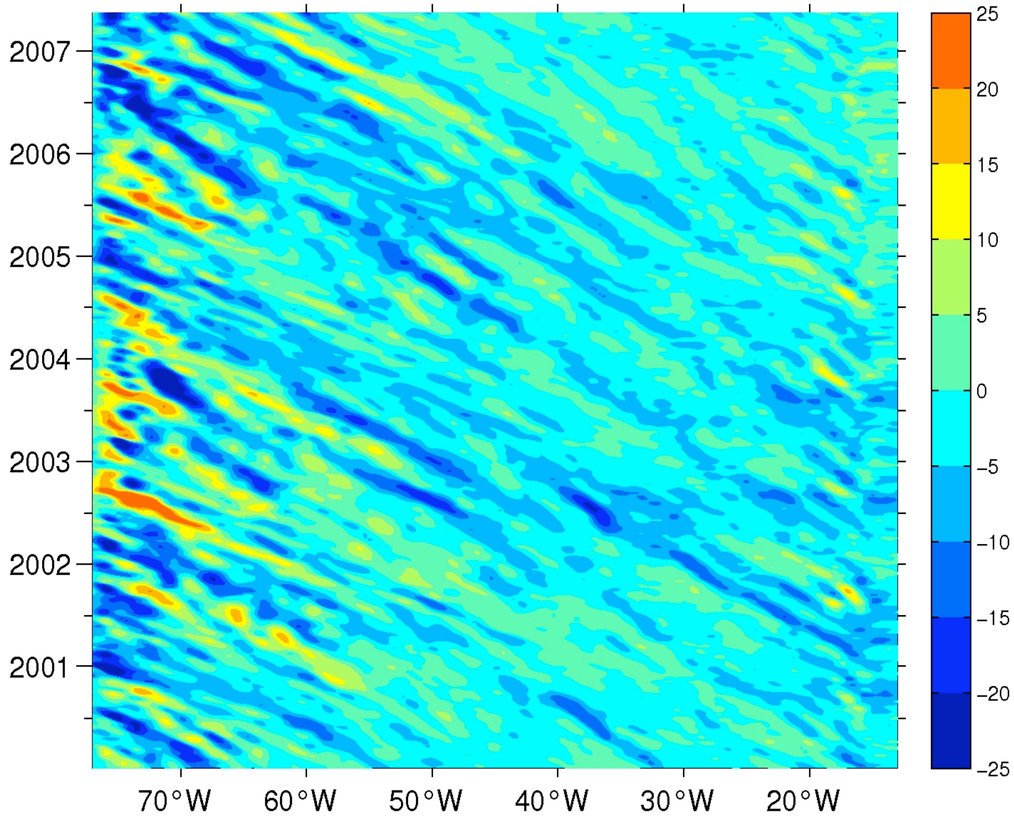
Basin-wide integrated volume transports in an eddy-filled ocean



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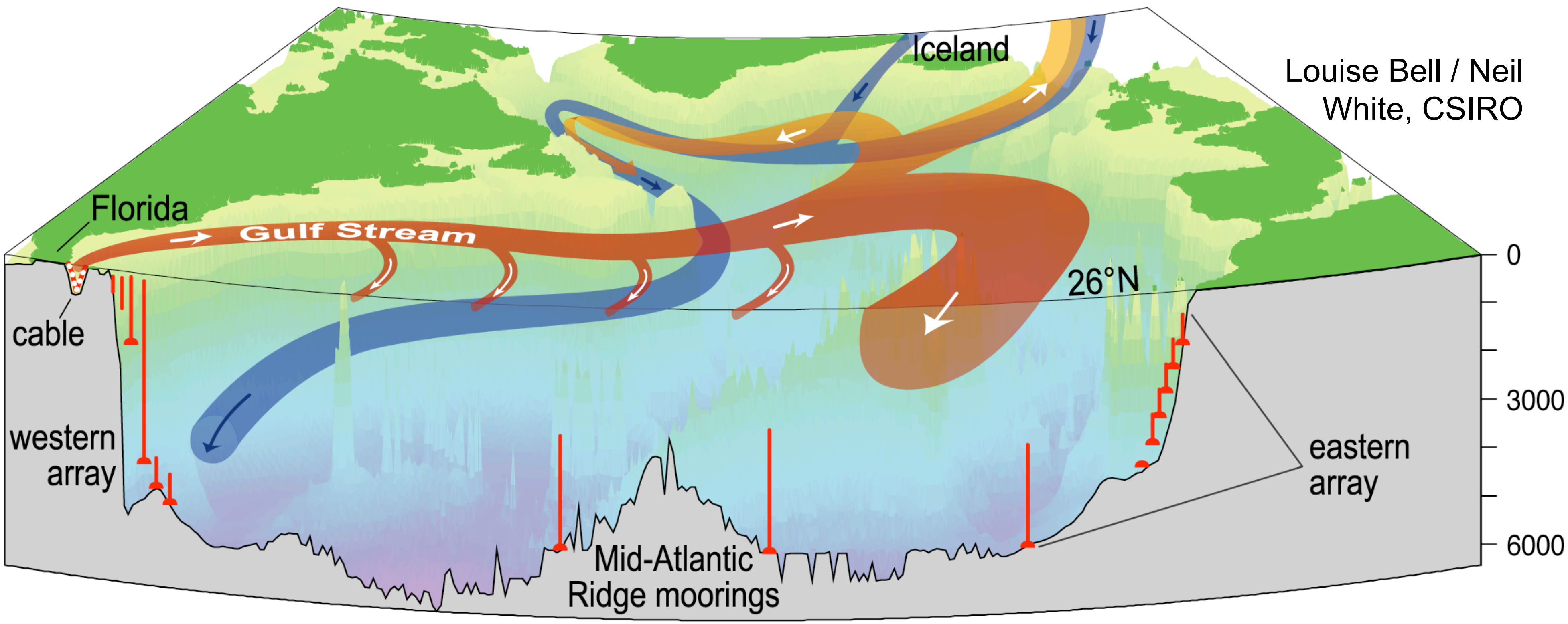
Rapid Climate Change

