

Introduction à l'Océanographie Physique

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Introduction:

- un coup d'œil général
- Les équations du mouvement et la géostrophie

I. Les réponses de l'océan au vent

- courants d'Ekman
- courants géostrophiques

II. Les réponses de l'océan aux flux thermo-halins atmosphériques:

- la formation des «masses» d'eau dans la couche de mélange
- la subduction, la circulation dans la thermocline et l'obduction
- la circulation profonde

III. Le rôle de l'océan dans le climat:

- les ondes océaniques
- la variabilité couplée o/a (El Niño,...)

Références:

- ***Introduction to physical oceanography***

R.H. Stewart

http://oceanworld.tamu.edu/resources/ocng_textbook/contents.html

- ***Physics of atmosphere and oceans***

J. Marshall et A. Plumb

<http://www-paoc.mit.edu/labweb/>

- **Descriptive Physical Oceanography**

George L. Pickard, William J. Emery et Lynne D. Talley

http://www-pord.ucsd.edu/~ltalley/sio210/pickard_emery

Introduction:

Cours 1

Un coup d'œil général

Estimations océanographiques de surface

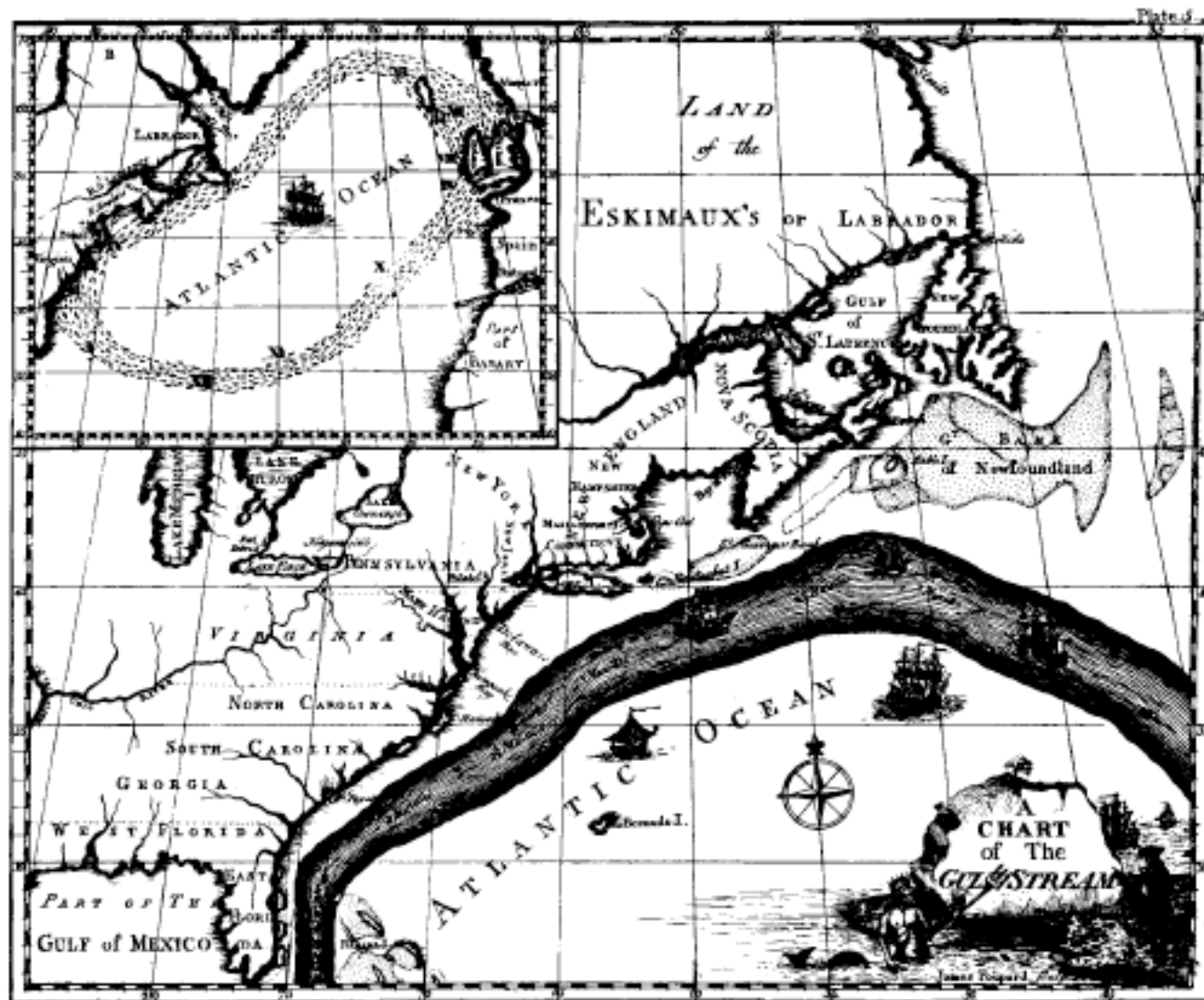
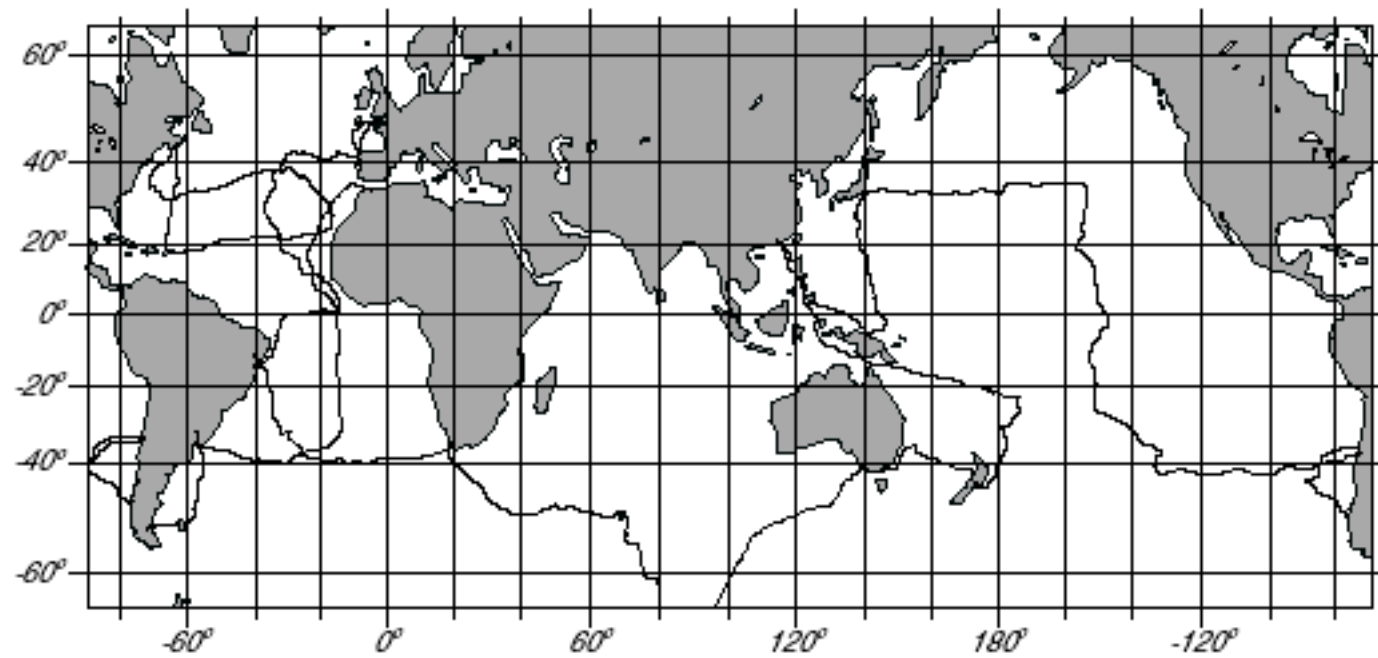


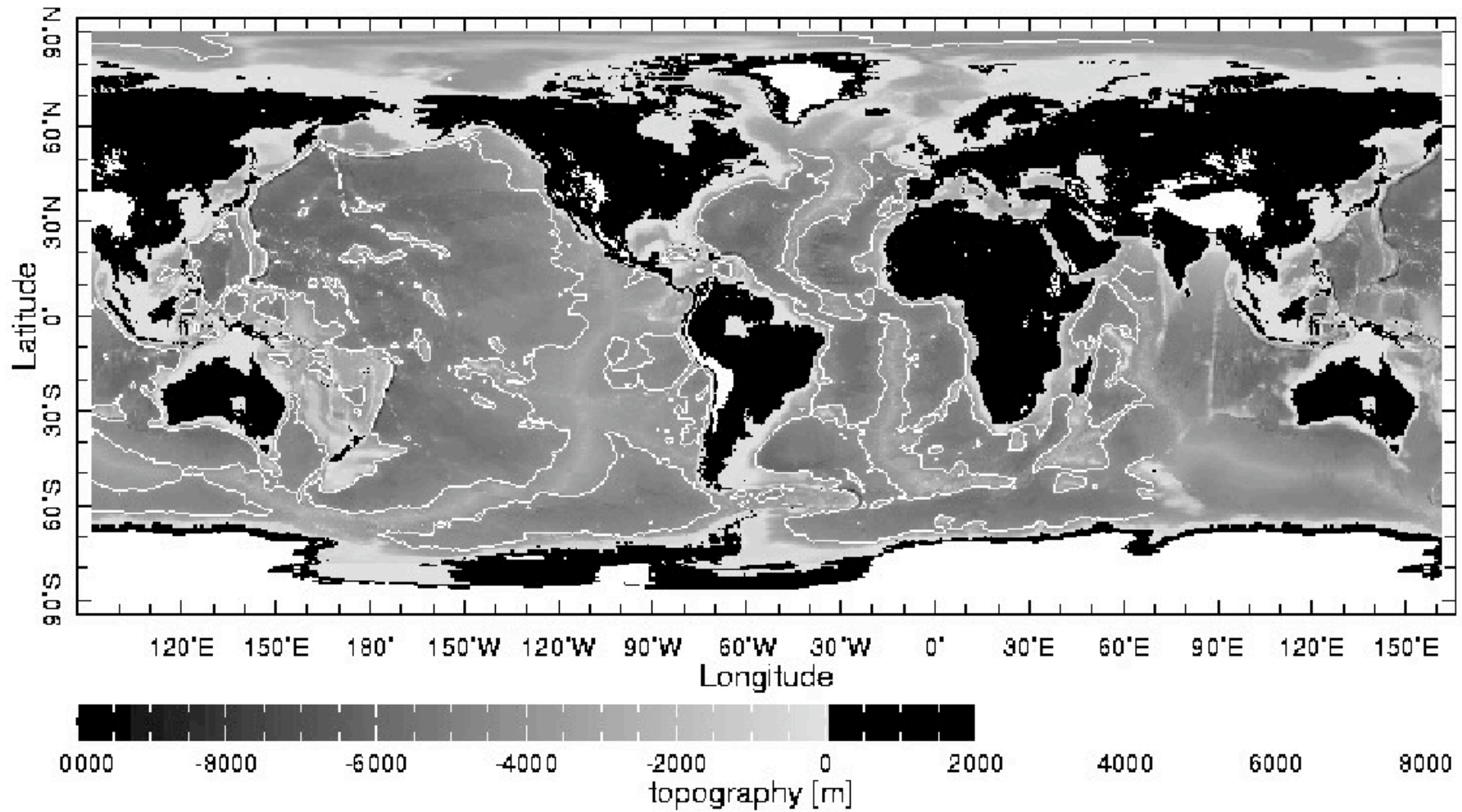
Figure 2.7 The 1786 version of Franklin-Folger map of the Gulf Stream.

1eres explorations océanographiques

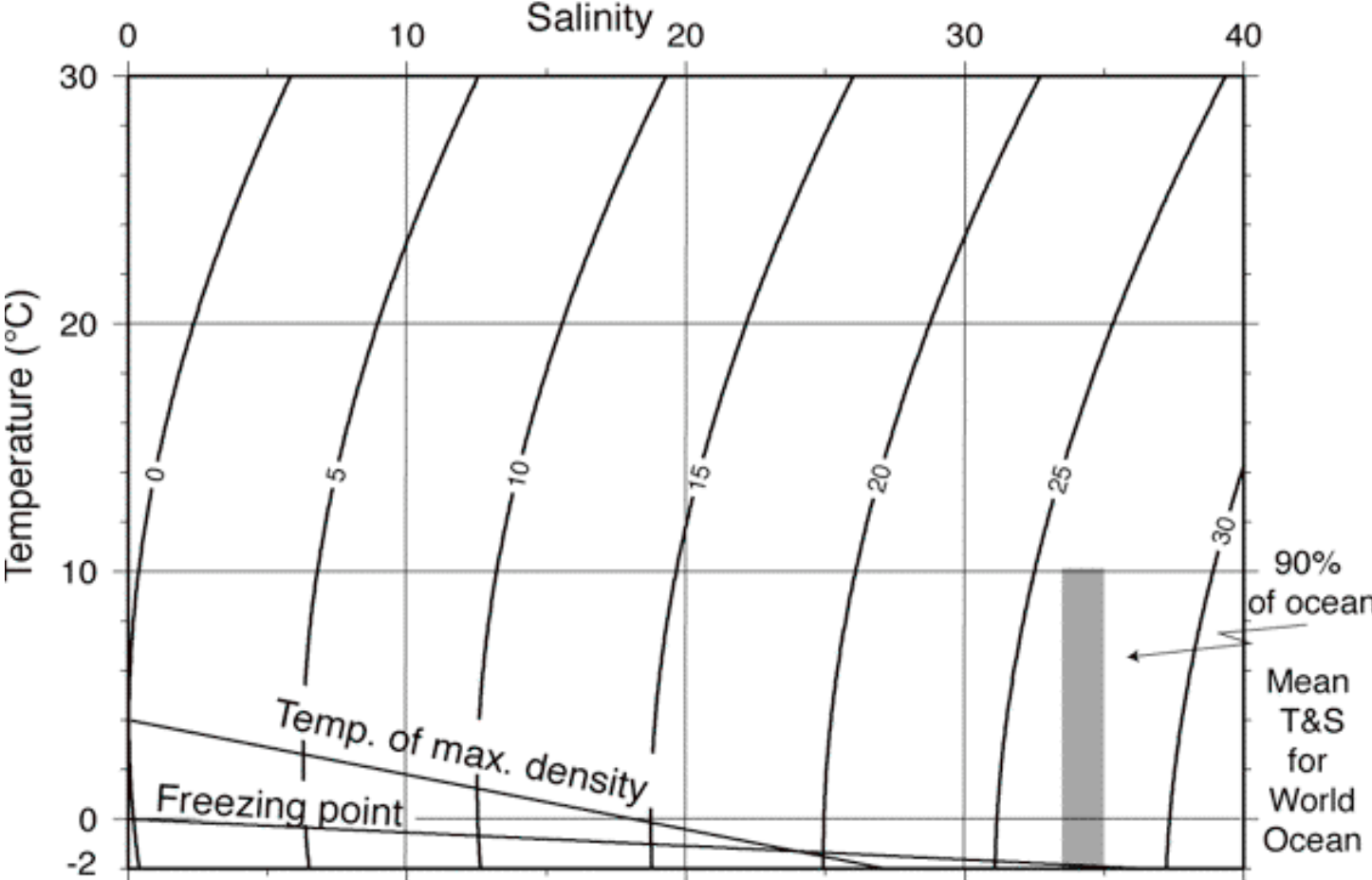


British Challenger Expedition 1872–1876. After Wüst (1964).

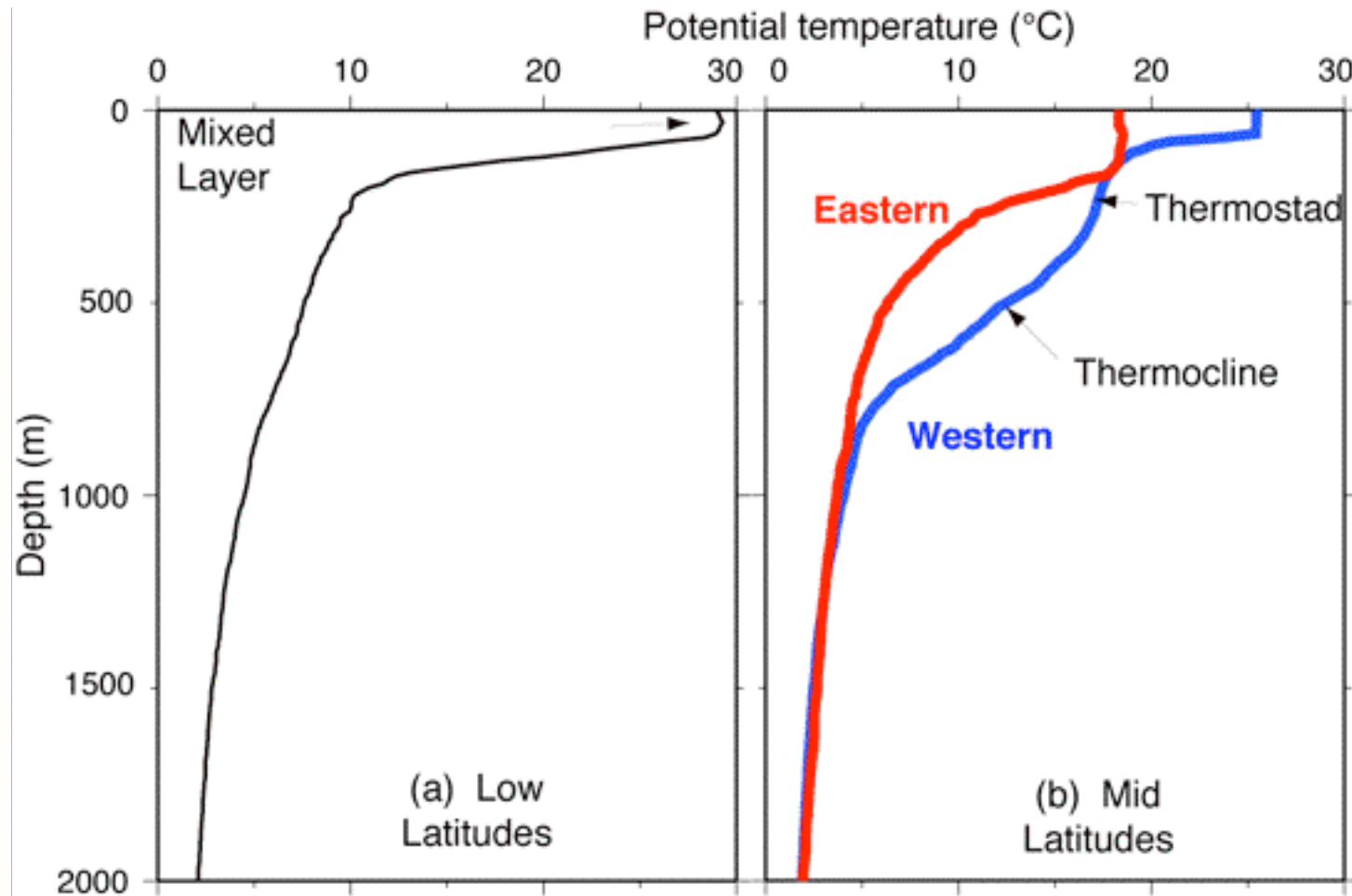
Topographie et rapport d'aspect de l'océan



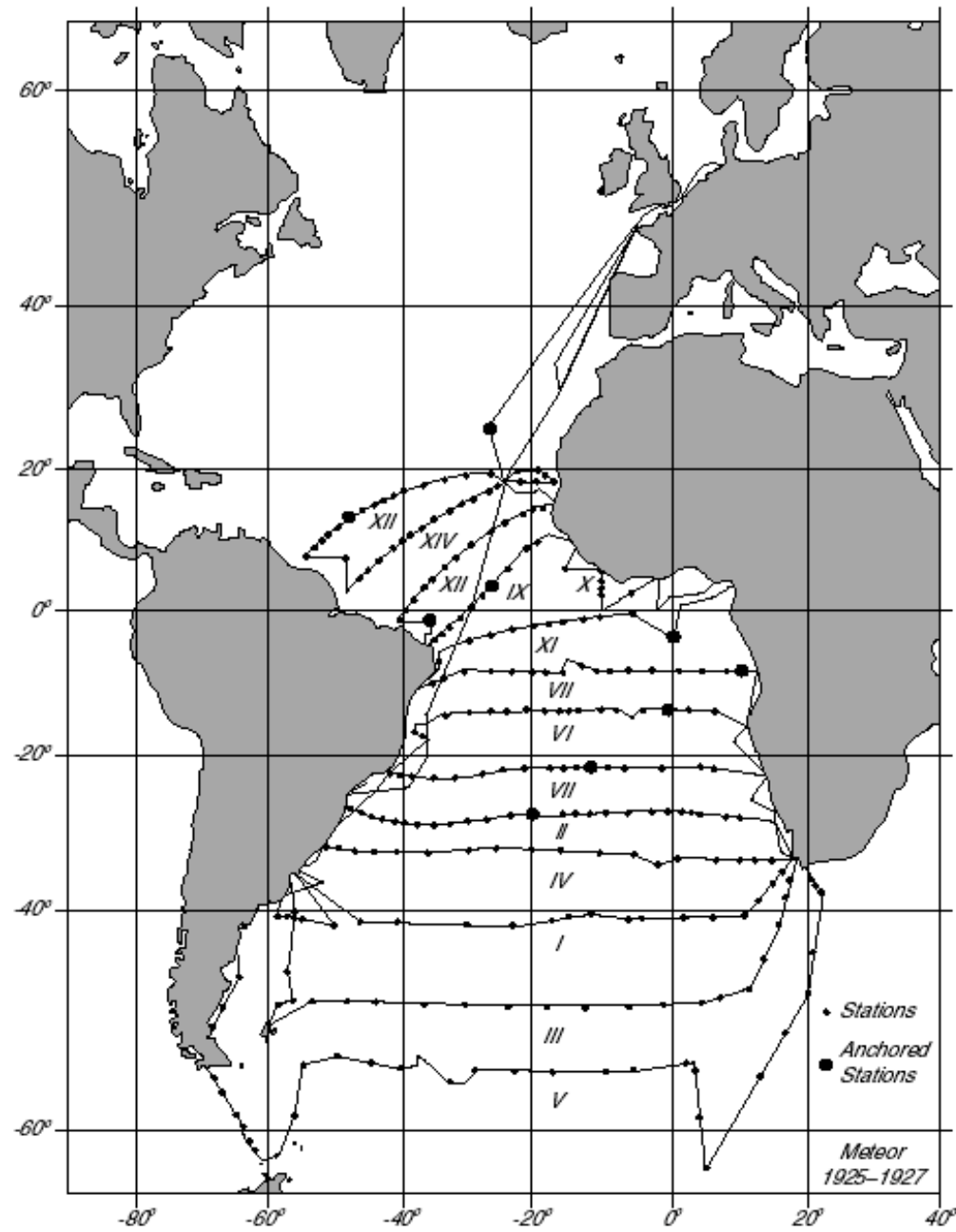
Apperçu des distributions et structures associées aux océans



