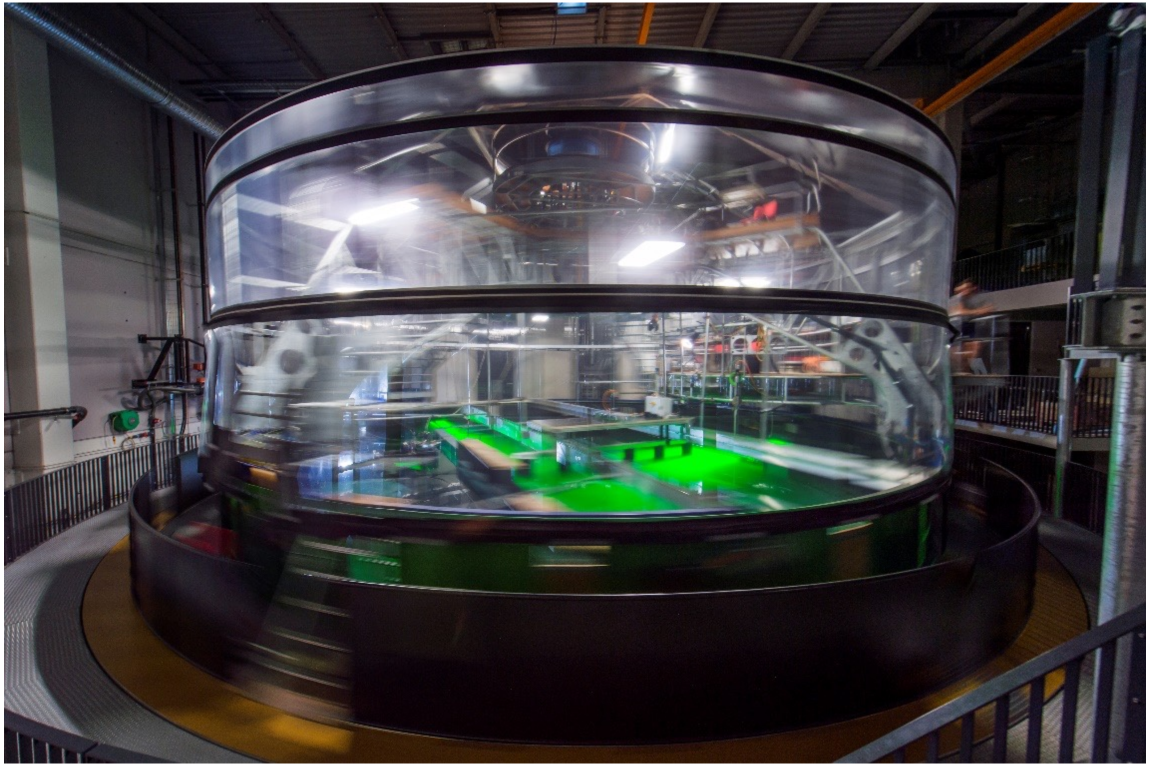
# TP-04 : Coriolis Platform

## The case of the intensified western boundary current (Gulf Stream)



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# Overview

- Some theoretical background
  - Wind-driven circulation
  - Sverdroup Theory
  - Stommel's theory: westerly currents
  - Munk transport
  - Atlantic Circulation and The Gulf Stream
- Coriolis Platform
  - Some history and technical aspects
  - Experimental Techniques
  - Gapwebs Project and dataset

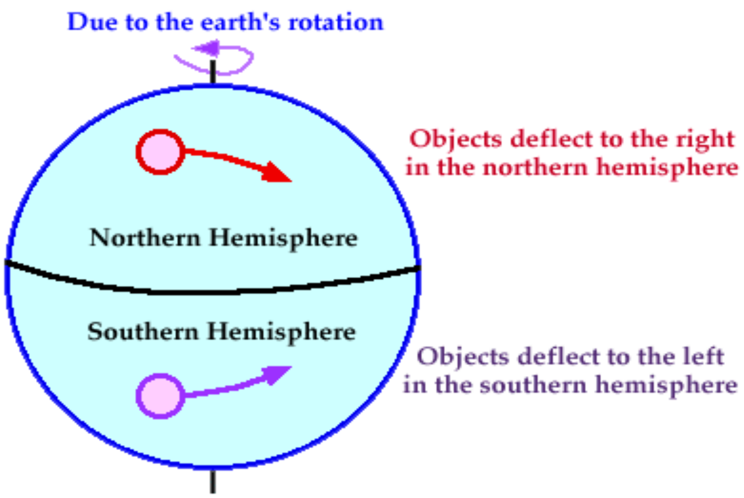
# Oceanic circulation

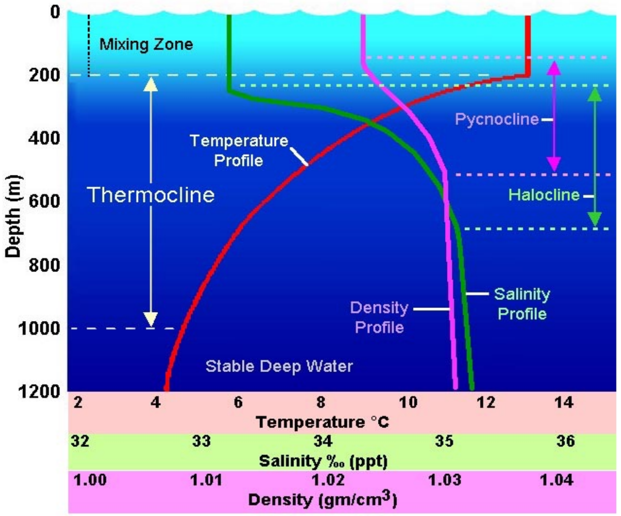
Oceanic circulation is driven by three main forcings:

1) by  the wind at the surface,

2) by the Earth's rotation, which makes the flow deviate to the right in the North emisphere and to the left in the Southern emisphere, and

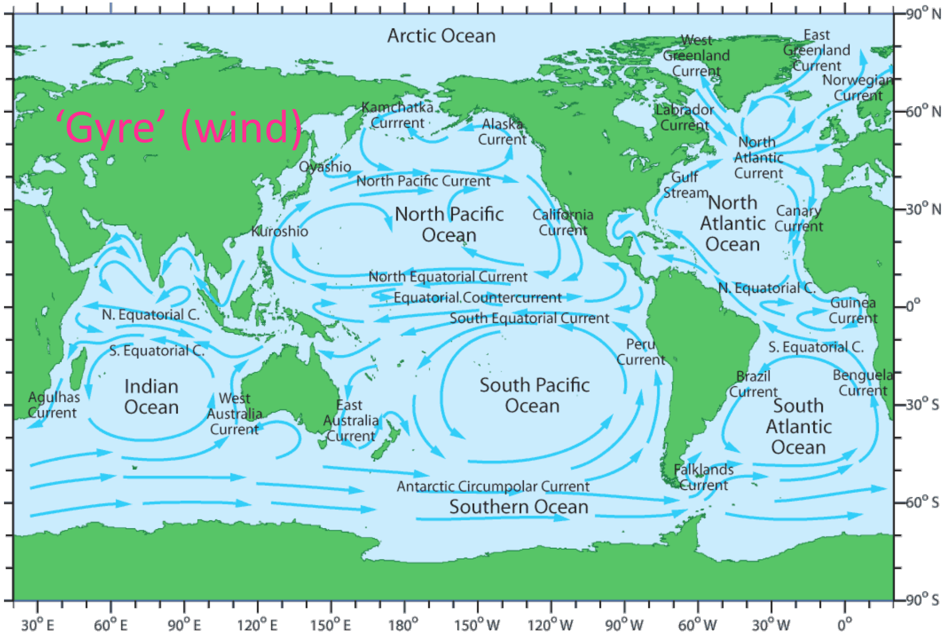
3) by the density variations induced by different temperatures or salinity.



# Oceanic circulation

The ocean circulation can be represented by two types of circulation: *the gyre circulation* on the surface (1km), mainly induced by wind (rapid and localized) and *the thermohaline circulation (conveyor belt)* (slow and deep) induced by density differences and **turbulent mixing.** This latter is important for climate.



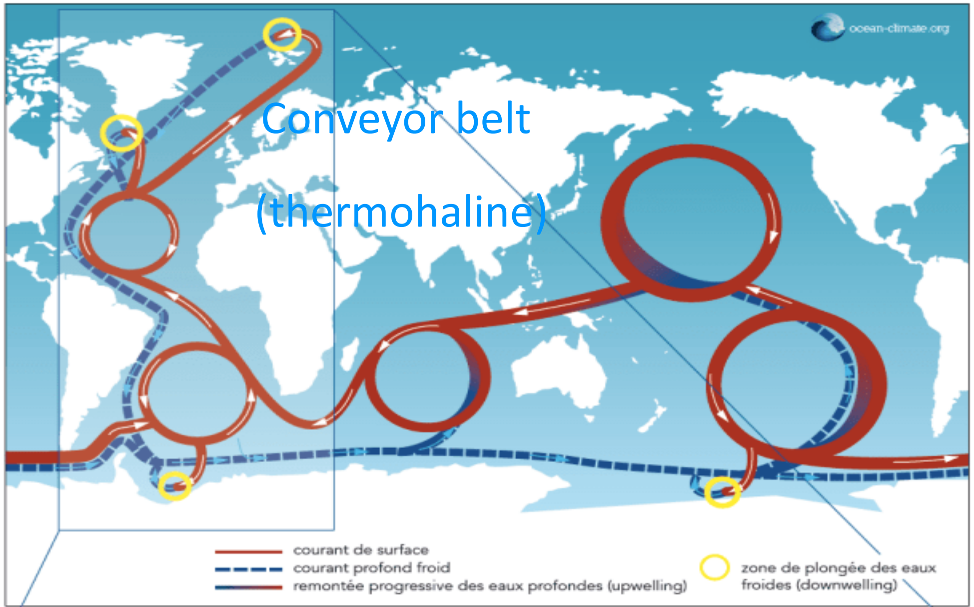



Schéma simplifié de la circulation océanique de grande échelle

# Gyre Wind driven circulation in the ocean

## What forces ocean currents?

As a first answer, we can say that the wind drives the ocean circulation.

Spanish navigators in the 16th century along the coast of Florida noticed a strong northward current that appeared to be unrelated to the wind.

How could this be explained? And, why were strong currents found off the eastern coasts and not off the western ones?

The answers to these questions can be found in a series of three seminal papers published from 1947 to 1951 that laid the foundations of modern ocean circulation theory.

# Wind driven circulation in the ocean

The foundations of modern ocean circulation theory :

- **Harald Sverdrup (1947)** showed that circulation in the first kilometre depth of the ocean is directly linked to the wind stress curl.

- **Henry Stommel (1948)** showed that the circulation of a 'gyre' ocean is asymmetrical because the Coriolis force varies with the latitude.

- **Walter Munk (1950)** added eddy viscosity and calculated the circulation of the surface layers of the Pacific.

# Sverdrup Theory

Mr Sverdrup assumed *stationary flow*, *horizontal turbulent friction zero* and *negligible molecular viscosity*. He assumed also that the baroclinic flow and the wind-driven circulation disappear at a certain 'no-motion' depth.

The horizontal components of the momentum equations are:

$$\frac{\partial p}{\partial x} = \quad f\,\rho\,v + \frac{\partial}{\partial z}\left(A_z \frac{\partial u}{\partial z}\right)$$

$$\frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial}{\partial z}\left(A_z \frac{\partial v}{\partial z}\right)$$

The right terms represent the wind stress tensor, *Tx* and *Ty*.

# Sverdrup Theory

$$\frac{\partial p}{\partial x} = f\,\rho\,v + \frac{\partial}{\partial z}\left(A_z \frac{\partial u}{\partial z}\right) \qquad \frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial}{\partial z}\left(A_z \frac{\partial v}{\partial z}\right)$$

Integrating these equations from the surface to a depth **- *D*** equal to or greater than the depth at which the horizontal pressure gradient becomes zero, we obtain:

$$\frac{\partial P}{\partial x} = \int_{-D}^{0} \frac{\partial p}{\partial x}\,dz, \qquad \frac{\partial P}{\partial y} = \int_{-D}^{0} \frac{\partial p}{\partial y}\,dz,$$

$$M_x \equiv \int_{-D}^{0} \rho\,u(z)\,dz, \qquad M_y \equiv \int_{-D}^{0} \rho\,v(z)\,dz,$$

where ***Mx*** , ***My*** are the mass transports in the wind forced layer extending to the ‘no-motion’ depth in the zonal (east-west) and meridional (north south) directions, respectively.

