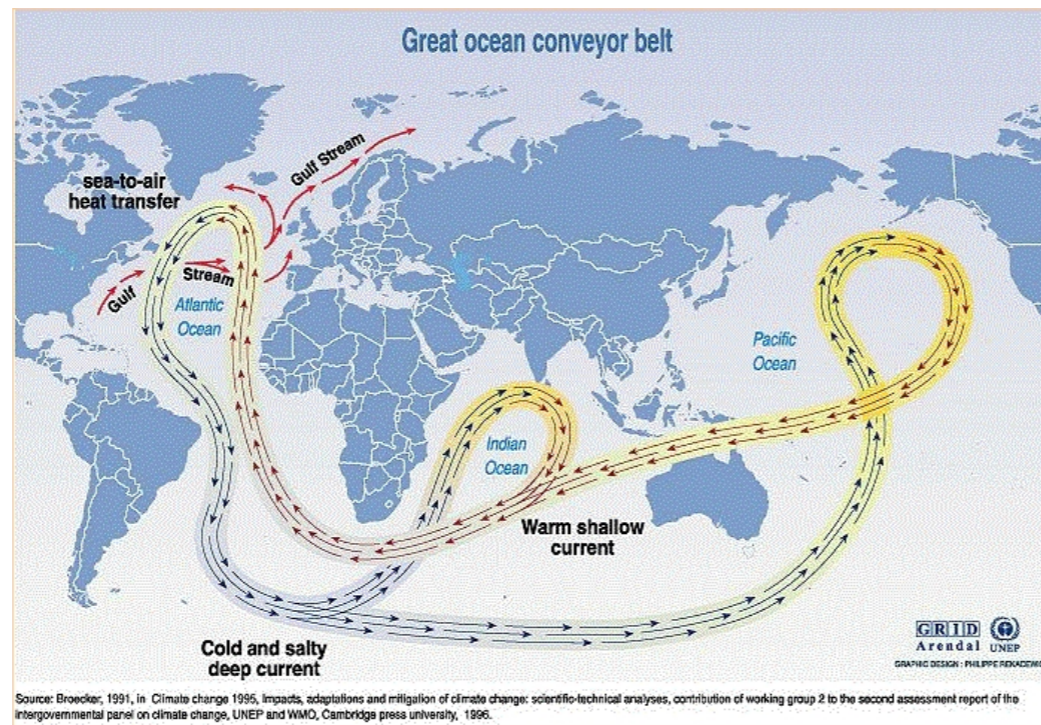


## Variations of the large-scale ocean circulation since 1950

Thierry Huck, Alain Colin de Verdière, Richard Schopp,  
Philippe Estrade, Thierry Begot

Laboratoire de Physique des Océans, Brest, FRANCE



[Broecker 1991]

## MOTIVATIONS

- variations of the large-scale ocean circulation may have climatic impact through poleward heat transport
  - what are these variations over the last decades in relation with natural variability, global warming, North Atlantic Oscillation, ...?
  - traditionnally investigated through sparse and synoptic hydrographic sections, and with realistic ocean models forced through surface fluxes (large uncertainty, relaxation to surface observations)
- ⇒ why not use directly insitu TS measurements: NODC CTD and XBT database, Argo profiling floats, ...?

## GOAL

Reconstruct large-scale ocean circulation (steady, geostrophic) from density field (T S) and surface wind stress

- diagnostic planetary geostrophic dynamics: Sverdrup balance + thermal wind!

## DATA

- **T+S**: pentadal temperature and salinity anomalies down to 3000 m (28 levels) from 1955-59 to 1994-98, superposed to mean climatology down to the bottom (WOD2004, Boyer et al. 2005)

also annual temperature anomalies 0-700 m (16 levels) from 1955-1945 to 2003 (WOD2004, Levitus et al. 2005, Ishii et al. 2003)

- **Wind stress**: ECMWF ERA40 1958–2001, 5-year averaged  
also NCEP 1948–now

## METHOD: SVERDRUP + THERMAL WIND

planetary-geostrophic diagnostic inviscid dynamics in spherical coordinates

$$-fv = -\frac{1}{a \cos \theta \rho_0} \partial_\phi P + \frac{\tau^x}{\rho_0 h_E} H(z + h_E)$$

$$fu = -\frac{1}{a \rho_0} \partial_\theta P + \frac{\tau^y}{\rho_0 h_E} H(z + h_E)$$

$$\partial_z P = -\rho g$$

$$\nabla \cdot \mathbf{u} = \frac{1}{a \cos \theta} \partial_\phi u + \frac{1}{a \cos \theta} \partial_\theta (\cos \theta v) + \partial_z w = 0$$

equation of state for seawater insitu density  $\rho(T, S, P)$

$T$  potential temperature,  $S$  salinity,  $P$  pressure,  $(u, v, w)$  velocity,  $\theta$  latitude,  $\phi$  longitude,  $z$  height,  $a$  Earth radius,  $H$  Heaviside step function,  $h_E$  surface Ekman layer thickness,  $f = 2\Omega \sin \theta$  Coriolis parameter and  $\beta$  its meridional gradient,  $(\tau^x, \tau^y)$  surface wind stress