The RAPID / MOCHA observing system at 26.5°N

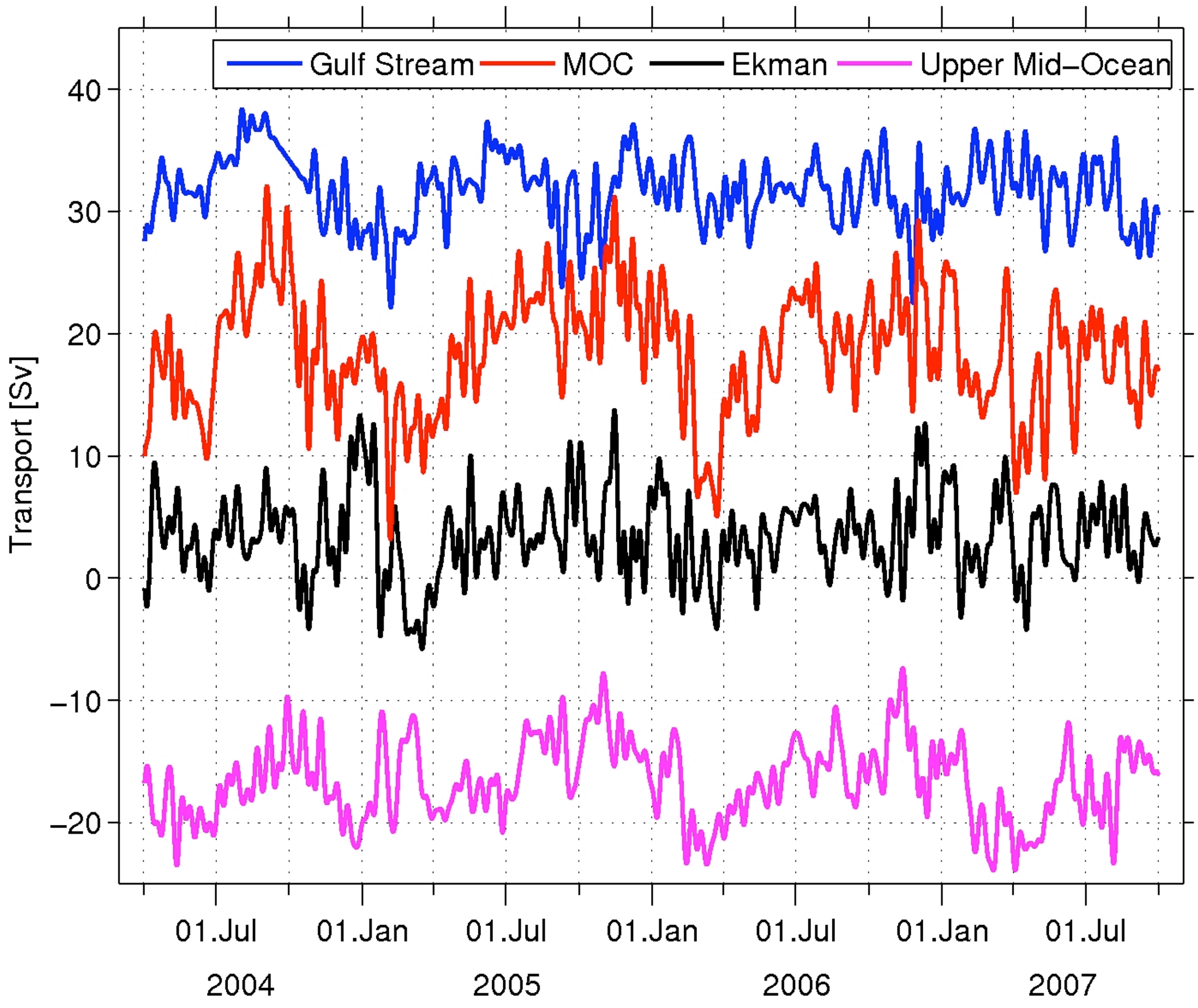


Gulf stream transport through Florida Straits - telephone cable

Ekman transport - scatterometer data

Mid-ocean transport - density structure and current meters

Variability in the MOC and its components



Statistics

Gulf Stream
+31.7 ± 2.8 Sv

MOC
+18.5 ± 4.9 Sv

Ekman
+ 3.5 ± 3.4 Sv

Upper Mid-Ocean
-16.6 ± 3.2 Sv

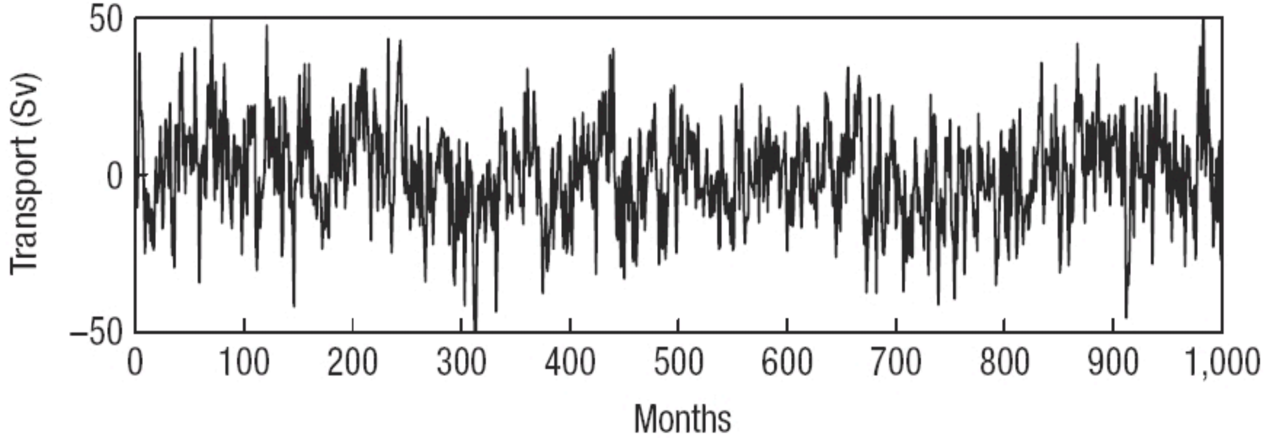
The net upper mid-ocean transport is a reliable measure of the mid-ocean component of the MOC:

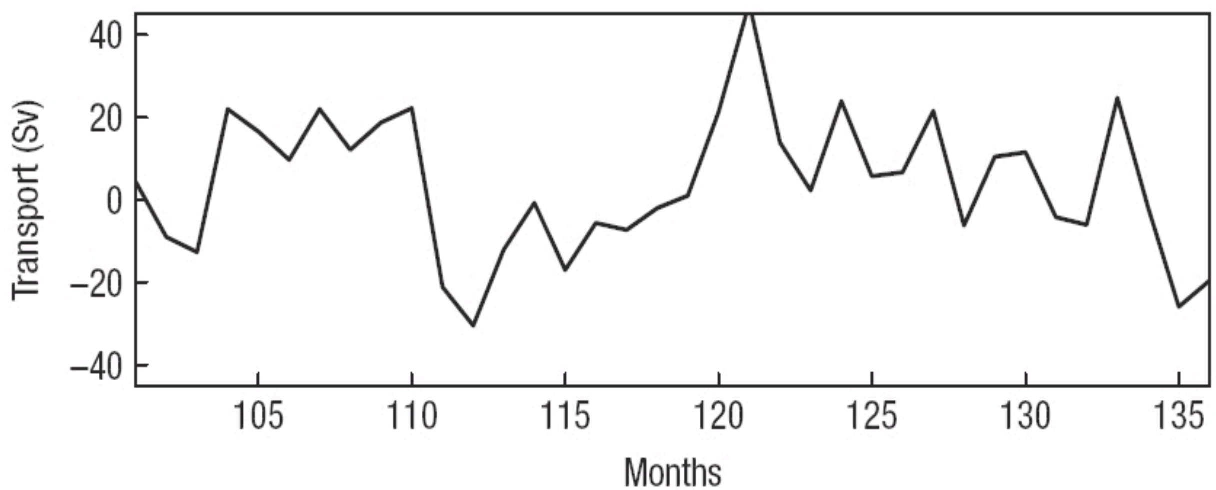
$$U_{MO} = \int_{Z=1000m}^{Z=0} T_{MO}\, dz$$

Sea surface height (η) variability = ± 16 cm rms near the western boundary.