# Sverdrup Theory

The boundary condition at the sea surface is the wind stress and that at the depth -D is zero because the current tends to zero

$$\left(A_z \frac{\partial u}{\partial z}\right)_0 = T_x \qquad \left(A_z \frac{\partial u}{\partial z}\right)_{-D} = 0$$

$$\left(A_z \frac{\partial v}{\partial z}\right)_0 = T_y \qquad \left(A_z \frac{\partial v}{\partial z}\right)_{-D} = 0$$

where ***Tx*** , ***Ty*** are the components of the wind stress.

$$\frac{\partial P}{\partial x} = \int_{-D}^{0} \frac{\partial p}{\partial x}\, dz, \qquad \frac{\partial P}{\partial y} = \int_{-D}^{0} \frac{\partial p}{\partial y}\, dz,$$

$$M_x \equiv \int_{-D}^{0} \rho\, u(z)\, dz, \qquad M_y \equiv \int_{-D}^{0} \rho\, v(z)\, dz,$$

$$\frac{\partial p}{\partial x} = f\,\rho\, v + \frac{\partial}{\partial z}\left(A_z \frac{\partial u}{\partial z}\right)$$

$$\frac{\partial p}{\partial y} = -f\,\rho\, u + \frac{\partial}{\partial z}\left(A_z \frac{\partial v}{\partial z}\right)$$

$$\frac{\partial P}{\partial x} = f\, M_y + T_x$$

$$\frac{\partial P}{\partial y} = -f\, M_x + T_y$$

# Sverdrup Theory

Integrating the continuity equation, assuming the vertical velocity zero at the top and bottom boundaries

$$\frac{\partial M_x}{\partial x} + \frac{\partial M_y}{\partial y} = 0$$

$$\frac{\partial P}{\partial x} = f\,M_y + T_x$$
$$\frac{\partial P}{\partial y} = -f\,M_x + T_y$$

Differentiating, subtracting and using continuity

$$\beta\,M_y = \frac{\partial T_y}{\partial x} - \frac{\partial T_x}{\partial y}$$
$$\beta\,M_y = \mathrm{curl}_z(T)$$

$\boldsymbol{\beta = \partial f/\partial y}$ is the derivative of the Coriolis parameter with respect to latitude

$$\frac{\partial f}{\partial x} = 0$$

$$\beta \equiv \frac{\partial f}{\partial y} = \frac{2\,\Omega\cos\varphi}{R}$$

$$\frac{\partial^2 f}{\partial y^2} = -\frac{f}{R^2}$$

***$curl_z(T)$*** is the vertical component of the curl of the wind stress.

***f*** vaires with the latitude ; ***R*** is the Earth’s radius ; ***phi*** is the latitude

# Sverdrup Theory

$$\beta\, M_y = \text{curl}_z(T)$$

$\boldsymbol{\beta = \partial f / \partial y}$ is the derivative of the Coriolis parameter with respect to latitude

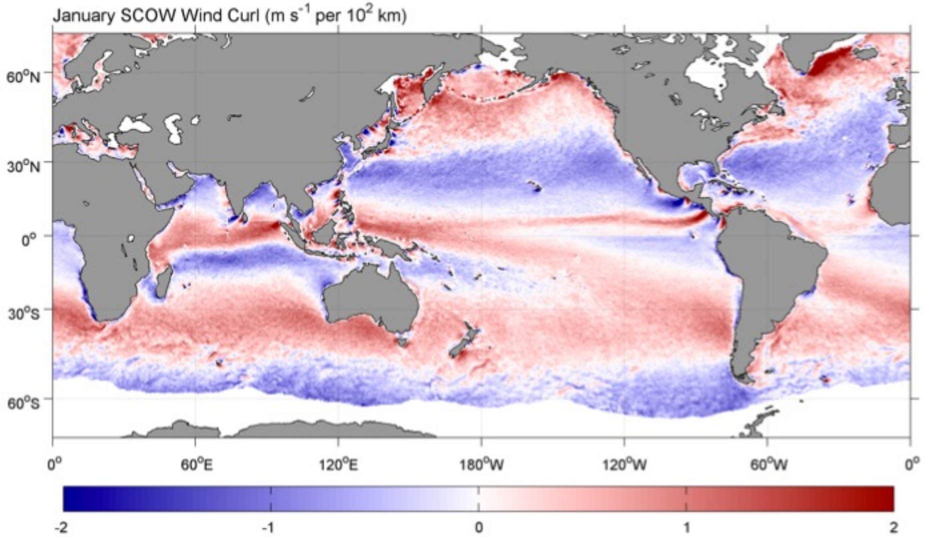
***$curl_z(T)$*** is the vertical component of the curl of the wind stress.

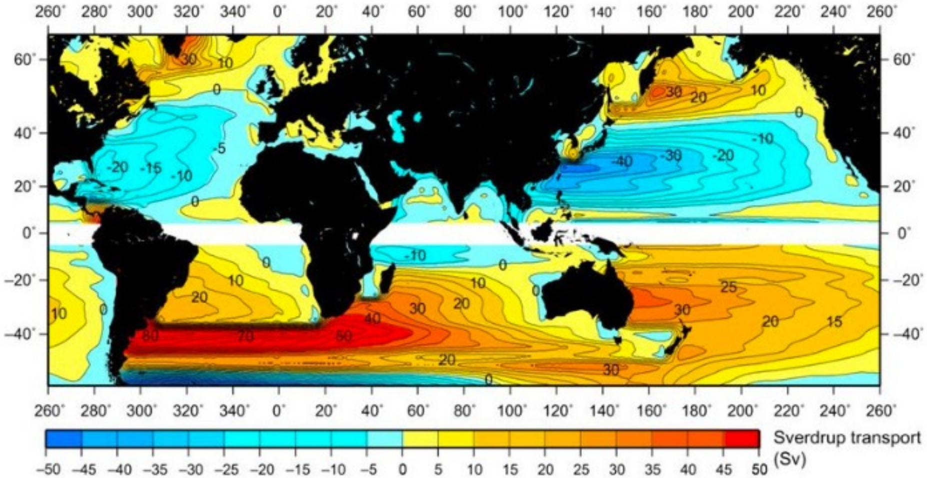
**The northward mass transport of wind-forced ocean currents is equal to the curl of the wind stress**

# Sverdrup Theory

$$\beta M_y = \text{curl}_z(T)$$

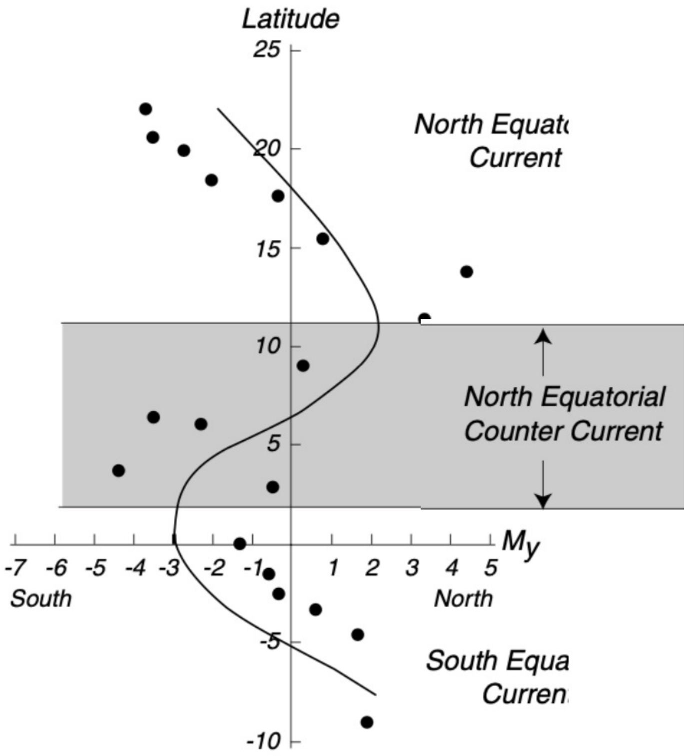
**The northward mass transport of wind-forced ocean currents is equal to the curl of the wind stress**




**Left:** Wind curl (figure from Risien and Chelton 2008). **Right:** Sverdrup circulation (figure from Talley, Descriptive oceanography)

# Sverdrup Theory




*The transport is in tonnes per second in a one-metre wide section extending from the sea surface to a depth of one kilometre. Note the difference between the My and Mx scales. From Reid (1948).*

To check his theory, Mr Sverdrup compared transports calculated from known winds in the tropical eastern Pacific with transports from hydrological data collected by Carnegie and Bushnell in October and November 1928, 1929, and 1939 between 22°N and 10°S along 80°W, 87°W, 108°W, and 109°W.

The comparison shows not only that transports can be accurately calculated from the wind, but also that the theory predicts upwind currents.

# Sverdrup Theory

## Limitations of the theory

1. Mr Sverdrup made simplifications: i) the flow within the ocean is geostrophic; ii) there is a uniform depth of 'no motion'; and iii) the Ekman transport is correct. But in reality it is not.
2. The solutions are limited to the eastern side of the oceans.
3. Only one boundary condition can be satisfied, i.e. no flow through the eastern edge. A more complete description of the flow requires more complete equations.
4. The solutions give no information on the vertical distribution of the current.
5. Results were based on data from two campaigns plus averaged wind data, assuming a steady state and setting the level of 'no motion' at 500 m. If another depth is chosen, the results are not as good.

# Stommel's Theory

At the same time that Mr Sverdrup was beginning to understand the circulation in the eastern Pacific, Mr Stommel was beginning to understand why intensified currents existed in the west.

To study the circulation of the North Atlantic, Mr Stommel (1948) used essentially the same equations used by Mr Sverdrup but added a simple bottom stress proportional to the velocity.

$$\left(A_z \frac{\partial u}{\partial z}\right)_0 = -T_x = -F\cos(x\,b/y) \qquad \left(A_z \frac{\partial u}{\partial z}\right)_D = -R\,u$$

$$\left(A_z \frac{\partial v}{\partial z}\right)_0 = -T_y = 0 \qquad \left(A_z \frac{\partial v}{\partial z}\right)_D = -R\,v$$

Where ***F*** and ***R*** are constants.

Recall Sverdrup theory

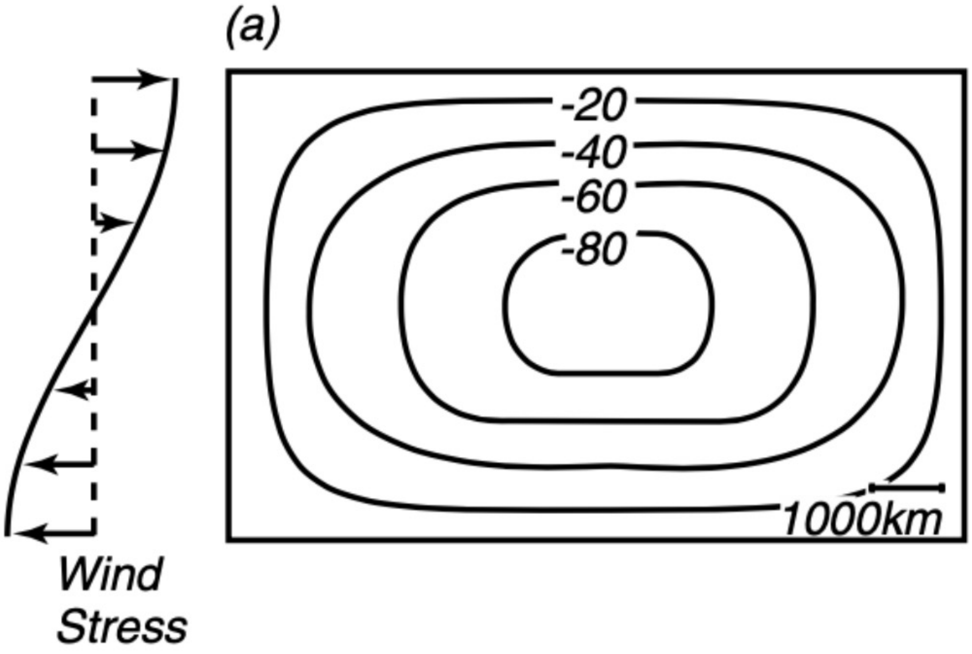
$$\left(A_z \frac{\partial u}{\partial z}\right)_0 = T_x \qquad \left(A_z \frac{\partial u}{\partial z}\right)_{-D} = 0$$

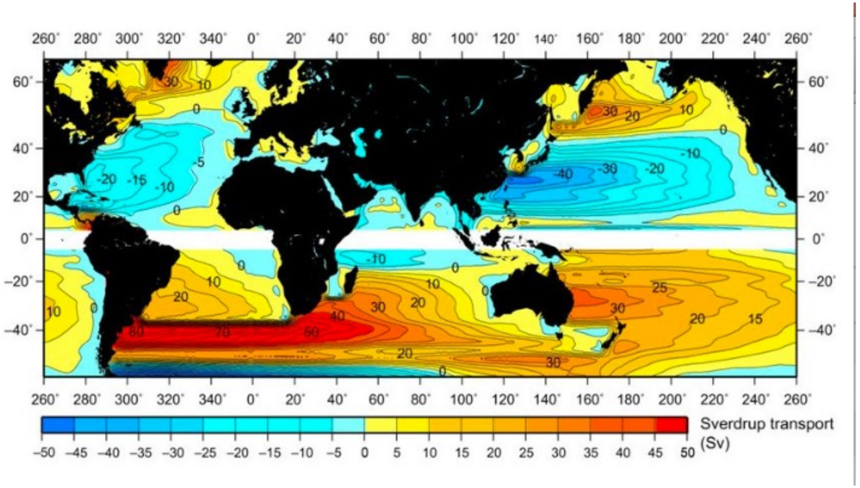
$$\left(A_z \frac{\partial v}{\partial z}\right)_0 = T_y \qquad \left(A_z \frac{\partial v}{\partial z}\right)_{-D} = 0$$

# Stommel's Theory

Mr Stommel calculated the stationary solutions for a flow inside a rectangular basin $0 \leq y \leq b$, $0 \leq x \leq \lambda$ with constant depth D filled with water of constant density. His first solution was for a non-rotating Earth.

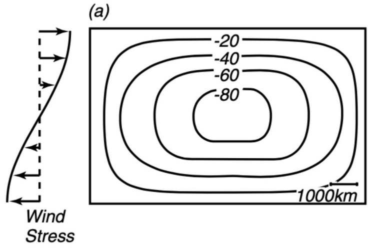
**This solution has a symmetrical flow with no western boundary current!**



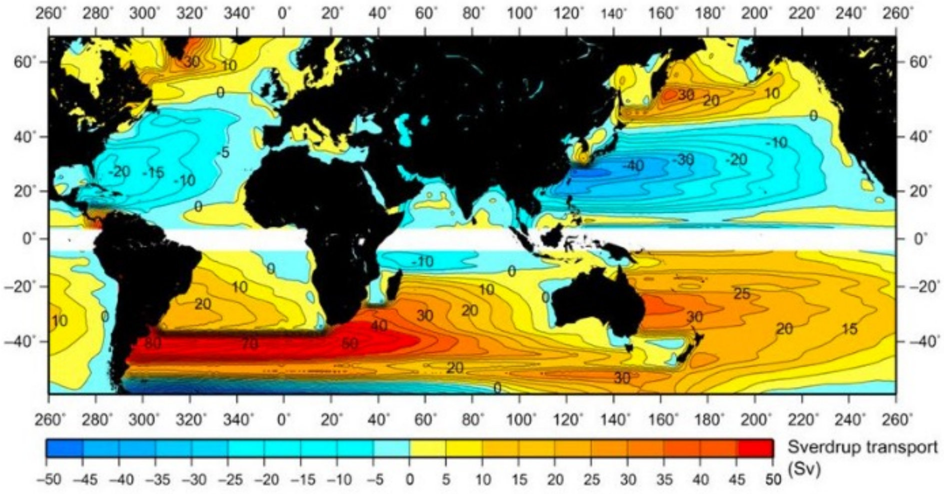



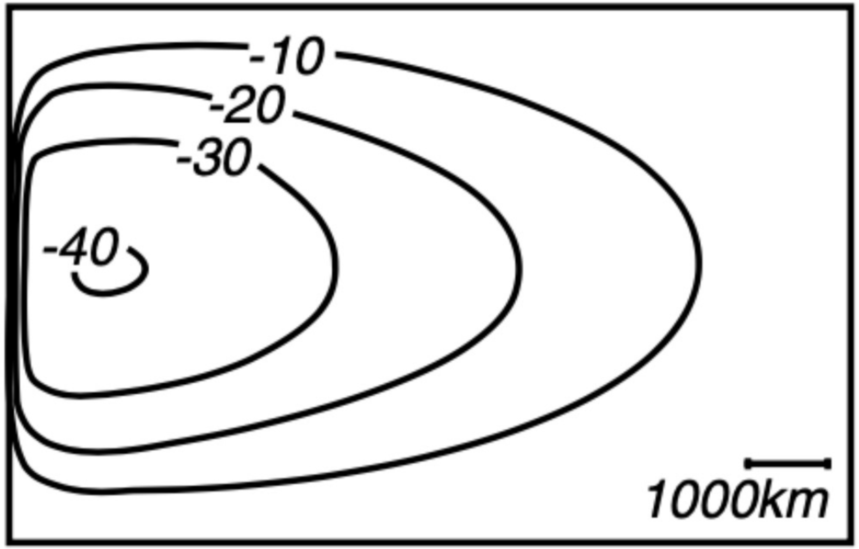
# Stommel's Theory

Then, Stommel assumed a constant rotation, which again led to a symmetrical solution with no western boundary current.