NB wind stress is uniformly distributed in Ekman layer to avoid resolving explicitly this frictional boundary layer

Velocities are split in barotropic and baroclinic components:

$$\mathbf{u}(\phi, \theta, z) = \bar{\mathbf{u}}(\phi, \theta) + \mathbf{u}'(\phi, \theta, z)$$
$$\bar{\mathbf{u}}(\phi, \theta) = \frac{1}{h(\phi, \theta)} \int_{-h(\phi, \theta)}^0 \mathbf{u}(\phi, \theta, z) dz$$

⇒ Sverdrup balance for barotropic meridional velocity:

$$\beta \bar{v} = \frac{1}{h \rho_0 a \cos \theta} [\partial_\phi \tau^y - \partial_\theta (\tau^x \cos \theta)] - \frac{f w_B}{h}$$

**Fundamental Hypothesis: vertical velocities are assumed to cancel at the bottom** (this does not mean flat bottom!) ⇒  $w_B = 0$

- western boundary currents (westernmost grid point of each basin) are set to satisfy mass conservation at every latitude in each basin (⇒ some continuity problems with latitude when number of basins vary...)
- zonal barotropic transport is determined through mass conservation (through barotropic streamfunction)

Baroclinic component satisfies momentum equations, once subtracted the barotropic component, hence has a zero vertical integral:

$$\begin{aligned}
 -fv' &= -\frac{1}{a \cos \theta \rho_0} \partial_\phi P' + \frac{\tau^X}{\rho_0} G(z), \\
 fu' &= -\frac{1}{a \rho_0} \partial_\theta P' + \frac{\tau^Y}{\rho_0} G(z)
 \end{aligned}$$

where  $G(\phi, \theta, z) = H(z + h_E)/h_E - 1/h(\phi, \theta)$ , has a zero vertical integral, and baroclinic hydrostatic pressure  $P'$  verifies

$$\partial_z P' = -\rho g, \quad \int_{-h(\phi, \theta)}^0 P' dz = 0$$

NB this baroclinic component may not be valid in frictional western boundary currents, but we have tried to reduce deviations from geostrophy

## TRADITIONAL DIAGNOSTICS

- meridional overturning streamfunction

$$\psi(\theta, z) = \int_{\phi_W}^{\phi_E} \int_z^0 v dz a \cos \theta d\phi$$

- poleward heat transport

$$\text{PHT}(\theta) = \int_{\phi_W}^{\phi_E} \int_{-h(\phi, \theta)}^0 v T dz a \cos \theta d\phi$$