Projected onto vertical modes of horizontal velocity:

- ±16 Sv transport fluctuations above 1000m
- substantial intraseasonal to decadal transport variations





Wunsch (2008)

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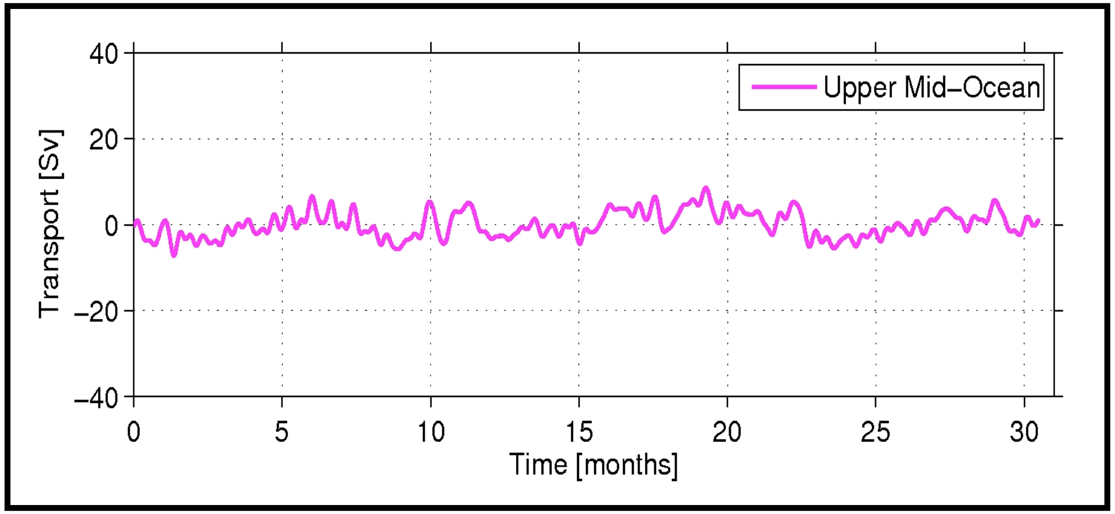
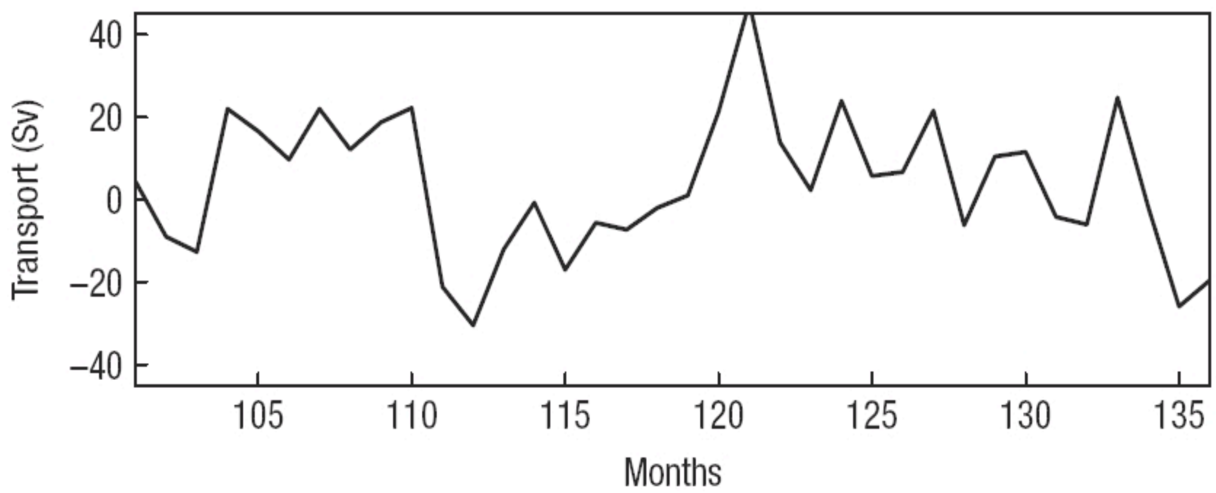
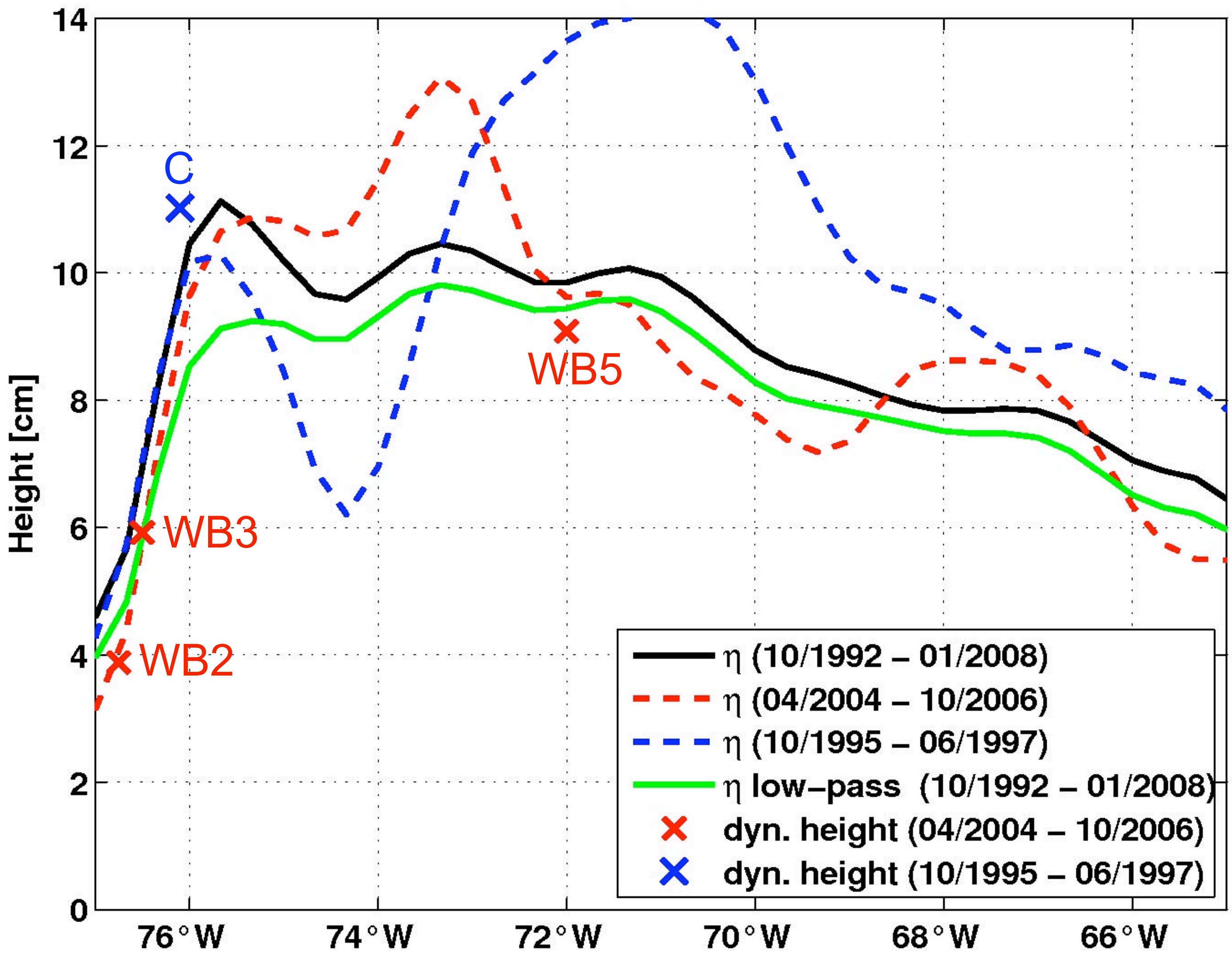
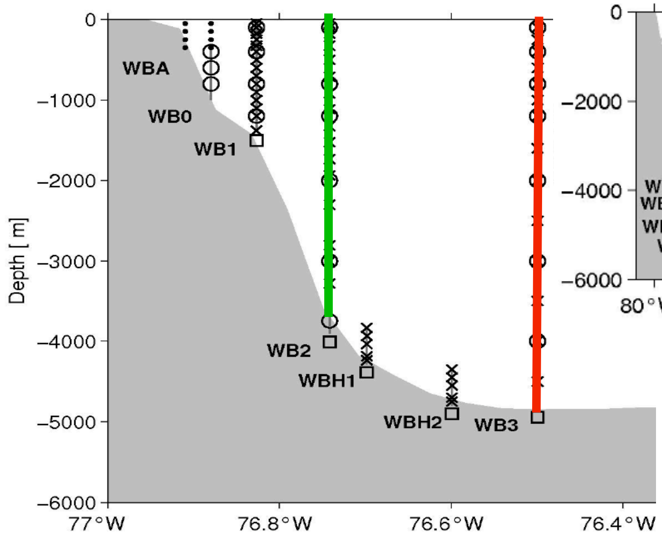