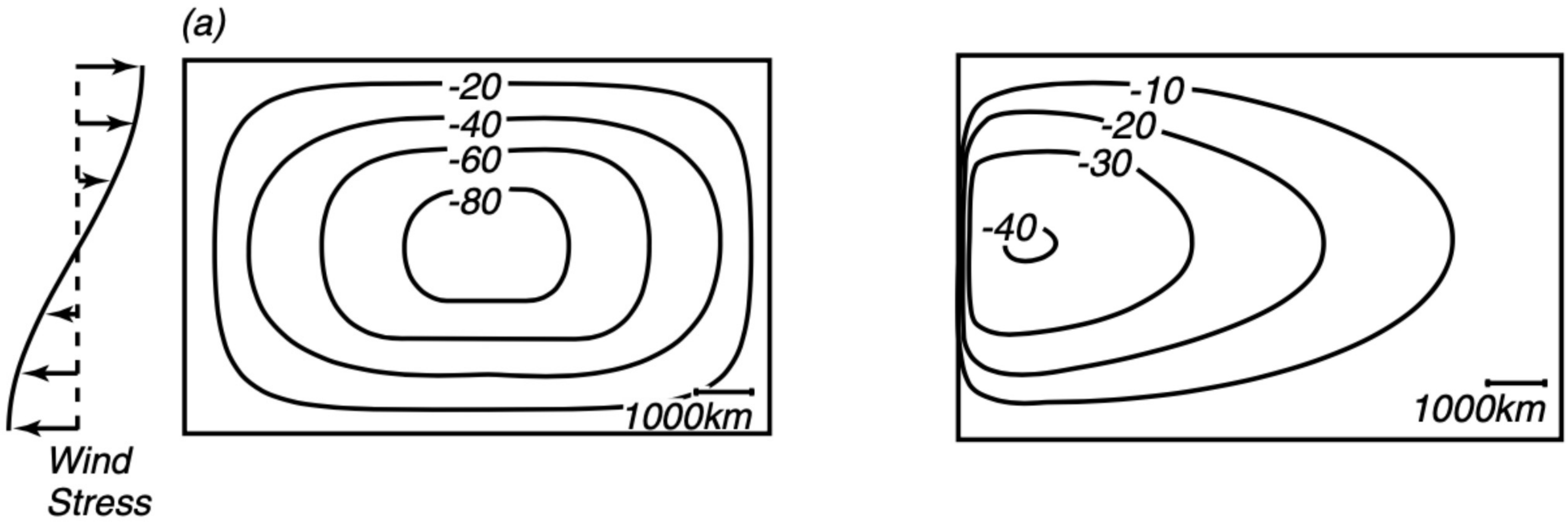



Finally, he assumed that the **Coriolis force varies with latitude**. This led to a solution with a western intensification.



# Stommel's Theory




Stream function for a basin flow calculated by Stommel (1948). **Left:** Flow for a non-rotating basin and one with a rotating constant rotation. **Right:** Flow with rotation varying linearly with y.

**The existence of the western boundary intensified currents is due to the variation of the Coriolis paramenter with latitude !**

# The concept of potential vorticity and its conservation

The sum of the relative ζ and the planetary f vorticity gives the $\boxed{\text{Absolute Vorticity} \equiv (\zeta + f)}$

We can obtain an equation for the absolute vorticity by simply manipulating the equations for frictionless motion (ocean interior).

$$\frac{Du}{Dt} - f\,v = -\frac{1}{\rho}\frac{\partial p}{\partial x}$$
$$\frac{Dv}{Dt} + f\,u = -\frac{1}{\rho}\frac{\partial p}{\partial y}$$

Expanding the total derivative, and subtracting ∂ /∂y of the former from ∂ /∂x of the latter, we obtain after a few algebraic steps

$$\boxed{\frac{D}{Dt}(\zeta + f) + (\zeta + f)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0}$$

Potential vorticity=

(relative+planetary)/depth

$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0$$

# The concept of potential vorticity and its conservation

The conservation of potential vorticity combines together the variations of depth, relative vorticity, and latitude. All three interact:

1. Changes in flow depth (H) cause changes in the relative vorticity. The concept is analogous with the way dancers on ice decrease their rotational speed by extending their arms and legs. This action increases their moment of inertia and decreases their rotational speed.

$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0$$

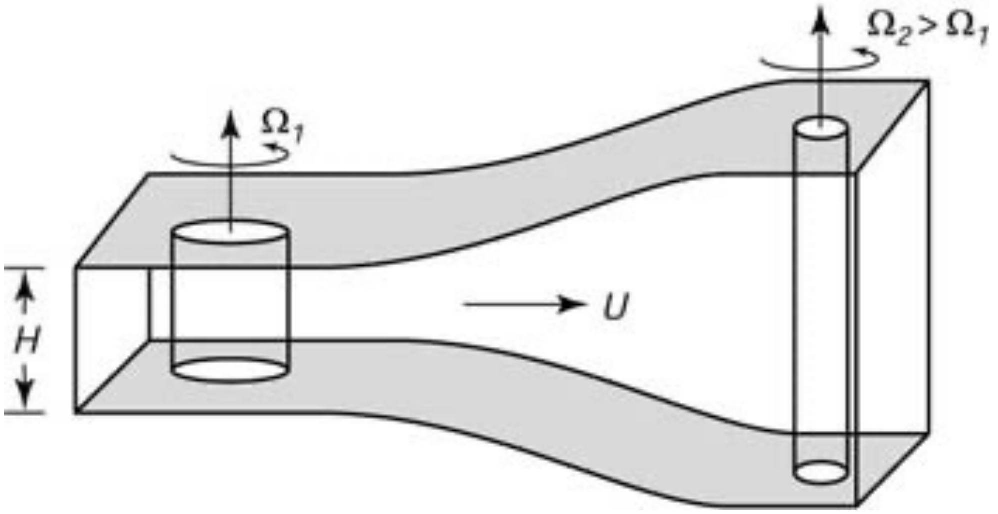



Diagram of the variation of relative vorticity in relation to variations in the height of a column of fluid. When a column of fluid moves from left to right, the vertical elongation reduces the moment of inertia of the column, causing it to rotate faster.

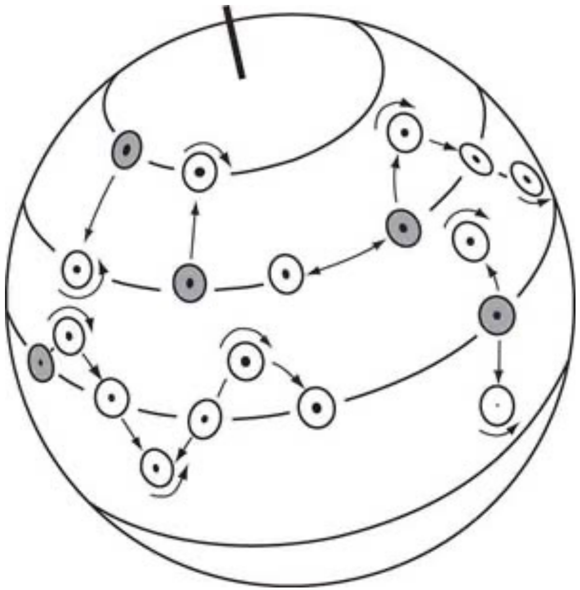
# The concept of potential vorticity and its conservation

The conservation of potential vorticity combines together the variations of depth, relative vorticity, and latitude. All three interact:

$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0$$

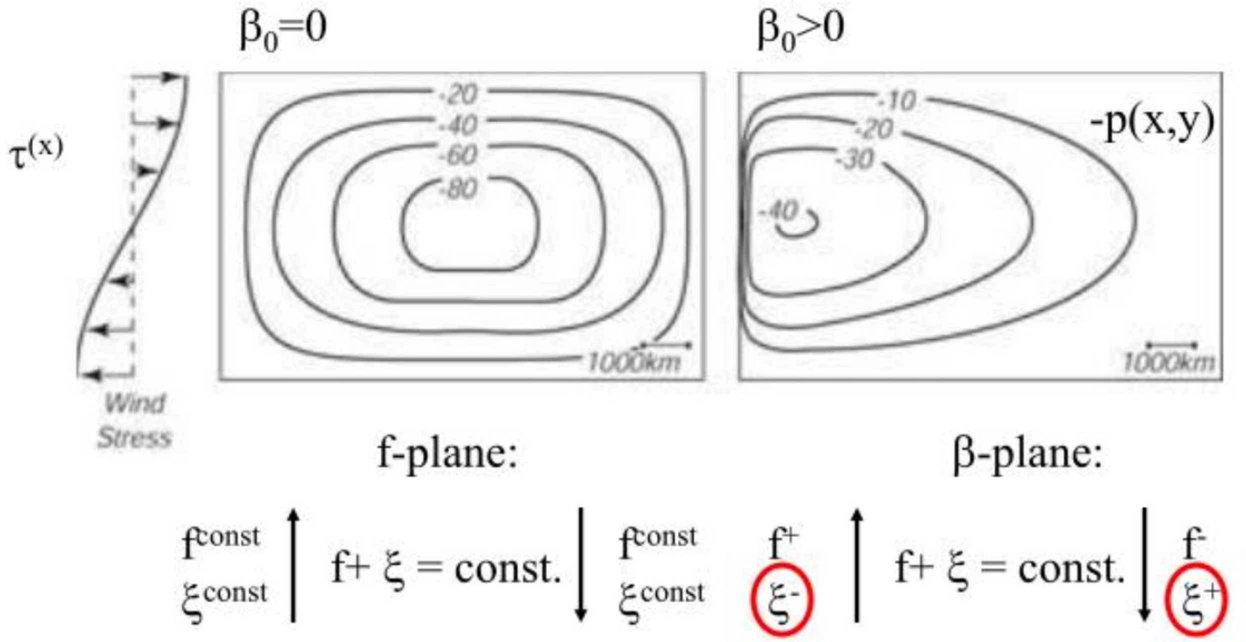
2. Variations in latitude require a corresponding variation in ζ. When a water column moves toward the equator, f diminishes and so ζ must increase.

Angular momentum has the tendency to be conserved when the water column changes latitude.



# Stommel's Theory

Stommel's results can also be explained in terms of vorticity; the wind produces a clockwise torque (vorticity), which must be balanced by an anti-clockwise torque at the western edge.




$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0$$

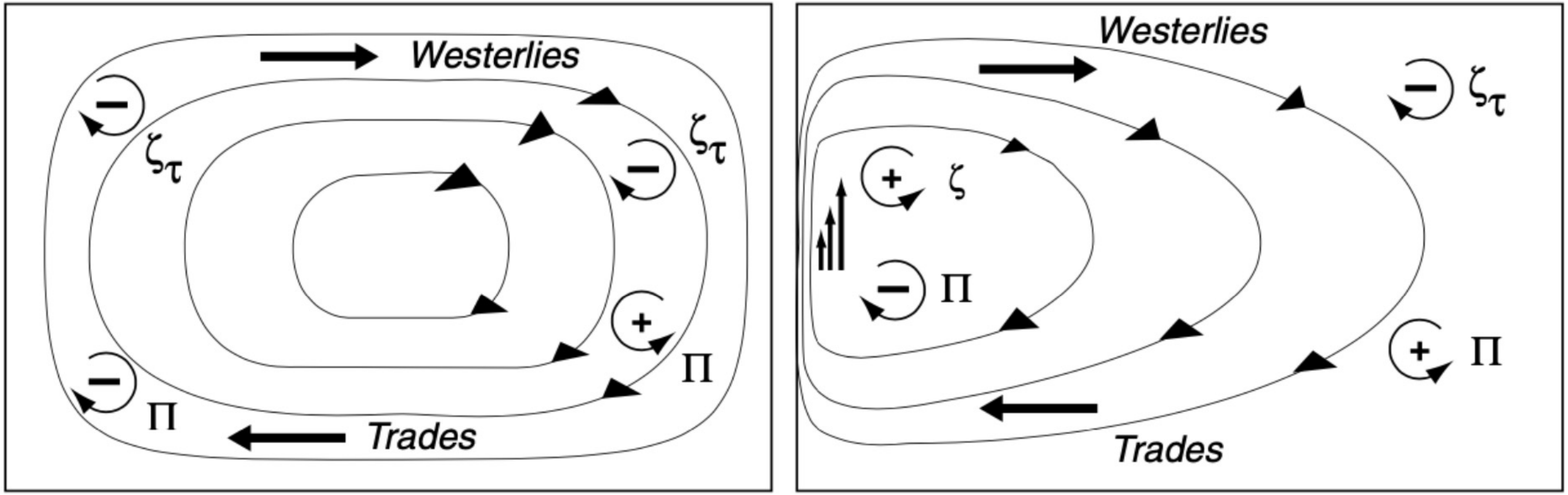
# Western boundary currents (WBC)

**The vorticity balance provides an alternative explanation for the existence of the gyre-scale westerly currents.** Consider the gyre-scale flow in the North Atlantic from 10°N to 50°N.

The wind blowing over the Atlantic adds negative vorticity. As water flows around the gyre, the gyre vorticity must remain almost constant, otherwise the flow should either increase or decrease. The input of negative vorticity must be balanced by a source of positive vorticity.