Apperçu des distributions et structures associées aux océans

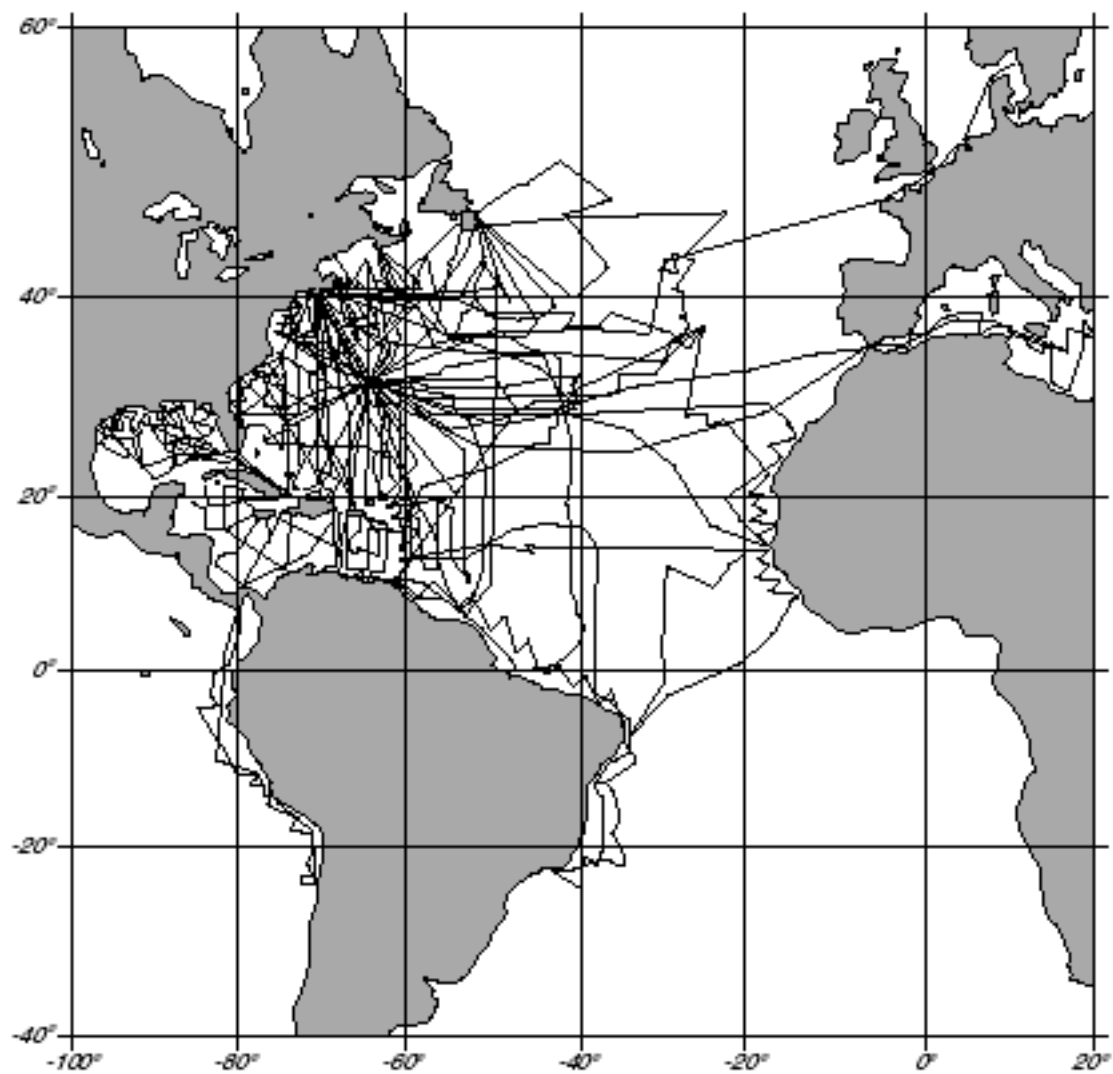


Campagnes systématiques



Campagnes systématiques

Développement des instruments de mesure



Campagnes de coopérations internationales

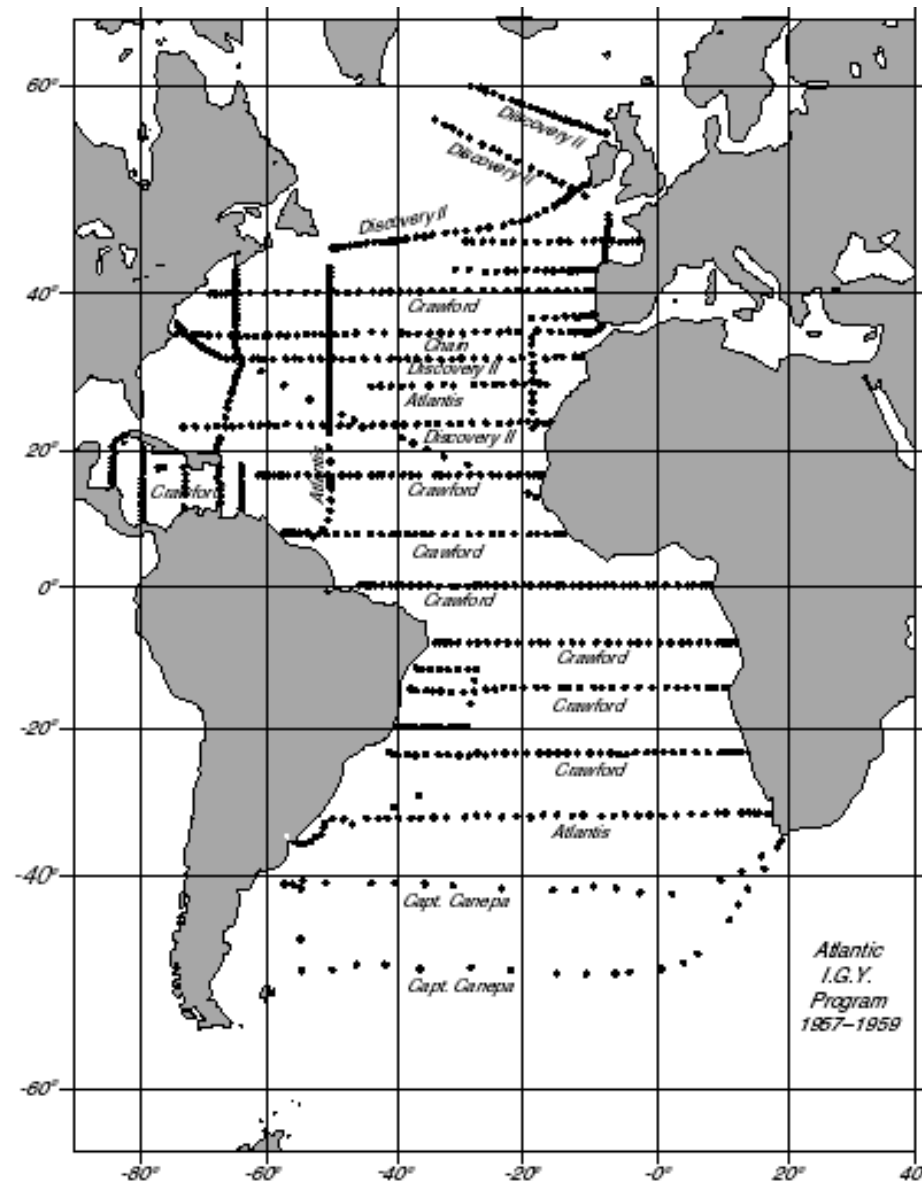
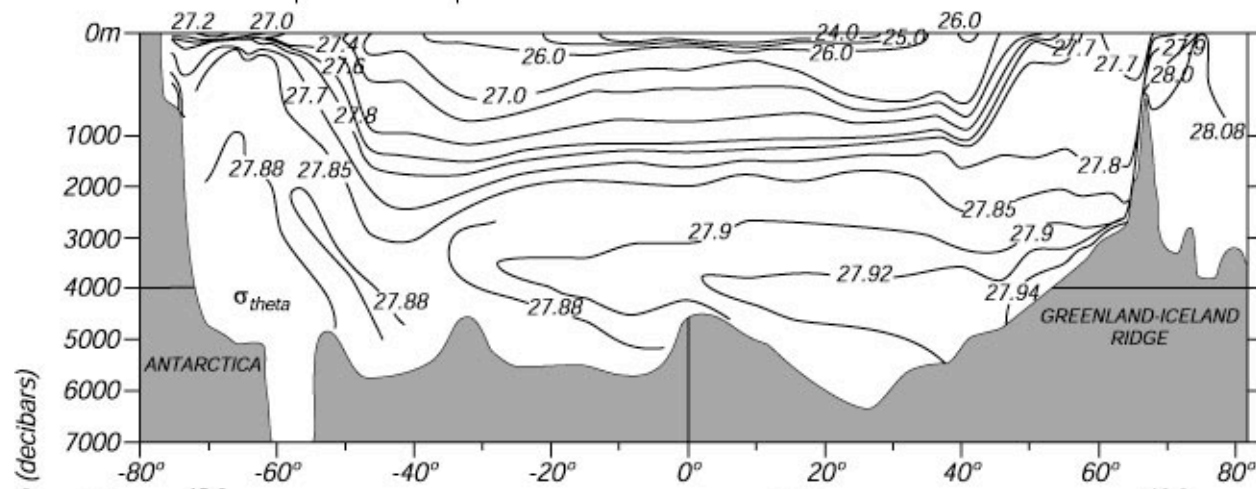
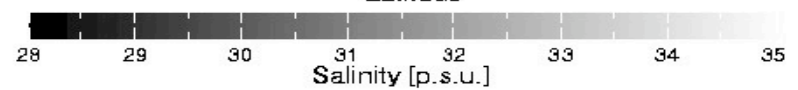
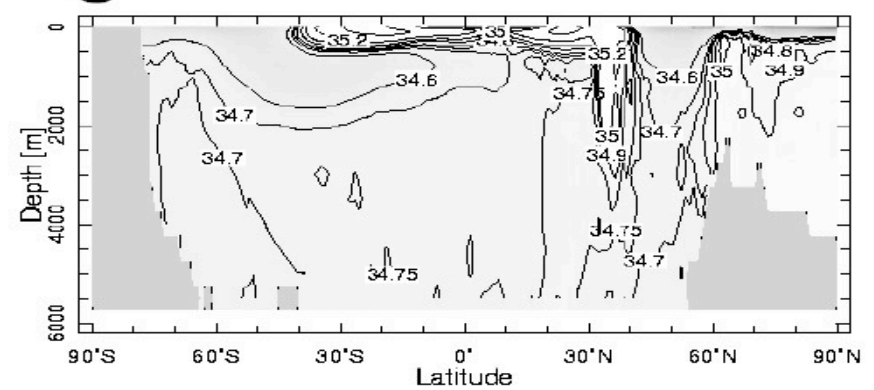
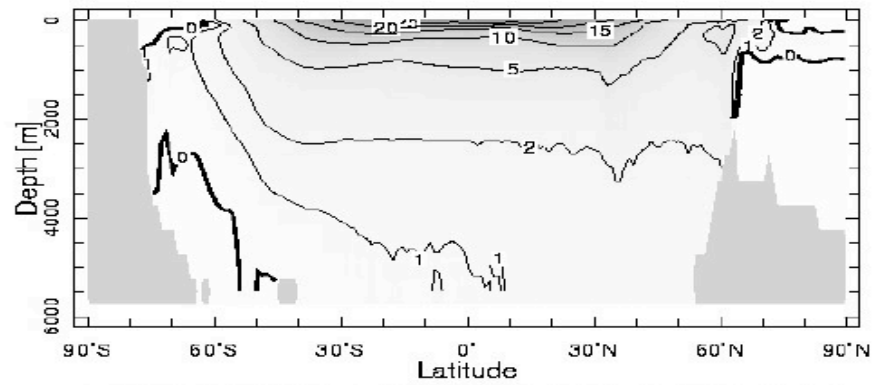
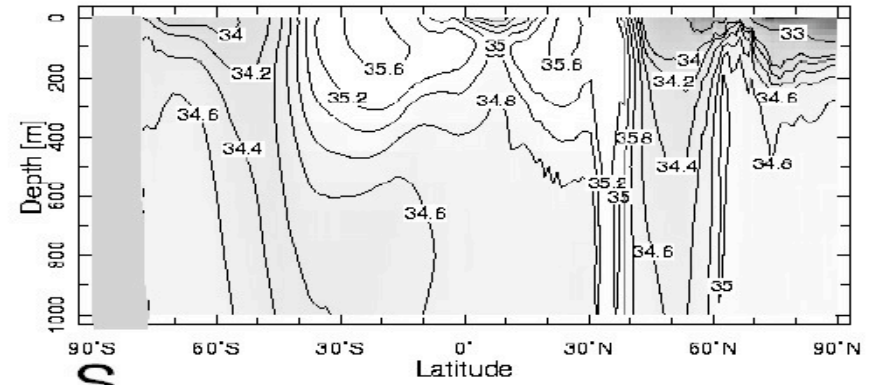
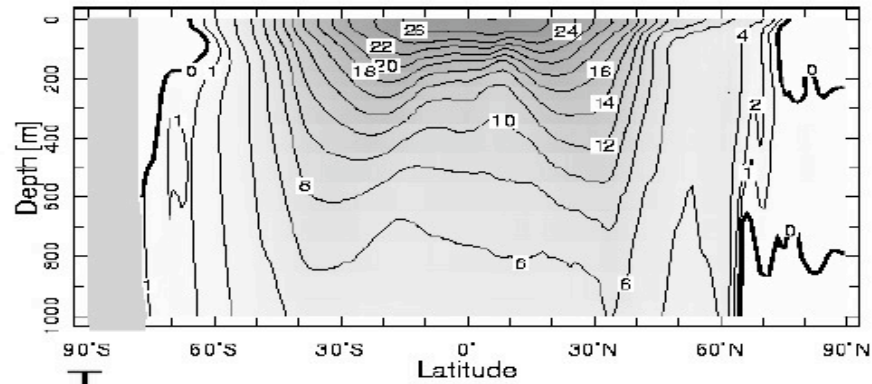
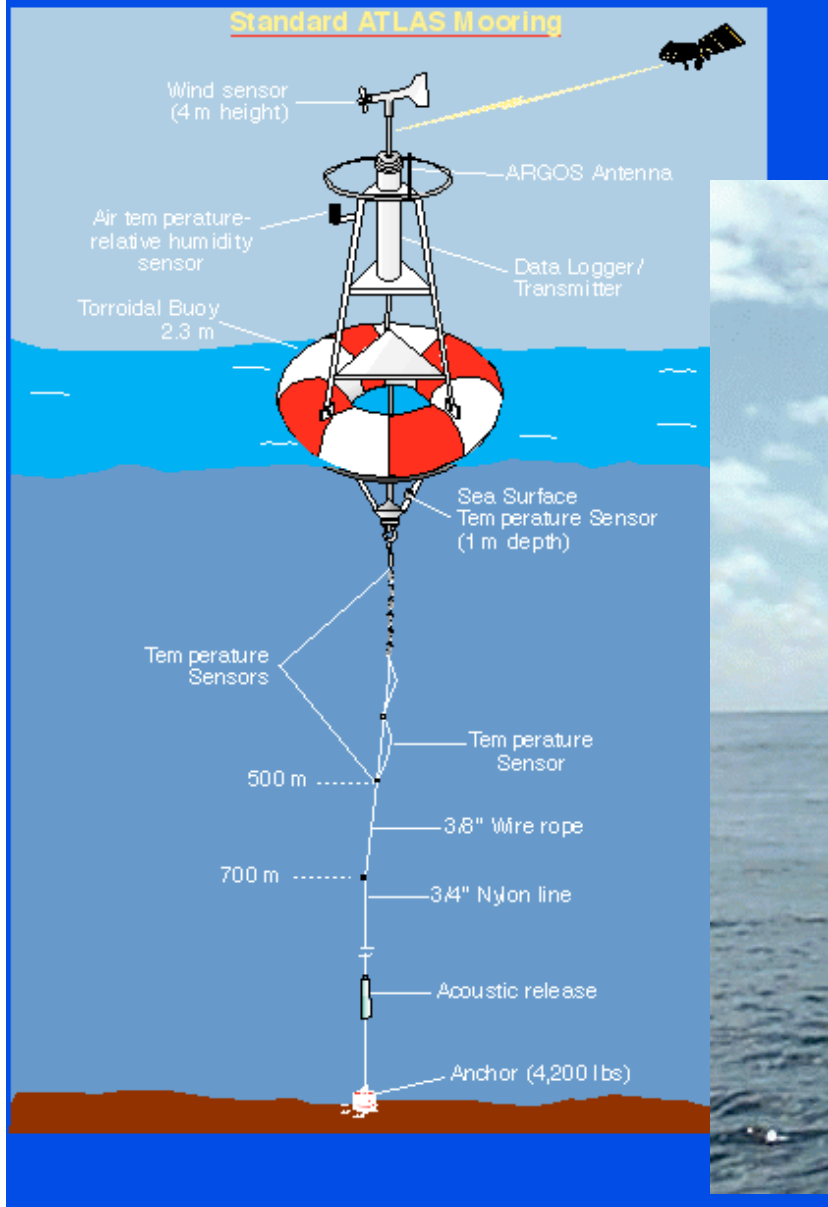


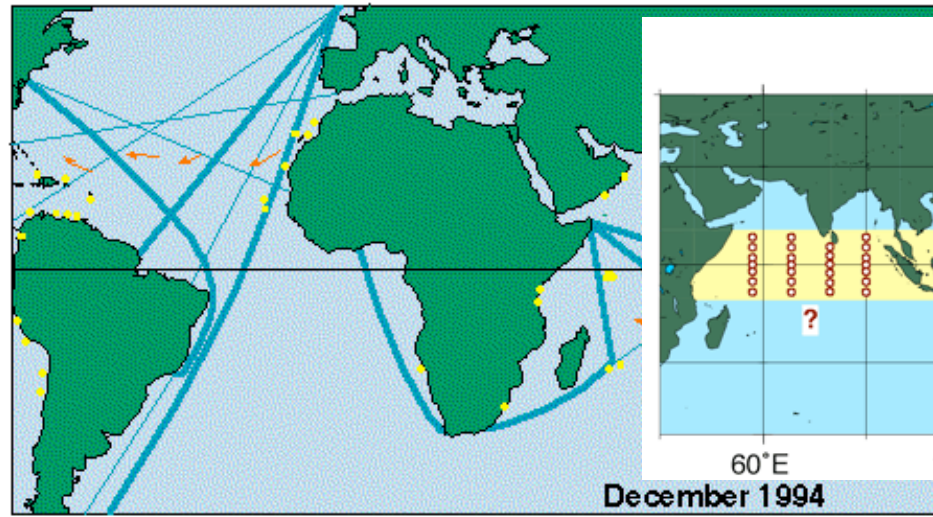
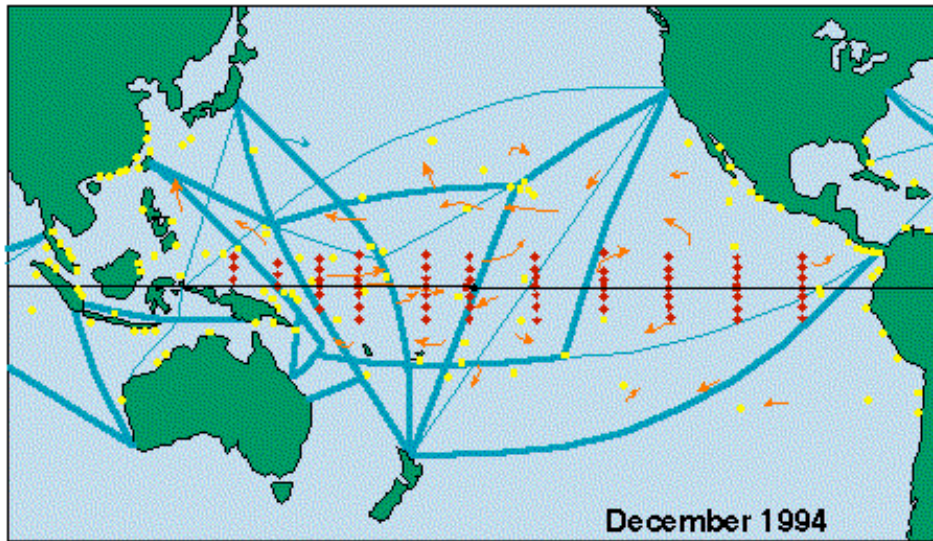
Figure 2.4 Example from the era of international cooperation . Sections measured by the International Geophysical Year Atlantic Program 1957-1959. After Wüst (1964).



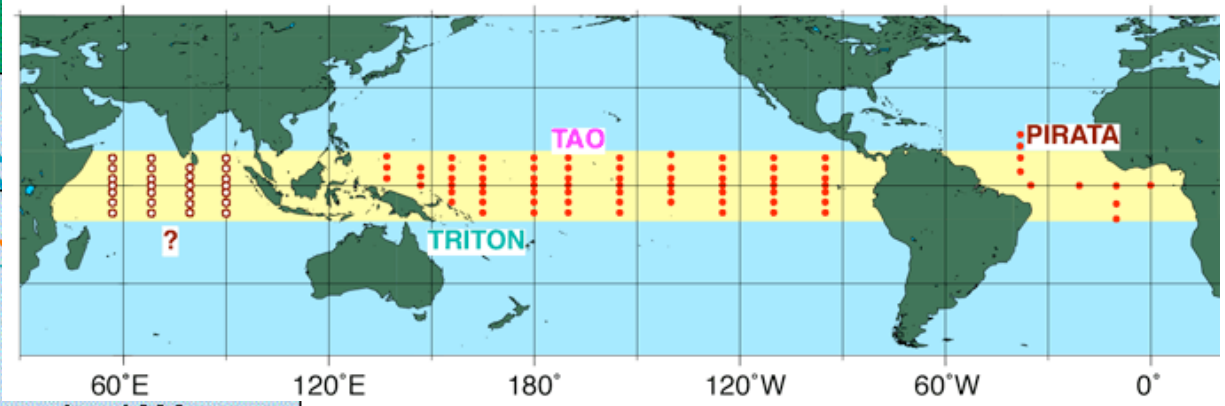
Standard ATLAS Mooring



TOGA In Situ Ocean Observing System Global Tropics



Global Tropical Moored Buoy Array



- 

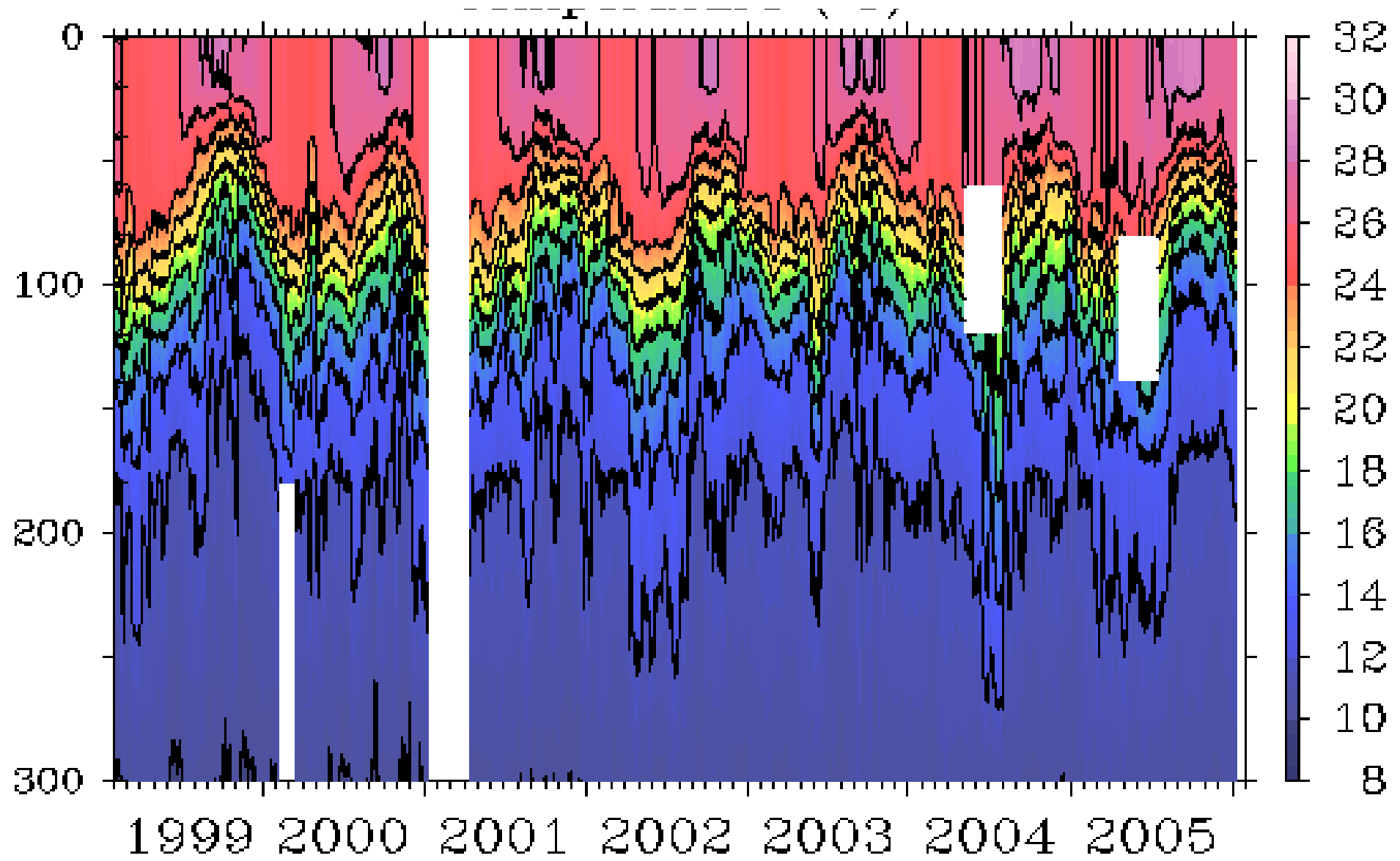
moored buoy
- 

tide gauge station
- 

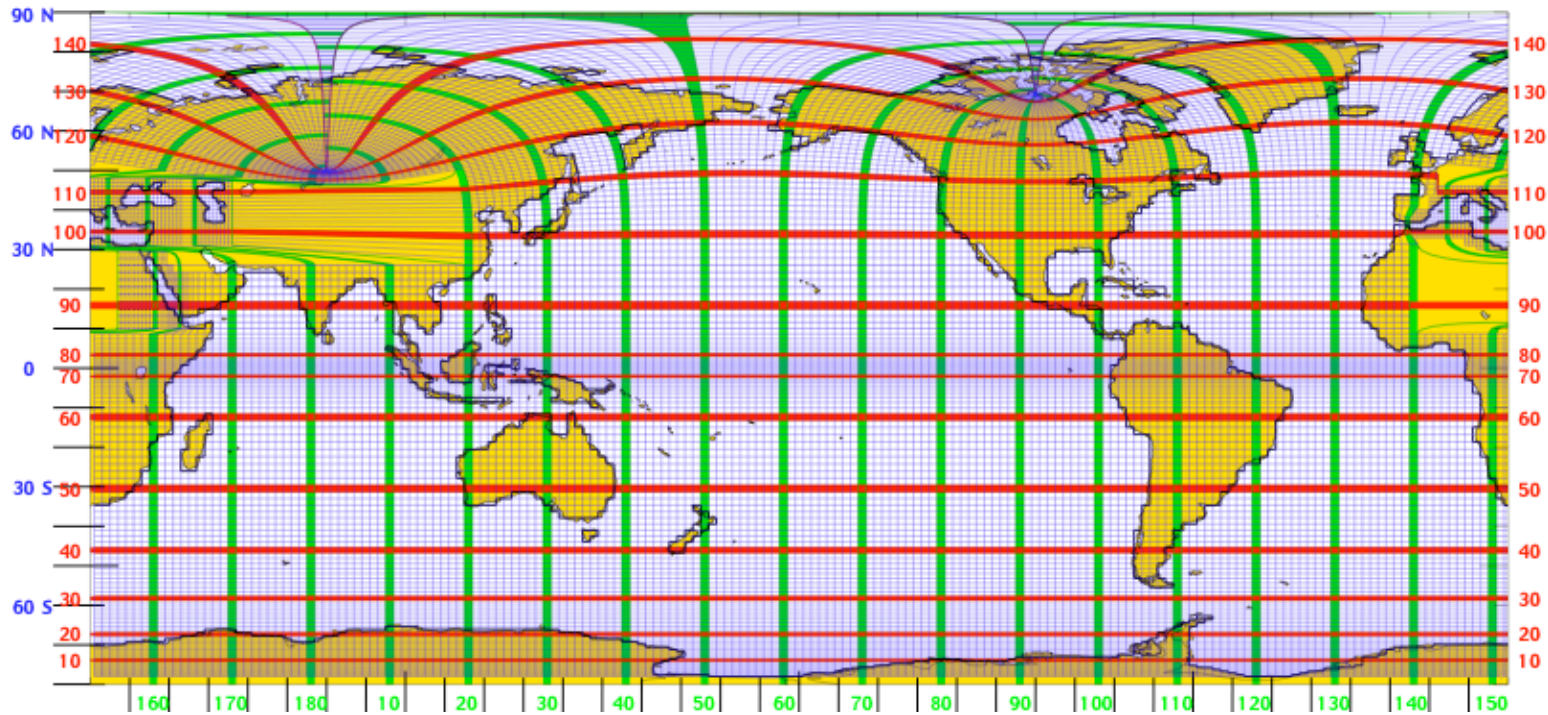
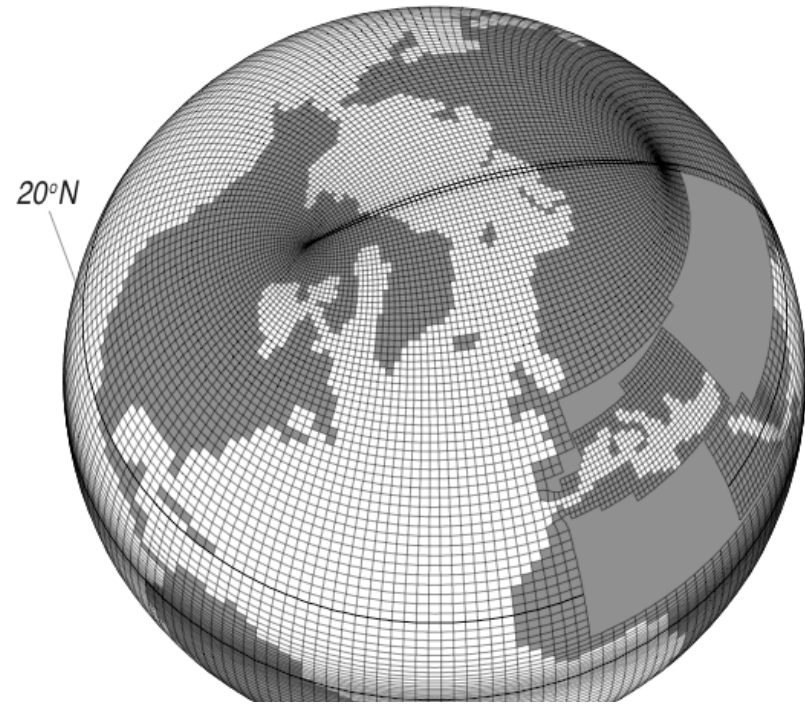
satellite data relay
- 

