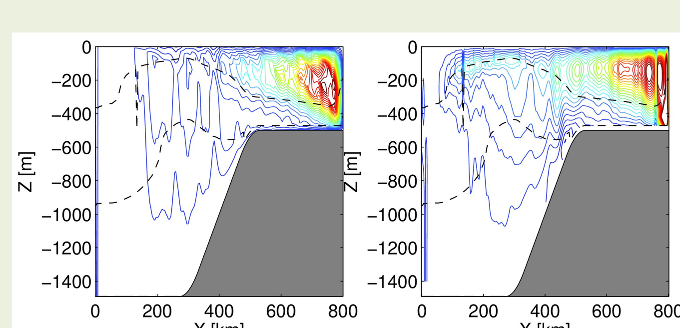


Fig. 5: Eddy overturning in an eddy-resolving model: (left) diagnosed, (right) parametrized.

Realistic diffusivities

- Classic (Stone, 1972) *K=Vtw*Ld1* appears not to work.
- Diffusivities (and eddy length scales) sensitive to bottom topography.

$$\kappa = \alpha_S V_{tw} L_d \qquad \kappa = \alpha_S \frac{g}{|f|}\left(\overline{\eta^2}\right)^{1/2}$$

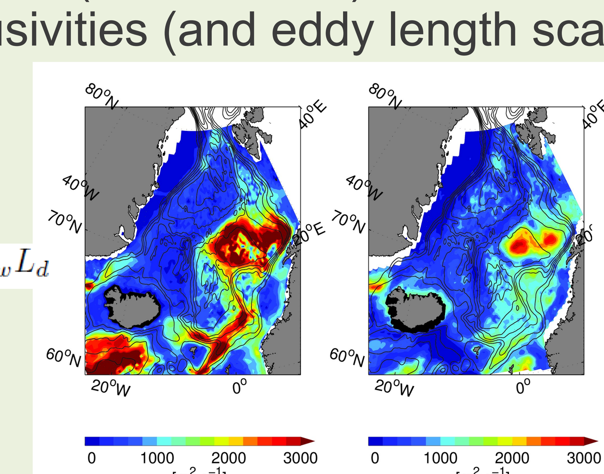


Fig. 6: Diffusivitiy estimates: (left) Stone (1972) and (right) Holloway (1986).

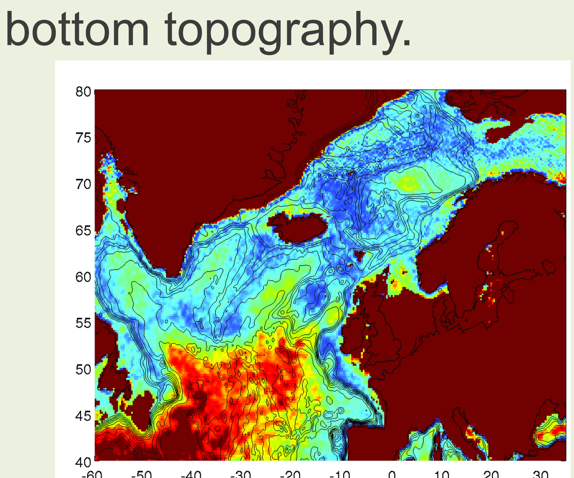


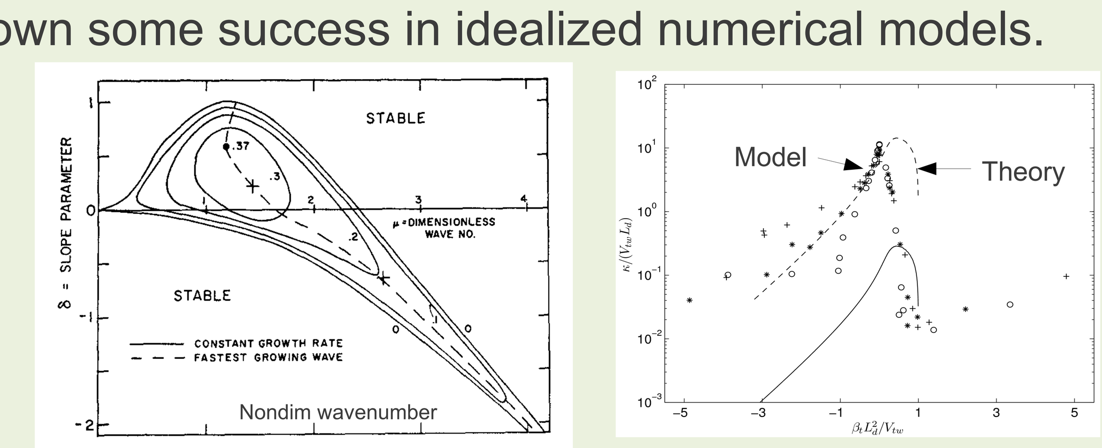
Fig. 7: Estimates of eddy length scales from along-track altimetry data.

- Modified Eady theory has shown some success in idealized numerical models.

$$\delta = \frac{h_x}{-\rho_x/\rho_z} \sim -\frac{(f/H)\cdot h_x \cdot L_d^2}{V_{tw}}$$

Topographic slope; Isopycnal slope

Fig 8: Modified Eady theory (Blumsack and Gierasch, 1972): (left) growthrates and (right) diffusivities from theory and diagnosed in eddy-resolving model (Isachsen, 2011).