The source of positive vorticity must be the edge current: the flow forced by the wind is barocline, which is weak near the bottom, so bottom friction cannot transfer vorticity out of the ocean. So, we have to decide which boundary contributes. The flows tend to be zonal (along east-west), and east-west boundaries cannot solve the problem. In the east, potential vorticity is conserved: the input of negative relative vorticity is balanced by a decrease in potential vorticity when the flow turns south. Only in the west is the vorticity not balanced, so a source of positive vorticity must exist. Vorticity is provided by the shear current in the western boundary when the current flows against the coast producing a northward velocity which is zero at the coast. In this example, friction transfers angular momentum from the wind to the ocean and turbulent viscosity -friction- transfers angular momentum from the ocean to the land.

# Western boundary currents (WBC)




The balance of potential vorticity can clarify why western edge currents are necessary. **Left:** The wind vorticity input $\zeta_t$ balances the relative vorticity change $\zeta$ in the east when the flow moves south and $f$ decreases; but the two do not balance each other in the west where $\zeta$ must decrease when the flow moves north and $f$ increases. **Right:** The vorticity in the west is balanced by the relative vorticity $\zeta_b$ generated by the shear in the western flow.

$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0$$

# Munk's solution

The works of Mr Sverdrup and Mr Stommel suggest the dominant processes of the wind-forced circulation at basin level. Starting from previous theories, Mr Munk (1950) added turbulent lateral viscosity $A_H$, to obtain a solution of the circulation within an ocean basin.

Munk used Sverdrup's idea of vertically integrated mass transport, flowing over a deeper layer that does not move. This simplifies the mathematical problem and is more realistic. The ocean currents are concentrated in the first 1000 m depth of the ocean, are not barotropic and are depth dependent. To include friction, Munk used the lateral eddy viscosity with constant value $A_H = A_x = A_y$

$$\frac{\partial p}{\partial x} = f\,\rho\,v + \frac{\partial}{\partial z}\left(A_z \frac{\partial u}{\partial z}\right) + A_H \frac{\partial^2 u}{\partial x^2}$$

$$\frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial}{\partial z}\left(A_z \frac{\partial v}{\partial z}\right) + A_H \frac{\partial^2 v}{\partial y^2}$$

# Munk's solution

$$\frac{\partial p}{\partial x} = f\,\rho\,v + \frac{\partial}{\partial z}\left(A_z\frac{\partial u}{\partial z}\right) + A_H\frac{\partial^2 u}{\partial x^2}$$

$$\frac{\partial p}{\partial y} = -f\,\rho\,u + \frac{\partial}{\partial z}\left(A_z\frac{\partial v}{\partial z}\right) + A_H\frac{\partial^2 v}{\partial y^2}$$

Mr Munk integrated the equations from depth -D to the surface at $z = z_0$ assuming that currents cancel at depth -D and applied viscosity with constant $A_H$ at the bottom, at the sea surface and along the horizontal edge.

To simplify the equations, Mr Munk used the mass transport stream function and continued along Sverdrup's lines. He eliminated the pressure term by deriving the first equation in *x* and the second in *y* and obtained the mass transport equation:

$$\underbrace{A_H\nabla^4\Psi}_{\text{Friction}} \underbrace{- \beta\frac{\partial\Psi}{\partial x} = -\,\text{curl}_z}_{\text{Sverdrup Balance}}$$

$$\nabla^4 = \frac{\partial^4}{\partial x^4} + 2\frac{\partial^4}{\partial x^2\,\partial y^2} + \frac{\partial^4}{\partial y^4}$$

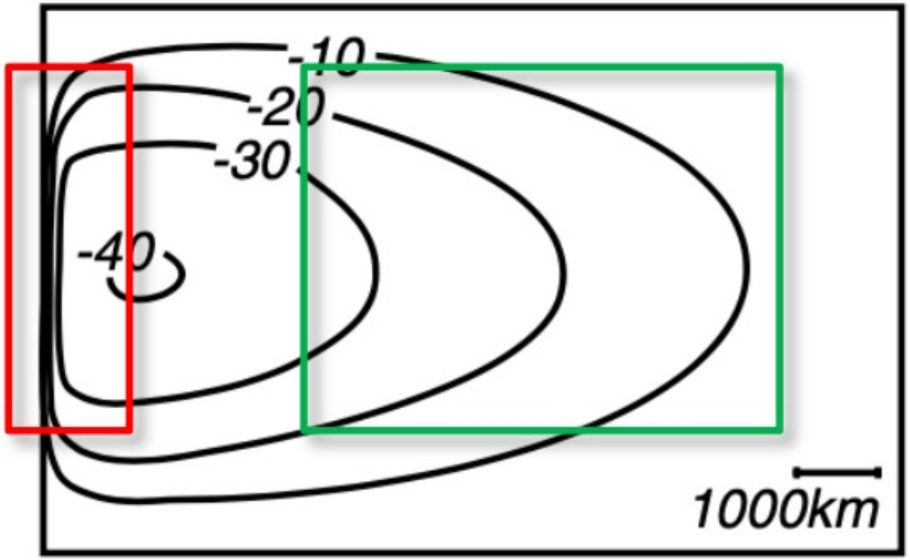
(Bi-harmonic operator)

# Munk's solution

$$\underbrace{A_H \nabla^4 \Psi}_{\text{Friction}} - \underbrace{\beta \frac{\partial \Psi}{\partial x} = -\,\mathrm{curl}_z}_{\text{Sverdrup Balance}}$$

$$\nabla^4 = \frac{\partial^4}{\partial x^4} + 2\frac{\partial^4}{\partial x^2\,\partial y^2} + \frac{\partial^4}{\partial y^4}$$ (Bi-harmonic operator)

The equation is the same as that obtained by Sverdrup, but with the added lateral friction term $A_H$.
The friction term is important near the **lateral boundary** where the horizontal derivatives of the velocity field are large and is small in the **oceanic interior**. Thus, in the interior, the balance of forces is the same as in *Sverdrup*'s solution.



$$\underbrace{A_H \nabla^4 \Psi}_{\text{Friction}} - \underbrace{\beta \frac{\partial \Psi}{\partial x} = -\,\mathrm{curl}_z}_{\text{Sverdrup Balance}}$$

$$\nabla^4 = \frac{\partial^4}{\partial x^4} + 2\,\frac{\partial^4}{\partial x^2\,\partial y^2} + \frac{\partial^4}{\partial y^4}$$

# Munk's solution

Considering that near the boundary one can neglect the wind forcing and that it is the meridional velocity component $v_B(x)$ that changes faster in the zonal x-direction, the dominant balance reduces to

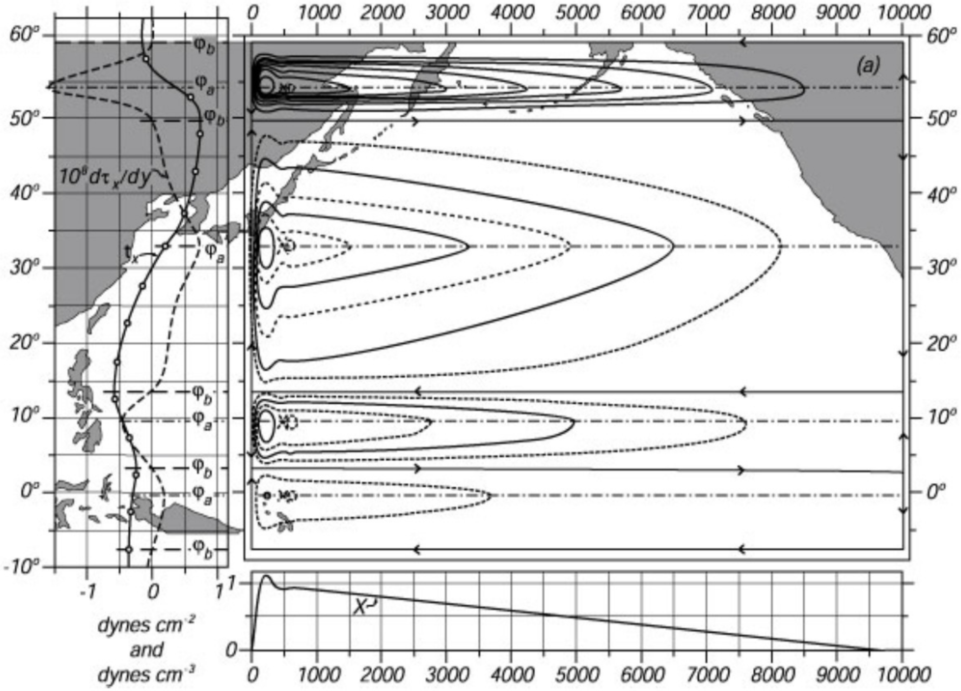
$$\beta v_B = A_H \partial_{xxx} v_B$$

Which admits the following solution for the meridional velocity $v_B$ along x

$$v_B = \exp - x/r* \left(C_1 \cos(-x/r*) + C_2 \sin(-x/r*)\right), \quad r^{\star} = (A_H/\beta)^{1/3}/(2\sqrt{3})$$

# Munk's solution

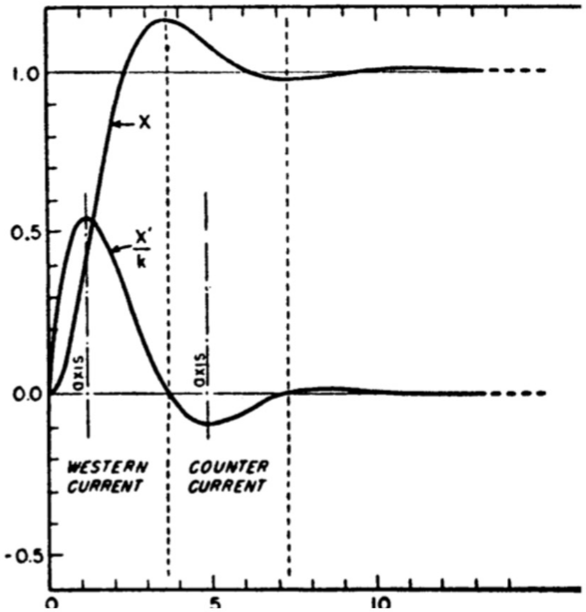
Munk's solution shows the dominant features of the gyre-scale circulation of an ocean basin. The circulation is similar to that of Sverdrup in the eastern parts of the basin but there is a strong westerly edge current in the west.

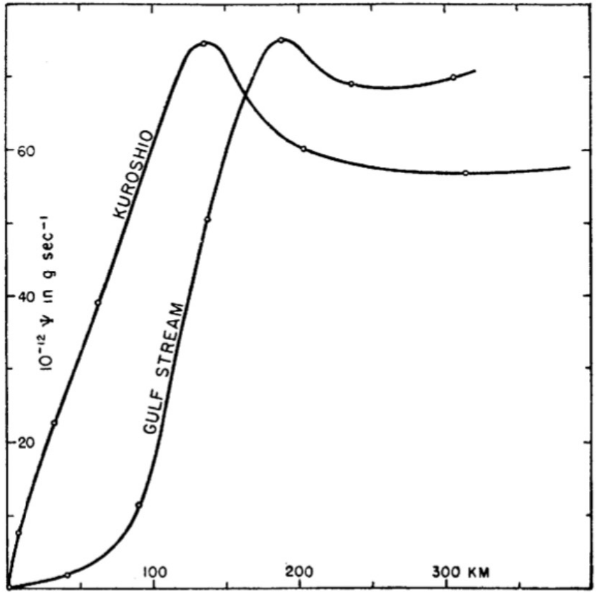


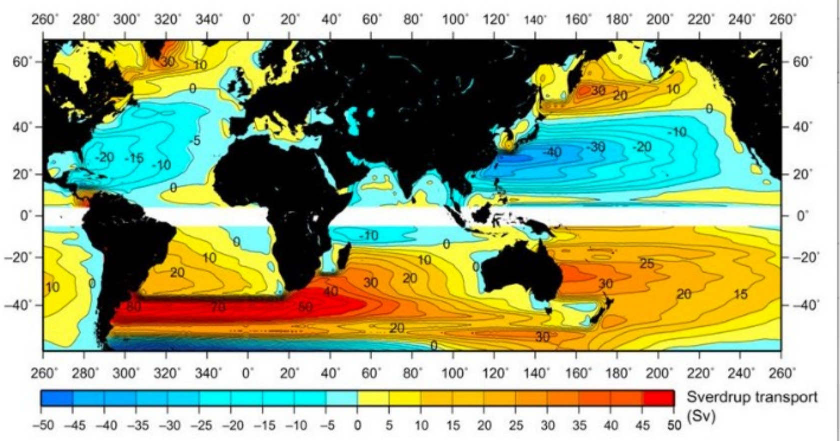

**Top Left**: The annual mean wind stress $T_x(y)$ over the Pacific and the wind stress rotor. $j_b$ are the north and south edges of the gyre, where $M_y$ = 0 and $curl_t$=0. $j_0$ is the centre of the gyre. **Top Right:** The mass transport current function for a rectangular basin calculated by Munk (1950) using the observed wind stress over the Pacific. The interval between isolines is 10Sv. The total transport between the coast and each point (x, y) is $M_y$(x, y). The transport in the narrow northern section is strongly exaggerated.

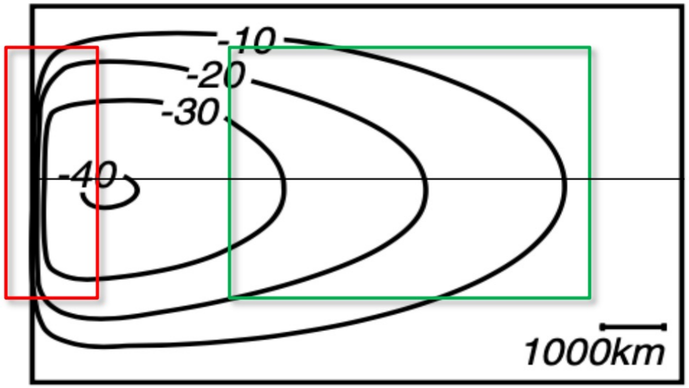
**Bottom Right:** North-south component of mass transport. From Munk (1950).