drifting buoy
- 

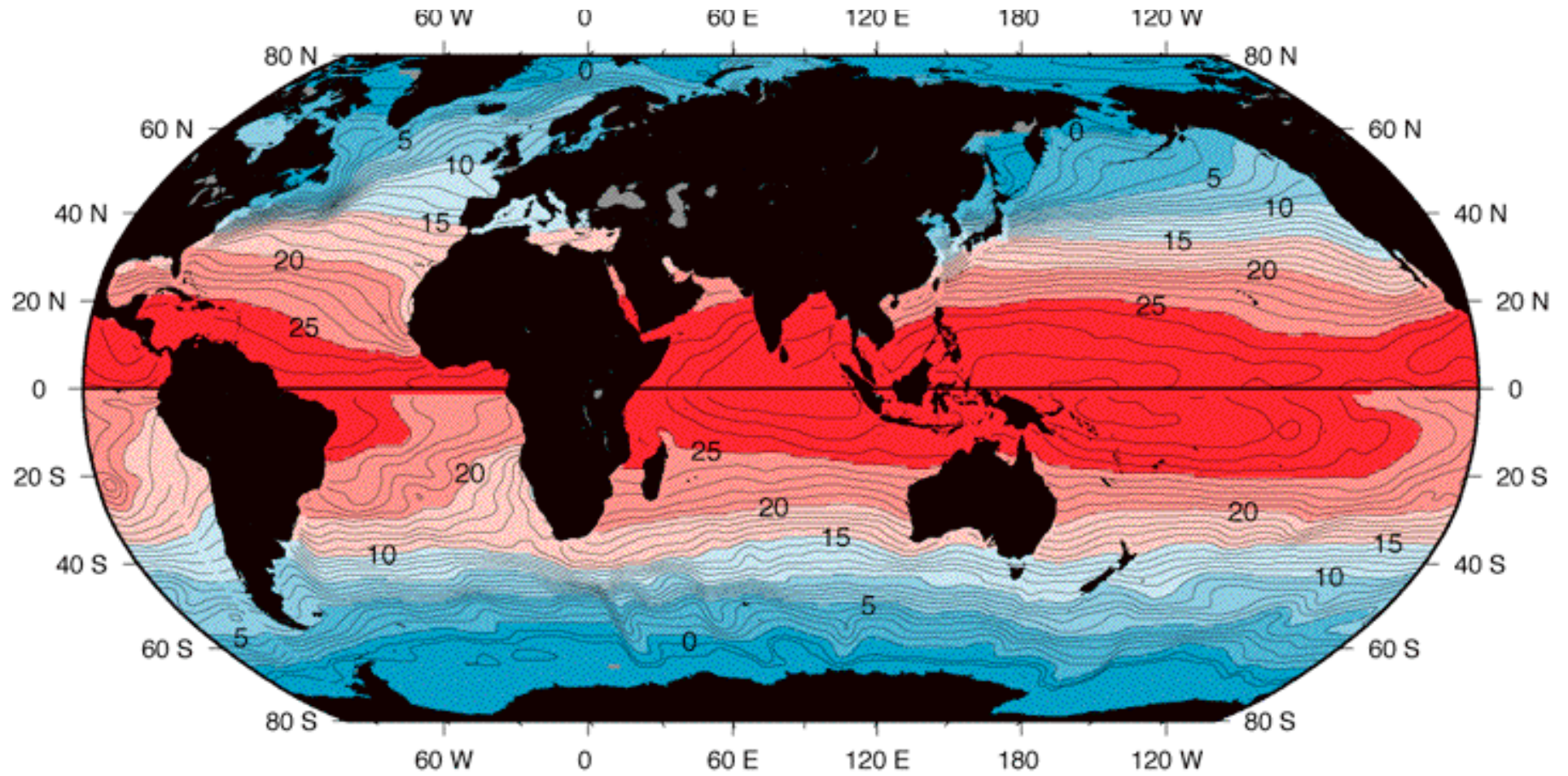
vessel carrying an SVP



L'apparition des MCGO

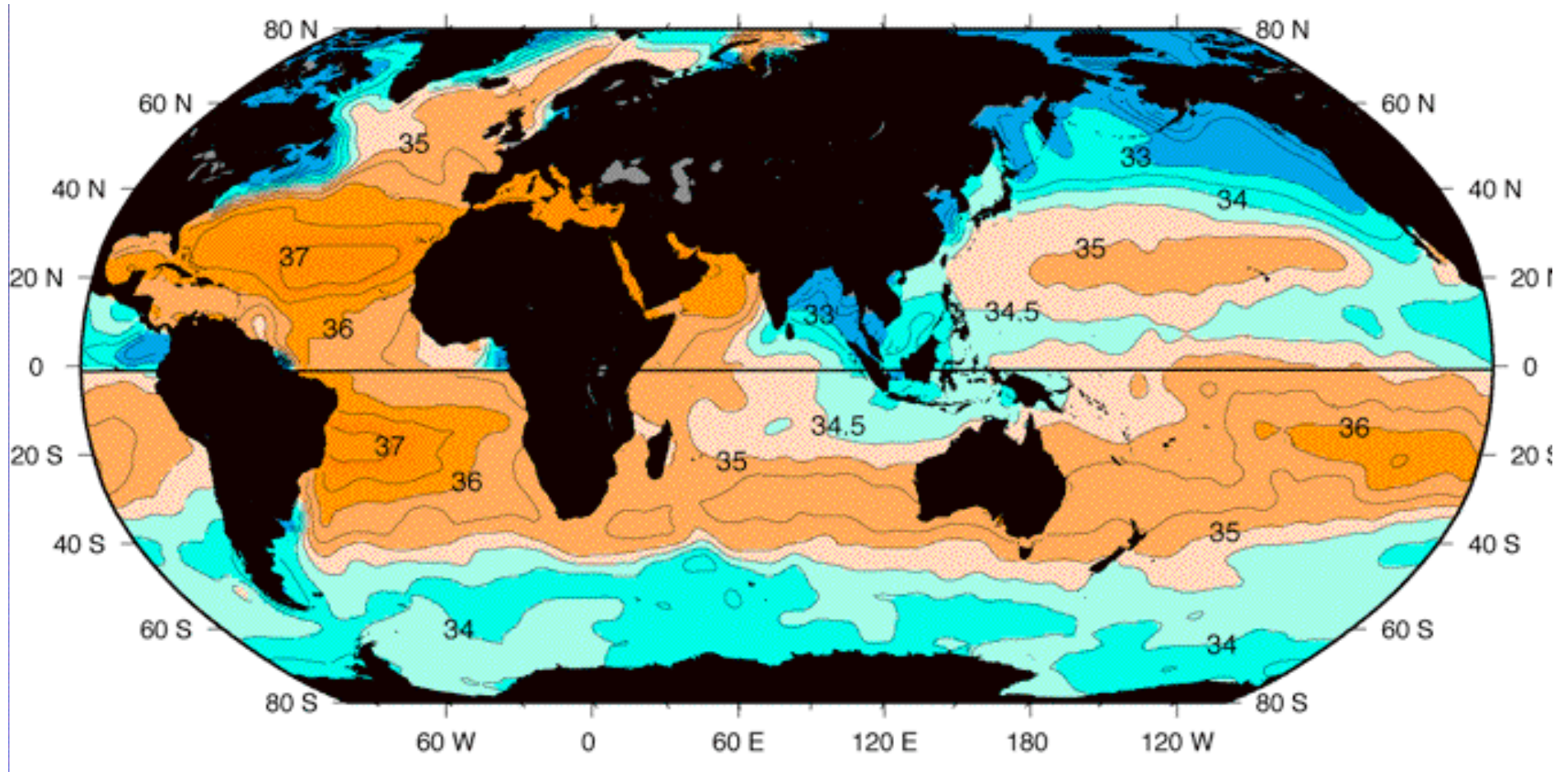


Apperçu des distributions et structures associées aux océans



SST

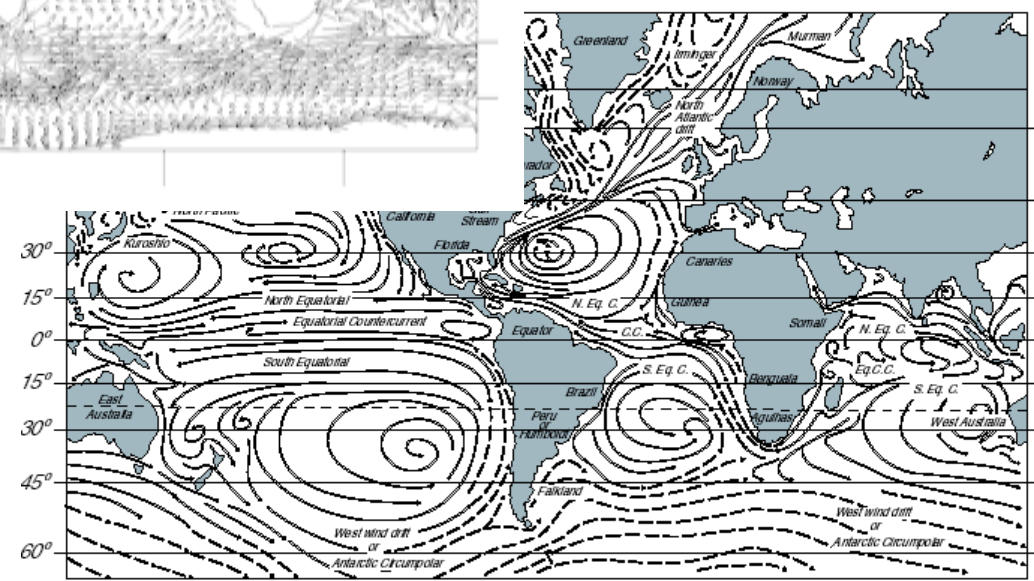
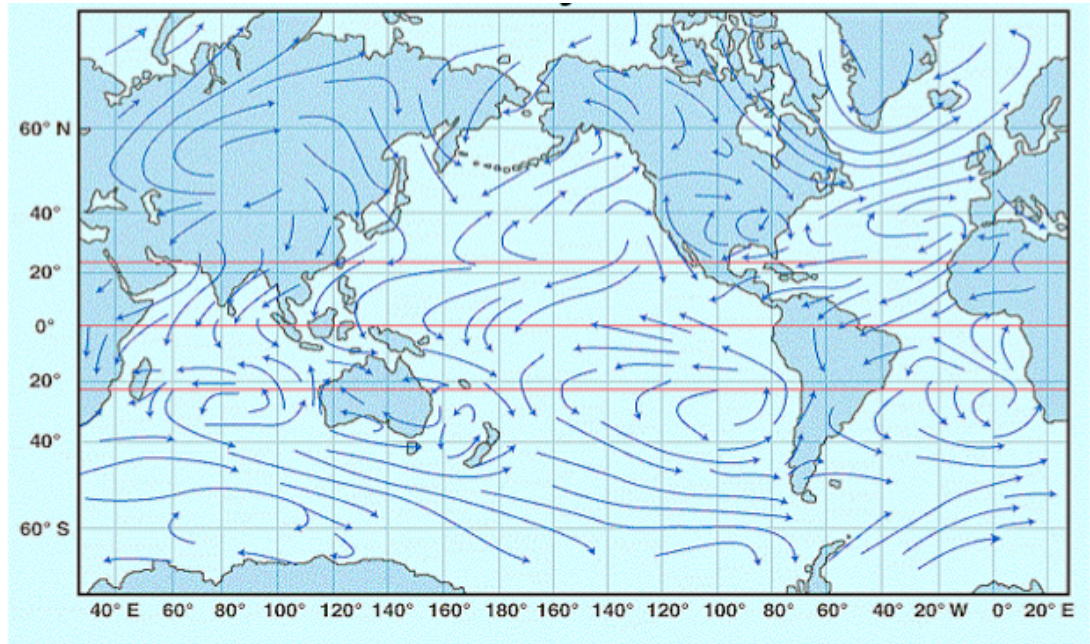
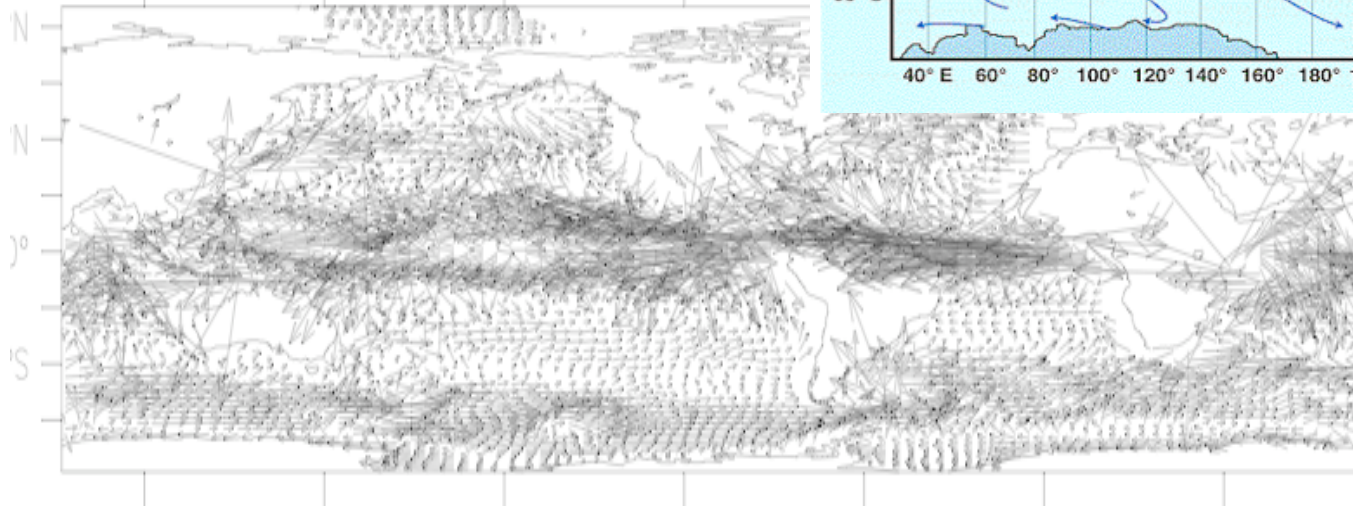
Apperçu des distributions et structures associées aux océans