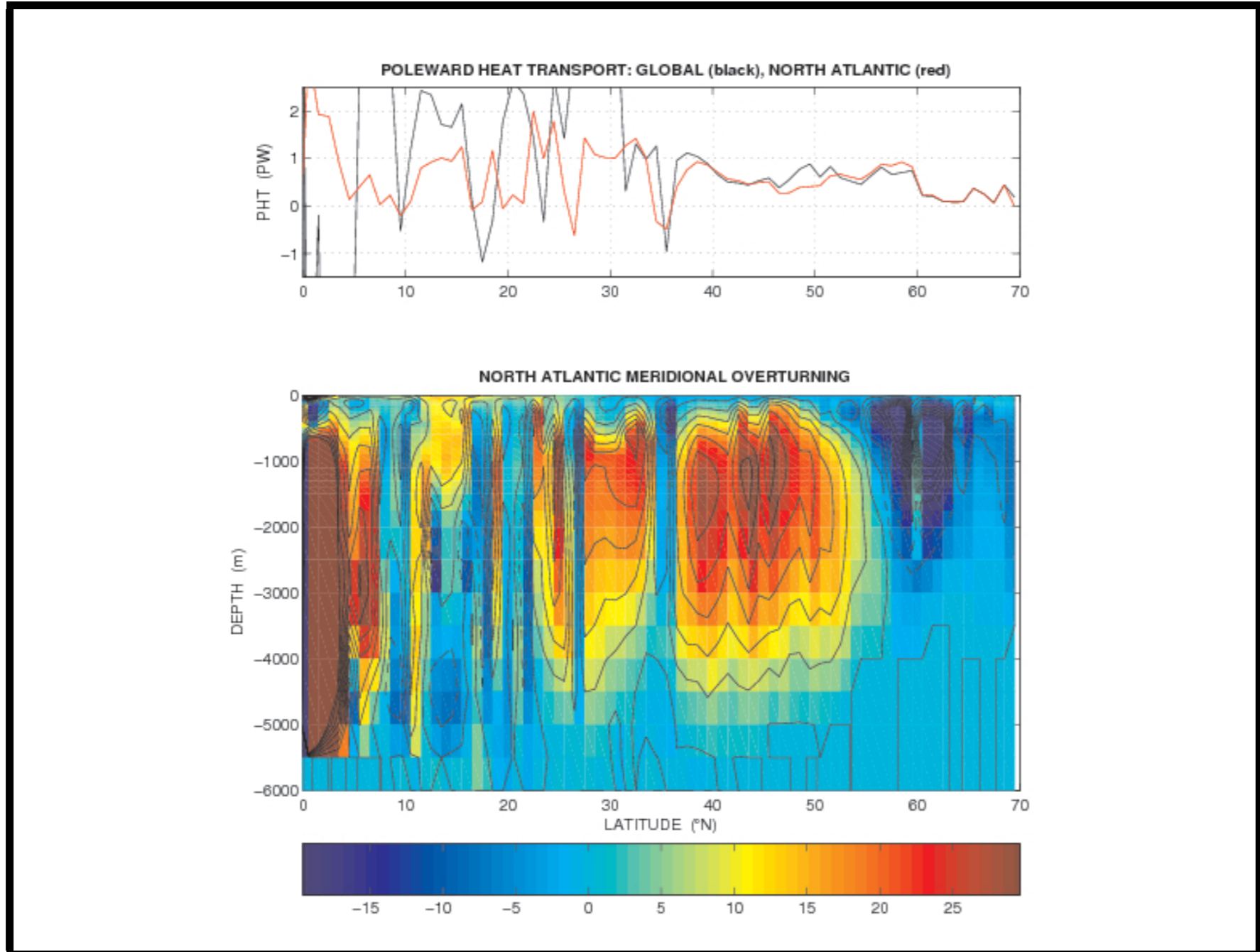
assuming poleward mass transport through the section cancels:

$$\int_{-h(\phi, \theta)}^0 v dz a \cos \theta d\phi \equiv 0$$

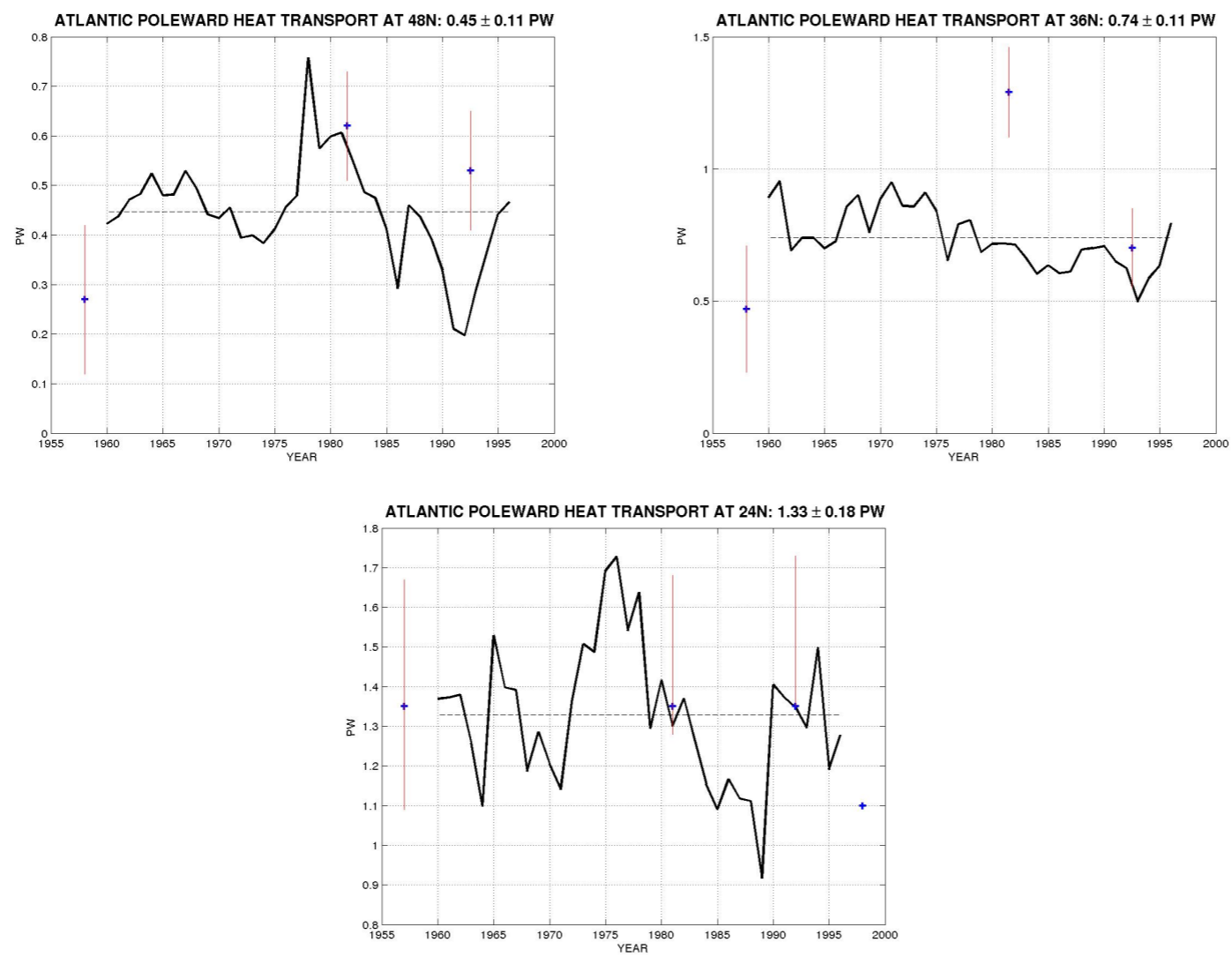
hence  $\psi(\theta, 0) = 0$  at the surface  $\Rightarrow \psi(\theta, -h(\phi, \theta)) = 0$  at the bottom

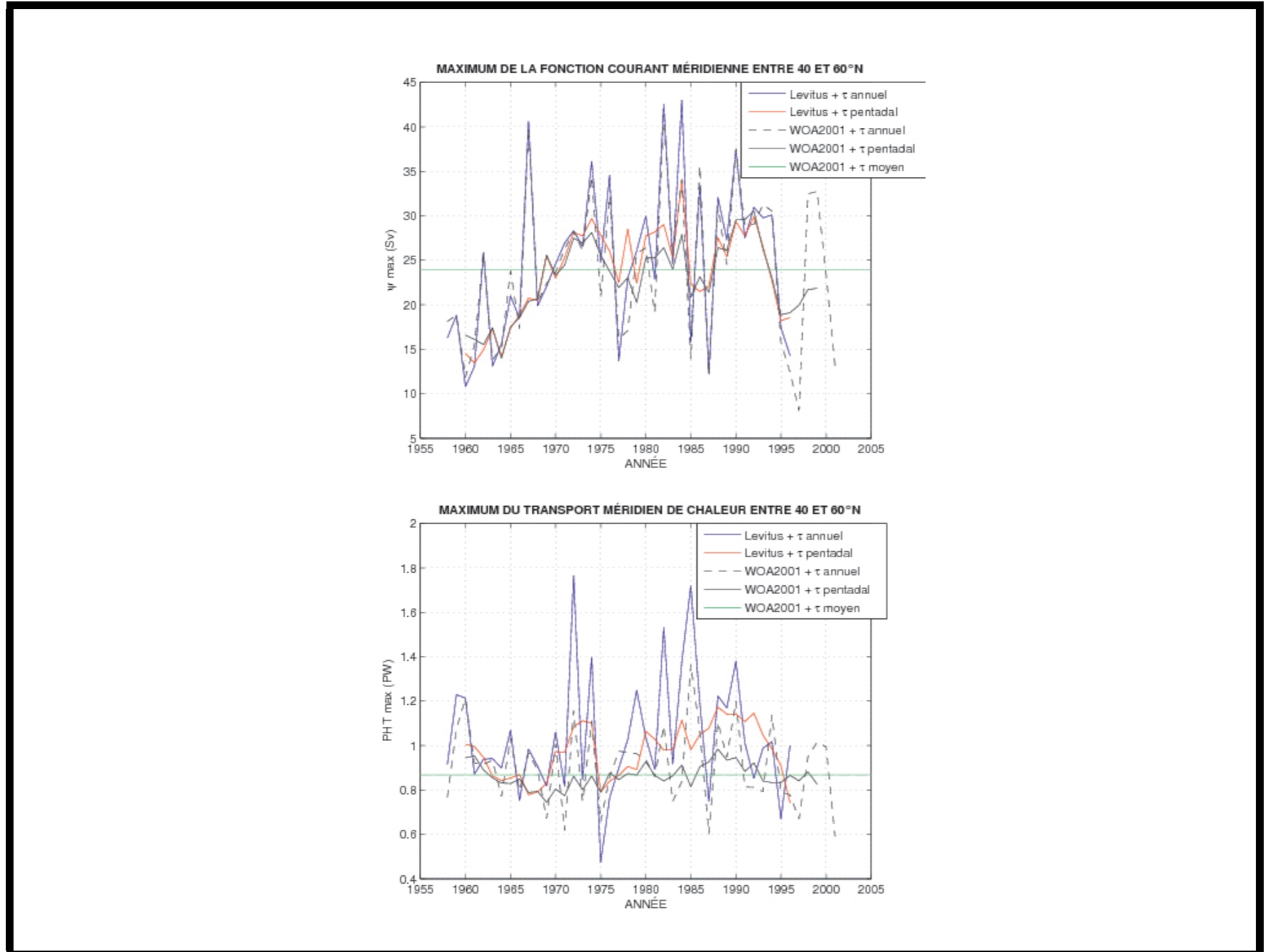
★ **It is verified on seasonal data that annual mean poleward heat transport is accurately diagnosed through annual mean data**