# Munk's solution



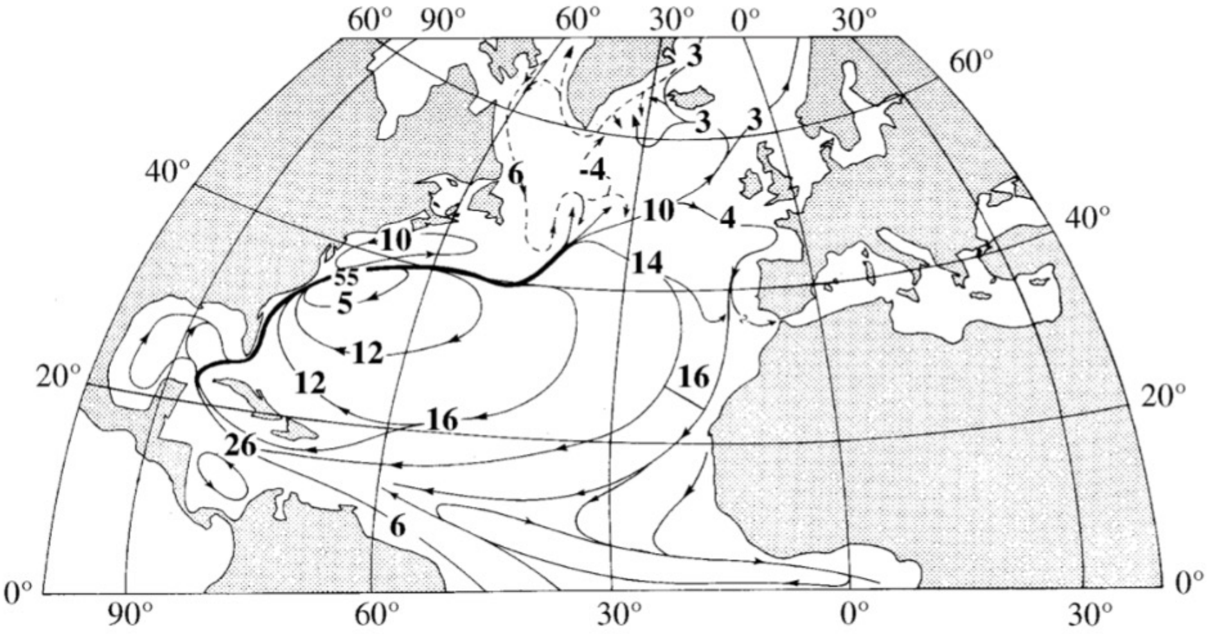
# Munk's solution

The transport in the western edge currents is independent of the $A_H$, but depends on the basin width, the wind stress curl, and **beta**. Using the best available estimate of wind stress, Munk calculated that the Gulf Stream should have a transport of 36 Sv and that for the Kuroshio it was 39 Sv. The values are about half of the measured values.
Using $A_H$ = 5 × 103 m2/s results in an edge current about 225 km thick with a shape similar to the flow observed in the Gulf Stream and Kuroshio.

This is in good agreement, considering that the wind stress was not well known.

The theories of Sverdrup, Munk, and Stommel describe a very simplified flow, but the ocean is much more complicated.
Look at the Atlantic Ocean, for example.

# The Atlantic Ocean

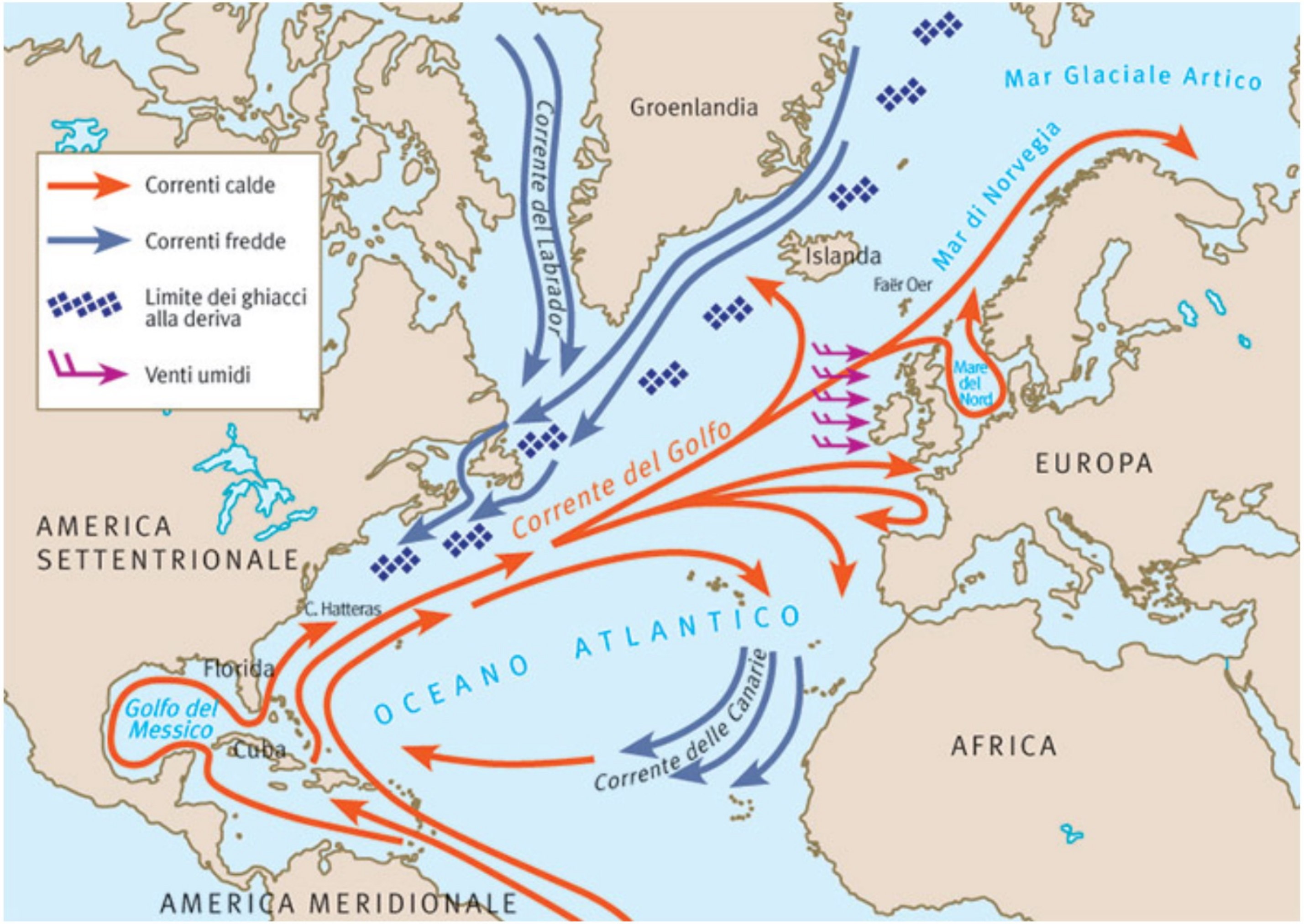



Mid-latitude processes in the Atlantic are similar to those in other oceans, the Gulf Stream is the westward intensifying current in the Atlantic Ocean.

A large mid-latitude, basin-scale 'gyre' as we expect it to be described by Sverdrup's theory is present.

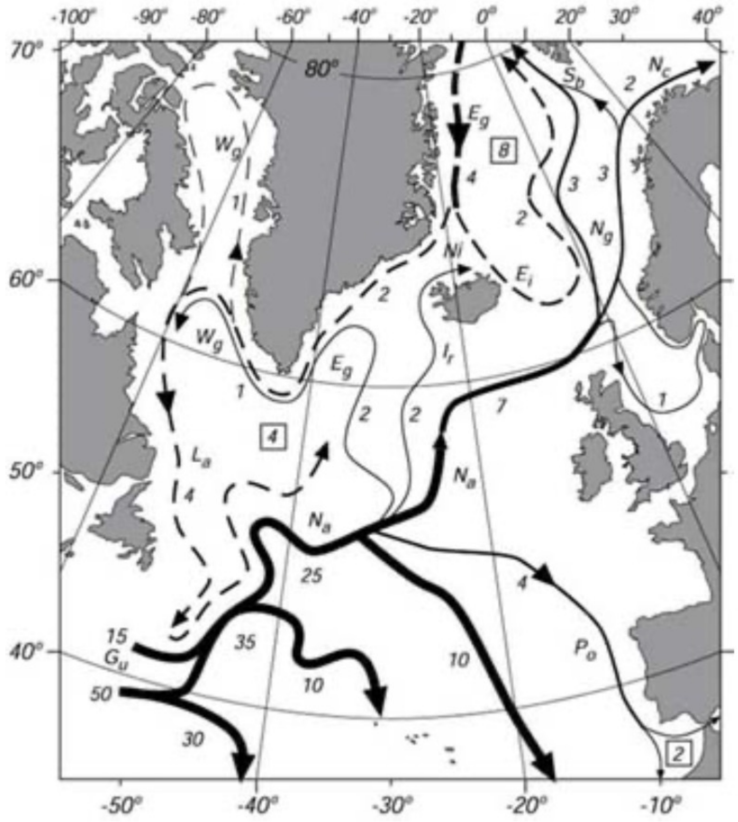
In the west, a western boundary current, the Gulf Stream, completes the 'gyre'. In the north, a subpolar 'gyre' includes the Labrador Current. A current and countercurrent system is seen in the low latitudes with a Pacific-like flow. Of note, however, is the large flow across the equator along the north-east coast of Brazil towards the Caribbean.

# The Gulf Stream



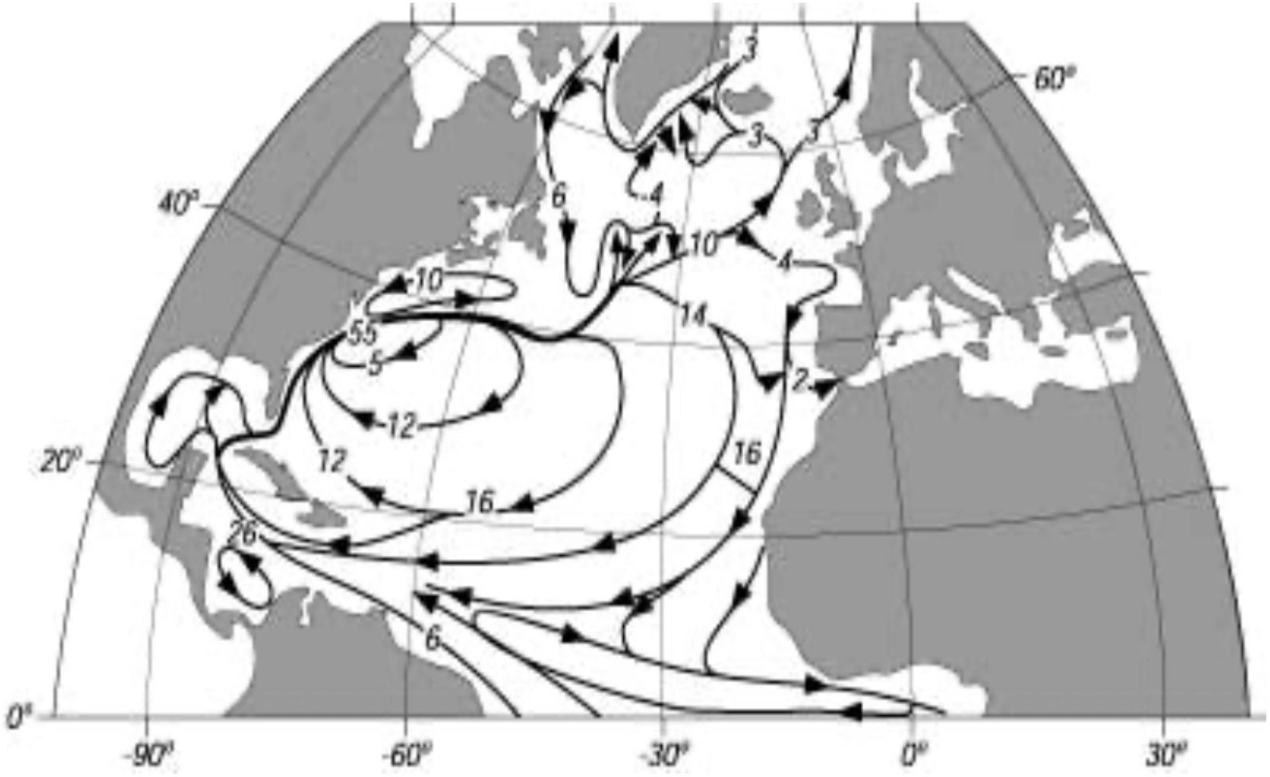
# The Atlantic Ocean

If we look more closely at the North Atlantic, we see that the flow is even more complex. This figure includes much more detail due to the many hydrological observations because the region is important for fishing and trade.



Detailed diagram of currents in the North Atlantic showing the major surface currents. The numbers in the squares give the transport in units of 106 m3/s from the surface to a depth of 1000 m. **Eg:** East Groenland current; **Ei:** East Island current; **Gu:** Gulf Stream; **Ir:** Irminger current; **La:** Labrador current; **Na:** North Atlantic current; **Nc:** Cape North current; **Ng:** Norwegian current; **Ni:** North Island Curr. **Po**: Portugal current; **Sp**: Spitsbergen current; **Wg:** West Groenland current. Solid lines: warm waters. Dashed lines: cold currents. From Dietrich, *et al*. (1980).