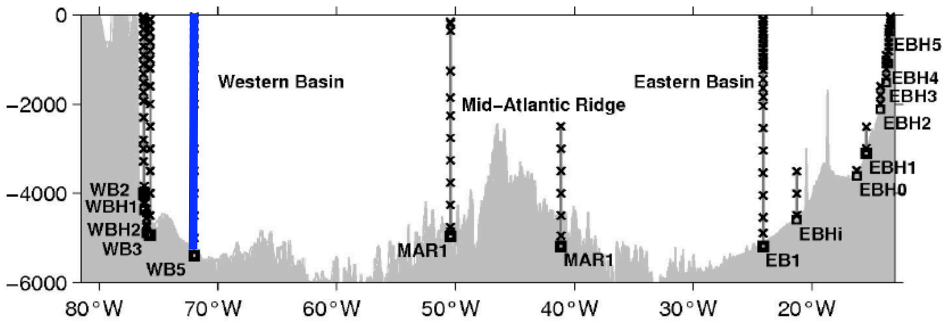


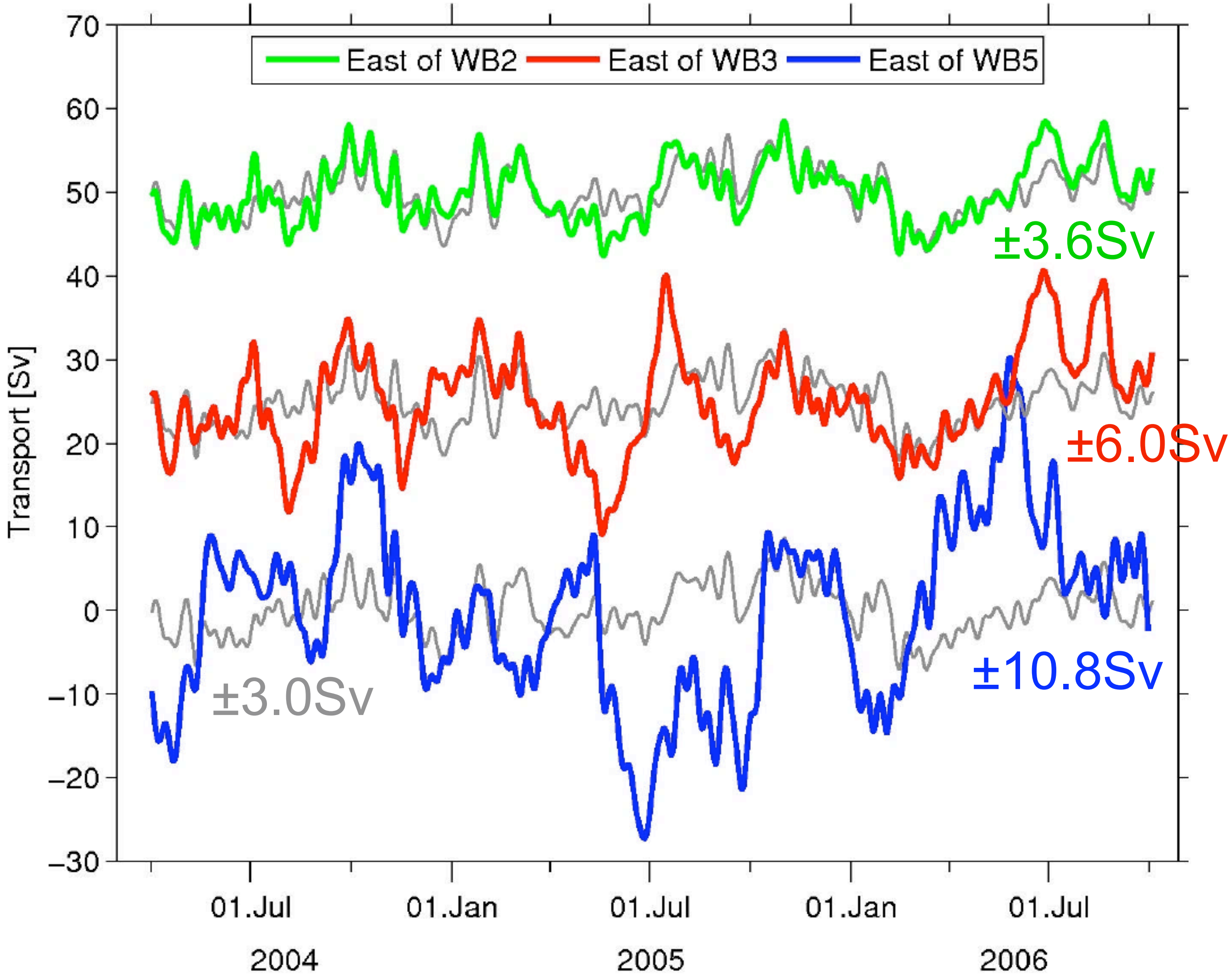
Wunsch (2008)

