Variations of the large-scale ocean circulation since 1950

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[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)
• 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

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

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

\[\partial_z P = -\rho g\]

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

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:

\[ u(\phi, \theta, z) = \bar{u}(\phi, \theta) + u'(\phi, \theta, z) \]

\[ \bar{u}(\phi, \theta) = \frac{1}{h(\phi, \theta)} \int_{-h(\phi, \theta)}^{0} u(\phi, \theta, z) \, dz \]

\[ \Rightarrow \text{Sverdrup balance for barotropic meridional velocity:} \]

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

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

- western boundary currents (westernmost grid point of each basin) are set to satisfy mass conservation at every latitude in each basin
  \( \Rightarrow \) 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:

\[-fv' = -\frac{1}{a \cos \theta \rho_0} \partial_\phi P' + \frac{x^X}{\rho_0} G(z),\]

\[fu' = -\frac{1}{a \rho_0} \partial_\theta P' + \frac{x^Y}{\rho_0} G(z),\]

where \(G(\phi, \theta, z) = H(z + h_E)/h_E - 1/\rho_0 G(\phi, \theta),\) has a zero vertical integral, and baroclinic hydrostatic pressure \(P'\) verifies

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

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
  \[ PHT(\theta) = \int_{\phi_W}^{\phi_E} \int_{-h(\phi, \theta)}^{0} v T \, dz \, a \cos \theta d\phi \]
  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
  \[ PHT(\theta) = \int_{\phi_W}^{\phi_E} \int_{-h(\phi, \theta)}^{0} v T \, dz \, a \cos \theta d\phi \simeq \int_{\phi_W}^{\phi_E} \int_{-h(\phi, \theta)}^{0} v T \, dz \, a \cos \theta d\phi \]
VARIATIONS OF POLEWARD HEAT TRANSPORT AT 24N, 36N AND 48N VS. HYDROGRAPHIC SECTIONS

ATLANTIC POLEWARD HEAT TRANSPORT AT 48N 1.18 ≤ 0.11 PM

ATLANTIC POLEWARD HEAT TRANSPORT AT 36N 0.78 ≤ 0.11 PM

ATLANTIC POLEWARD HEAT TRANSPORT AT 24N 0.68 ≤ 0.11 PM

Thierry Huck et al., OCEAN SCIENCES 2006
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) ⇒ 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)
RÉFÉRENCES


