Ocean Circulation from remote sensing observation

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Ocean circulation

1. The main components of the ocean circulation
   - Geostrophic currents
   - Ekman currents
   - Stoke drifts
   - Inertial oscillations

2. Observing the ocean circulation from space
   - Geostrophic currents from altimetry
   - Wind driven currents from altimetry, drifters and scatterometters
   - Higher resolution currents from sensor synergy

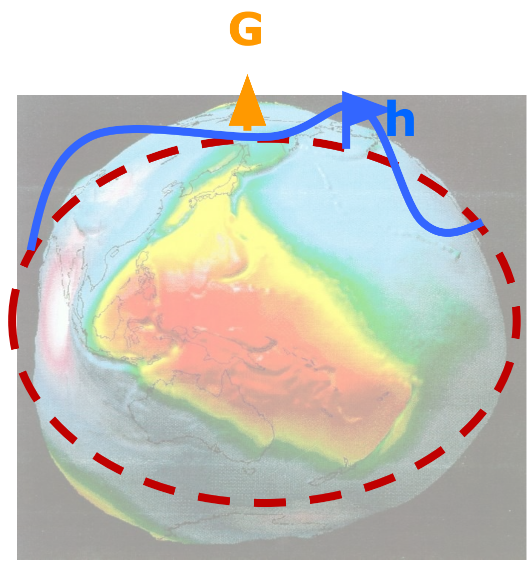
3. Practicals: from the observation to the current
   - Reconstruct Geostrophic current from altimetric SSH
   - Reconstruct wind driven component from ECMWF wind
   - Lagrangian advection and comparison with in-situ data

Ocean circulation

The surface of an ocean of homogeneous density covering an Earth at rest would coincide with an Earth Gravity Equipotential surface called GEOID

Thermal forcing

Gravity forces generating tides



Wind effects

Variations of the Atmospheric pressure

E : Reference Ellipsoid
Equipotential of the gravity field associated to an hypothetically *homogeneous* distribution of *mass* within the *earth*

Coriolis Force due to the Earth Rotation

Hydrological Cycle

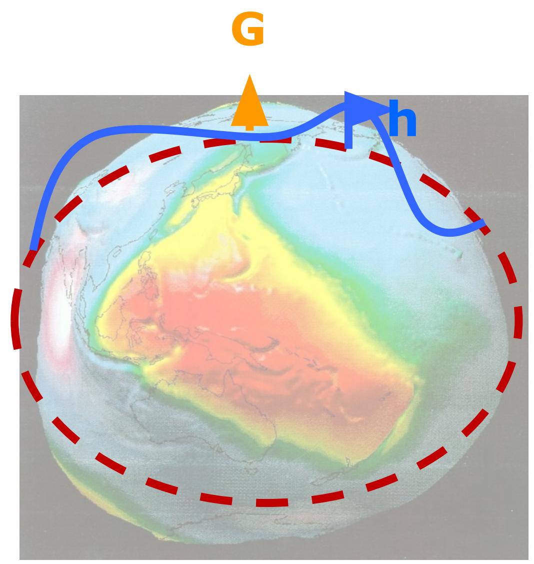
Ocean circulation

As a consequence, at a given time, at a given place, the sea level differs from its position at rest, the geoid. The difference between the two positions is the **ocean dynamic topography h**

Geostrophic currents

Tidal currents

Ekman currents



E : Reference Ellipsoid
Equipotential of the gravity field associated to an hypothetically *homogeneous* distribution of *mass* within the *earth*

Stokes drift

Internal waves

Surface gravity waves

Inertial oscillations

Equation of motions

The Lagrangian acceleration of a fluid particule is due to 4 main forces:

Pressure gradient force

Friction forces

$$\frac{D\vec{u}}{Dt} = -2\vec{\Omega} \wedge \vec{u} - \frac{1}{\rho}\vec{\nabla}p - \vec{g} + \nu\vec{\nabla}^2\vec{u}$$

Coriolis force (due to earth rotation)

Gravitation force

Different approximations can be done depending on the relative order of magnitude of these 4 forces
R_0=(non linear terms)/(Coriolis term)   E=(friction term / Coriolis term)

Geostrophic currents

$$\cancel{\frac{D\vec{u}}{Dt}} = -2\vec{\Omega} \wedge \vec{u} - \frac{1}{\rho}\vec{\nabla}p - \vec{g} + \cancel{\nu\vec{\nabla}^2\vec{u}}$$

Geostrophic hypothesis: E<<1, R_0<<1 and w<<u,v
Valid away from the boundary layers, from the equator, and over large (> 50-100 km) spatial and long (>2-10 days) temporal scales.