SSS

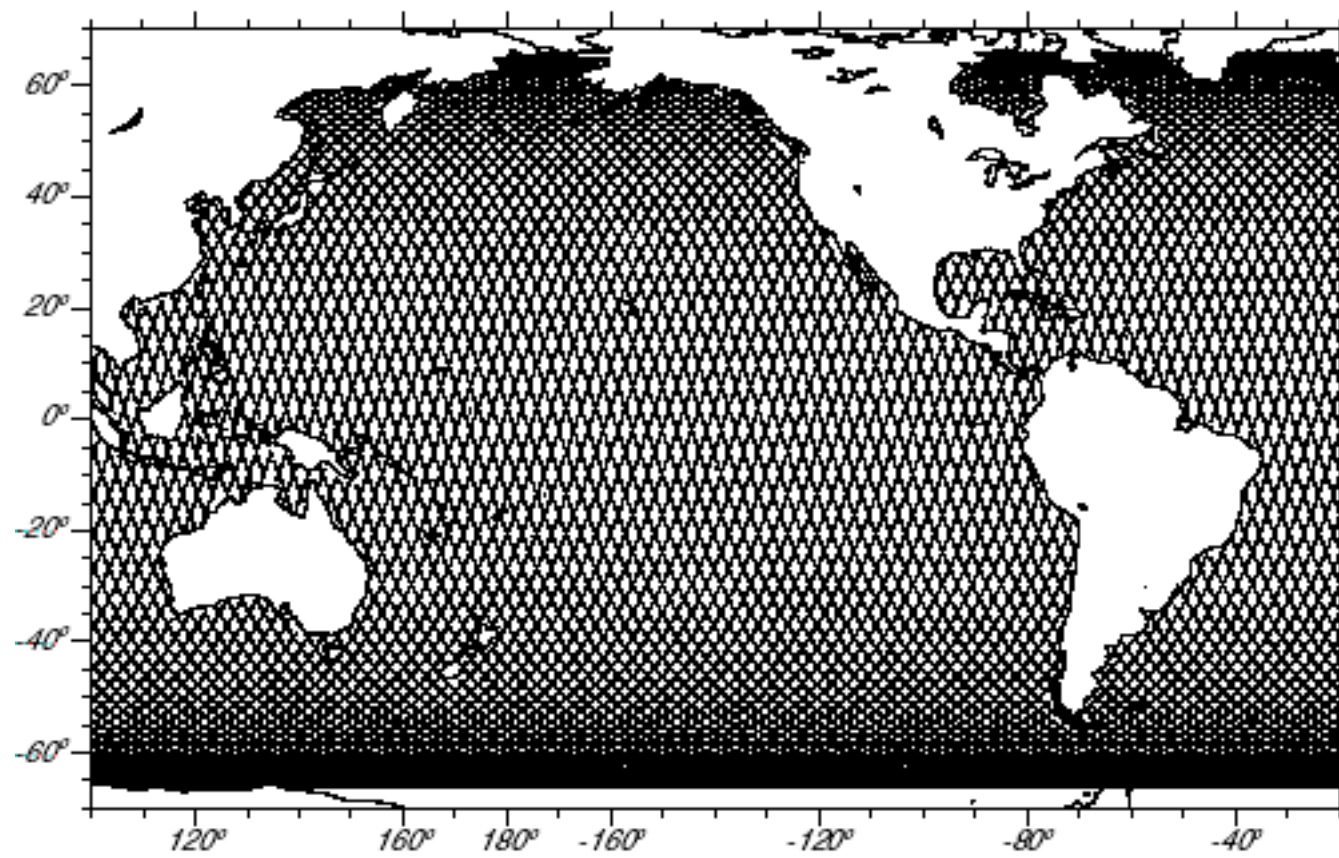
juillet

100 L 100 W → 0.200

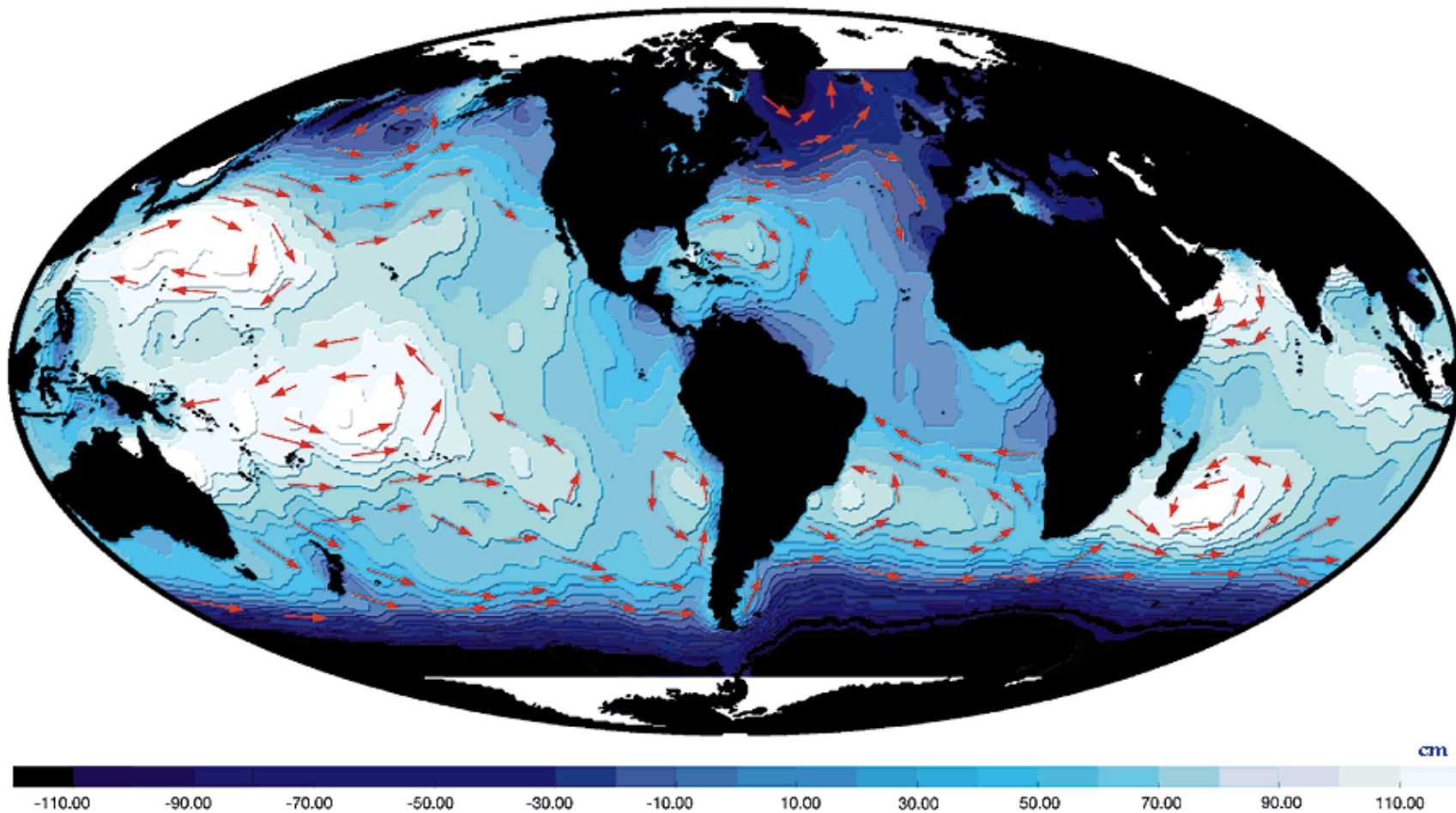


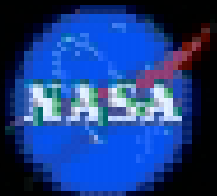
Satellites

1978 -

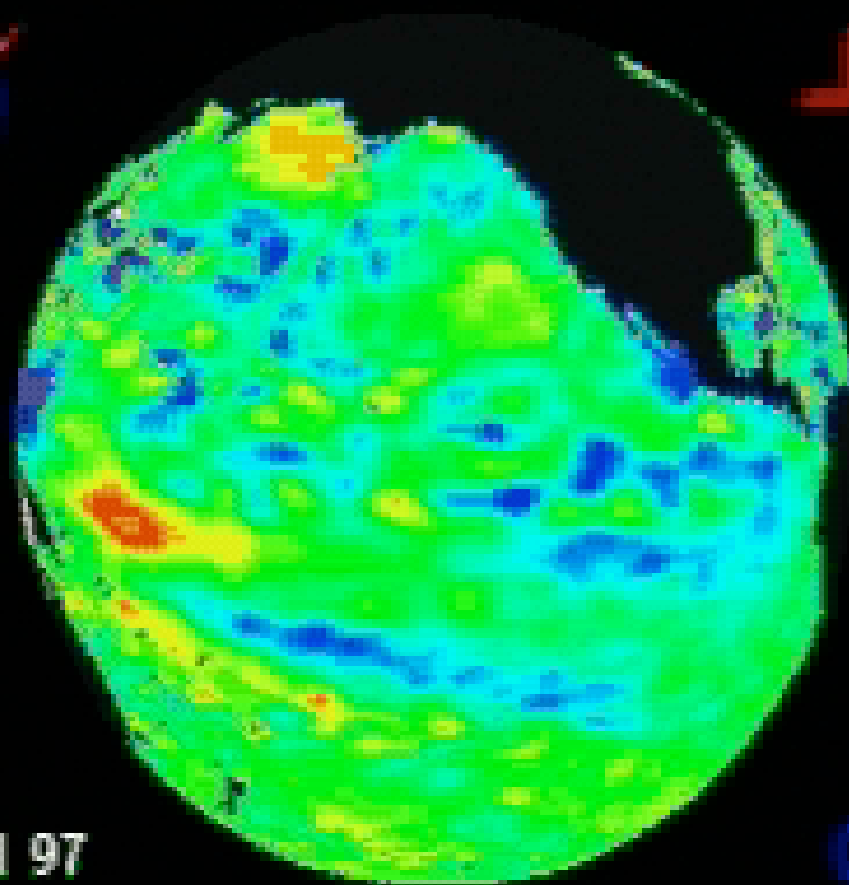


Topographie dynamique de la surface de la mer observée par TOPEX/POSEIDON



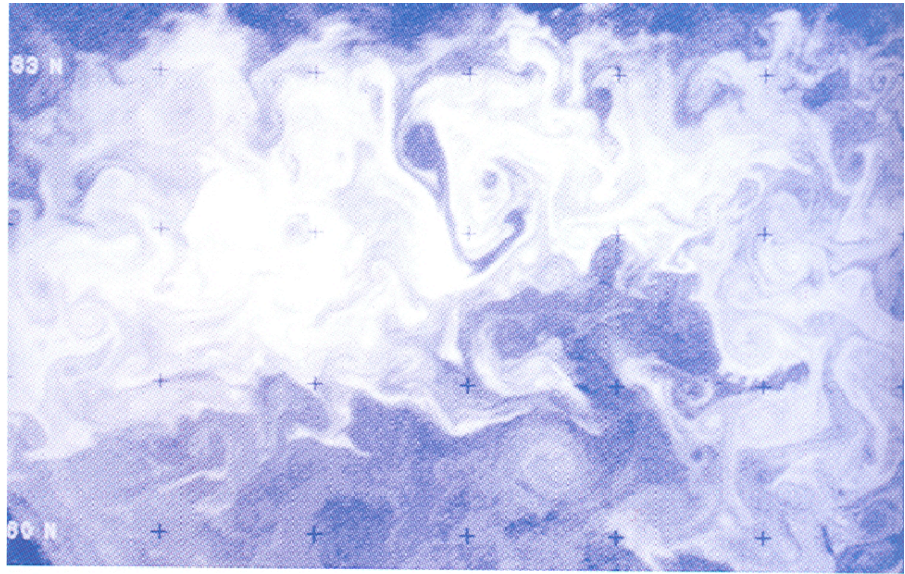


JPL

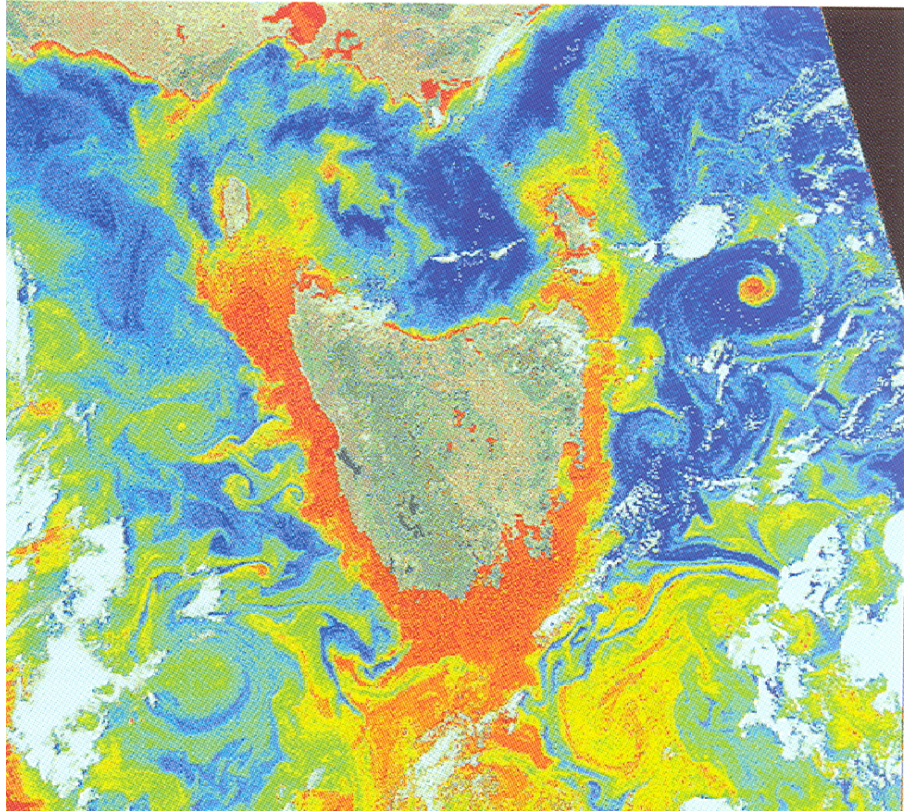


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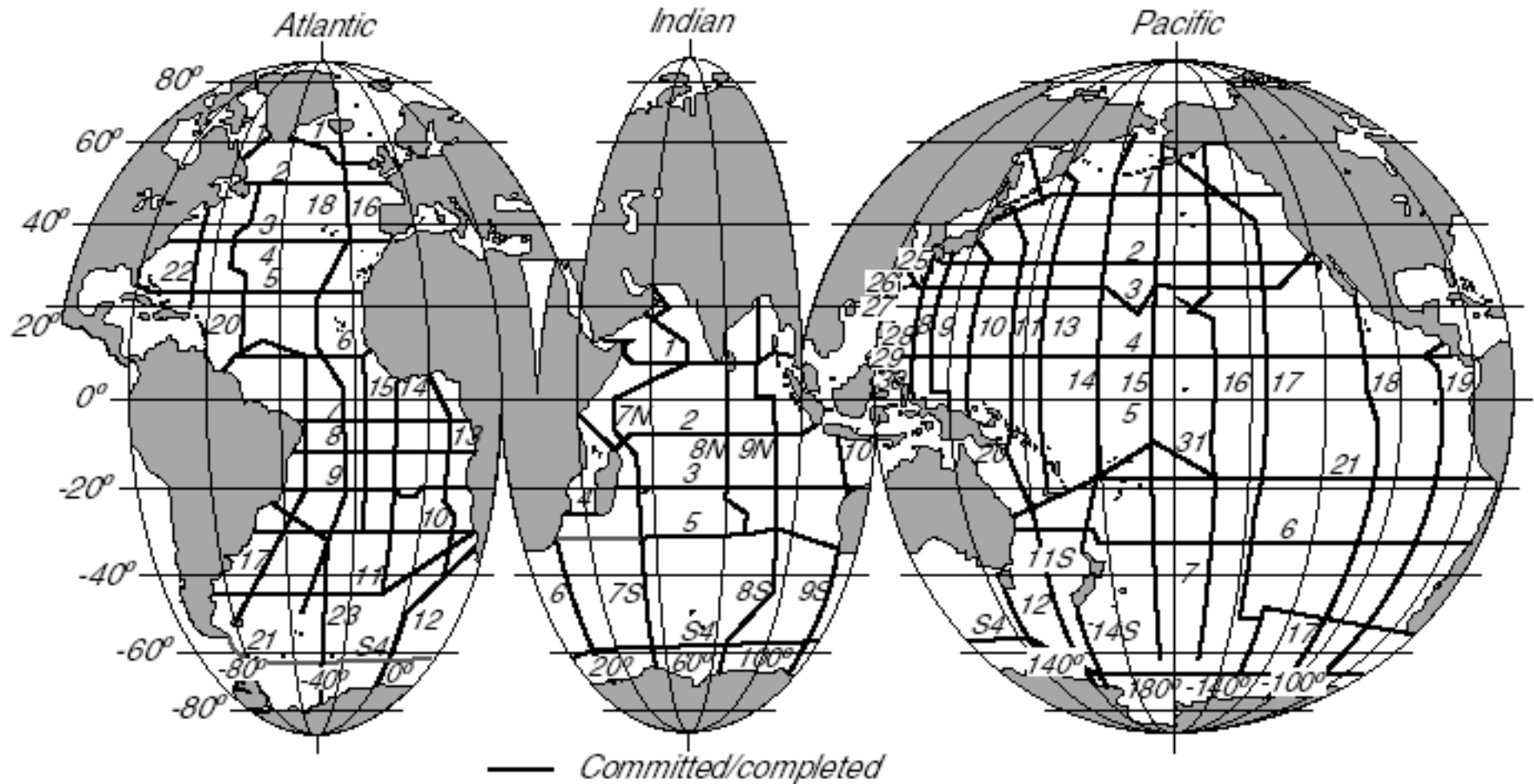




(b)



Campagnes de coopérations internationales

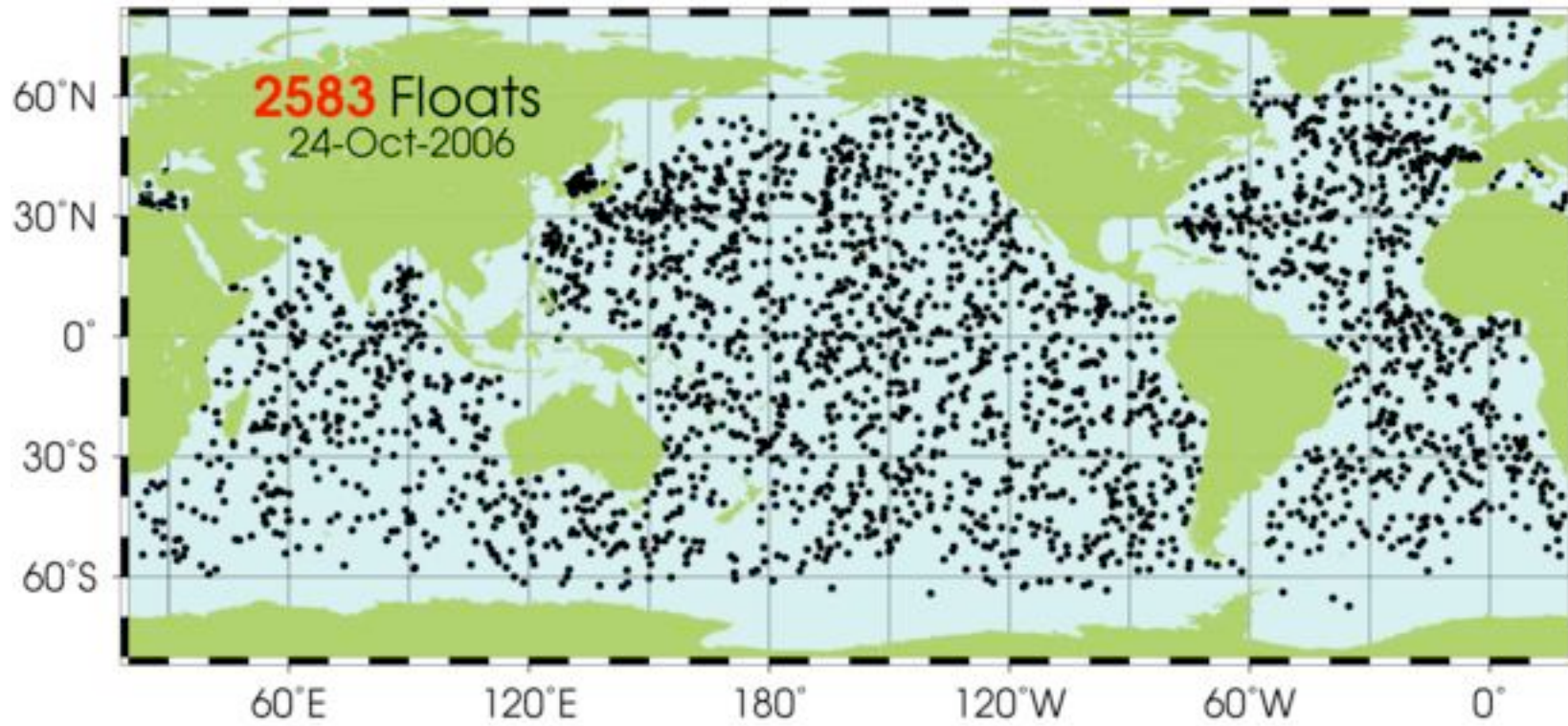


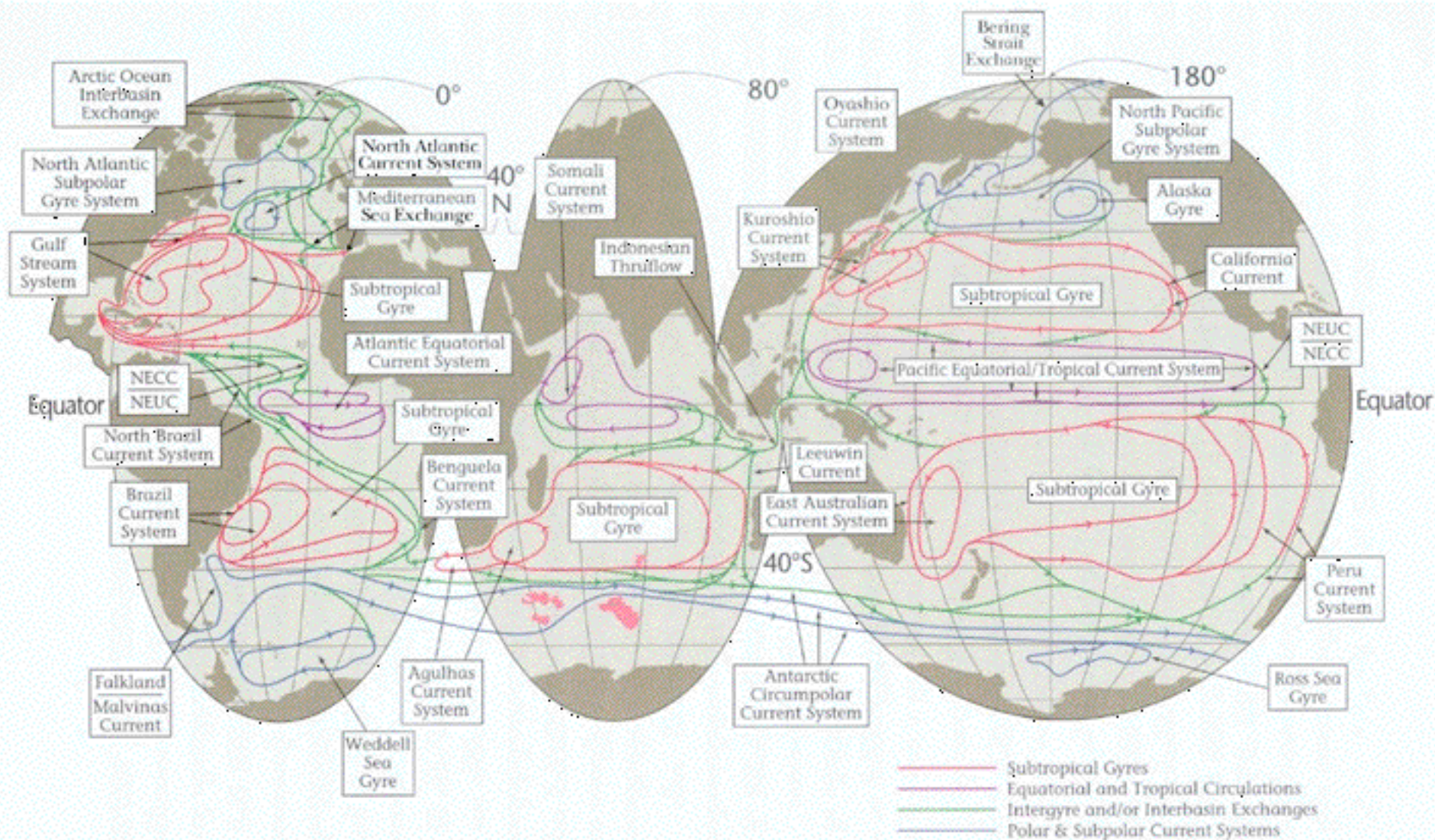
WOCE, mesures :1990-1997

ARGO
2000-



ARGO
2000-

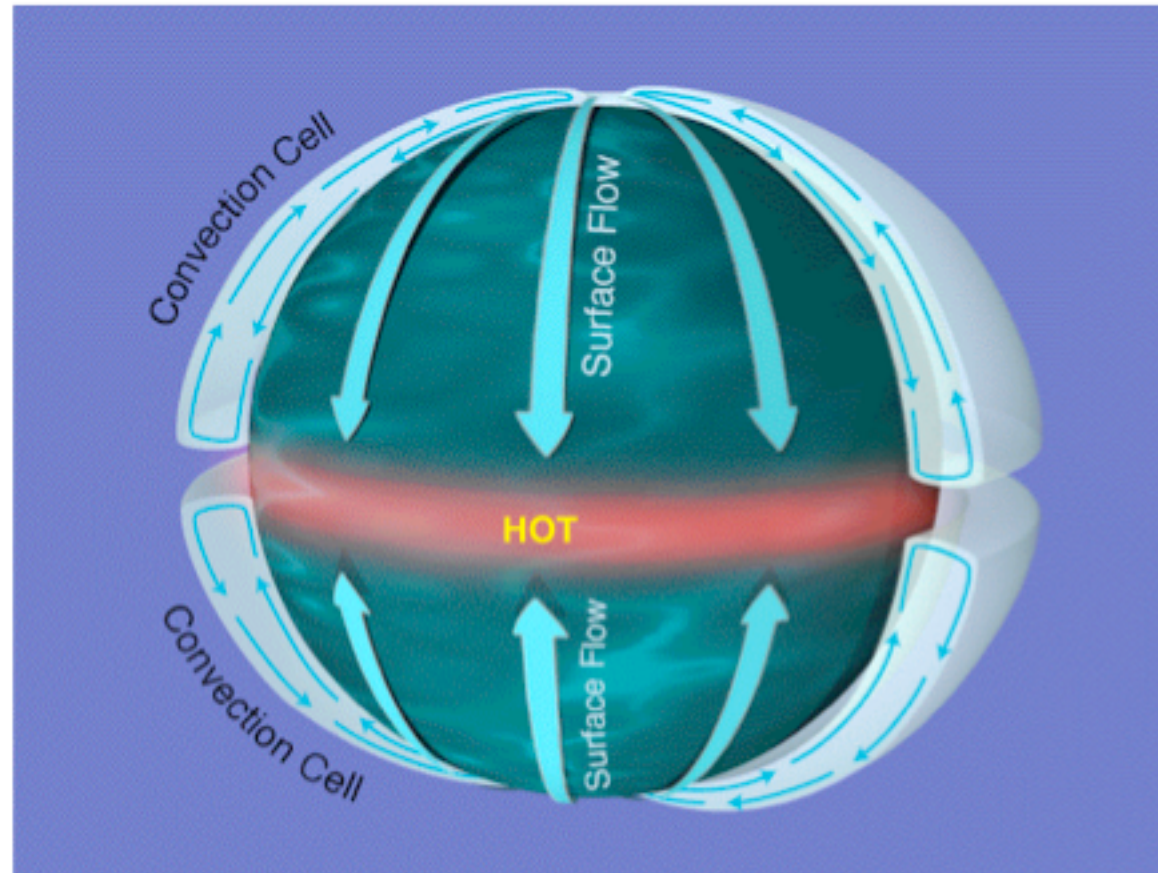




Cours II

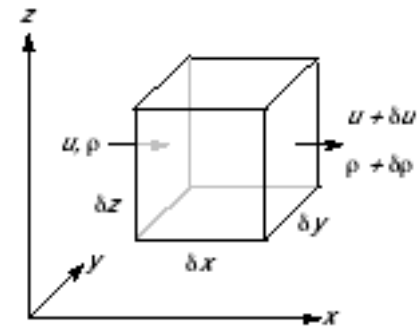
Les équations du mouvement & la géostrophie

Circulation attendue avec chauffage différentiel solaire
sans rotation ni continents



Équations du mouvement des fluides non tournant

$$\begin{aligned} \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} + \frac{1}{\rho} \frac{\partial p}{\partial x} &= \mathcal{F}_x \\ \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + \frac{1}{\rho} \frac{\partial p}{\partial y} &= \mathcal{F}_y \\ \frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} + \frac{1}{\rho} \frac{\partial p}{\partial z} + g &= \mathcal{F}_z \end{aligned}$$



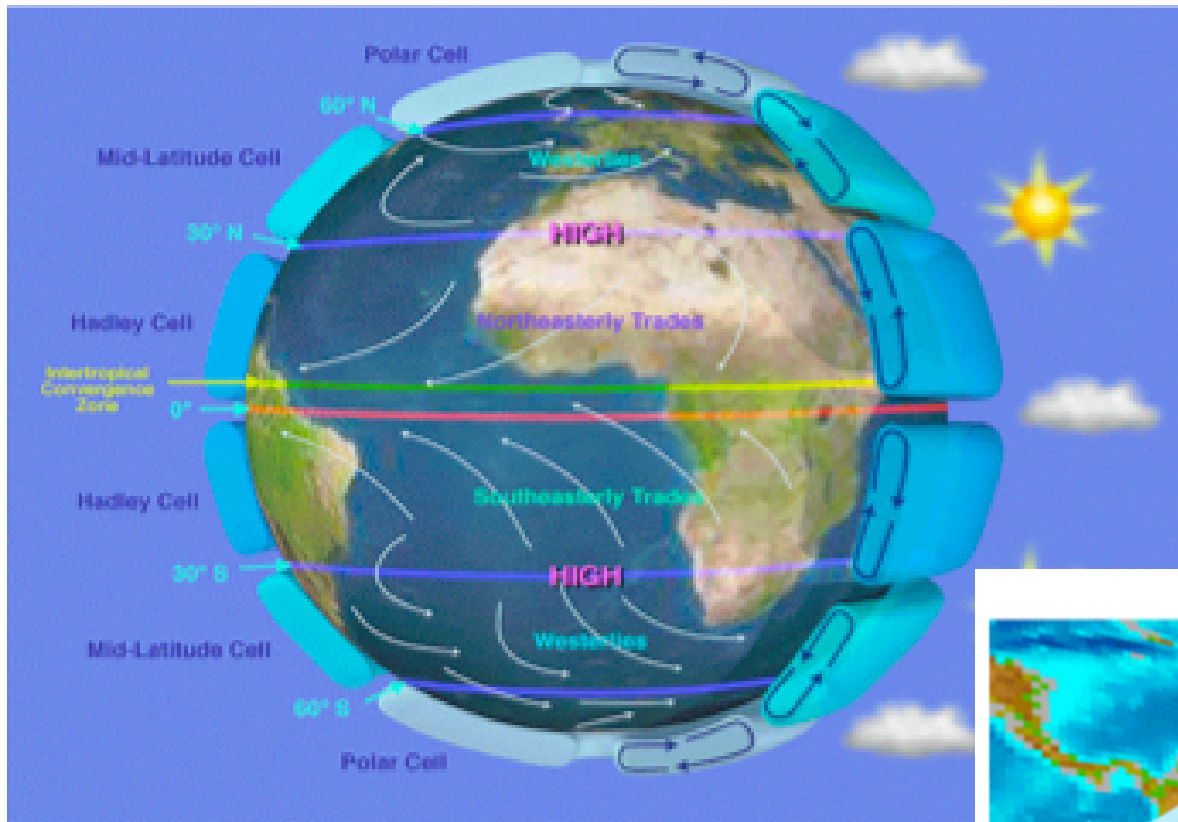
$$\rho \delta x \delta y \delta z \frac{D\mathbf{u}}{Dt} = \mathbf{F}_{\text{gravité}} + \mathbf{F}_{\text{pression}} + \mathbf{F}_{\text{friction}}$$

$$\frac{D\mathbf{u}}{Dt} = -\frac{1}{\rho} \nabla p - g\hat{\mathbf{z}} + \mathcal{F}$$

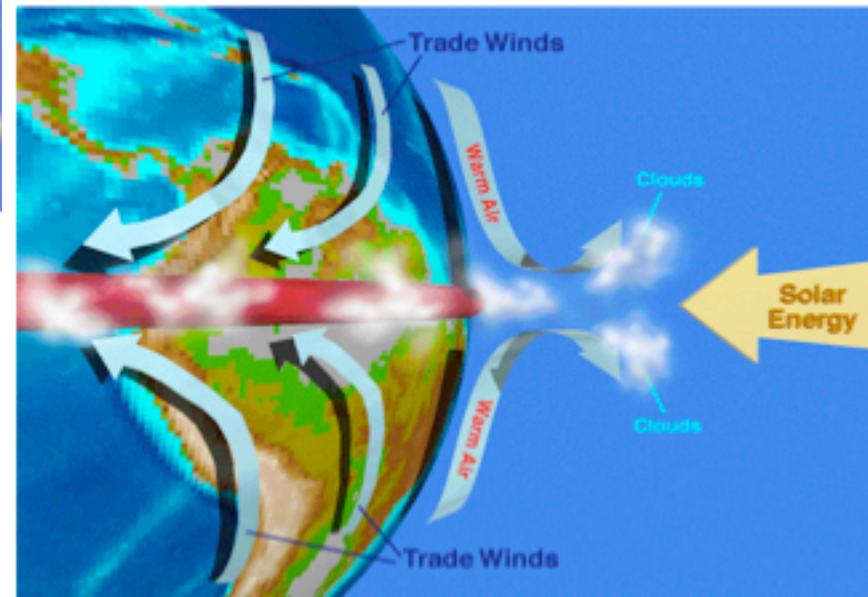
$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho \mathbf{u}) = 0 \quad \frac{D\rho}{Dt} + \rho \nabla \cdot \mathbf{u} = 0$$