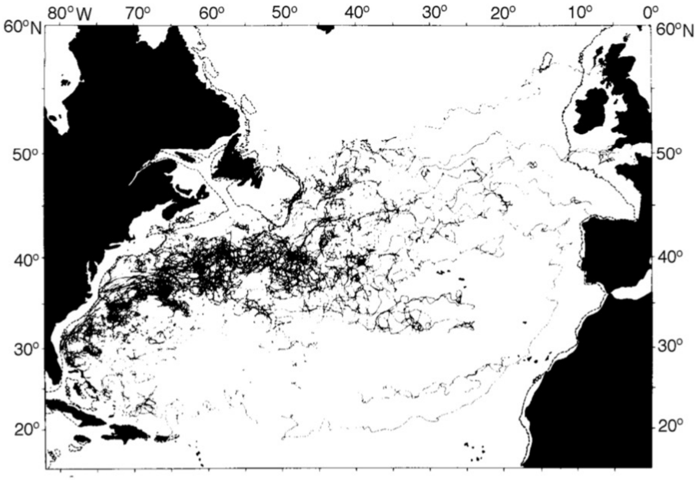
# The Atlantic Ocean



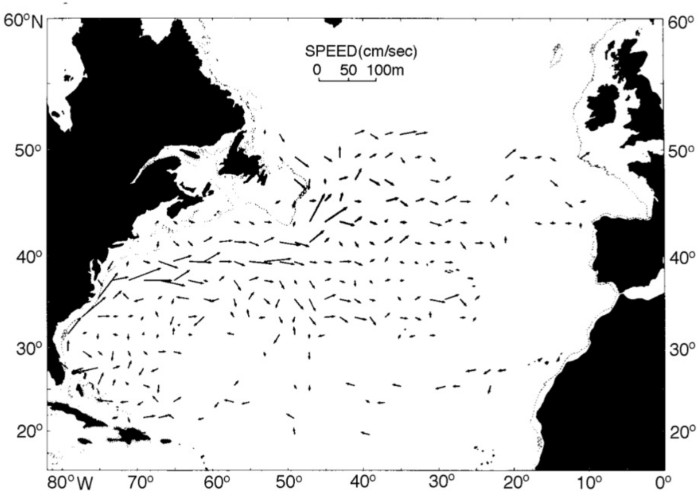

**If we launch a drifter into the Atlantic, will it follow the current lines shown in the figure?**

# The Atlantic Ocean

Track of 110 drifters launched in north-west Atlantic



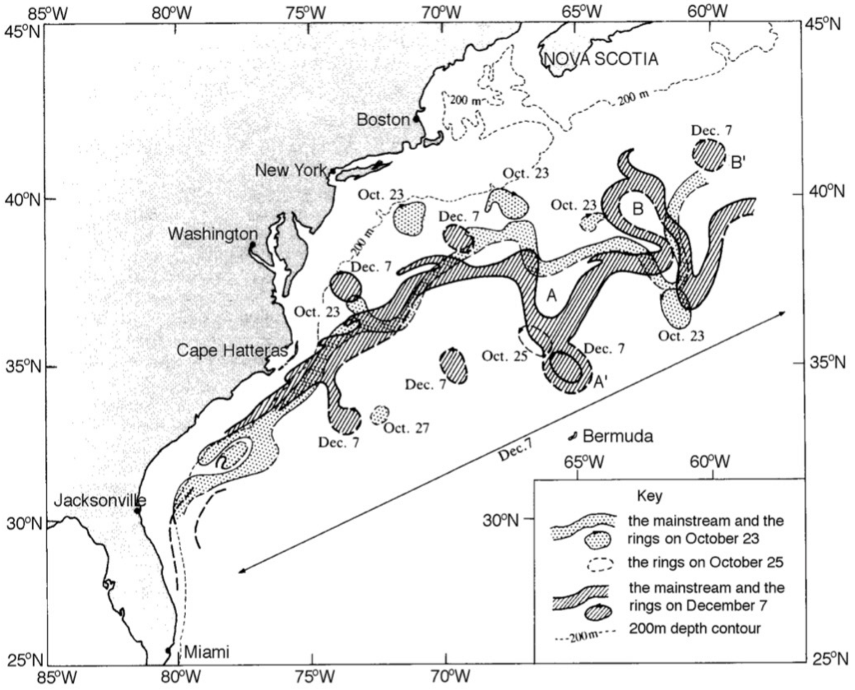

The Current obtained from the measurements of drifters




The flow is so variable that the average is not stable. Forty or more observations do not give a stable average value. Overall, Richardson found that the kinetic energy of the eddies is 8 to 37 times greater than that of the average flow. Thus, ocean turbulence is very different from that of the laboratory. In the laboratory, the mean flow is typically much faster than the eddies.

# The Gulf Stream

The shear of the intense current on the western edge causes a meander. The meander intensifies, and eventually breaks away from the main current, forming rings. Those on the south side deviate to the southwest, and eventually re-enter the current many months later. The process occurs throughout the recirculation region and satellite images show almost a dozen gyres (rings) north and south of the current.



# The Gulf Stream

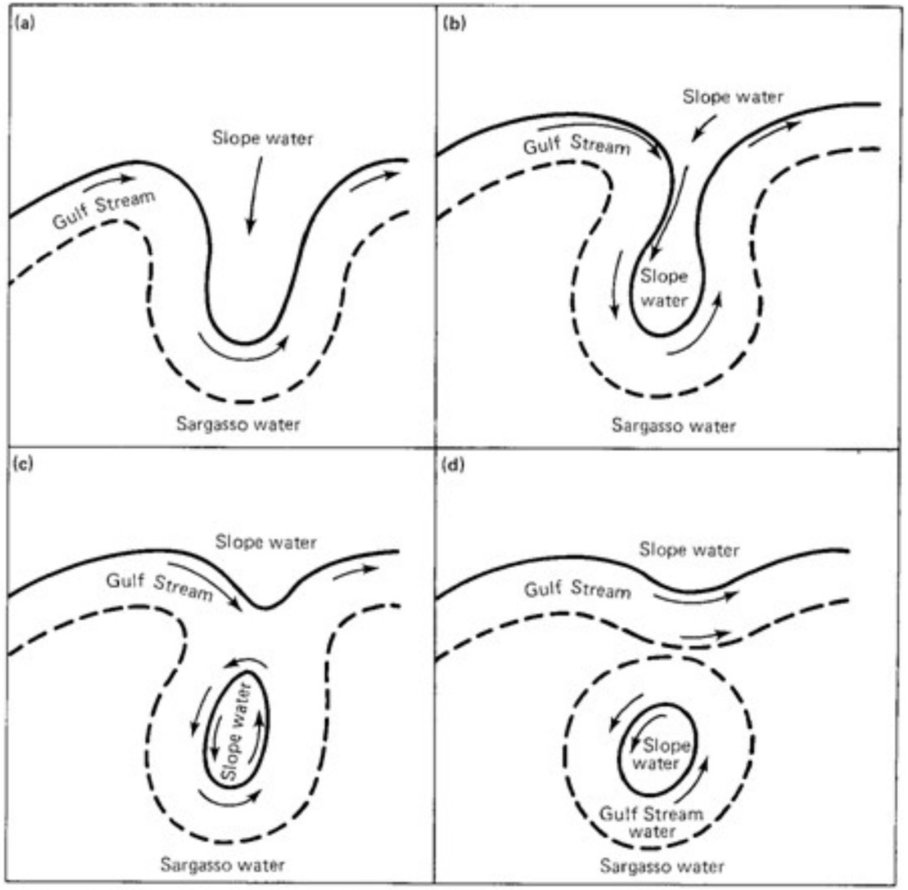
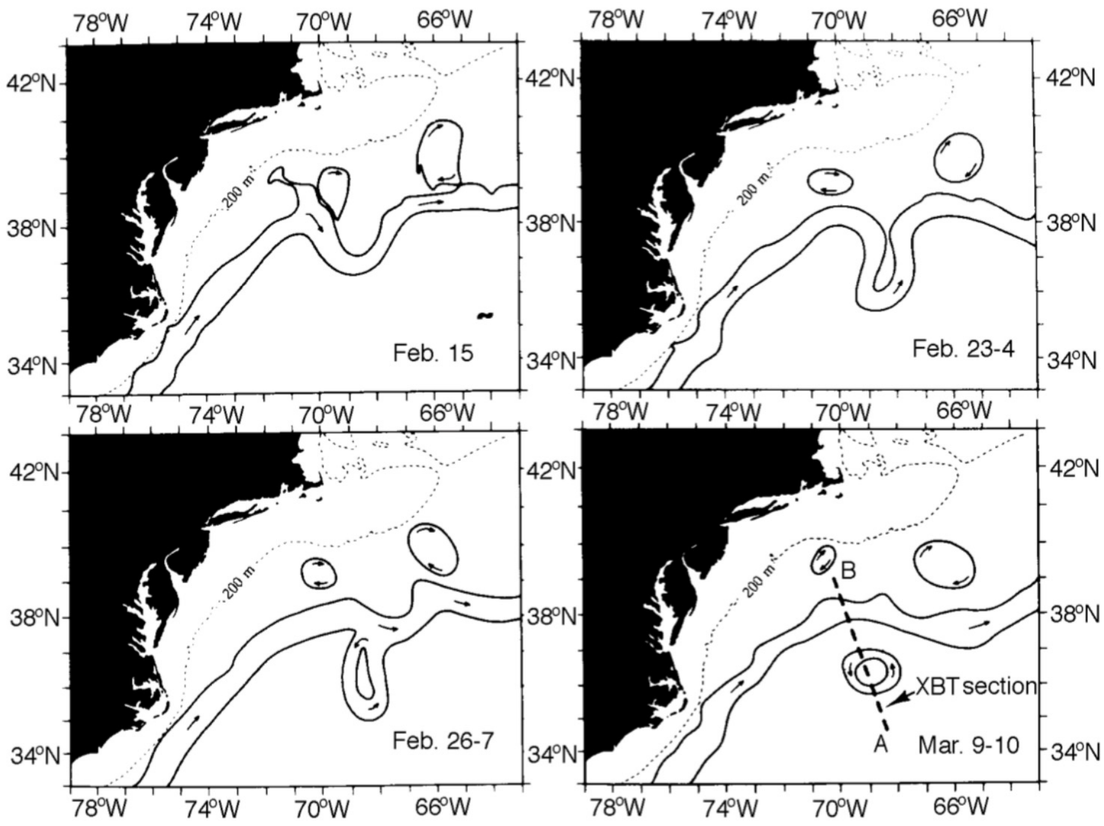



**Fig. 6.43** Diagram of separation of cold-core, cyclonic ring from the Gulf Stream starting with (a) meander motion, (b) extreme nonlinear meander, (c) detachment of ring and rejoining of main current system, and (d) independent ring. [From Parker, C. E., *Deep-Sea Res.* (1971).]

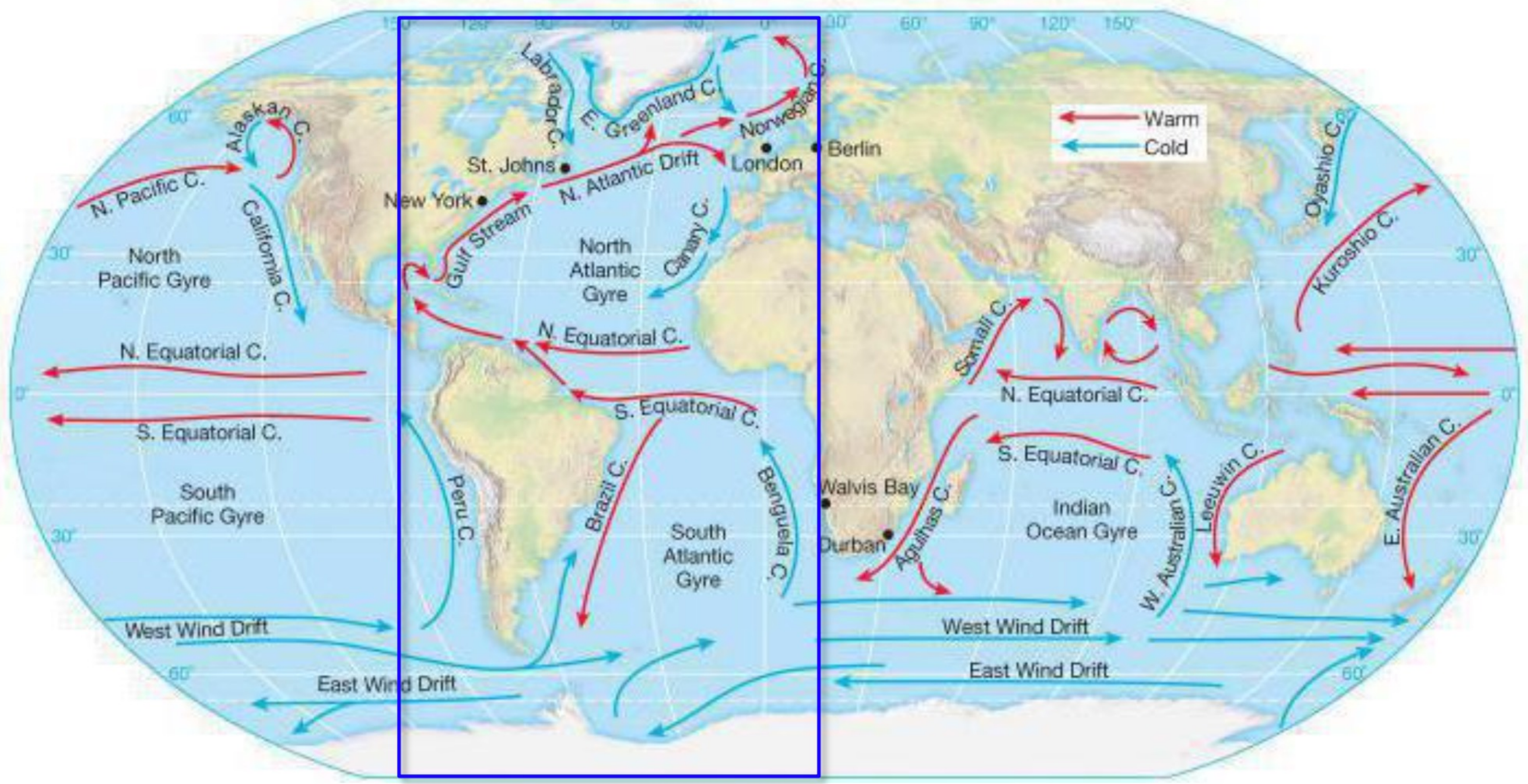
# The Gulf Stream




The meanders of the Gulf Stream lead to the formation of rotating vortices, the rings.

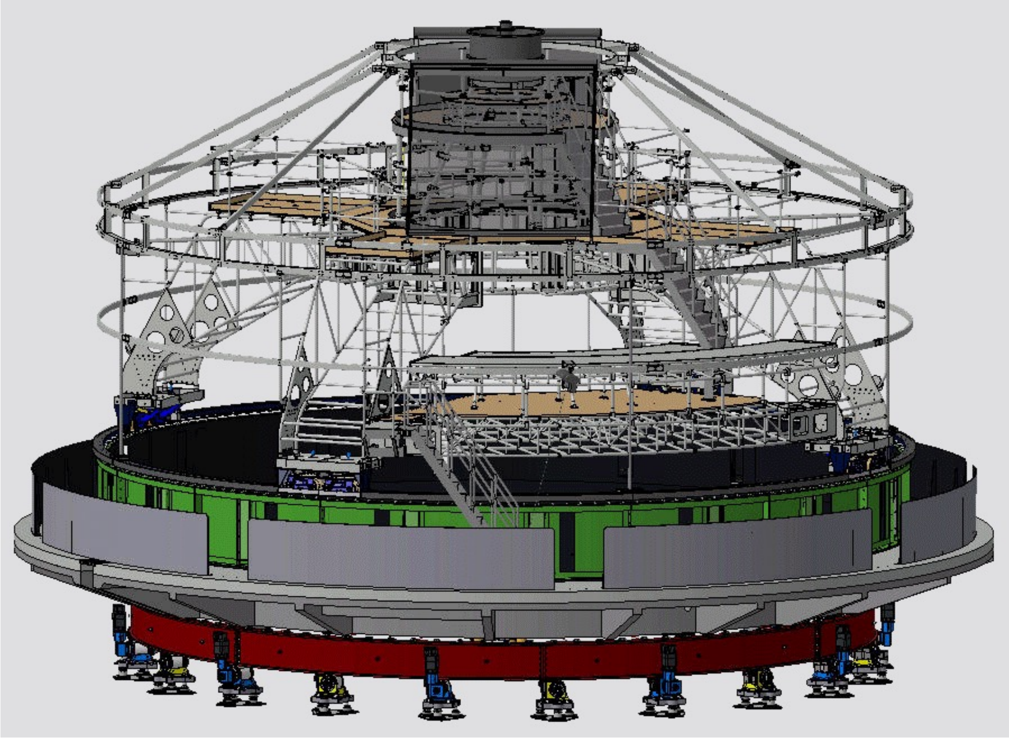
# The Atlantic Ocean

In the South Atlantic, there is another WBC, the Brazil Current, which completes the circulation in this basin. Between the North and South Atlantic Current lies the *equatorial circulation* similar to the Pacific.