RMS amplitude of sea surface height and dynamic height along 26.5°N



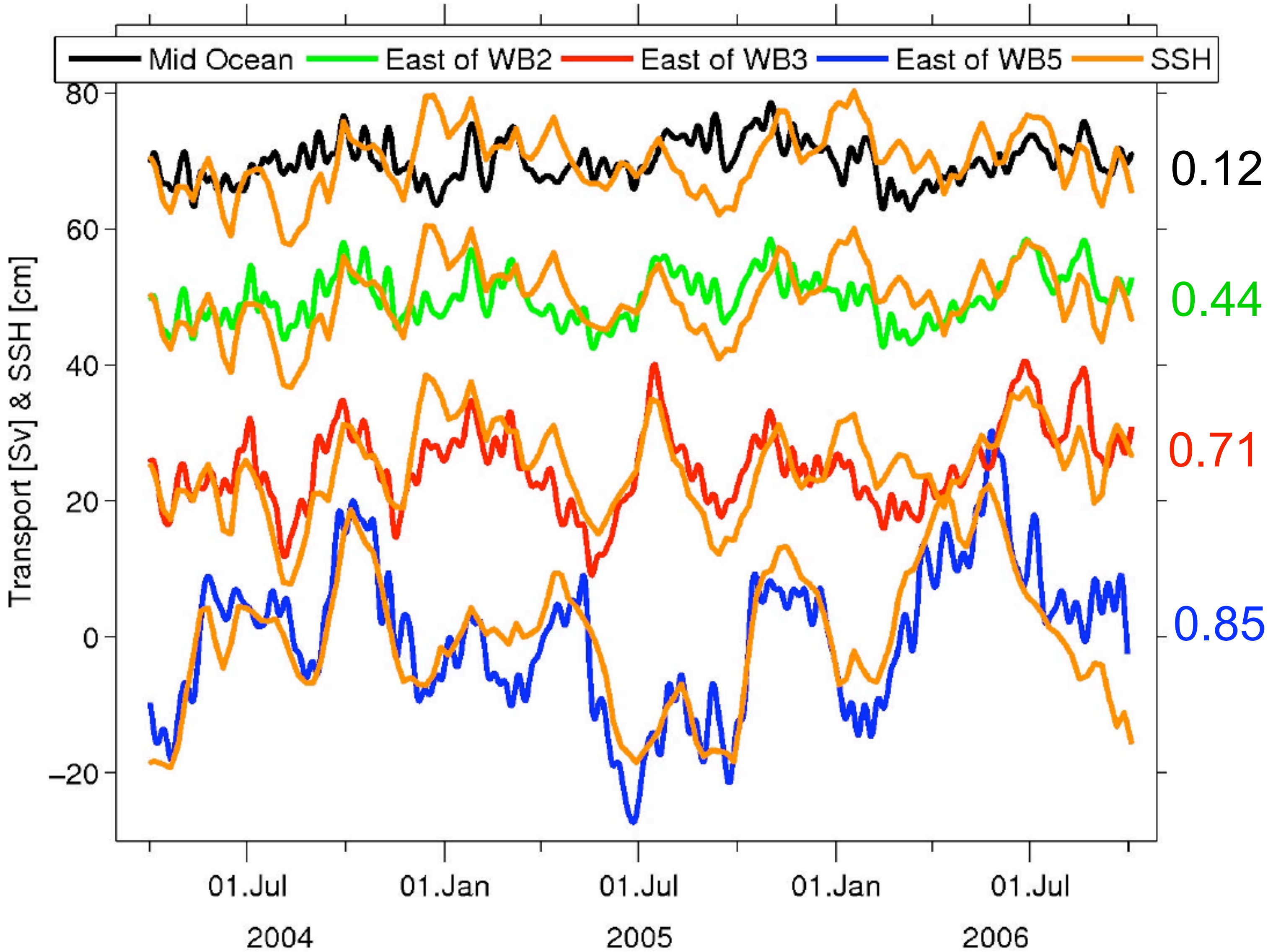






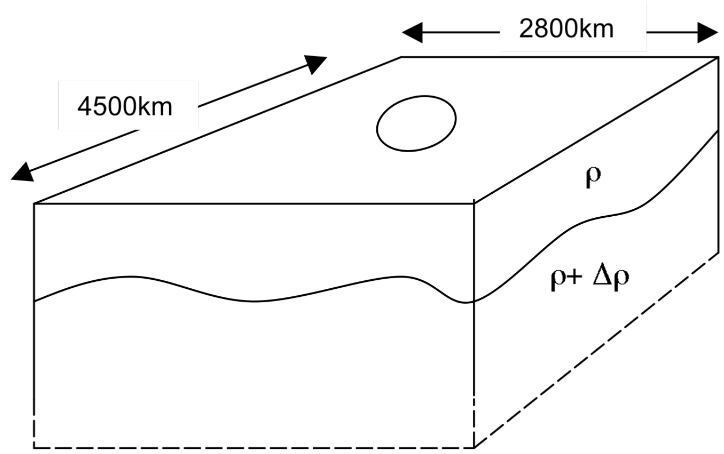
Northward transport fluctuations (Sv) above 1000m and east of WB2, WB3 and WB5.

Can sea surface height difference be used to estimate upper ocean transport?



Correlation of sea surface height difference between eastern boundary and each mooring with the zonally-integrated transport to the east.

What happens to an eddy as it approaches the western boundary? Reduced-gravity model experiments on a mid-latitude β-plane.



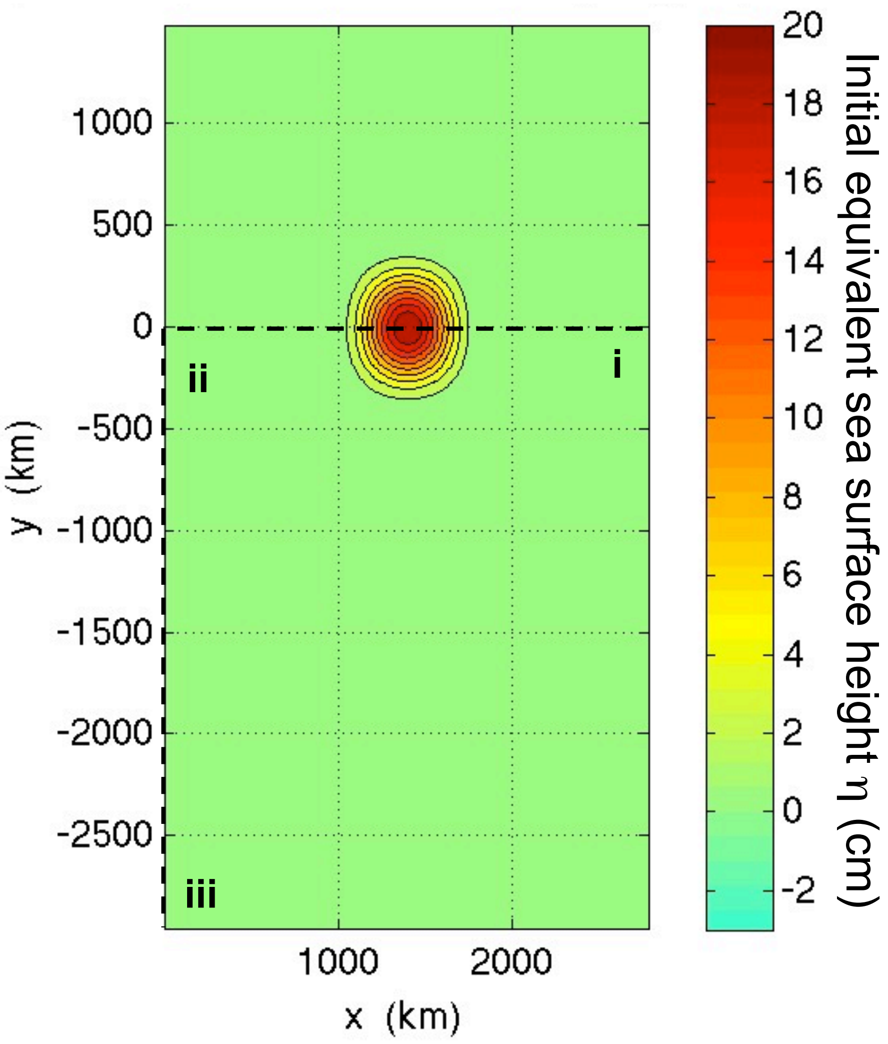
Δx = 14km (1/8°)