Incompressibilité : $\nabla \cdot \mathbf{u} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$

Circulation schématique à grande-échelle observée



Interpétation:
les cellules et le champ de pression

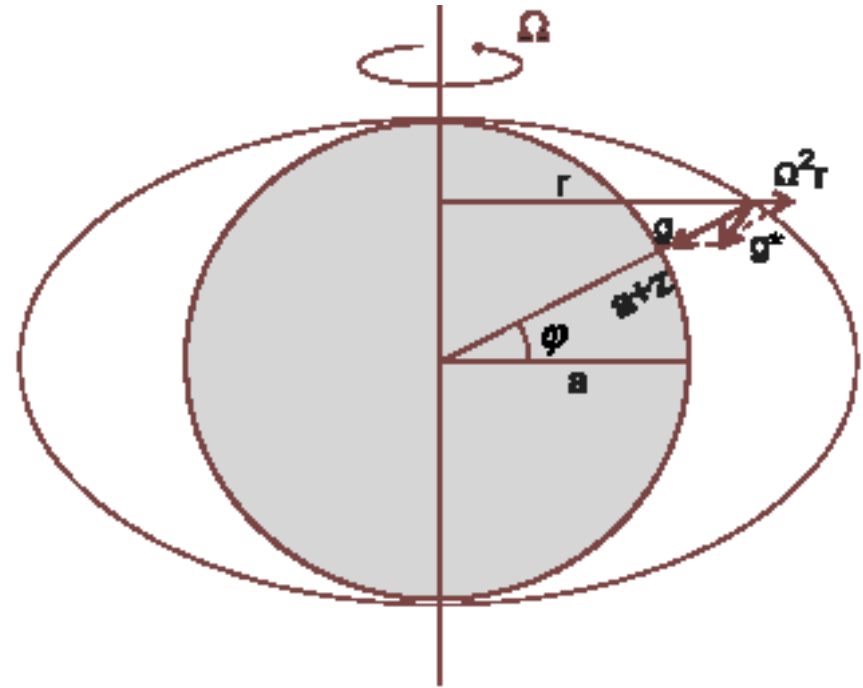


Équations du mouvement des fluides géophysiques

$$\frac{D\mathbf{u}}{Dt} + \frac{1}{\rho}\nabla p + g\hat{\mathbf{z}} = \underbrace{-2\boldsymbol{\Omega} \times \mathbf{u}}_{\text{Coriolis accel}^n} + \underbrace{-\boldsymbol{\Omega} \times \boldsymbol{\Omega} \times \mathbf{r}}_{\text{Centrifugal accel}^n} + \mathcal{F}$$

$$\frac{D\mathbf{u}}{Dt} + \frac{1}{\rho}\nabla p + \nabla\phi = -2\boldsymbol{\Omega} \times \mathbf{u} + \mathcal{F}$$

$$\phi = gz - \frac{\Omega^2 r^2}{2}$$

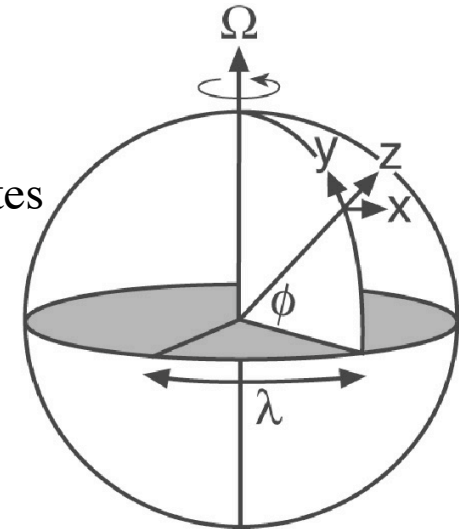


Effet de la rotation terrestre

Equilibre Géostrophique

Expérimentation et Interprétation

- argument de conservation du moment angulaire et limites
=> le plan tangent
- Expérience sur plateau tournant
- Analyse en ordres de grandeur
Interprétation et ex.



mouvements océaniques de grande échelle :
Simplification de l'équation sur la verticale

$$\begin{aligned} \frac{Du}{Dt} + \frac{1}{\rho} \frac{\partial p}{\partial x} - fv &= \mathcal{F}_x, \\ \frac{Dv}{Dt} + \frac{1}{\rho} \frac{\partial p}{\partial y} + fu &= \mathcal{F}_y, \\ \frac{1}{\rho} \frac{\partial p}{\partial z} + g &= 0, \end{aligned}$$

équilibre géostrophique :

$$\frac{D\mathbf{u}}{Dt} + f\hat{\mathbf{z}} \times \mathbf{u} = \frac{\partial \mathbf{u}}{\partial t} + \underbrace{\mathbf{u} \cdot \nabla \mathbf{u}}_{\frac{u^2}{L}} + \underbrace{f\hat{\mathbf{z}} \times \mathbf{u}}_{fu}$$

$$f\hat{\mathbf{z}} \times \mathbf{u} + \frac{1}{\rho} \nabla p = 0.$$

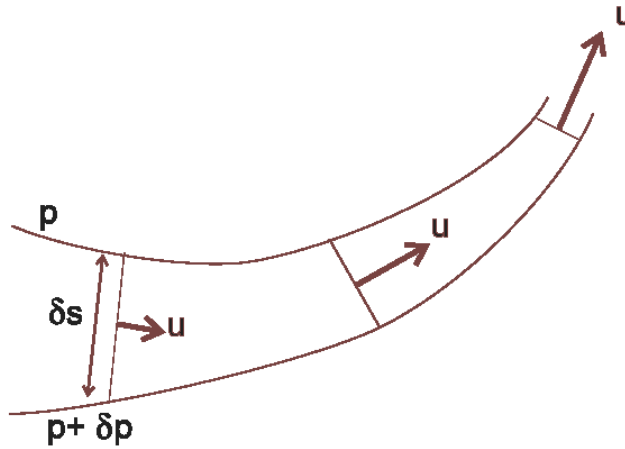
latitude	$f (\times 10^{-4} \text{ s}^{-1})$	$\beta (\times 10^{-11} \text{ s}^{-1} \text{ m}^{-1})$
90°	1.46	0
60°	1.26	1.14
45°	1.03	1.61
30°	0.73	1.98
10°	0.25	2.25
0°	0	2.28

Applications :

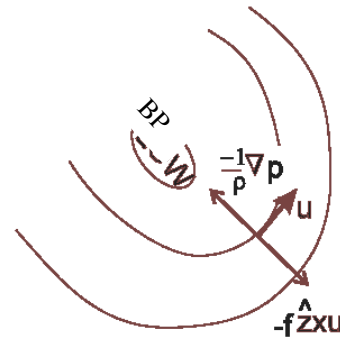
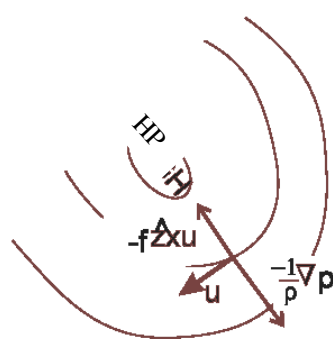
Circulation géostrophique dans
Hémisphère nord

$$\mathbf{u}_g = \frac{1}{f\rho} \hat{\mathbf{z}} \times \nabla p$$

$$(u_g, v_g) = \left(-\frac{1}{f\rho} \frac{\partial p}{\partial y}, \frac{1}{f\rho} \frac{\partial p}{\partial x} \right)$$

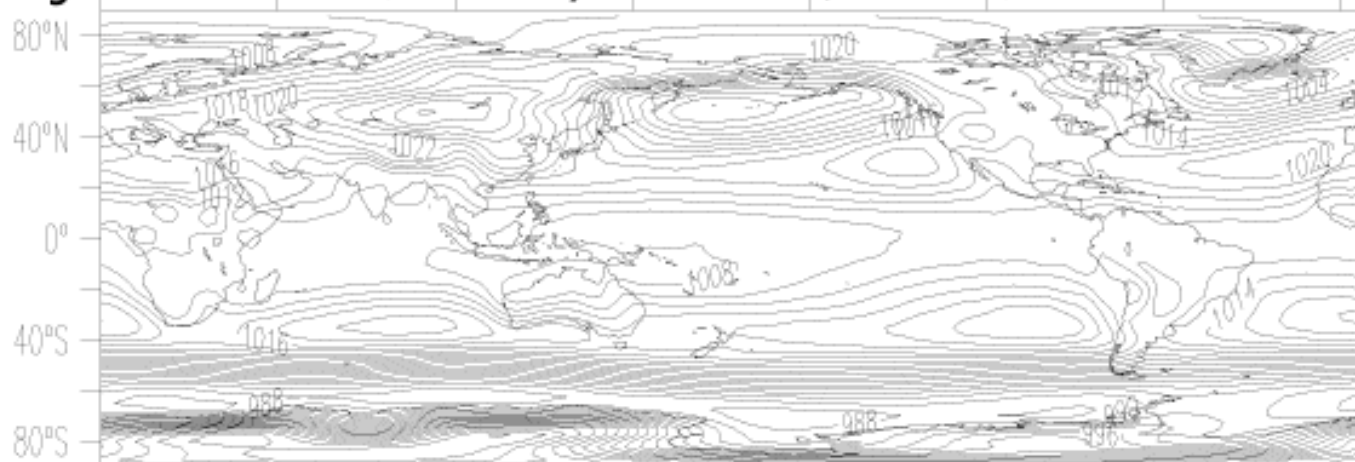


$$|\mathbf{u}_g| = \frac{1}{f\rho} |\nabla p| = \frac{1}{f\rho} \frac{\delta p}{\delta s}$$



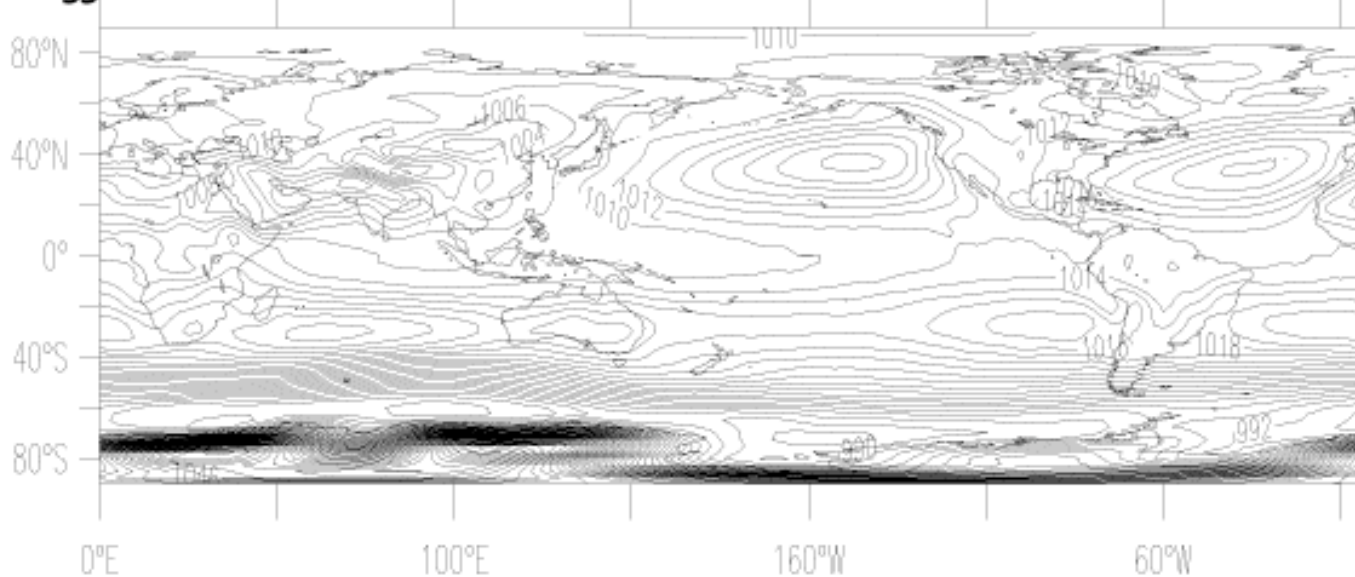
Pression à la surface de la mer (réanalyse NCEP, 1998)

JFM



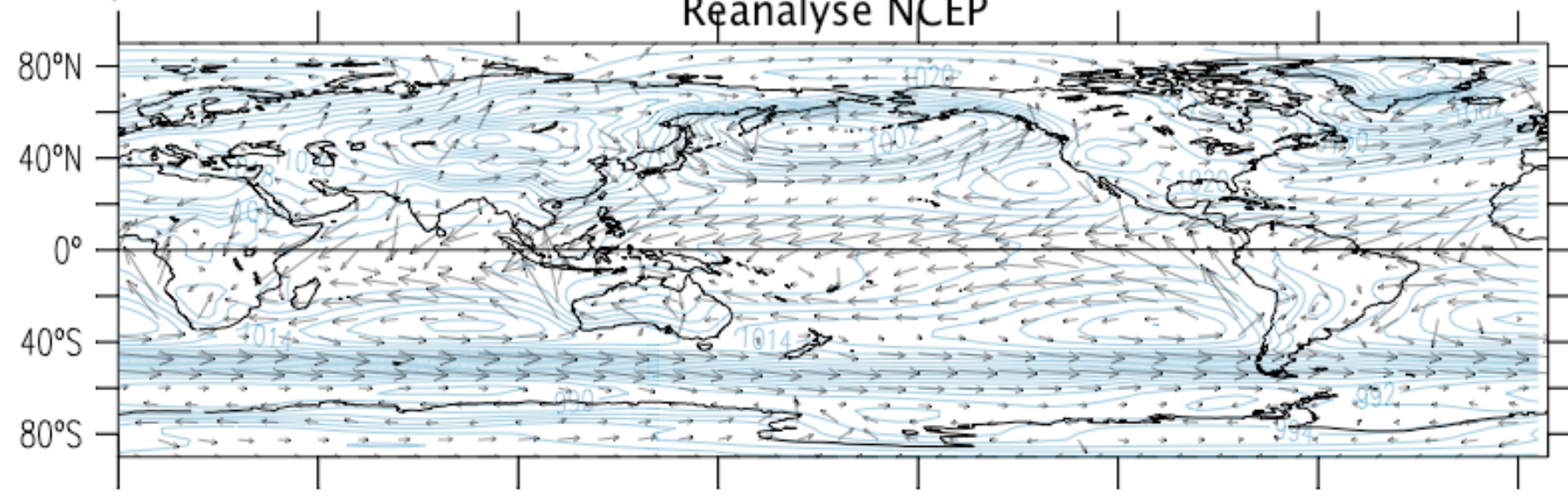
Exercice: Tracer les vecteurs vent géostrophique de surface avec précision (direction, taille)
Discuter des écarts attendus à la géostrophie et de la variabilité saisonnière

JJA



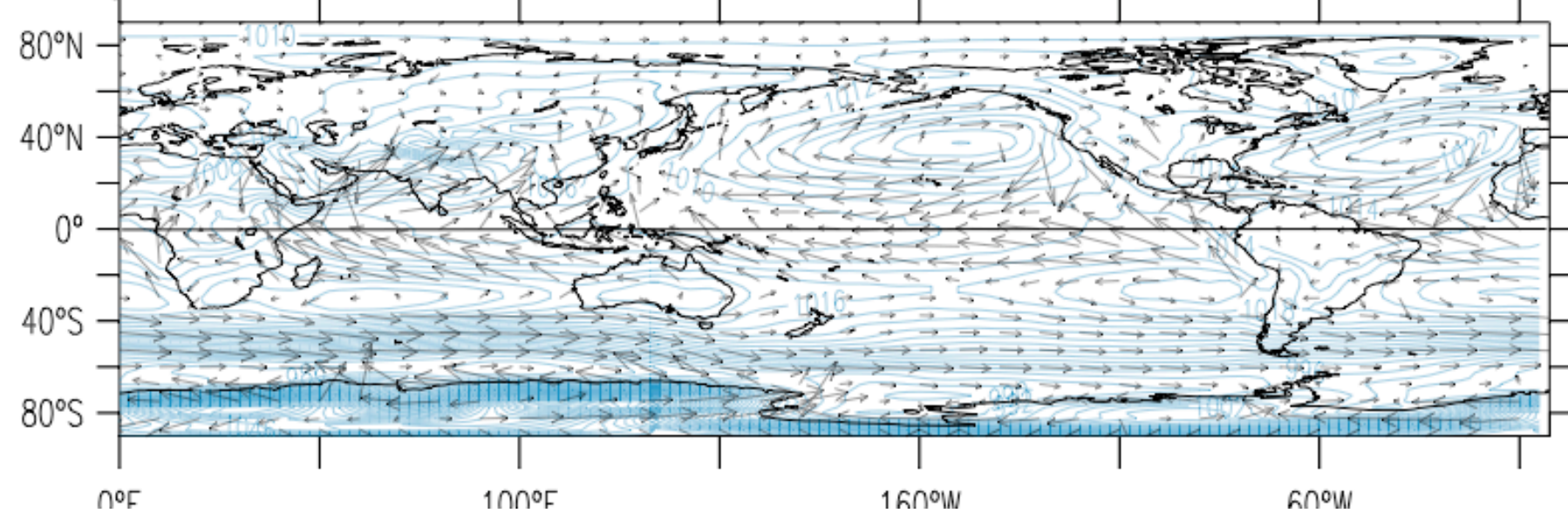
janvier

Régime de surface des vents et des pressions Réanalyse NCEP



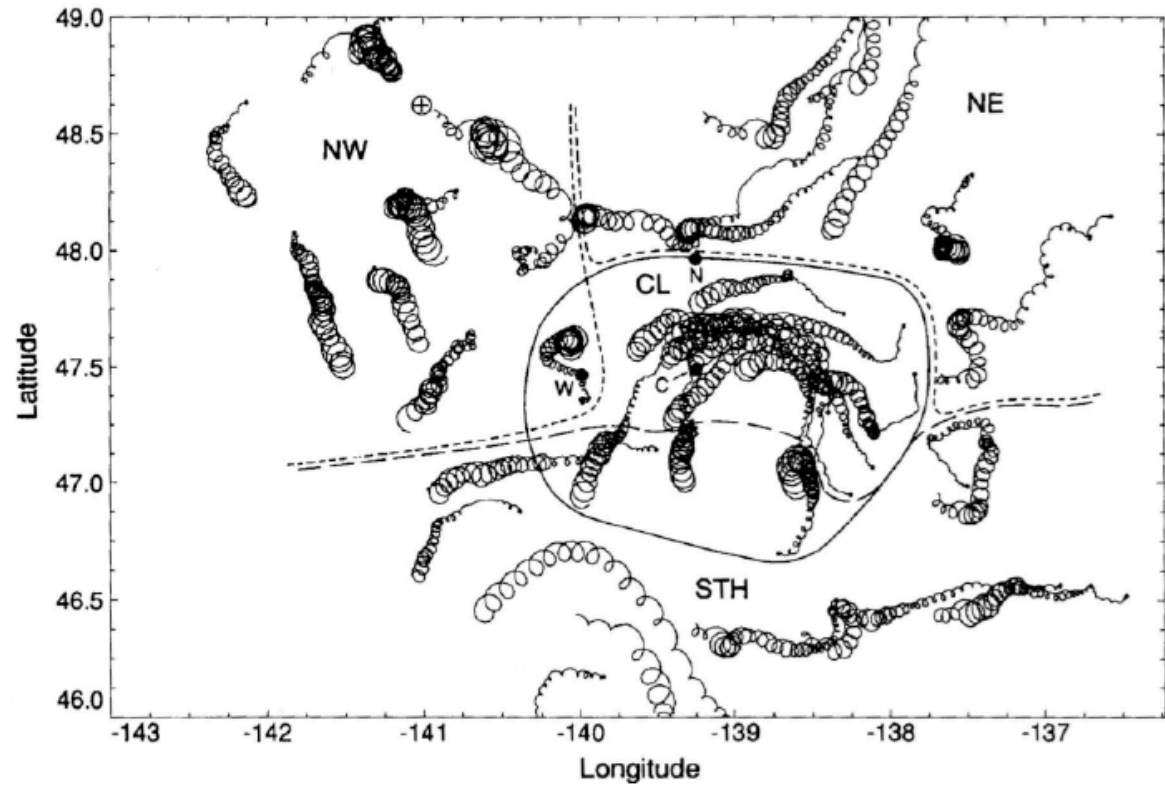
juillet

8 m/s



oscillations d'inertie

$$\frac{du}{dt} = -\frac{1}{f} \frac{d^2v}{dt^2} = fv$$



A.N.: V=20cm

Lat Période

Diamètre

90°

35°

10°

oscillations d'inertie

$$\frac{du}{dt} = -\frac{1}{f} \frac{d^2v}{dt^2} = fv$$

$$\frac{d^2v}{dt^2} + f^2v = 0$$

$$u = V \sin ft$$

$$v = V \cos ft$$

$$V^2 = u^2 + v^2$$

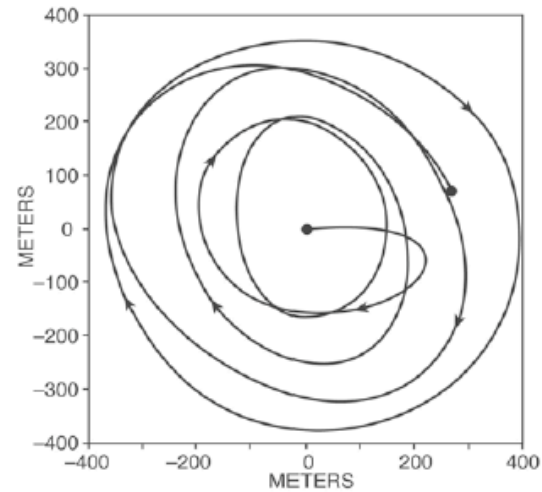


Figure 6.15: Inertial circles observed by a current metre in the main thermocline of the Atlantic Ocean at a depth of 500 m; 28°N, 54°W. Five inertial periods are shown. The inertial period at this latitude is 25.6 h and 5 inertial periods are shown. Courtesy of Carl Wunsch.

$$Ti=2\pi/f$$

$$Ri=V/f$$

V=20cm

Lat	Période(h)	Diamètre (km)
90°	11.97	2.7
35°	20.87	4.8
10°	68.93	15.8

Cours III

Friction et effet du vent

Force de friction turbulente dans le fluide

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + 2\Omega v \sin \vartheta + \frac{\partial}{\partial x} \left[\nu \frac{\partial u}{\partial x} \right] + \frac{\partial}{\partial y} \left[\nu \frac{\partial u}{\partial y} \right] + \frac{\partial}{\partial z} \left[\nu \frac{\partial u}{\partial z} \right] \quad (1)$$

$$u = U + u'; \quad v = V + v'; \quad w = W + w'; \quad p = P + p'$$

$$U = \langle u \rangle = \frac{1}{T} \int_0^T u(t) dt \quad \text{or} \quad U = \langle u \rangle = \frac{1}{X} \int_0^X u(x) dx$$

$$\begin{aligned} \left\langle (U + u') \frac{\partial(U + u')}{\partial x} \right\rangle &= \left\langle U \frac{\partial U}{\partial x} \right\rangle + \left\langle U \frac{\partial u'}{\partial x} \right\rangle + \left\langle u' \frac{\partial U}{\partial x} \right\rangle + \left\langle u' \frac{\partial u'}{\partial x} \right\rangle \\ \left\langle (U + u') \frac{\partial(U + u')}{\partial x} \right\rangle &= \left\langle U \frac{\partial U}{\partial x} \right\rangle + \left\langle u' \frac{\partial u'}{\partial x} \right\rangle \end{aligned}$$

$$\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \frac{\partial W}{\partial z} + \frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} = 0 \quad (2)$$

$$(2) - \langle (2) \rangle \Rightarrow \begin{aligned} \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \frac{\partial W}{\partial z} &= 0 \\ \frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} &= 0 \end{aligned} \quad \text{Équations de continuité à grande et petite échelle}$$

$\langle (1) \rangle \Rightarrow$

$$\begin{aligned} \frac{DU}{Dt} &= -\frac{1}{\rho} \frac{\partial P}{\partial x} + 2\Omega V \sin \varphi \\ &+ \frac{\partial}{\partial x} \left[\nu \frac{\partial U}{\partial x} - \langle u'u' \rangle \right] + \frac{\partial}{\partial y} \left[\nu \frac{\partial U}{\partial y} - \langle u'v' \rangle \right] + \frac{\partial}{\partial z} \left[\nu \frac{\partial U}{\partial z} - \langle u'w' \rangle \right] \end{aligned}$$

$$F_x = -\frac{\partial}{\partial x} \langle u'u' \rangle - \frac{\partial}{\partial y} \langle u'v' \rangle - \frac{\partial}{\partial z} \langle u'w' \rangle$$

Cas bi-dimensionnel stationnaire (x,z)

$$2\Omega V \sin \varphi + \frac{\partial}{\partial z} \left[\nu \frac{\partial U}{\partial z} - \langle u'w' \rangle \right] = 0$$