The largest terms in the equations of motion reduce to the Coriolis force and the pressure gradient.

$$0 = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv$$

$$0 = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu$$

Geostrophic balance

The ocean surface velocity field (u,v) is derived from the gradients of the sea level above the geoid.

$$u_{geo} = -\frac{g}{f}\frac{\partial h}{\partial y}$$

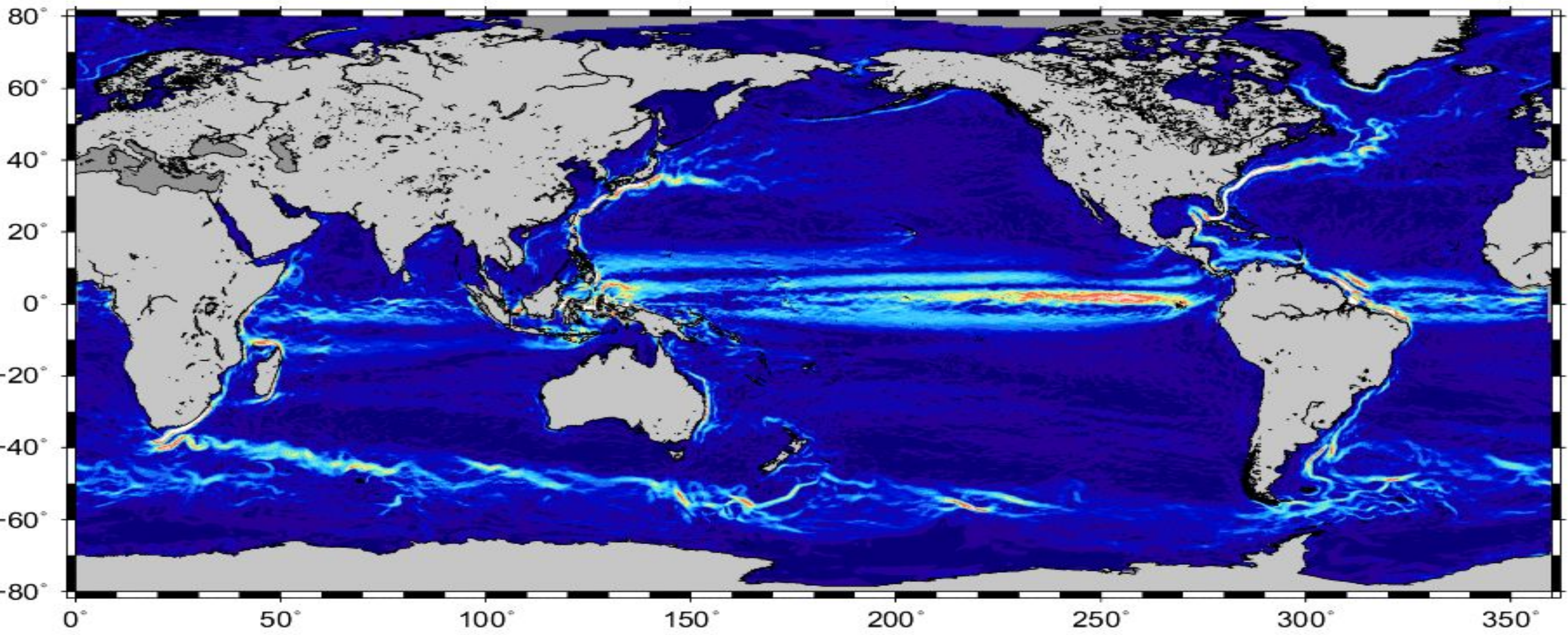
$$v_{geo} = \frac{g}{f}\frac{\partial h}{\partial x}$$

$$0 = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$$

Hydrodynamic equation

Geostrophic currents

Mean geostrophic currents speed
From Altimetry+GOCE+in-situ measurements



Ekman currents

$$\cancel{\frac{D\vec{u}}{Dt}} = -2\vec{\Omega} \wedge \vec{u} - \cancel{\frac{1}{\rho}\vec{\nabla} p} - \cancel{\vec{g}} + \nu \vec{\nabla}^2 \vec{u}$$