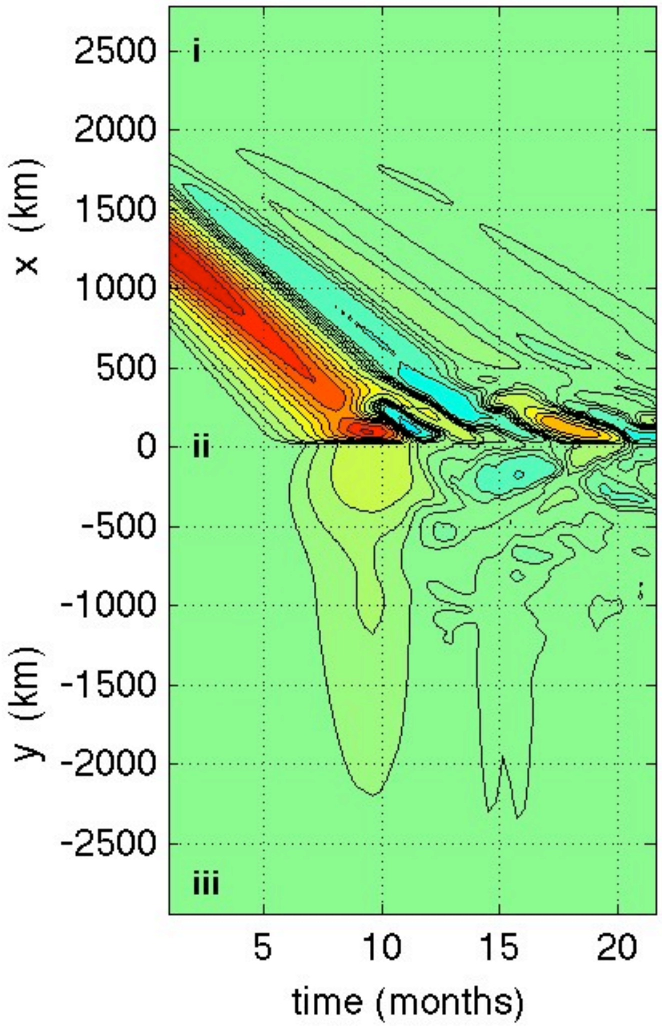
$f_0 = 0.65 \times 10^{-4}\ s^{-1}$

$g' = 0.015\ ms^{-1}$

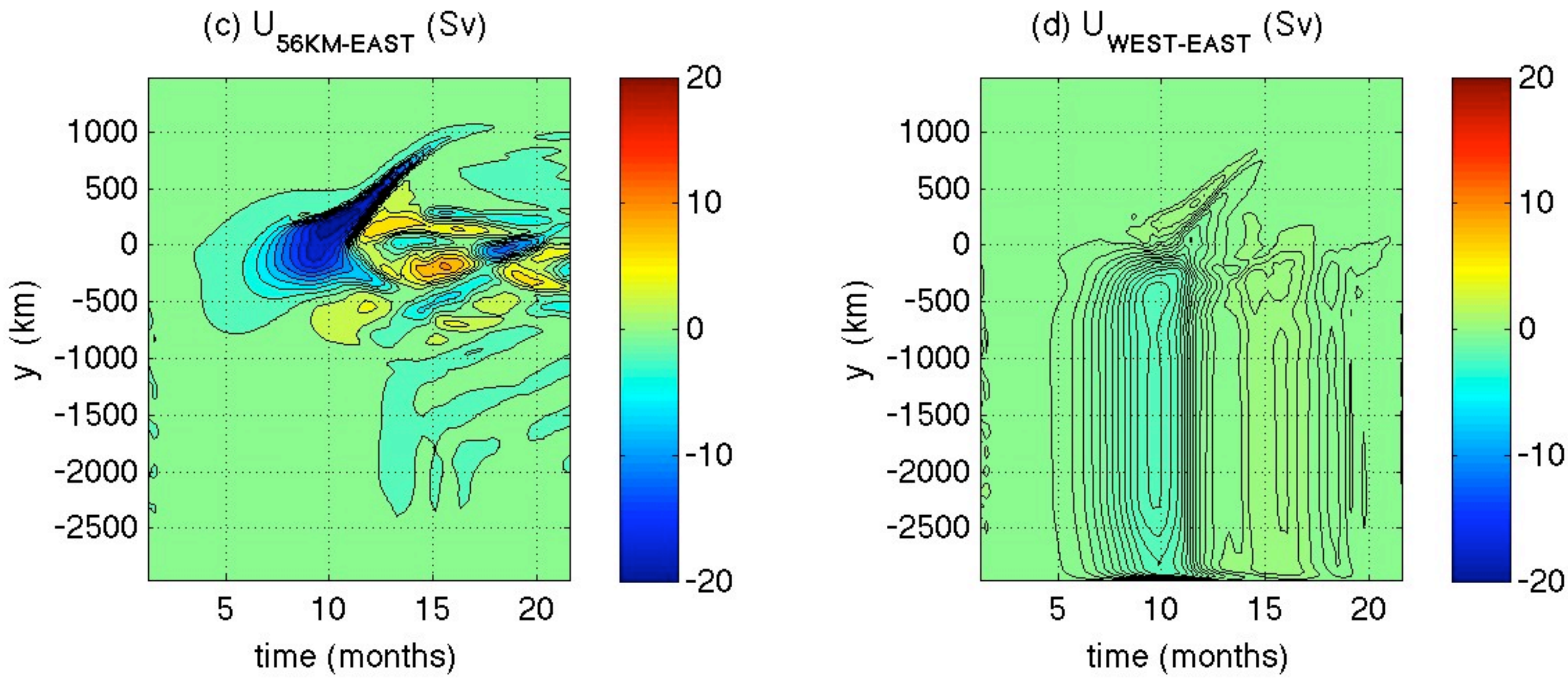
H = 750m

$A_0 = 30\ m^2 s^{-1}$

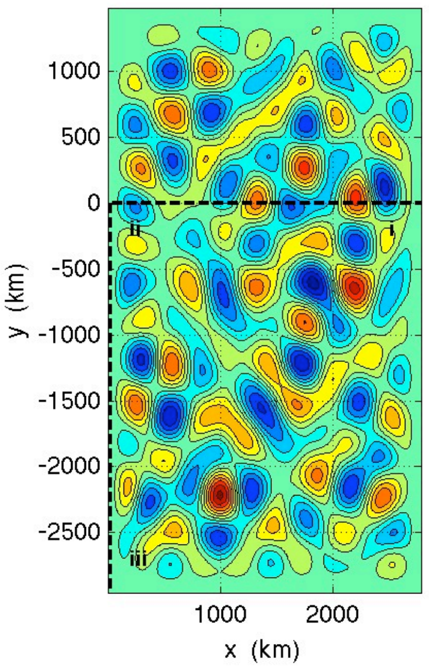




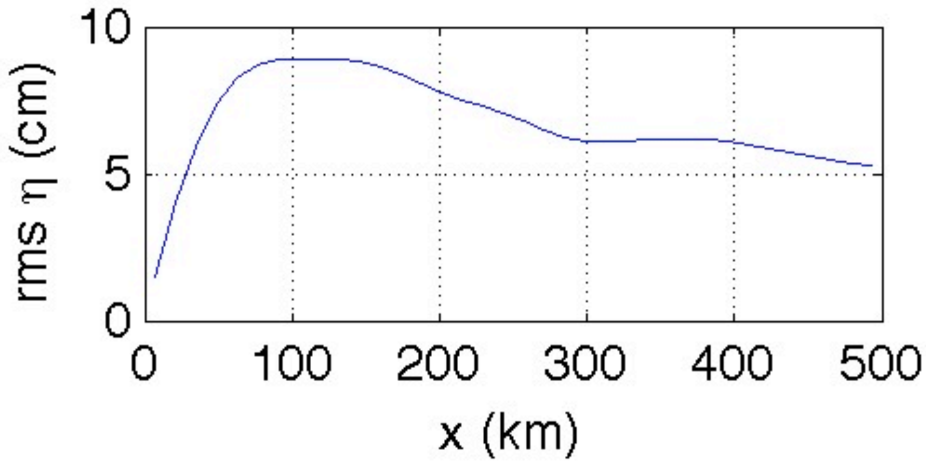
Net northward transport within the upper model layer