$$\overline{\text{PHT}(\theta)} = \overline{\int_{\phi_W}^{\phi_E} \int_{-h(\phi, \theta)}^0 v T dz a \cos \theta d\phi} \simeq \int_{\phi_W}^{\phi_E} \int_{-h(\phi, \theta)}^0 \overline{v T} dz a \cos \theta d\phi$$



## VARIATIONS OF POLEWARD HEAT TRANSPORT AT 24N, 36N, AND 48N VS. HYDROGRAPHIC SECTIONS





VARIATIONS OF NORTH ATLANTIC MERIDIONAL OVERTURNING  
AND POLEWARD HEAT TRANSPORT MAXIMUM 40N–60N

→ maximum meridional overturning varies roughly from 15 Sv in the 1960's to 30 Sv in the early 1990's

→ maximum poleward heat transport varies from 0.8 PW in 1965-70 to 1.2 PW in 1985-90

NB variations of overturning and heat transport maximum are not necessarily in phase!

→ 5-yr averaged winds considerably reduce the amplitude of interannual variations, especially for heat transport

→ low-frequency variations of overturning and heat transport are controlled by thermohaline variations, whereas interannual variations are of larger amplitude and controlled by the wind variations

## CONCLUSIONS

- annual mean T S data is sufficient to estimate annual mean poleward heat transport
- **low-frequency variations of overturning and heat transport are controlled by thermohaline variations**, whereas interannual variations are of larger amplitude and controlled by the wind variations (imprint of wind forcing in thermohaline structure?)
- **significant increase in maximum subpolar poleward heat transport from 1970-1975 to 1990-1995 (+25%), but weakening since then?**
- Levitus *et al.* (2005b) concluded "*Our results suggest that using 5-year or even 10-year running composites of ocean data to produce temperature anomaly fields is reasonable for estimating the interdecadal variability of ocean heat content*". Similar conclusions apply here to estimate the variations of the large-scale ocean circulation: large interannual wind variability (NAO)  $\Rightarrow$  large and noisy interannual variations of the general ocean circulation, whereas circulations diagnosed on 5-yr averaged T S and wind show more coherent variations  
[→ interpretation of the variations are underway...](#)

## DISCUSSION

- robustness should be estimated regarding climatologies, wind stress forcing, resolution of the analysis, and **uncertainties should be estimated**
- zero bottom vertical velocities hypothesis is important, and consequences are being verified in more sophisticated models
- **a better estimation of the barotropic flow will improve the mean values for poleward heat transport, but its variations are certainly well captured by the baroclinic flow**
- **variations of TS below 3000 m are not captured in this work** (Bryden et al. 2005) and may have significant impact on diagnosed quantities!
- products presently available on shelves (WOD2004) do not allow 'real time' analysis, for hydrographic cruise support for instance (Ovide repeated sections between Greenland and Portugal)  
→ tools are being built within the Coriolis/Argo french program (F. Gaillard and E. Autret, IFREMER)

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