Analogie avec la tension moléculaire $T_{xz} = \rho \nu \frac{\partial u}{\partial z}$

$$-\rho \langle u'w' \rangle = T_{xz} = \rho A_z \frac{\partial U}{\partial z}$$

$$\rho f V + \frac{\partial T_{xz}}{\partial z} = 0$$

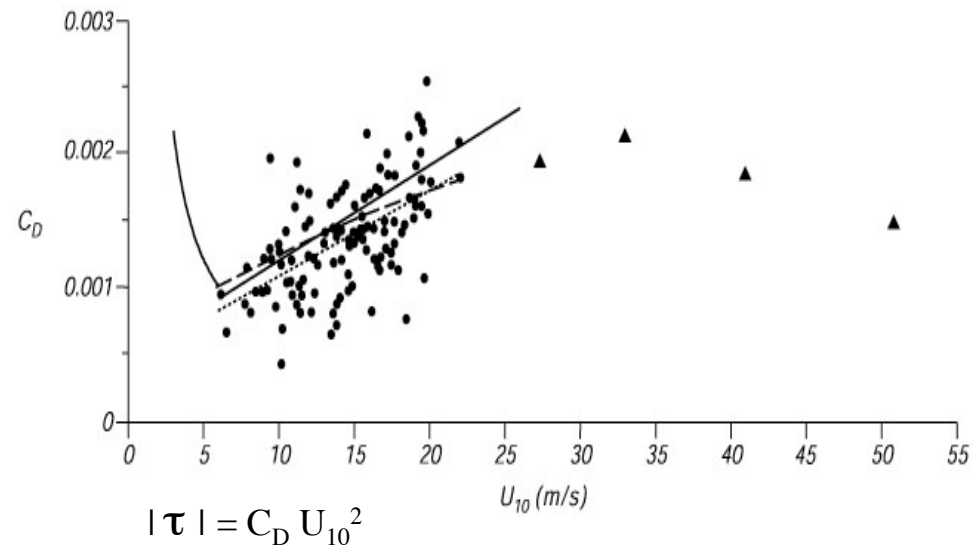
$$\rho f U - \frac{\partial T_{yz}}{\partial z} = 0$$

Dans la suite,
U est noté u

$$\begin{aligned} f v + A_z \frac{\partial^2 u}{\partial z^2} &= 0 \\ -f u + A_z \frac{\partial^2 v}{\partial z^2} &= 0 \end{aligned}$$

Couche Ekman $\langle A_z \rangle = 10^{-1} - 10^{-2} \text{ m}^2/\text{s}$

Océan intérieur $\langle A_z \rangle = 1.3 \times 10^{-4} \text{ m}^2/\text{s}$



Friction et courants d'Ekman

$$fv + A_z \frac{\partial^2 u}{\partial z^2} = 0$$

$$-fu + A_z \frac{\partial^2 v}{\partial z^2} = 0$$

Pour un vent orienté nord :

$$u = V_0 \exp(az) \cos(\pi/4 + az)$$

$$v = V_0 \exp(az) \sin(\pi/4 + az)$$

$$V_0 = \frac{T}{\sqrt{\rho_w^2 f A_z}} \quad \text{and} \quad a = \sqrt{\frac{f}{2A_z}}$$

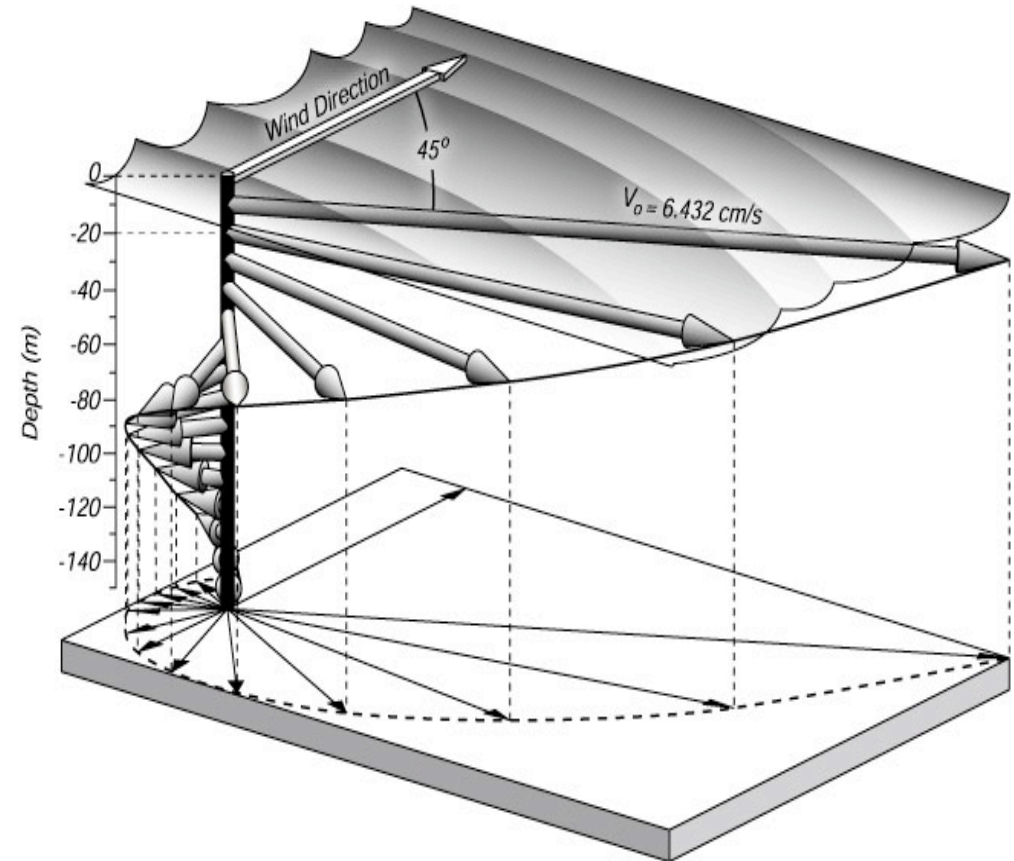
$$u(0) = V_0 \cos(\pi/4)$$

$$v(0) = V_0 \sin(\pi/4)$$

$$[u^2(z) + v^2(z)]^{1/2} = V_0 \exp(az)$$

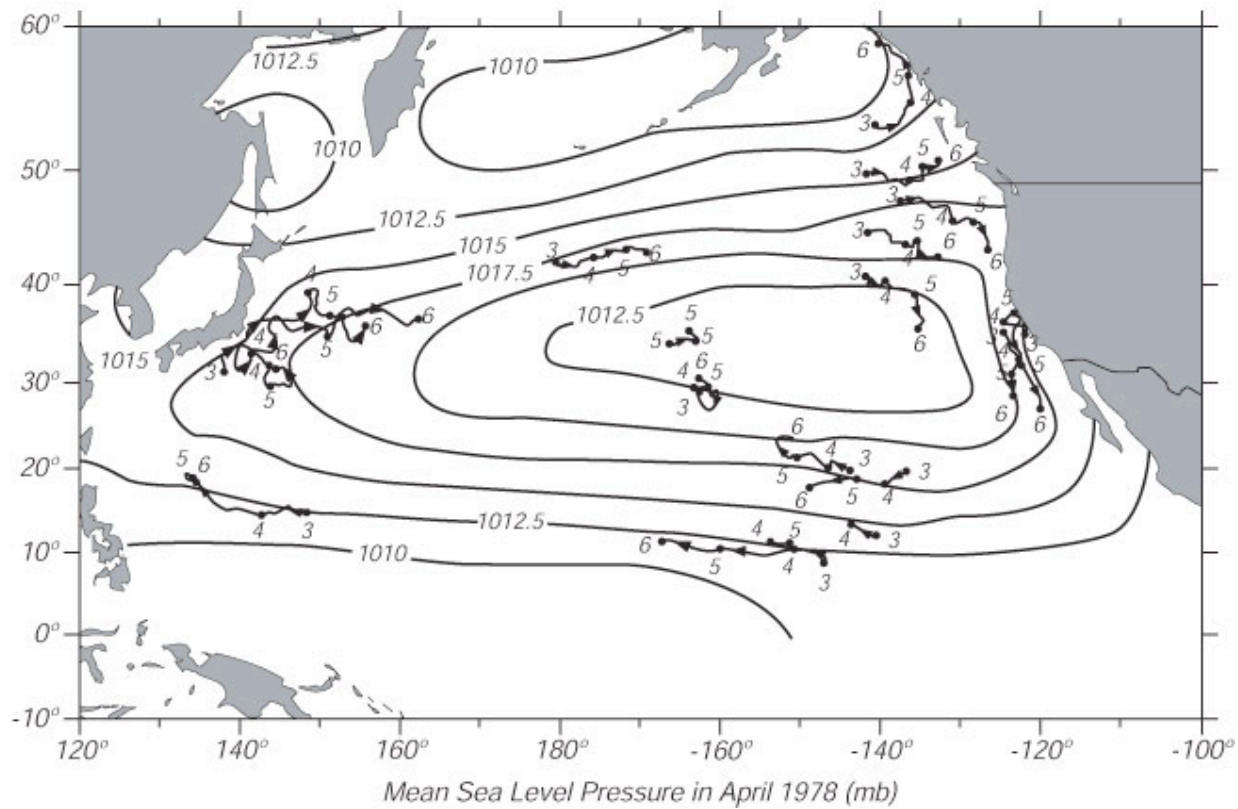
$$D_E = \sqrt{\frac{2\pi^2 A_z}{f}}$$

$$E_z = \frac{A_z \frac{\partial^2 u}{\partial z^2}}{fu}$$



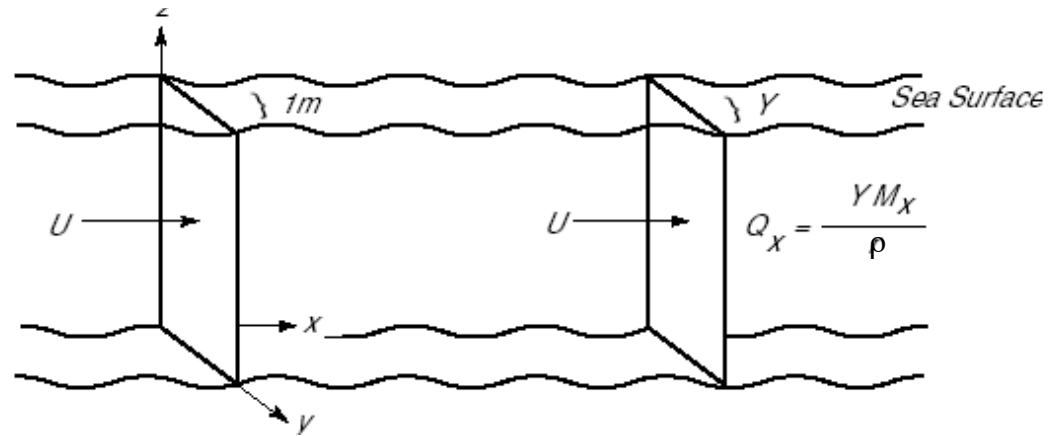
Spirale d'Ekman d'un vent de 10m/s à 35°N (Stewart)

A. N.



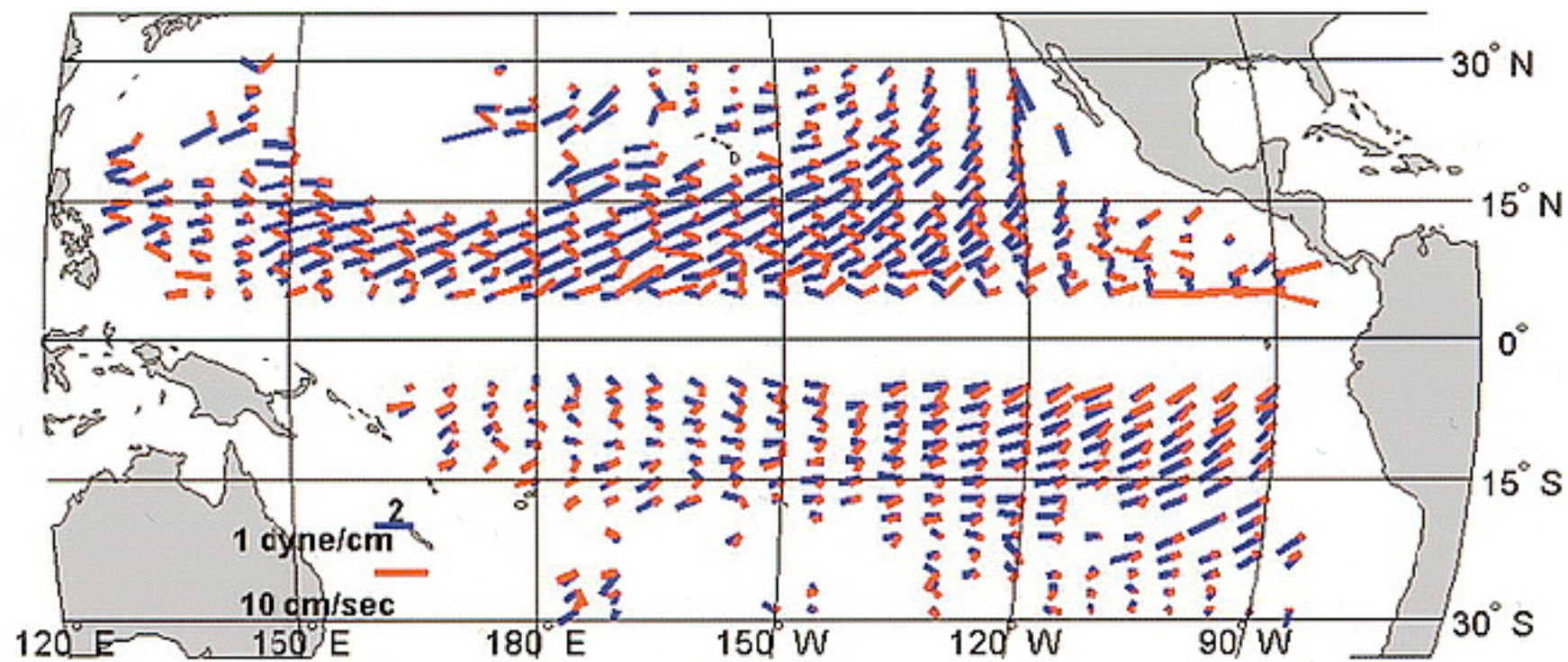
Transport d'Ekman

$$M_{Ex} = \int_{-d}^0 \rho U_E dz, \quad M_{Ey} = \int_{-d}^0 \rho V_E dz$$



$$M_{Ex} = \tau^y / f$$

$$M_{Ey} = -\tau^x / f$$

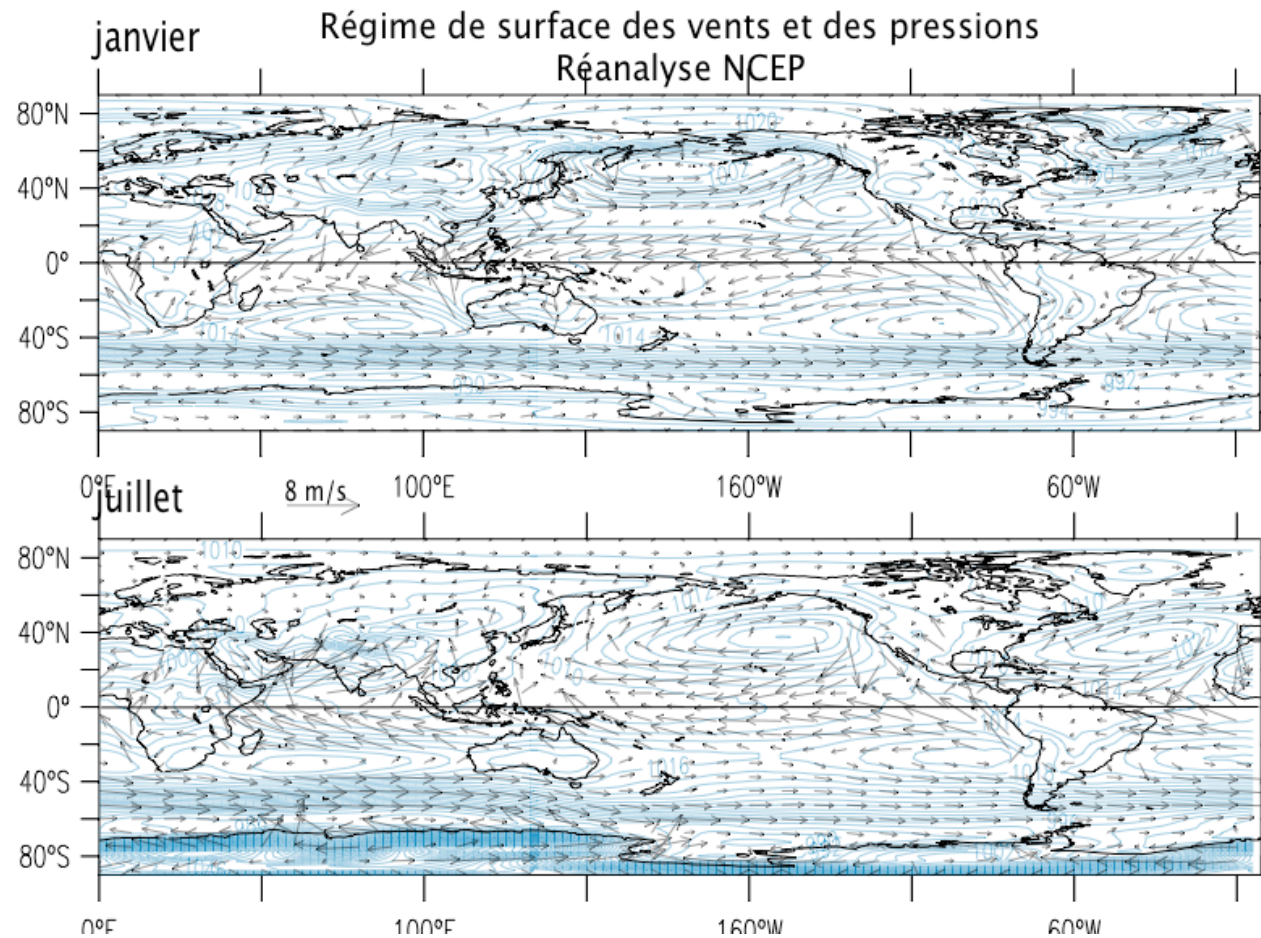


Application au calcul de Q_E , flux volumique d'Ekman

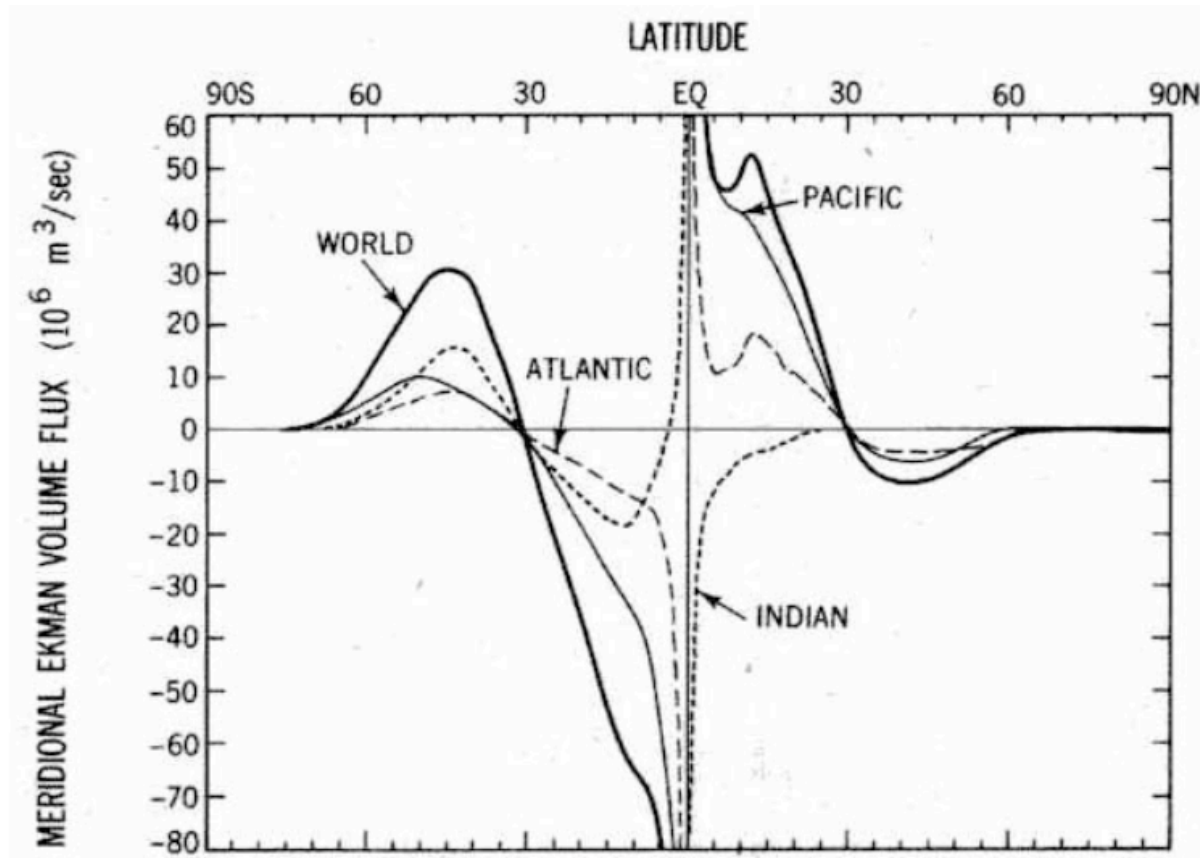
1) Modèle simplifié

$$\tau^x = \tau_0 \sin(\pi y/L) \quad 0 < y < 2L \quad L = 30^\circ$$

2) Transport estimés d'après les observations de vent



3) Transport calculé d'après les observations de vent



Interprétation des courants d'inertie

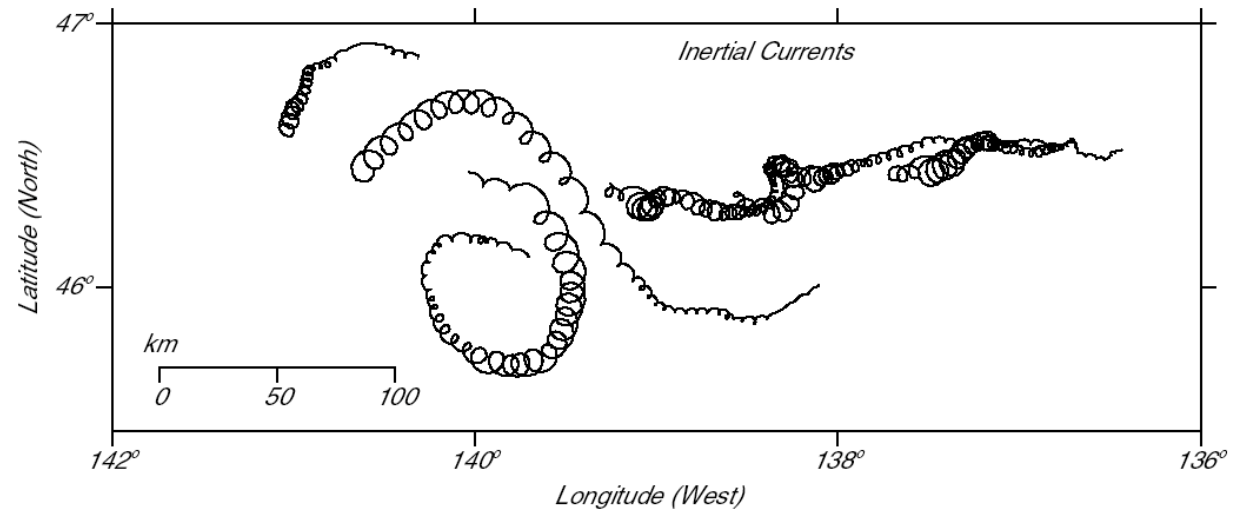
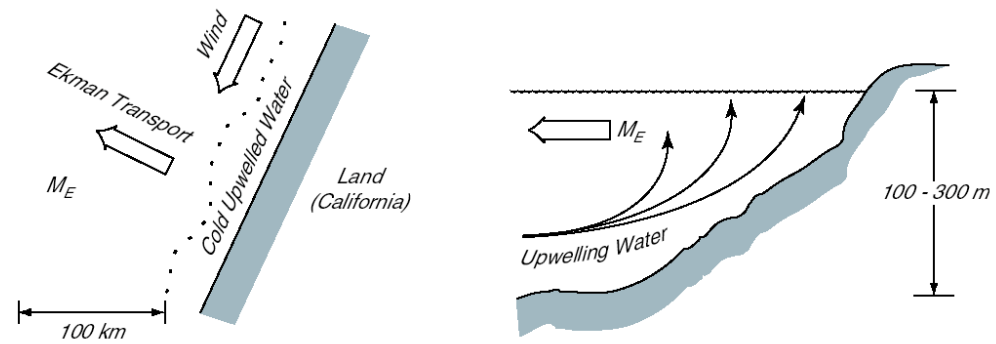