Assessments of "coarse-scale", "eddy-permitting" and "fully eddy-resolving" simulations (Work in progress)

- A comparison of eddy dynamics and transport in models at 20km, 4km and 800 m horizontal resolution (also with varying vertical resolution).
- Using ROMS primitive equation model (terrain-following vertical coordinates).
- Validation against in situ current and hydrographic observations.

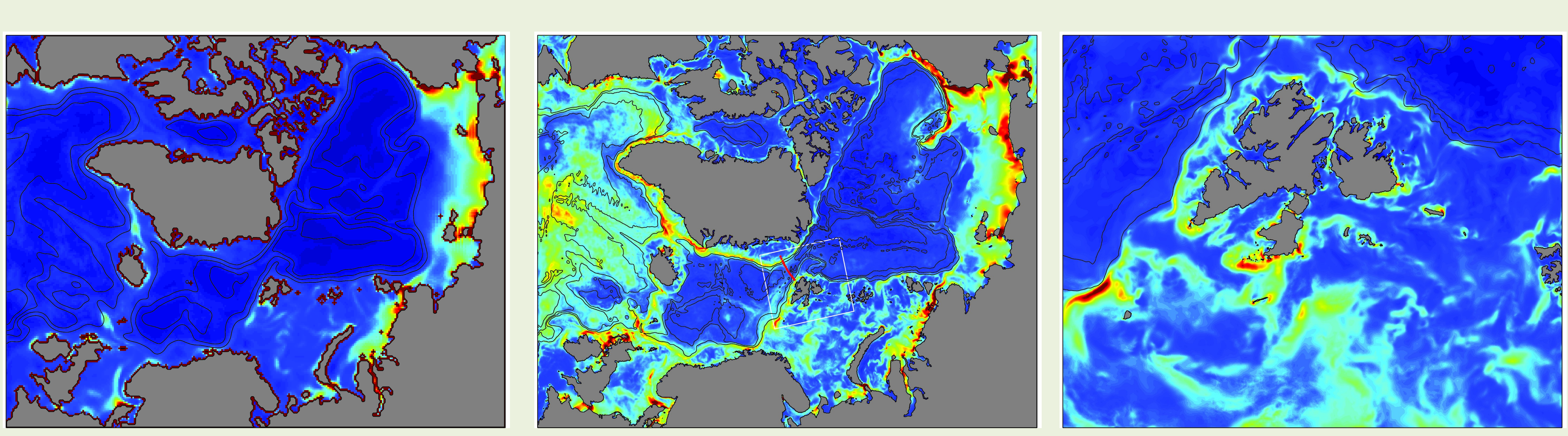


Fig. 8: Winter-time EKE in models with (left) 20km (middle) 4 km and (right) 800 m horizontal resolution. Fram Strait mooring array is shown with red dots in middle panel.

- Questions:
  - What resolution is adequate to resolve fluxes over cont. slope?
  - How important is Eady vs. non-Eady dynamics and associated transport?
  - How well do current parametrizations do? What modifications are crucial?

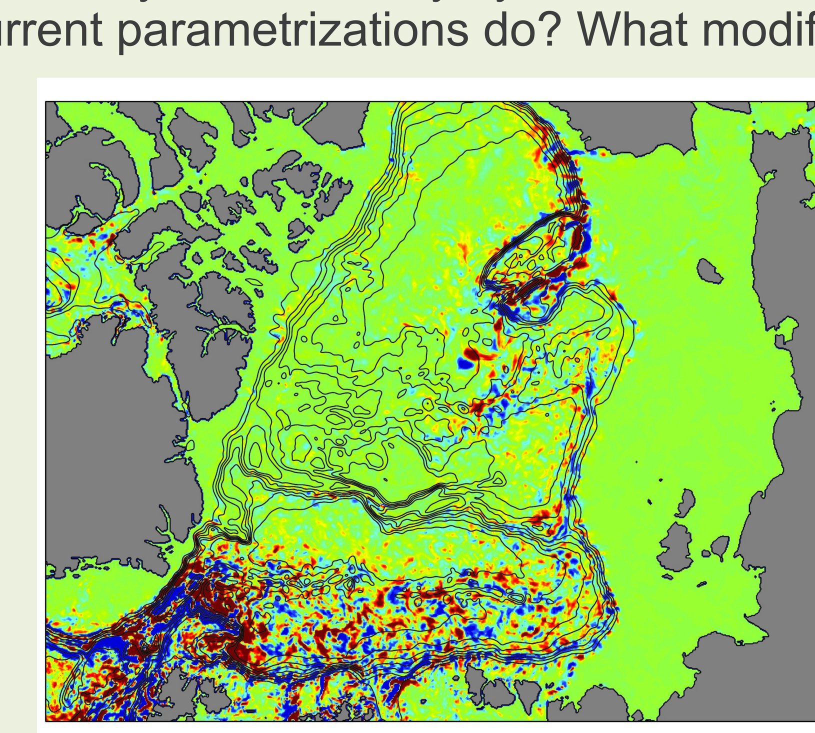


Fig. 9: Winter-time depth-integrated eddy temperature flux convergences in 4 km model (red=advective heating, blue=advective cooling).

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