# The Coriolis Platform

The Coriolis Rotating Platform is the largest Rotating Platform in the world, with its 13m diameter. It permits to study geophysical and environmental fluid mechanics problems taking into account the Earth's rotation, the stratification and the topography, approching the dynamical similarity of real flows.



Tecnical management

Samuel Viboud
(IR CNRS)

Thomas Valran
(IE CNRS)

Scientific coordination

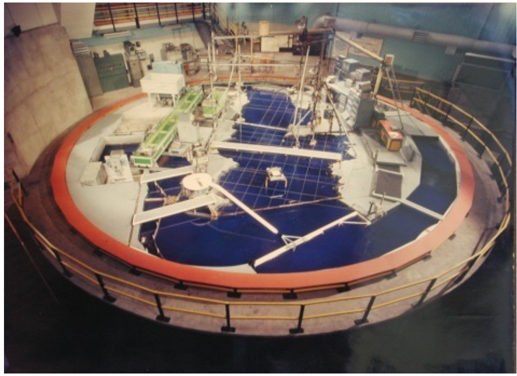
Eletta Negretti
(CR CNRS)

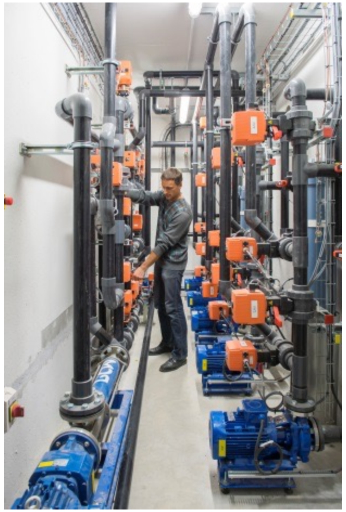
Joël Sommeria
(DR CNRS)

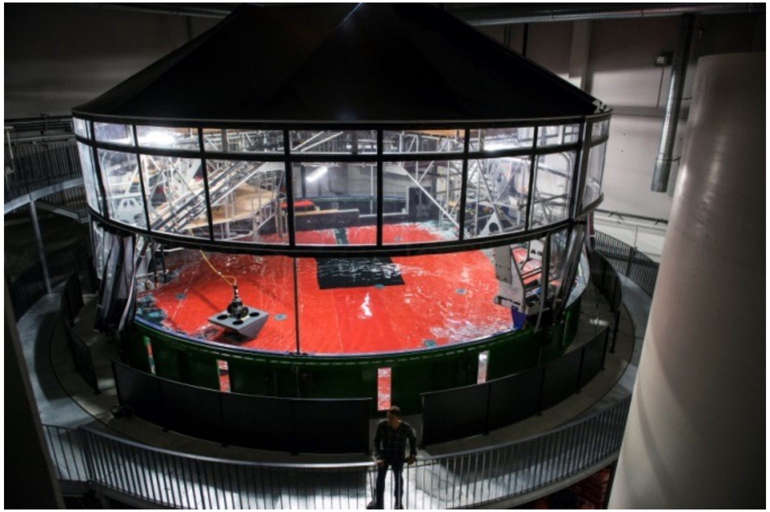
# The Coriolis Platform

A brief history

In 1960 the French national electricity company (EdF) asked for the construction of a model of the English channel for the simulation of tidal motions, aimed at the construction of a tidal power plant. In 1985, the Platform has been took in charge by the University and the CNRS, who financed the construction of the experimental tank and the in site portique. In 2011 it has been destroyed and reconstructed at LEGI.

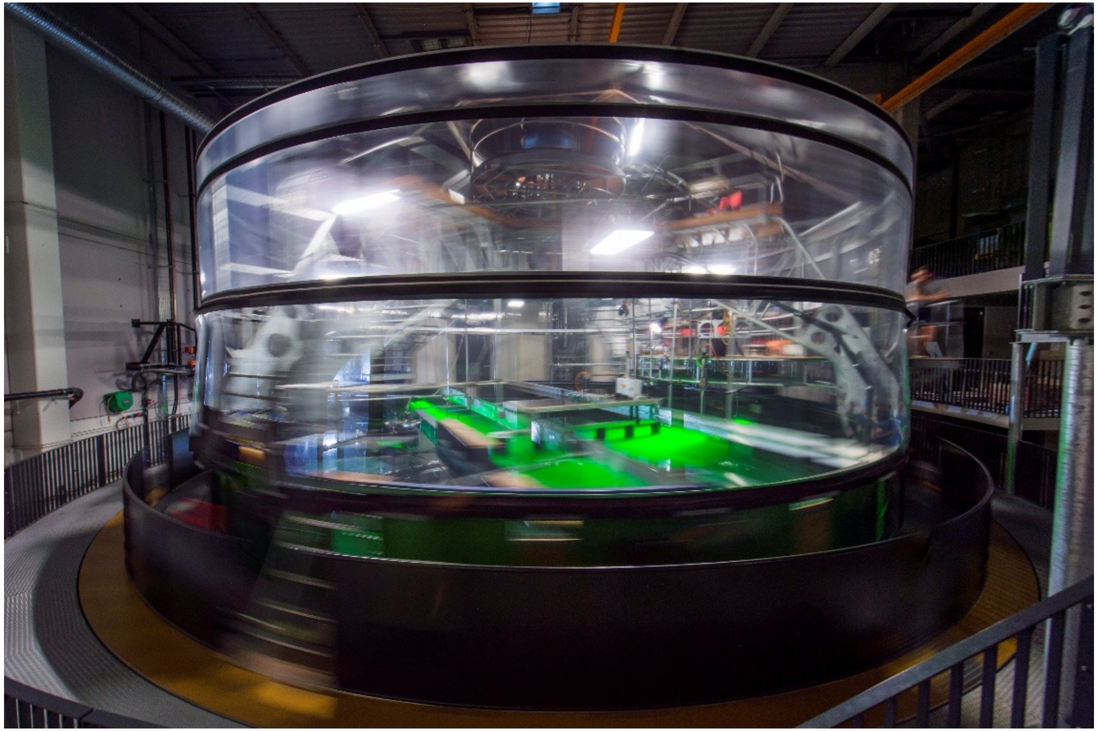






# The Coriolis Platform

## Technical aspects



### Caracteristics :

- 13 m de diamètre
- 350 Tons fully charged
- Maximum speed: 6 Tr/ min
- Maximum water depth: 1 m
- Global Cost: 6,5 M€

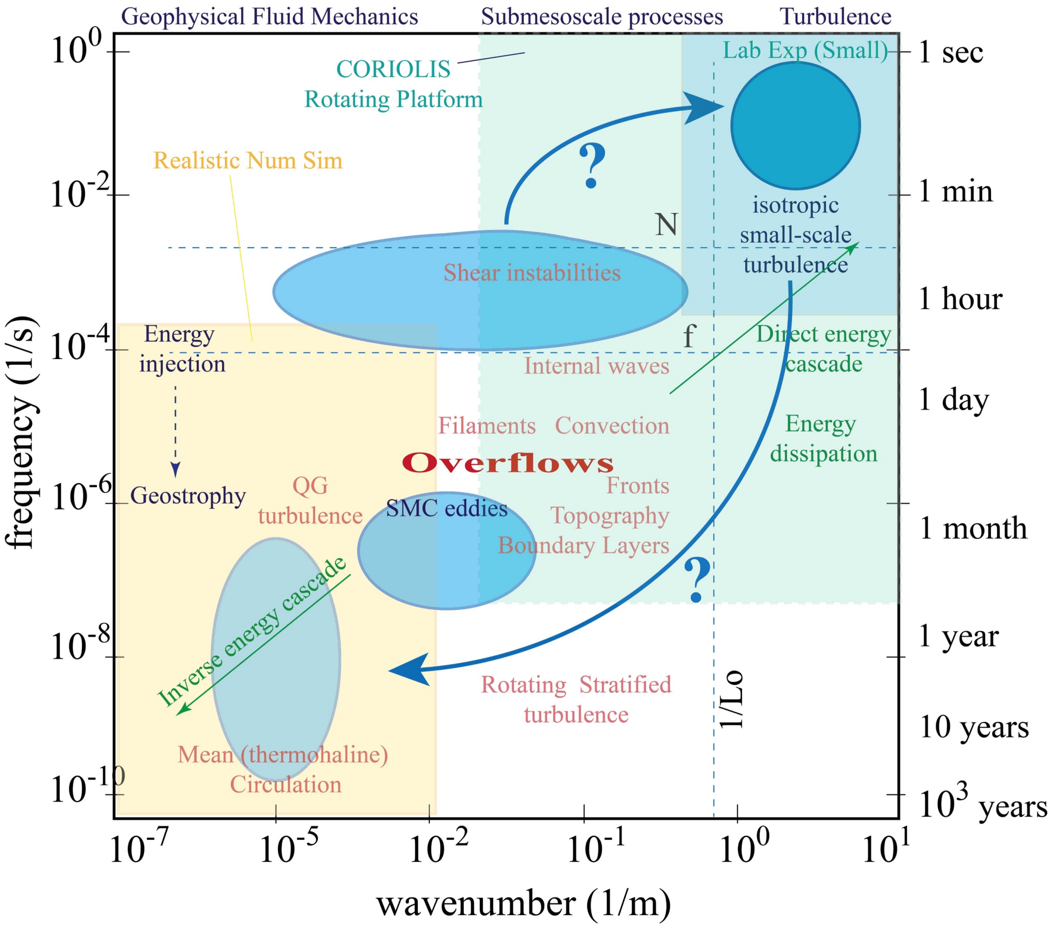
# International research groups

98 research projects in 22 years, 35 since 2014

# The Coriolis Platform

## Why is the CoriolisPlatform a unique tool for the simulation of geophysical flows?

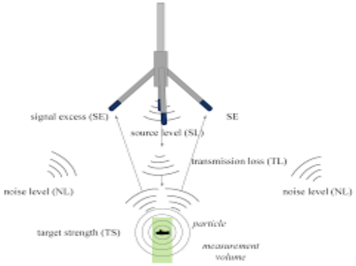
Geophysical fluid dynamics involves a large variety of spatial and temporal scales. It is impossible to resolve them in today's and near future numerical simulations. A parametrization of subgrid scale processes is needed for reliable predictions. For a correct parametrization, a good understanding of the physical processes is needed.



# Laboratory measurement techniques

Experimental techniques permit to measure the velocity fields (2D-3D, 2-3 velocity components) and the density fields, along with their fluctuations with high temporal and spatial resolution. These are the most important quantities on the base of which mean flow and turbulent characteristics can be determined. Experimental techniques are substantially based on

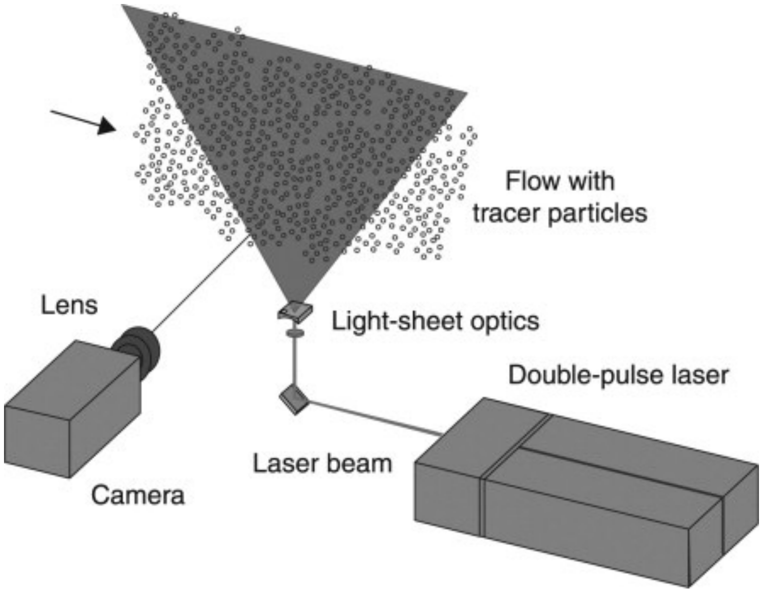
- Intrusive (point-wise, profiler): Acoustic doppler Velocimetry (ADV, for 3C velocities), Conductivity probes (measure of salinity and temperature for density)

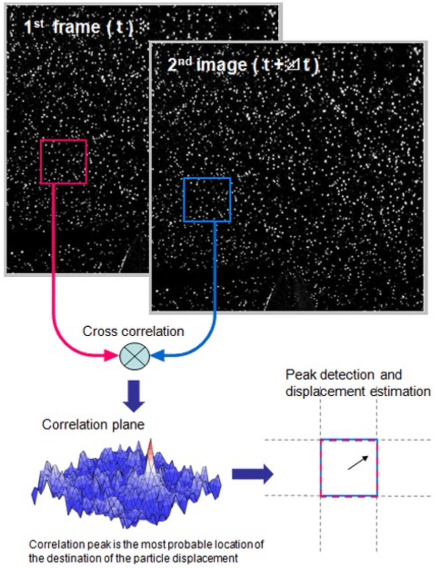



- Optical, non intrusive:
    - Particle Image Velocimetry (PIV), for 2D-2C velocities
    - Laser Induced Fluorescence (LIF)