Diagramme progressif

Effet de « pompage d'Ekman »



pompage d'Ekman

1) Théorie

$$\rho = \rho_0 = \text{cst}$$

$$d = \text{cst}$$

$$\rho \int_{-d}^0 \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) dz = 0$$

$$\frac{\partial}{\partial x} \int_{-d}^0 \rho u dz + \frac{\partial}{\partial y} \int_{-d}^0 \rho v dz = -\rho \int_{-d}^0 \frac{\partial w}{\partial z} dz$$

$$\frac{\partial M_x}{\partial x} T_x - \frac{\partial M_y}{\partial y} T_y = -\rho [w(0) - w(-d)]$$

$$\text{Si } f = \text{cste} \Rightarrow \nabla_H \cdot \mathbf{M}_g = 0 \Rightarrow \nabla_H \cdot \mathbf{M}_T = \nabla_H \cdot \mathbf{M}_E$$

$$\text{or l'interface}(z=0) \text{ immobile} \Rightarrow \nabla_H \cdot \mathbf{M}_E = 0 + \rho w(-d)$$

$$\Rightarrow w_E(-d) = \frac{1}{\rho} \left[\frac{\partial}{\partial x} \left(\frac{T_{yz}(0)}{f} \right) - \frac{\partial}{\partial y} \left(\frac{T_{xz}(0)}{f} \right) \right]$$

$$w_E(-d) = \text{curl} \left(\frac{\mathbf{T}}{\rho f} \right) \quad \text{C'est le pompage d'Ekman}$$

Mais $f=f(y)$ permet d'être plus rigoureux: \Rightarrow

$$\nabla_H \cdot \mathbf{M}_T = \rho [w_g(0) + w_E(0) + w_g(-d) + w_E(-d)]$$

$$\text{en } z \rightarrow d_E, \text{ les effets de la friction} \rightarrow 0 \Rightarrow \mathbf{U}_E(u, v, w)|_{-d} \rightarrow 0$$

$$\Rightarrow w_E(-d) = 0 \Rightarrow w_E(0) = \nabla_H \cdot \mathbf{M}_E = \text{rot}(\mathbf{T}/\rho f)$$

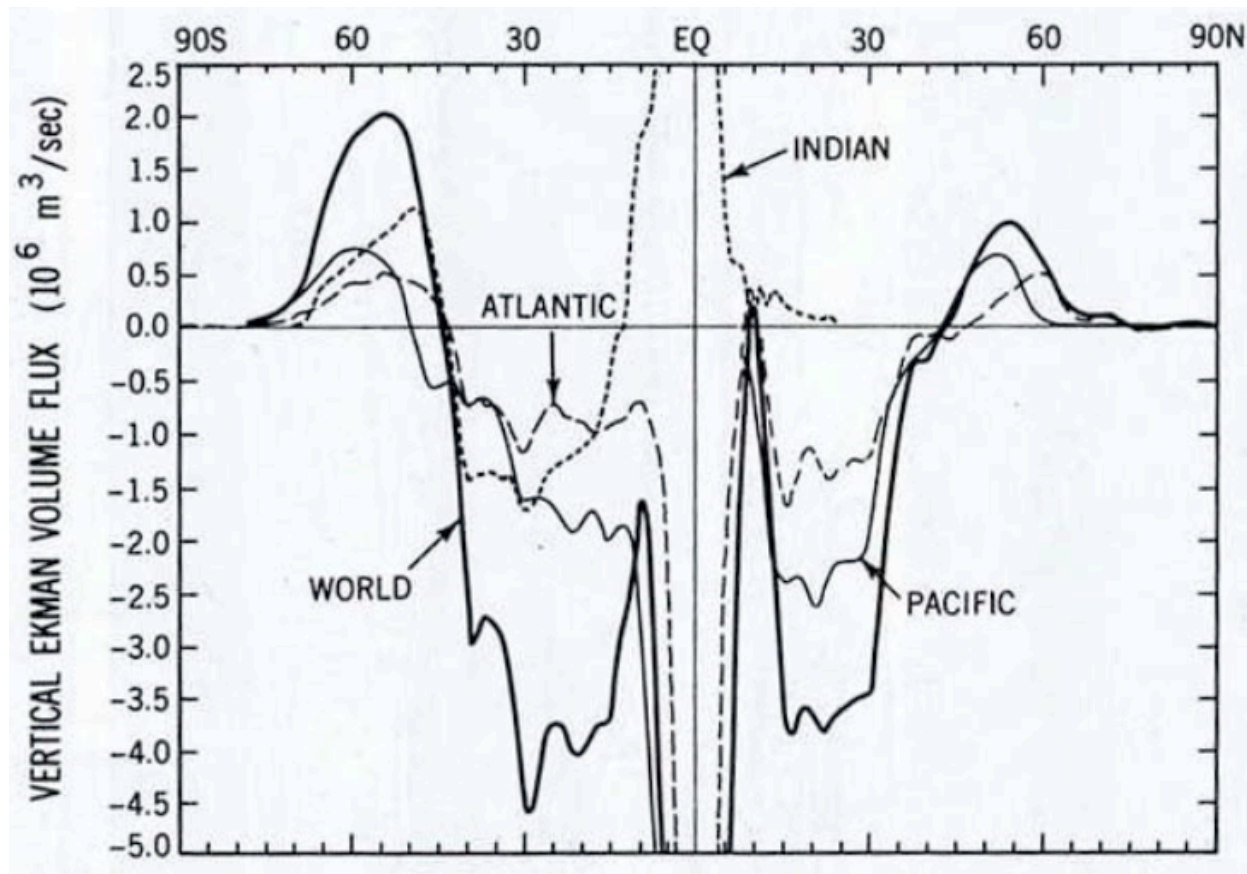
$$\text{or l'interface}(z=0) \text{ tjs immobile} \Rightarrow w_g(0) = -w_E(0)$$

$$\Rightarrow w(-d) = w_g(-d)$$

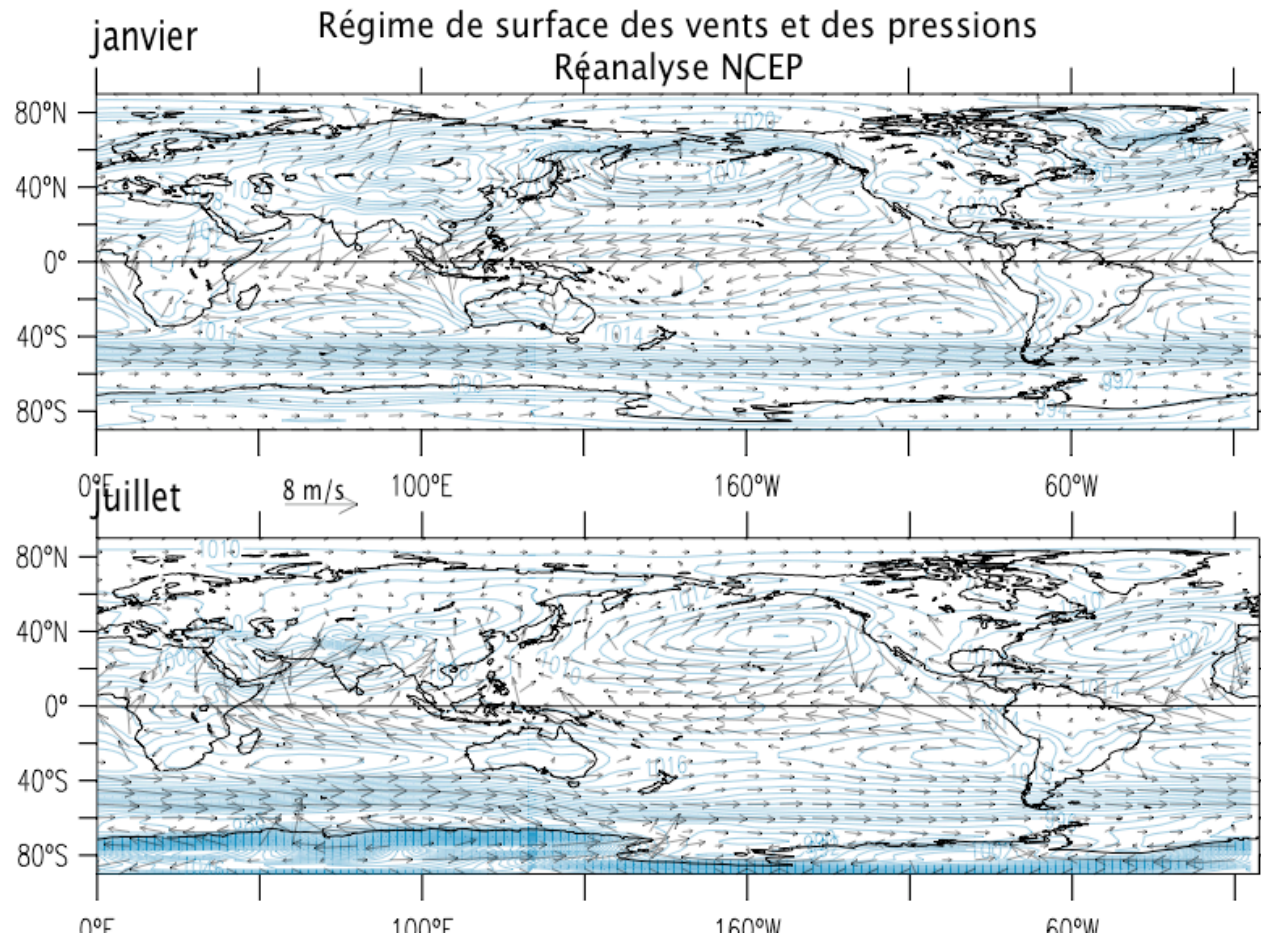
2) Calcul

a) Modèle simplifié

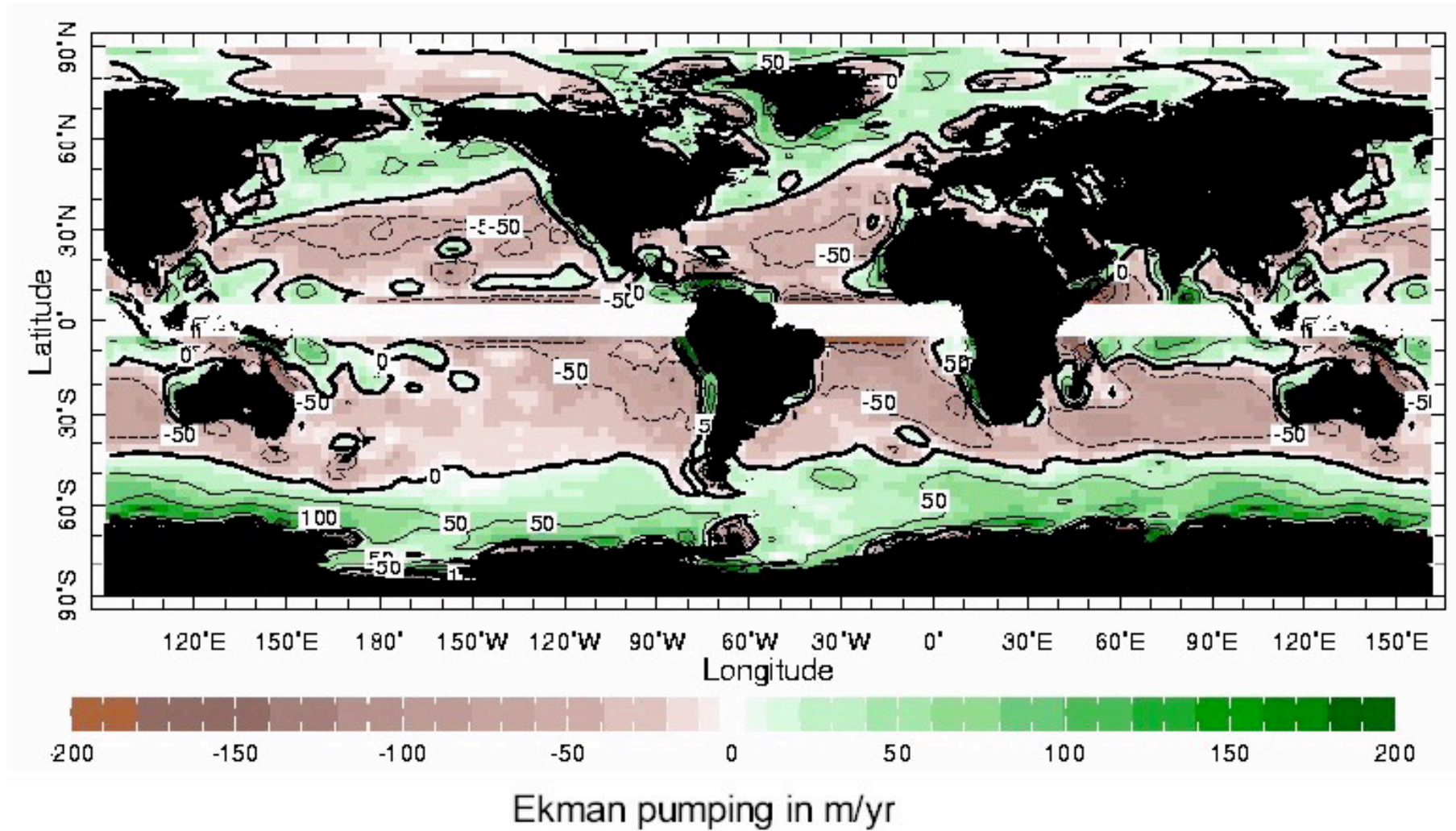
b) Transport calculé d'après les observations de vent



2) Estimation des distribution de We

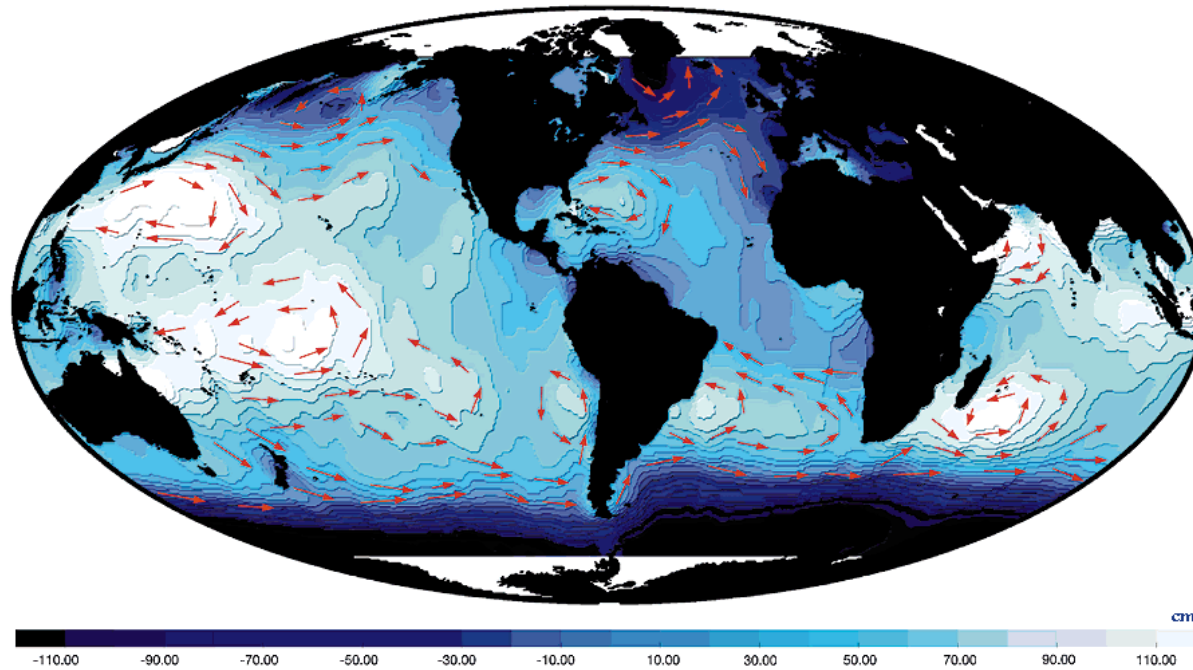


Distribution moyenne du Pompage d'Ekman

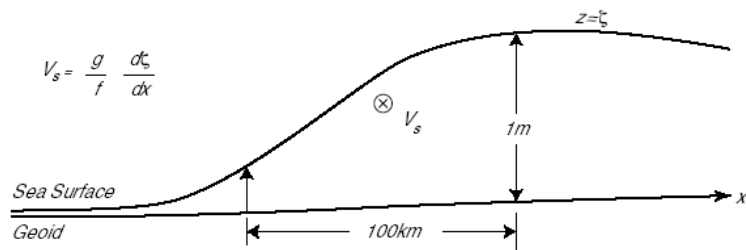


Hauteur dynamique de la mer et courants géostrophiques barotropes

Topographie dynamique de la surface de la mer observée par TOPEX/POSEIDON

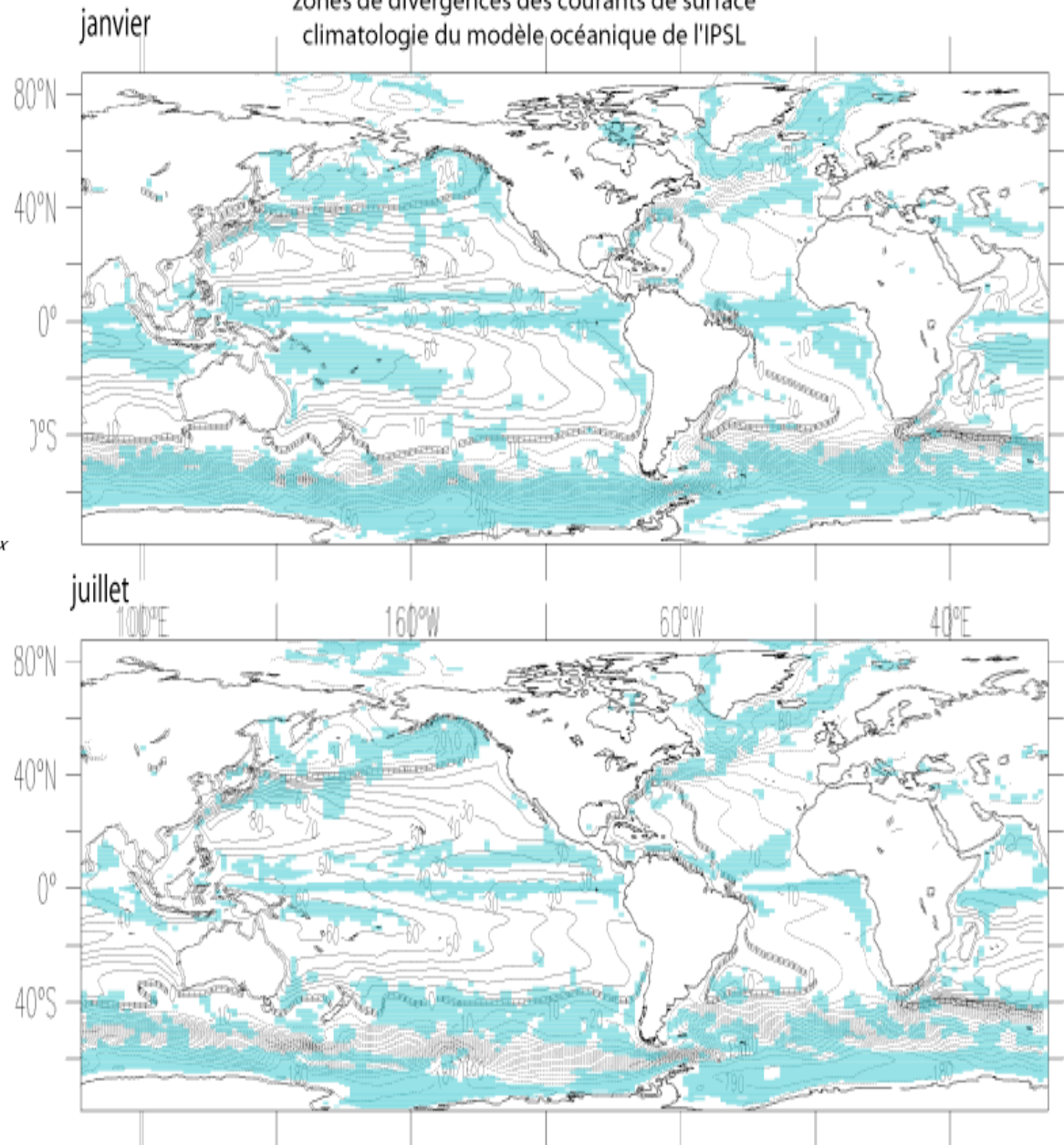


rho=cste : courants géostrophiques barotropes

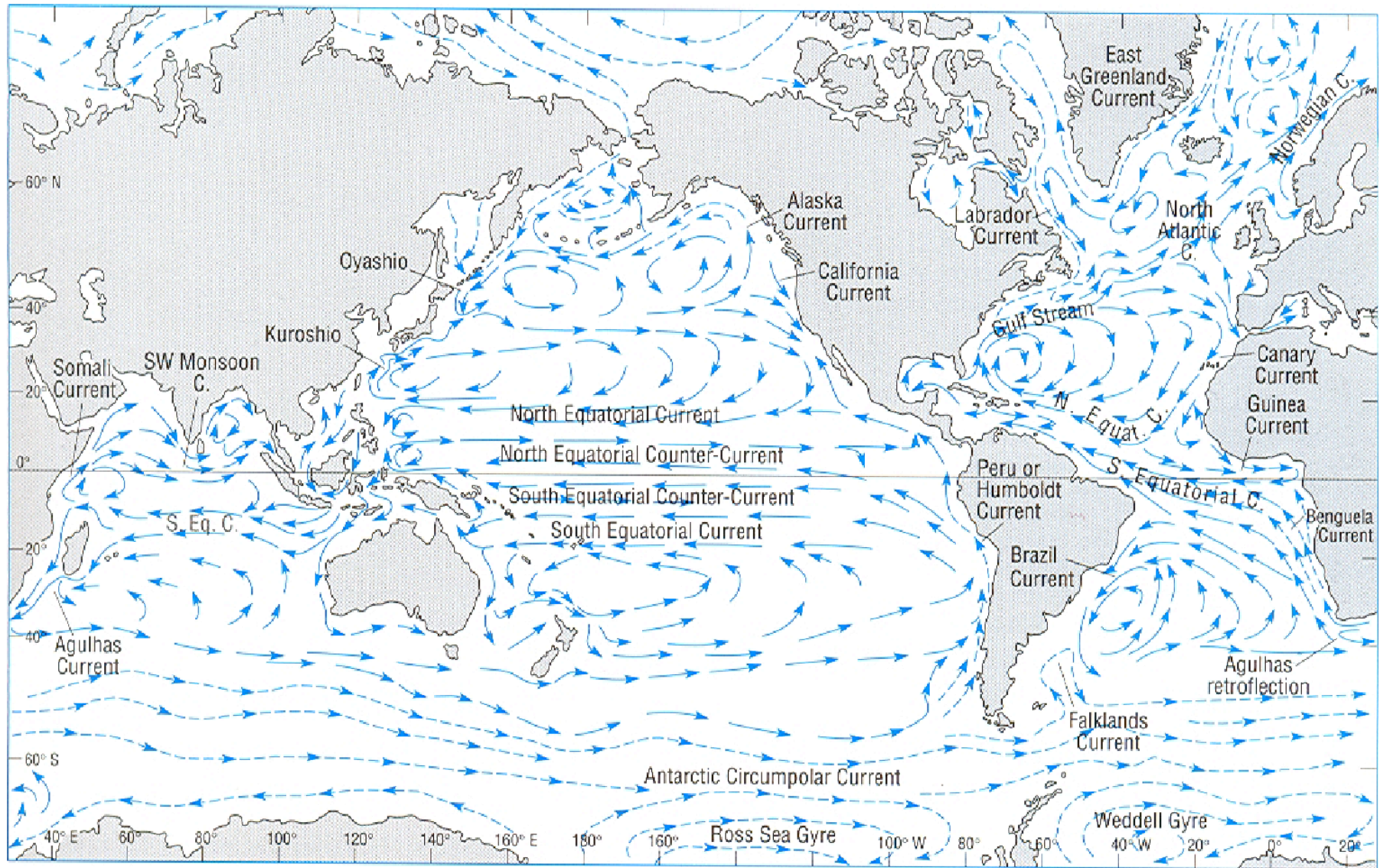


Application :

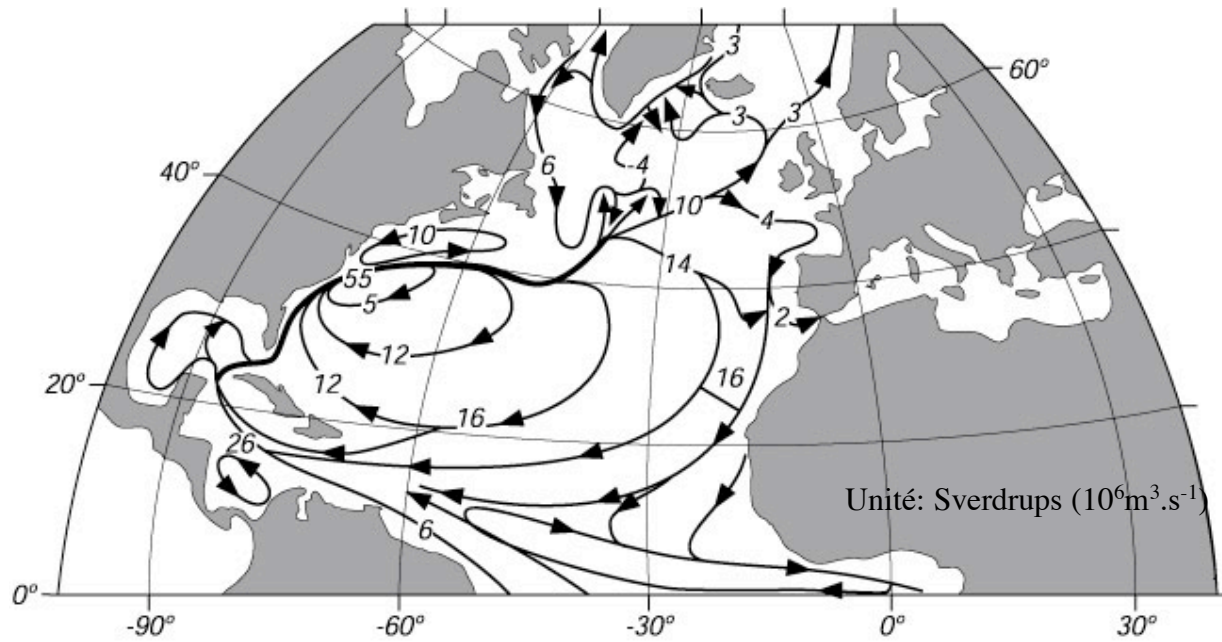
anomalie de hauteur de la mer et zones de divergences des courants de surface climatologie du modèle océanique de l'IPSL



Schématisation des courants de surface observés



Observations Nord Atlantique



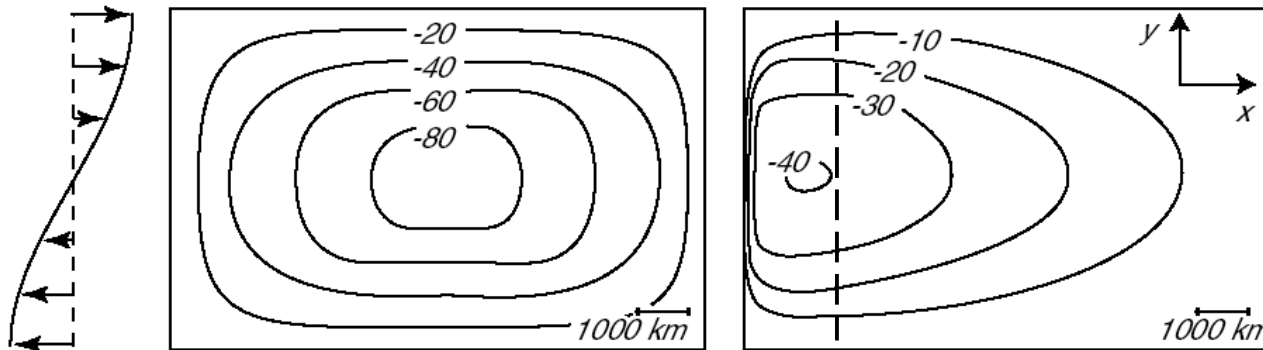
Circulation (intégrée) de Sverdrup

$$-fv + \frac{1}{\rho_{ref}} \frac{\partial p}{\partial x} = \frac{1}{\rho_{ref}} \frac{\partial \tau_{wind_x}}{\partial z} ; \quad fu + \frac{1}{\rho_{ref}} \frac{\partial p}{\partial y} = \frac{1}{\rho_{ref}} \frac{\partial \tau_{wind_y}}{\partial z}$$

Diff. croisée $\Rightarrow \beta v = f \frac{\partial w}{\partial z} + \frac{1}{\rho_{ref}} \frac{\partial}{\partial z} \left(\frac{\partial \tau_{wind_y}}{\partial x} - \frac{\partial \tau_{wind_x}}{\partial y} \right)$

A.N.