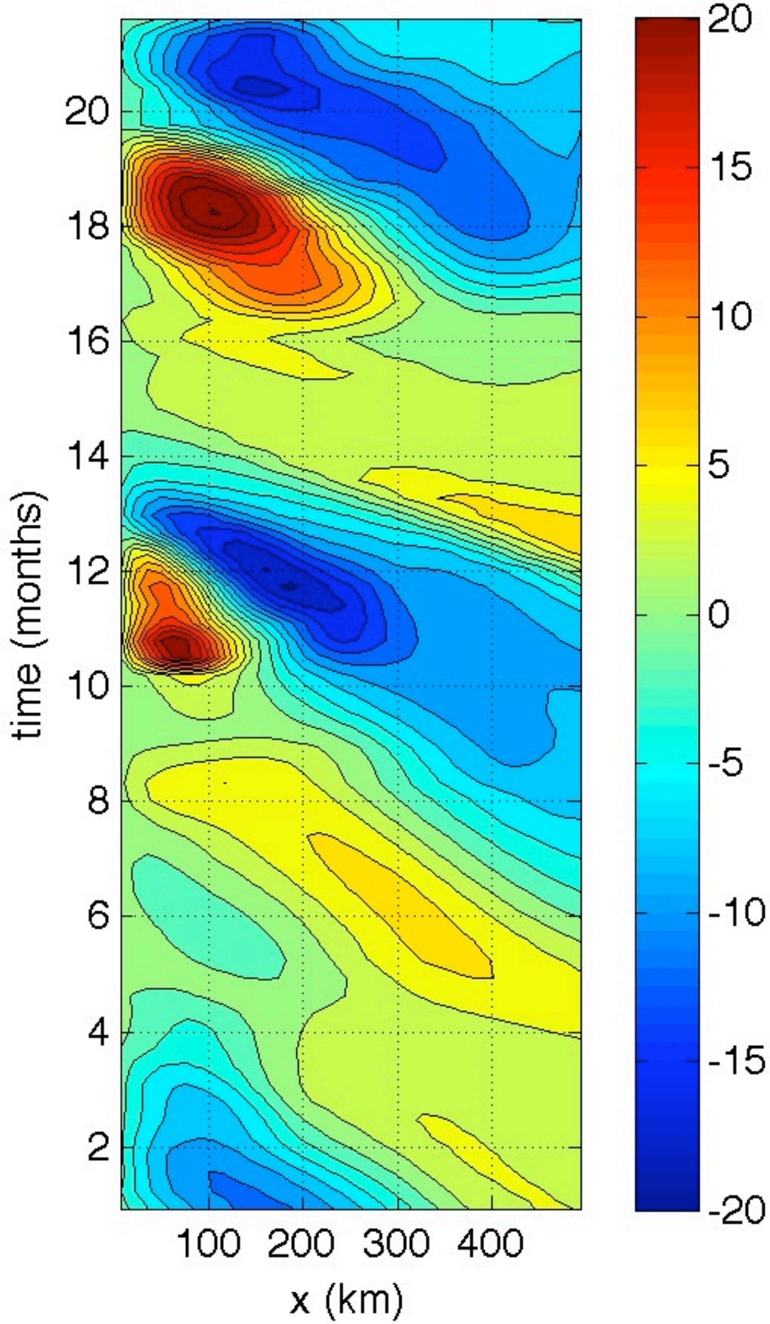


An eddy-filled ocean:



Variability in η near western boundary decays on approximately the same scale as seem in the altimetry and mooring data.





In a linear, reduced-gravity ocean:

$$1. \quad \frac{\partial u}{\partial t} - fv + g'\frac{\partial h'}{\partial x} = 0$$

$$2. \quad \frac{\partial v}{\partial t} + fu + g'\frac{\partial h'}{\partial y} = 0$$

$$3. \quad \frac{\partial h'}{\partial t} + H\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0$$

Low frequency limit ⇒ long and short Rossby wave solutions:

$$\frac{\omega}{k_l} = -\beta L_d^{\,2}$$

$$\omega k_s = -\beta$$

At the western boundary u=0 and layer thickness depends on balance between incoming long Rossby waves and reflected short Rossby waves:

$$h' = A_l(y)e^{i(k_l x - \omega t)} + A_s(y)e^{i(k_s x - \omega t)}$$

Substituting into 2 and noting $k_s >> k_l$ gives:

$$\frac{d}{dy}\left(\frac{A_l + A_s}{f}\right) \approx \frac{-\beta A_l}{f^2}$$

Assuming no wave disturbance on the boundary to the north:

$$A_l + A_s \sim \frac{\beta \Delta y}{f}\left|A_l\right|$$

Conclusions and implications for monitoring strategy

- Sea surface and dynamic height variability decline within roughly 100km of the Abaco shelf.

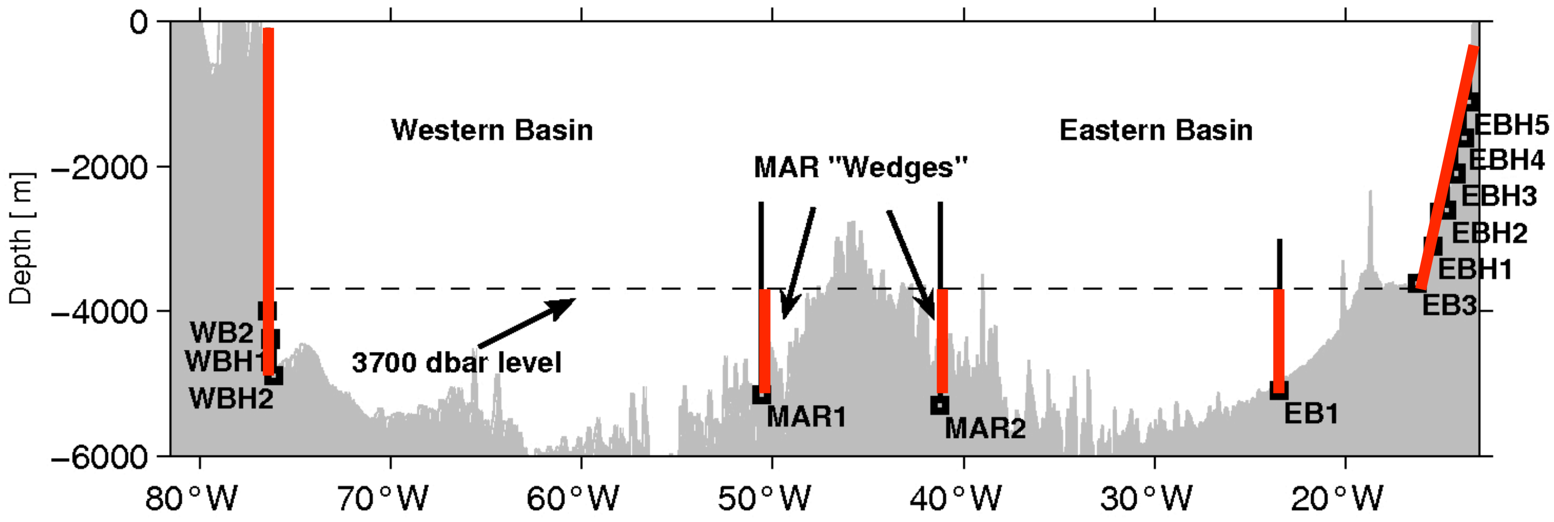
- Therefore, upper ocean transport integrated from boundary to boundary displays much lower variability than that integrated from any station close to <u>but not right at</u> the western boundary.
  $\Rightarrow$ To capture the variability of upper mid-ocean transport, measurements right at the boundary are required.

- The correlation between $\Delta\eta$ and transport decreases as the western boundary is approached due to changes in the vertical structure of the integrated flow.
  $\Rightarrow$ Sea surface height cannot be used to measure the time-variable strength of the upper mid-ocean transport.

- Simple model experiments imply that the reduction in η variability is due to the rapid propagation of pressure anomalies along the boundary as waves. Linear wave dynamics suggest that thermocline thickness (and η) anomalies on the western boundary are a factor $f / \beta\Delta y$ smaller than those in the basin interior.

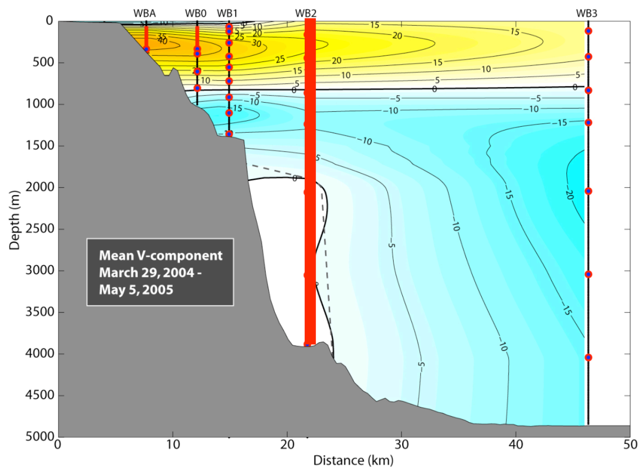
The eddy field at 26.5°N does not dominate MOC variability on interannual to decadal timescales, and does not pose a large signal-to-noise problem for RAPID measurements.

Mid-ocean transport - the hydrographic method on a grand scale!