# Particle Image Velocimetry

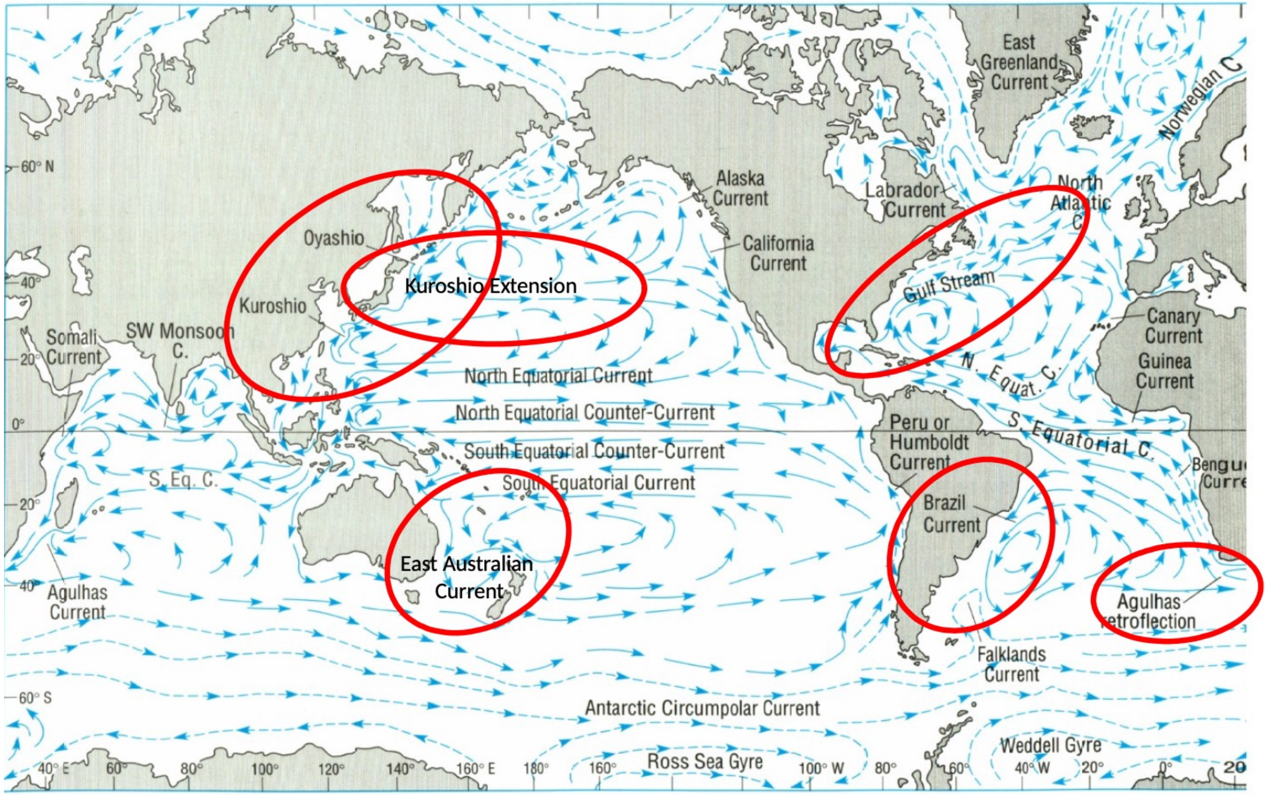
Used to obtain instantaneous velocity measurements and related properties in fluids. The fluid is **seeded with tracer particles** which, for sufficiently small particles, are assumed to faithfully follow the flow dynamics. The fluid with entrained particles is **illuminated with a laser** so that particles are visible. Images are captured using **scientific cameras** at a given frequency, **synchronized** with the laser. The motion of the seeding particles is used to calculate speed and direction, i.e. velocity field of the flow, via (auto/cross-) correlations between sub-areas of two successive images.



# Gapwebs project: WBC (the Gulf Stream)

WBCs important for climate because of their huge heat transports, corresponding air–sea interactions and the role they play in sustaining the global conveyor belt.




*(from "Ocean Circulation", The Open University Oceanography Course Team, 2001)*

# Gapwebs project: WBC (the Gulf Stream)

**How to produce the beta effect (meridional gradient of Coriolis parameter) in the lab ?**

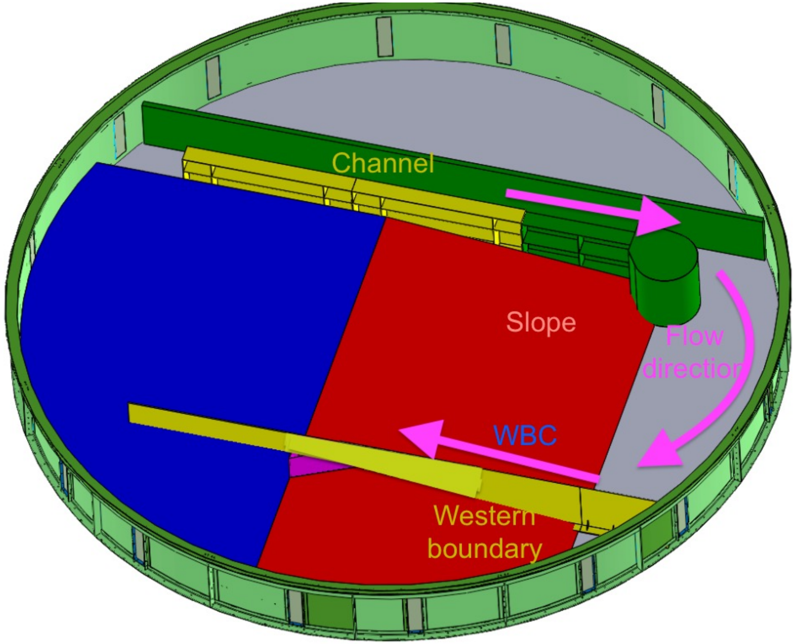
$$\frac{D}{Dt}\left(\frac{\zeta + f}{H}\right) = 0$$

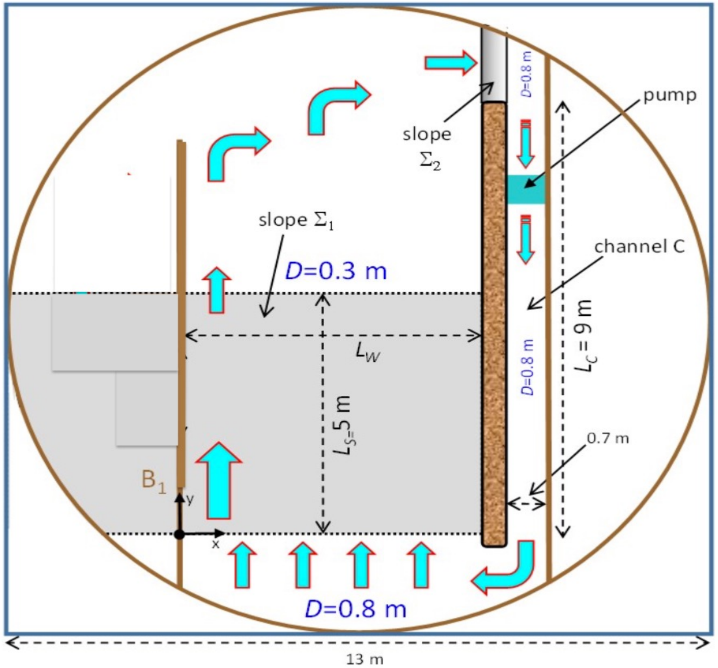
**The *topographic* beta effect**

By invoking the conservation of potential vorticity in the quasi-geostrophic approximation, one has $\zeta + f + \beta y = \text{const.}$
where $\zeta = v_x - u_y$ is the relative vorticity of the water column, $\beta = f\, s/D$ is the topographic $\beta$ simulating the corresponding planetary effect (**s** is the bottom topography, and **D** is the mean slope depth) and **y** is the "northward" direction. By moving northward, a water column experiences an induction of negative (clockwise) relative vorticity from the planetary vorticity: at the entrance of the slope (y=0), we find $\zeta = 0$, thus, at a given $y = y_0$ one has $\zeta = -\beta y_0$. Negative shear vorticity (i.e., the westward intensification) in a virtually rectilinear boundary current is therefore generated by the presence of the lateral boundary. Moreover, positive shear vorticity is present in a thin frictional boundary layer along the wall due to the no-slip boundary condition. The flux of negative vorticity input continuously provided by the northward flow is balanced by dissipation associated with bottom and lateral friction, so that a constant WBC is present.

# Gapwebs project: WBC (the Gulf Stream)

A pumping system located in the channel C produces a current of constant speed U at its "southern" border; a virtually unsheared flow at the southern entrance of the topographic slope **$\Sigma_1$=0.105** is thus formed. The bottom slope provides the topographic β -effect necessary for the intensification along the "western" boundary $B_1$ of a total length **$L_s$=5m**. The total water depth at the beginning of the slope till the end of the slope varies linearly from **D=0.8m** to **D=0.3m**, respectively.



# Gapwebs project: WBC (the Gulf Stream)

- The dataset in 22TPTMA\GAPWEBS\WBC\ containes **4 experiments** (23-26), whose characteristics are given in the table below.

| EXP | T | Q | Gape length |
|---|---|---|---|
| EXP21 | 30 | 6 | 1.2 |
| EXP22 | 30 | 8 | 1.2 |
| EXP23 | 30 | 2 | NO |
| EXP24 | 30 | 3 | NO |
| EXP25 | 30 | 6 | NO |
| EXP26 | 30 | 8 | NO |

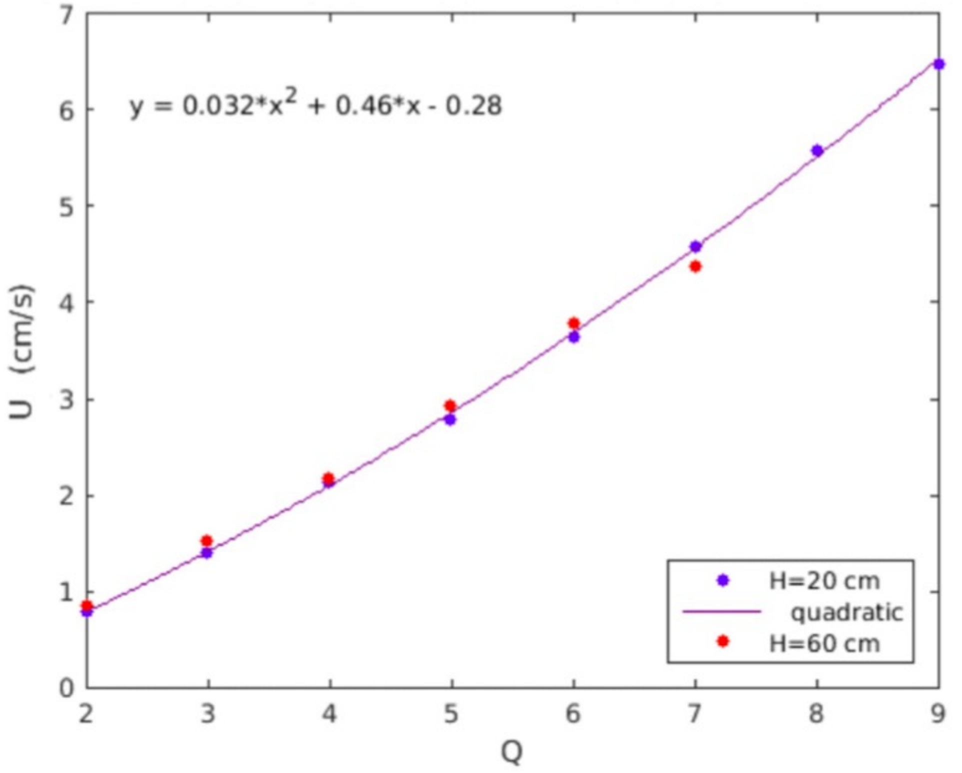
- In each experiment, **1000 *.nc files are present**, which contain the instantaneous two-dimensional velocity field (U,V) in the (x,y) plane obtained by PIV in a plane **10cm** below the free surface, where the origin of the axis are on the western boundary, at the beginning of the slope (see figure in the previous slide).

- The time shift between two successive velocity fields (img_i_1.nc - img_i+$1_1$) is **one second**.
- Note that f in the lab is calculated as f=4pi/T

# Gapwebs project: WBC (the Gulf Stream)

- The Flow rate injected can be estimated using the calibration curve below. Refer also to the sketchs of the set up presented in slide 24.

| EXP | T | Q | Gape length |
|---|---|---|---|
| EXP21 | 30 | 6 | 1.2 |
| EXP22 | 30 | 8 | 1.2 |
| EXP23 | 30 | 2 | NO |
| EXP24 | 30 | 3 | NO |
| EXP25 | 30 | 6 | NO |
| EXP26 | 30 | 8 | NO |



# To Do List

1) Determine the averaged velocity fields and the Reynolds shear stresses field for each experiment. Make a comparison and highlight the differences.
2) Determine the averaged potential vorticity field for each experiment. Make a comparison and highlight the differences.
3) Give significant snapshots of the instantaneous velocity and potential vorticity fields and comment them.
4) Plot the transverse meridional velocity profile v(x) and zonal velocity profile u(x) at three significant **y** positions. Do the temporal averaged and instantaneous profiles differ considerably ?
5) Estimate the Kolmogorov length scale and compare it to the PIV spatial resolution = 3mm.
6) In the outer part of the velocity profile v(x) at the three chosen ***y*** positions, calculate the turbulent lateral eddy viscosity coefficient $\mathbf{A_H}$, the vertical eddy viscosity $\mathbf{A_Z}$, the turbulent dissipation rate ***e*** and a characteristic mixing length $\mathbf{L_m}$. Compare between the different experiments.
7) Show that the outer part of the velocity profile can be predicted using the conservation of the potential vorticity equation ζ + f + βy = const., where ζ is approximately $v_x$ and β = f s/D (s is the slope of the bottom and f=4pi/T, with T being the rotation period in the experiment).
8) Calculate the following quantities characteristics of the WBC boundary layer : $\delta_I = \left(\frac{U}{\beta}\right)^{1/2}$ ; $\delta_M = \left(\frac{A_H}{\beta}\right)^{1/3}$ $\delta_* = \sqrt{\delta_I A_H / U}$

that are the inertial, viscous Munk and the thin viscous boundary layer widths, respectively. How does the width of the WBC scale with the inertial boundary layer ? And how do the viscous and viscous Munk boundary layer widths scale with the inertial boundary layer widths ? Are the conclusions similar for all ***y*** and for each experiment ? Could the turbulent eddy viscosity $\mathbf{A_H}$ be replaced by the kinematic viscosity of water?