Int. Z + c.l. $\Rightarrow \beta V = \frac{1}{\rho_{ref}} \left(\frac{\partial \tau_{wind_y}}{\partial x} - \frac{\partial \tau_{wind_x}}{\partial y} \right) = \frac{1}{\rho_{ref}} \hat{\mathbf{z}} \cdot \nabla \times \boldsymbol{\tau}_{wind}$



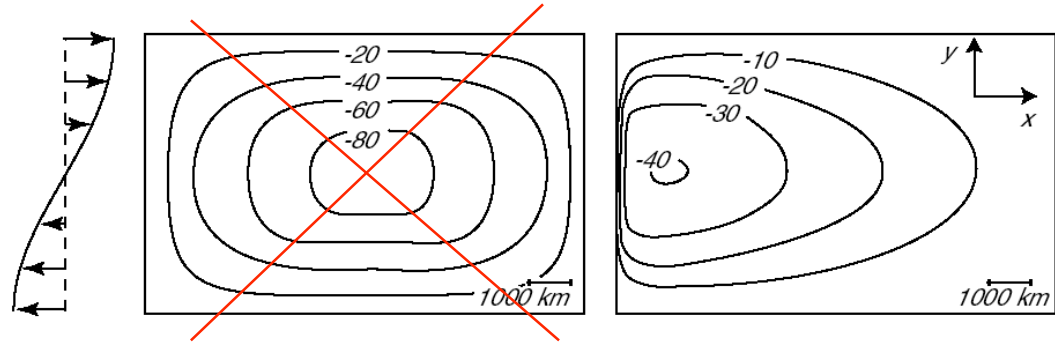
Pourquoi des courant de bord ouest:
La dissipation doit être compensée par le travail du vent

Interprétation des gyres

- Direction dans l'intérieur :

Divergence du courant géostrophique

$$\nabla_h \cdot \mathbf{u}_g + \frac{\partial w}{\partial z} = 0.$$



$$\beta v_g = f \frac{\partial w}{\partial z}$$

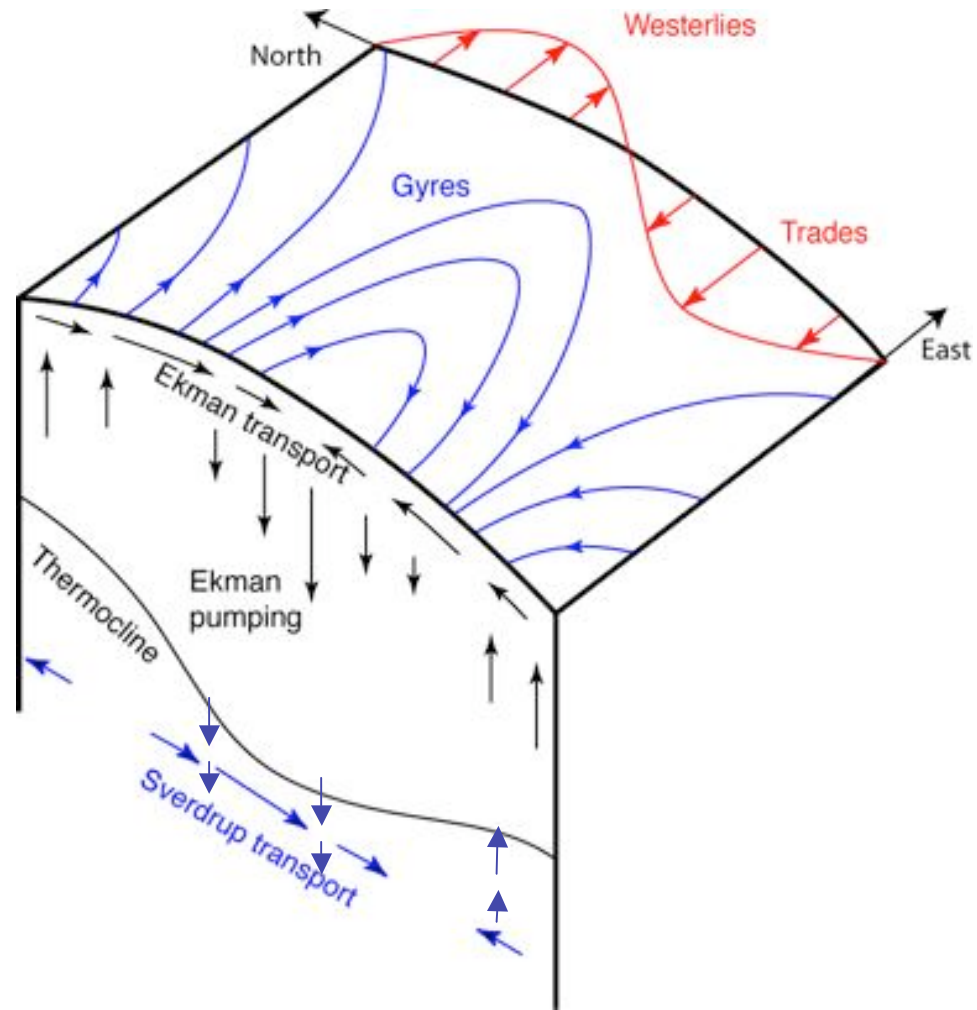
$$\beta = \frac{df}{dy} = \frac{1}{a} \frac{df}{d\varphi} = \frac{2\Omega}{a} \cos \varphi$$

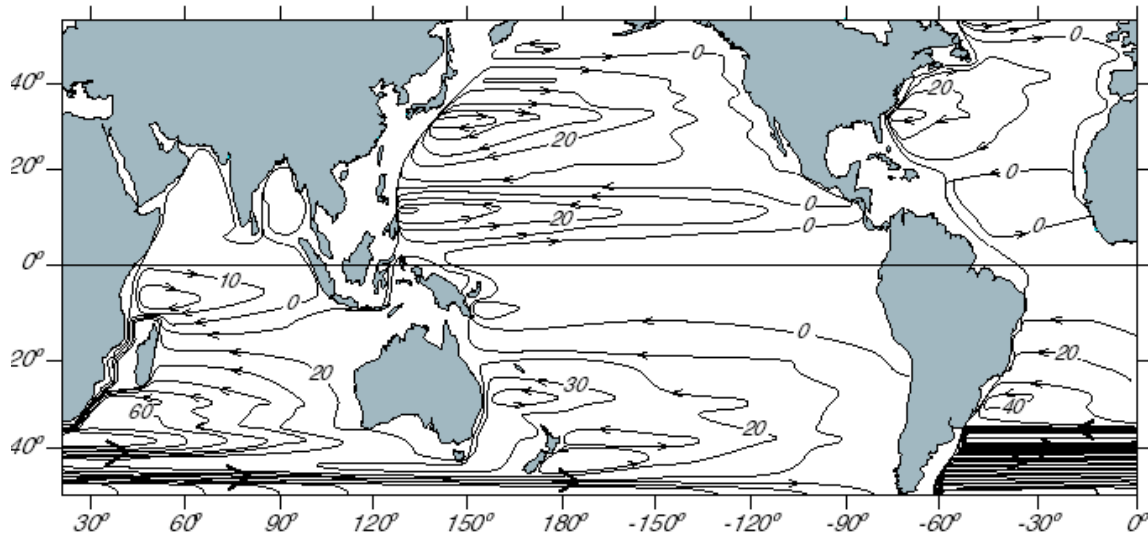
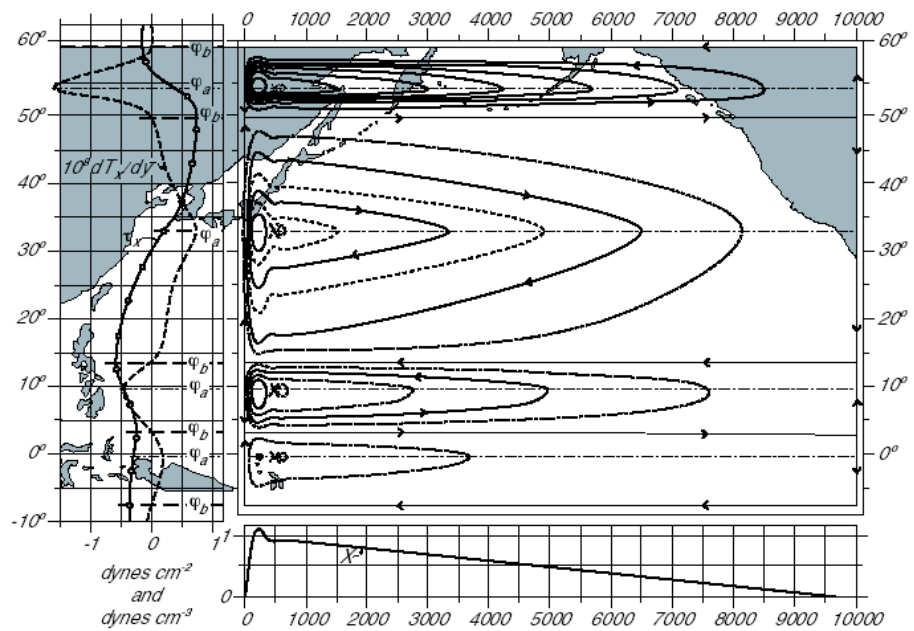
Nécessité d'une pente zonale de la surface de l'océan, car elle engendre un transport géostrophique divergent

- Flot de retour :

Pourquoi des courant de bord ouest?:
La dissipation doit être compensée par le travail du vent

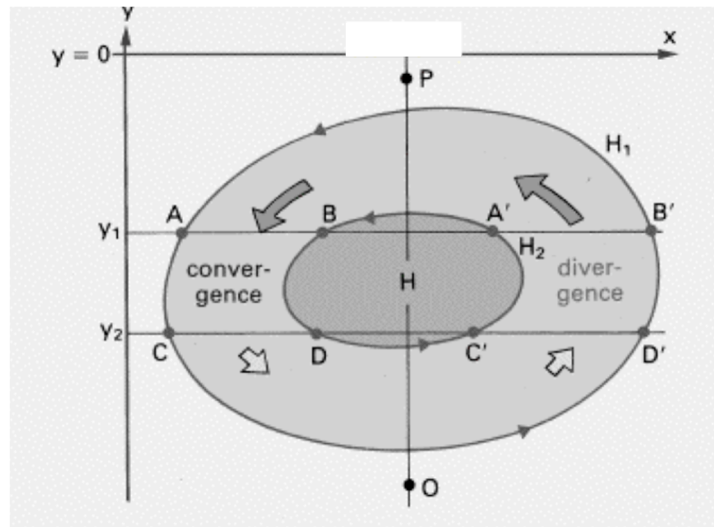
Modèle de Sverdrup





Etablissement de la pente zonale:
effets des ondes de Rossby

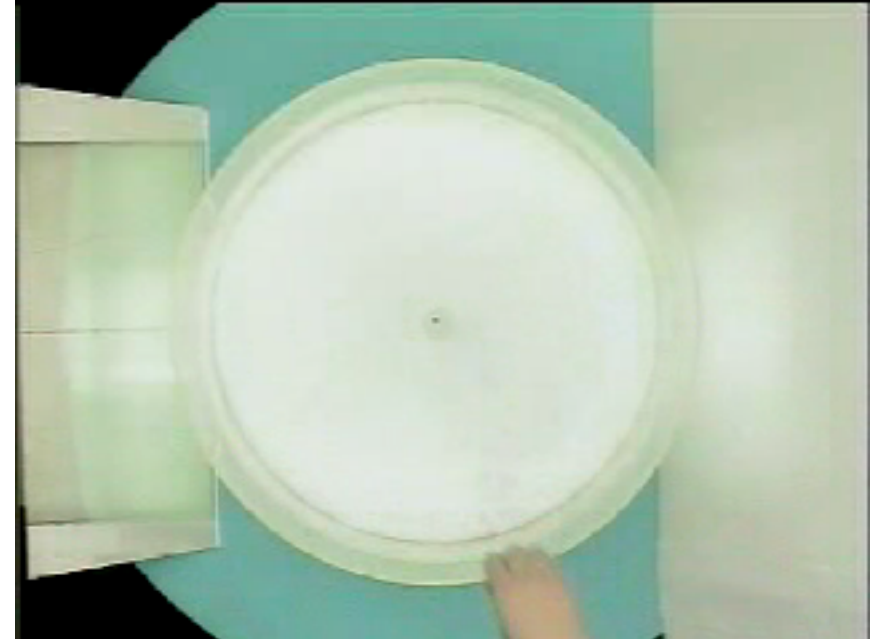
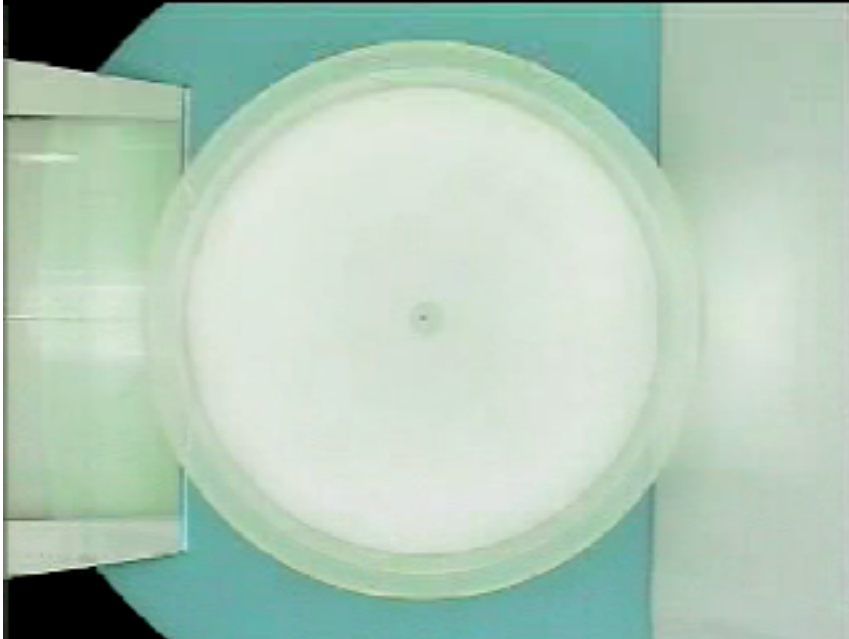
Exemple dans l'hémisphère sud (Tomczack et Godfrey)



$$C = -(\partial H / \partial t) / (\partial H / \partial x) = \frac{\beta g H}{f^2(y)}$$

La rigidité des fluides géostrophique

Taylor-Proudman
et leurs colonnes de densité homogène



$$\rho = \rho_0$$

$$f v = \frac{1}{\rho_0} \frac{\partial p}{\partial x}$$

$$f u = -\frac{1}{\rho_0} \frac{\partial p}{\partial y}$$

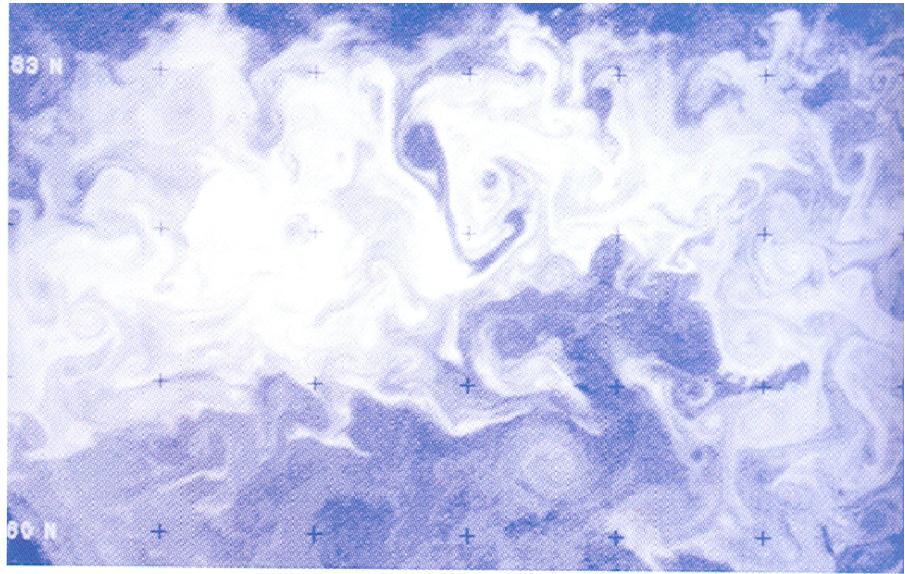
$$g = -\frac{1}{\rho_0} \frac{\partial p}{\partial z}$$

$$0 = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}$$

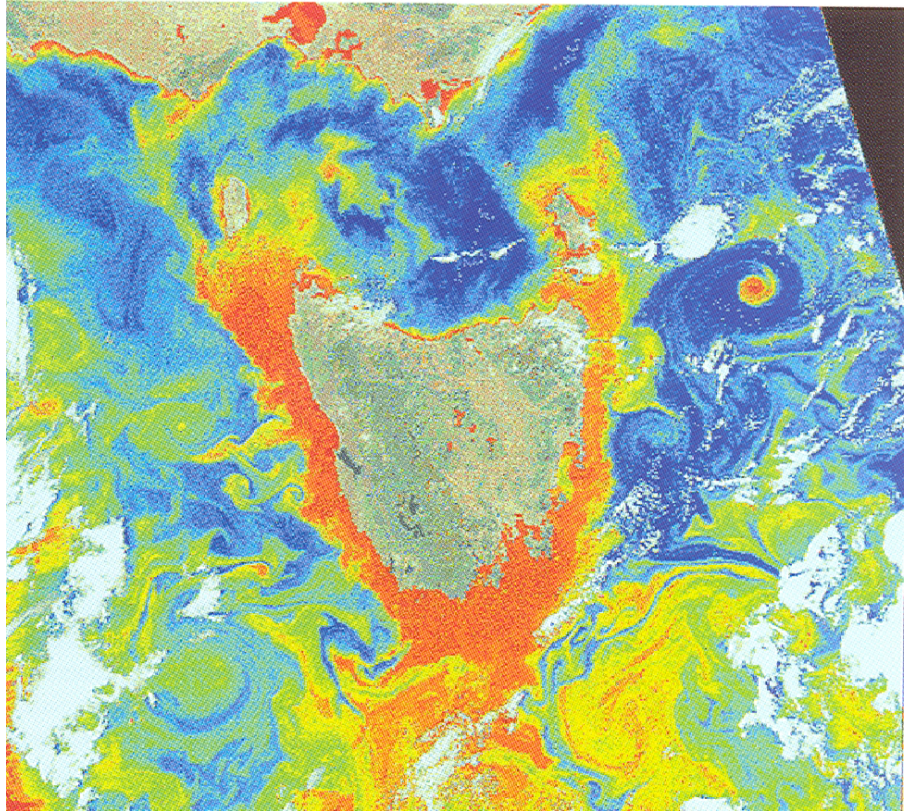
$$\frac{\partial u}{\partial z} = \frac{\partial v}{\partial z} = ?$$

Expériences de laboratoire issues du cours
« Physics of Atmospheres and Oceans »
de John Marshall and Alan Plumb
<http://www-paoc.mit.edu/labweb/>

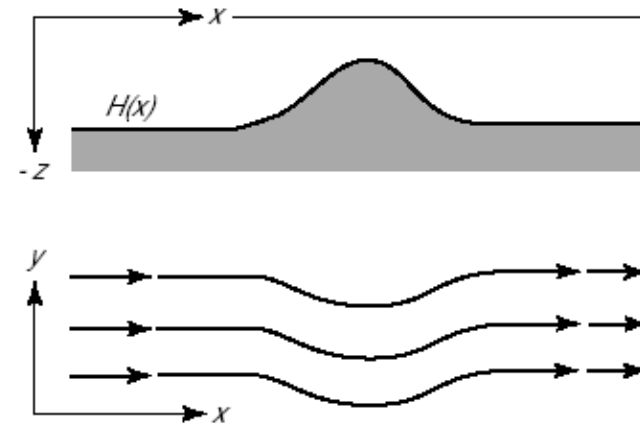
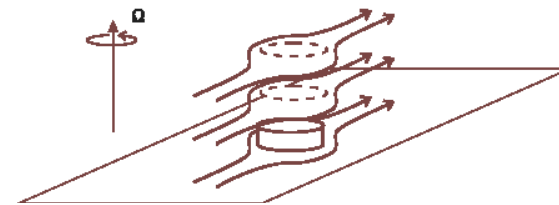
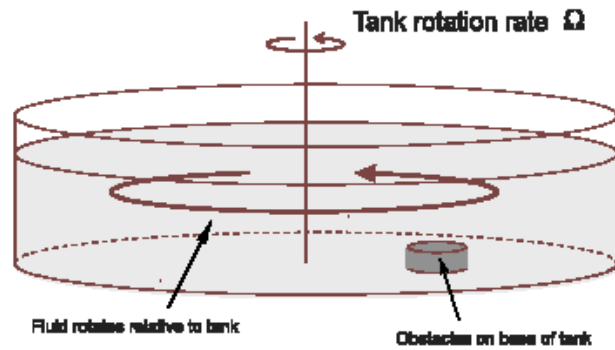
Quid de la relation du vent thermique ?



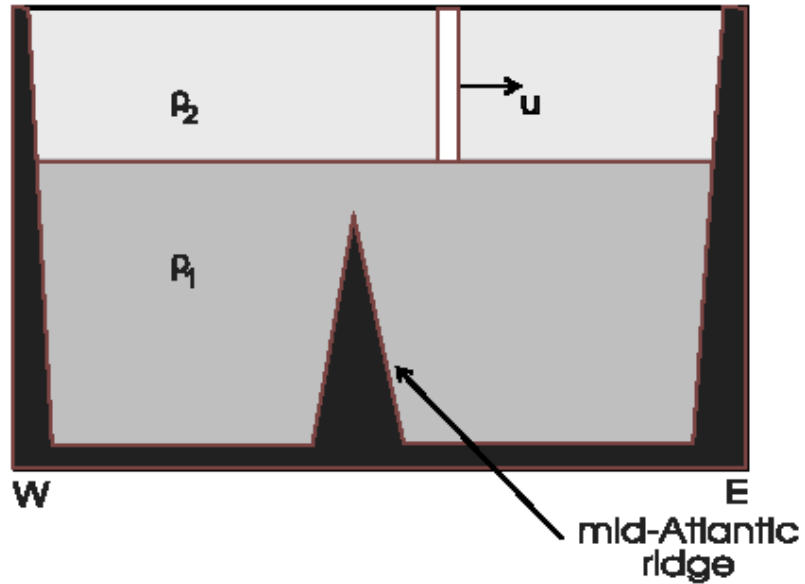
(b)



Si $f = f_0$ alors $\frac{\partial w}{\partial z} = 0$ Forçage topographique !



Effet de la stratification sur les colonnes



Dans l'océan, existence d'une grosse différence verticale de densité : la pycnocline

- Au sein d'une zone stratifiée (pycnocline) et géostrophique, chaque couche homogène abrite ses petites colonnes de T-P

-Cas plus réaliste de couches de fluide barotrope