Zonal density gradient gives basin-wide integrated internal transport (T_{INT}):

$$T_{INT}(z) = -\frac{g}{\rho f}\int_{Z_{REF}}^{Z}[\rho_{EAST}(z') - \rho_{WEST}(z')]\,dz$$



Current meter measurements give transport through western boundary wedge (T_{WBW}).

Johns et al. (2008)

Imposing zero net flow across the section:

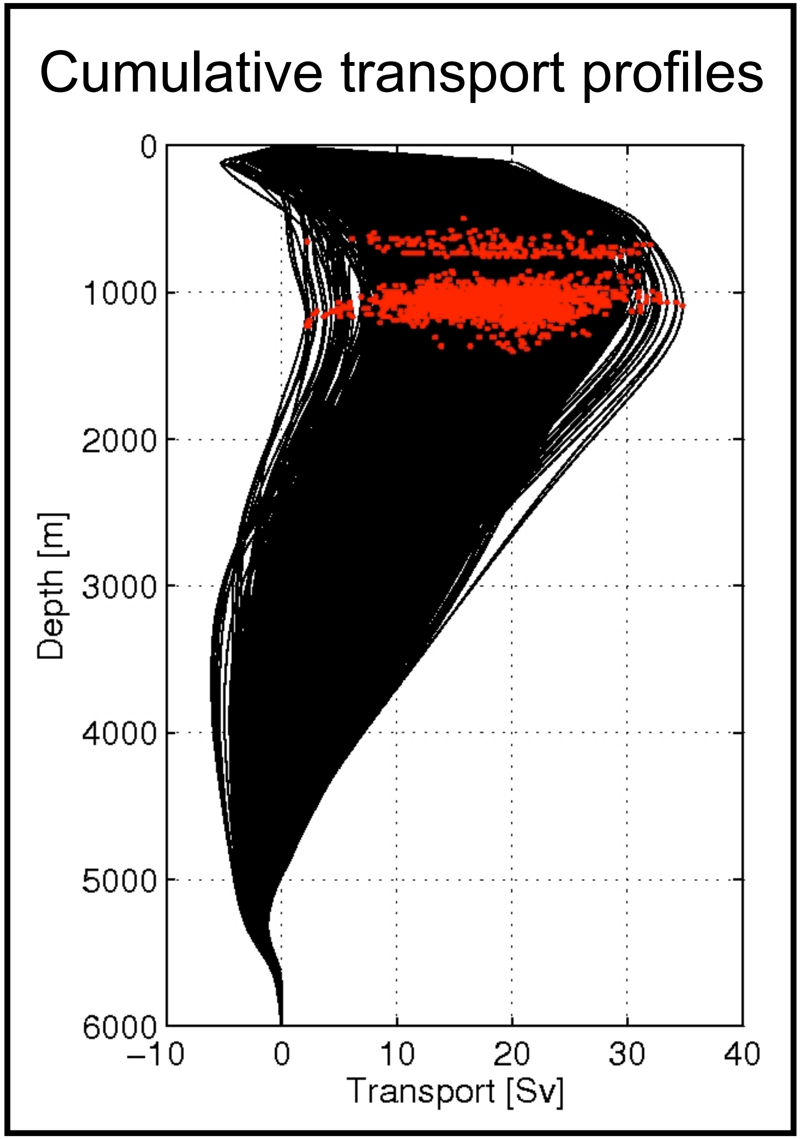
$$\int_{Z=-H}^{Z=0} \left[T_{EK}(z) + T_{GS}(z) + T_{MO}(z) \right] dz = 0$$

…allows a compensating transport (T_{COMP}) to be determined so that:

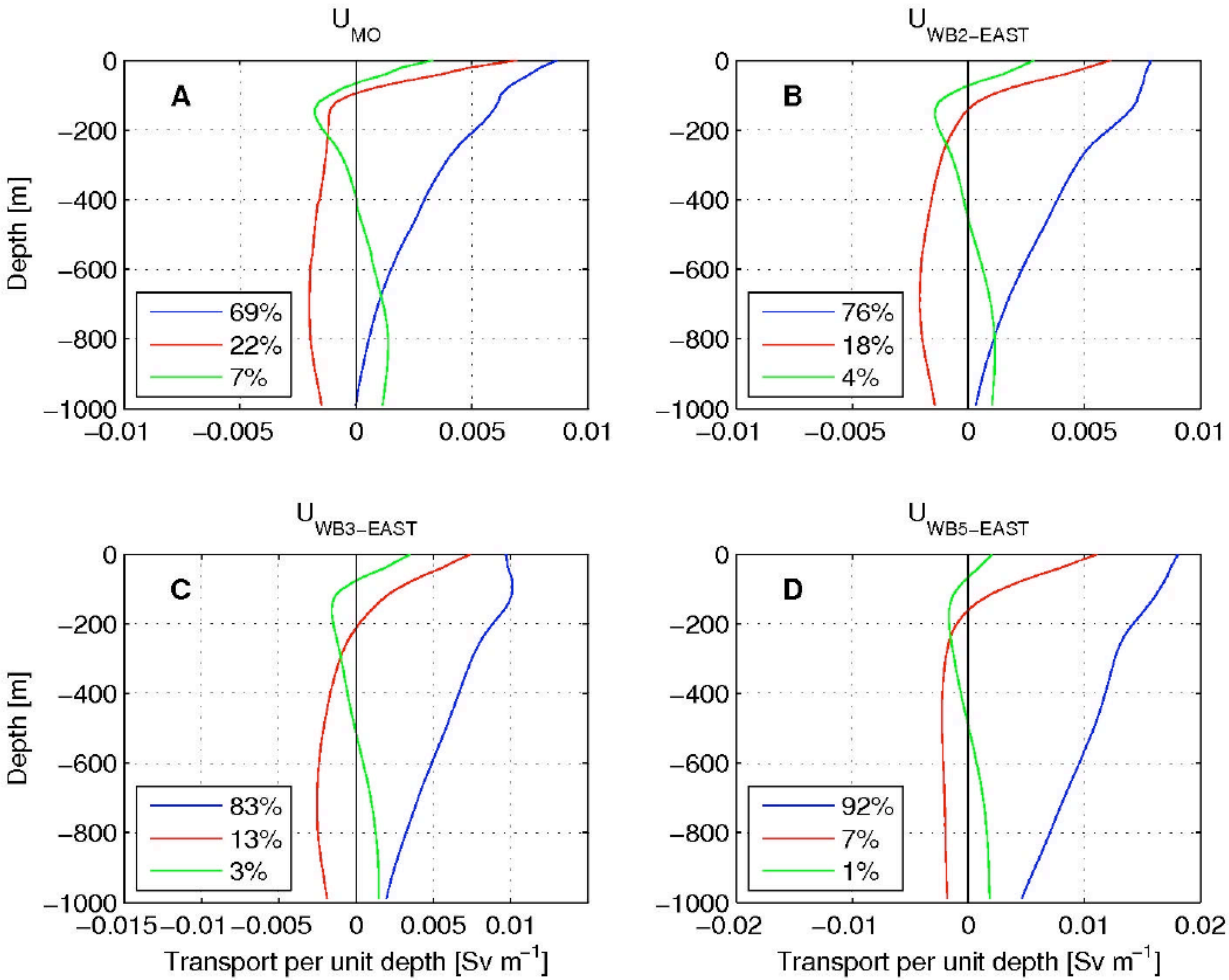
$$T_{MO}(z) = T_{WBW}(z) + T_{INT}(z) + T_{COMP}(z)$$

This gives total meridional transport as a function of depth:

$$T_{BASIN}(z) = T_{MO}(z) + T_{GS}(z) + T_{EK}(z)$$



…and hence the MOC, defined as the maximum northward transport.



First baroclinic mode explains less of the variance in upper ocean transport as western end-point of section gets closer to the boundary.