$R_0 << 1$, $E \sim 1$

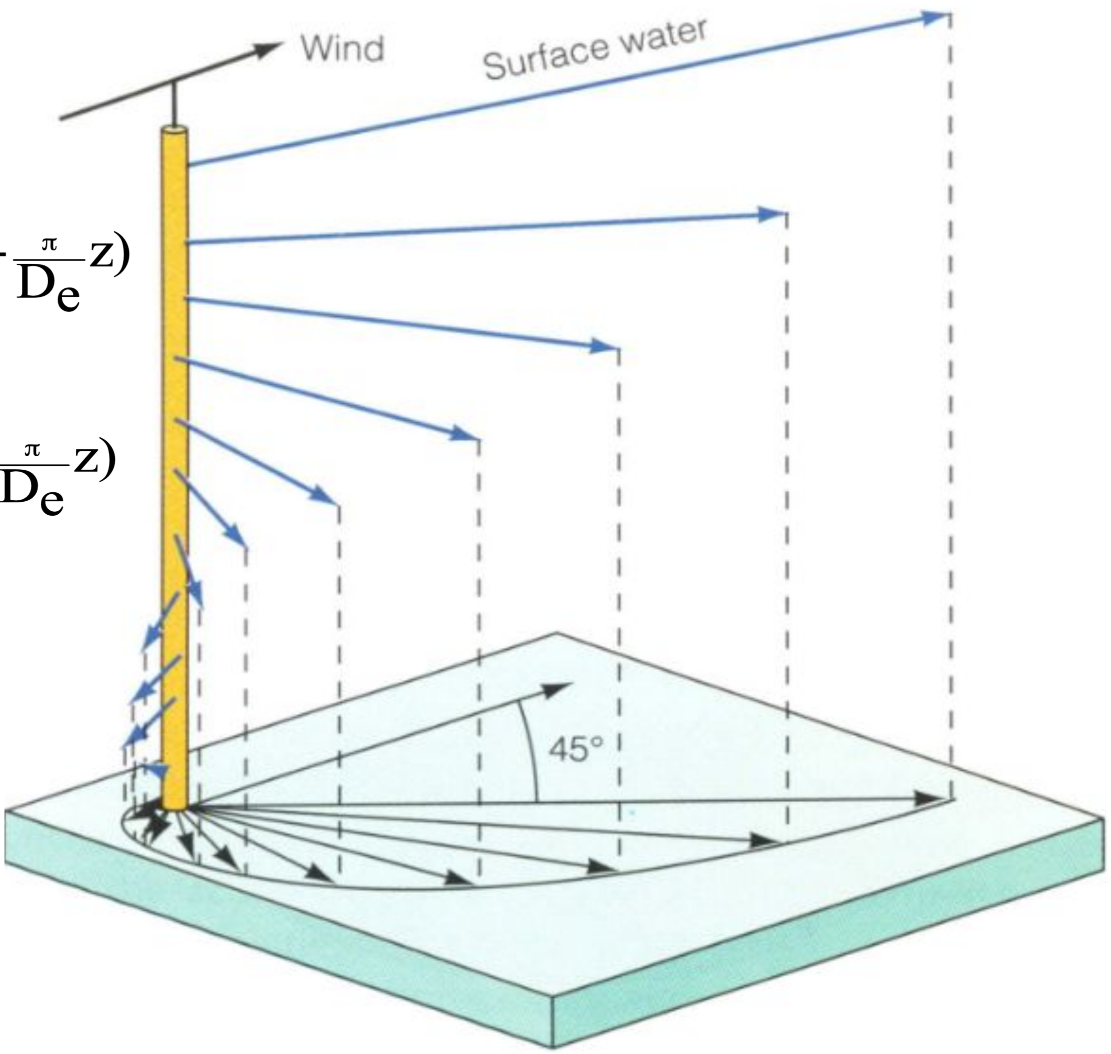
In an homogenous spatial area, and stability conditions, under a stationary temporal forcing, over a few inertial periods, the equilibrium between the Coriolis forces and the friction forces due to wind stress leads to the classical Ekman current formulation.

$$f u_E = A_z \frac{\partial^2 v_E}{\partial z^2} = \frac{1}{\rho} \frac{\partial \tau_y}{\partial z}$$

$$-f v_E = A_z \frac{\partial^2 u_E}{\partial z^2} = \frac{1}{\rho} \frac{\partial \tau_x}{\partial z}$$

$$u_e = \pm \frac{\pi\sqrt{2}}{\rho(f+w)D_e} e^{\frac{\pi}{D_e} z} * \tau_e * \cos\left(\frac{\pi}{4} + \frac{\pi}{D_e} z\right)$$

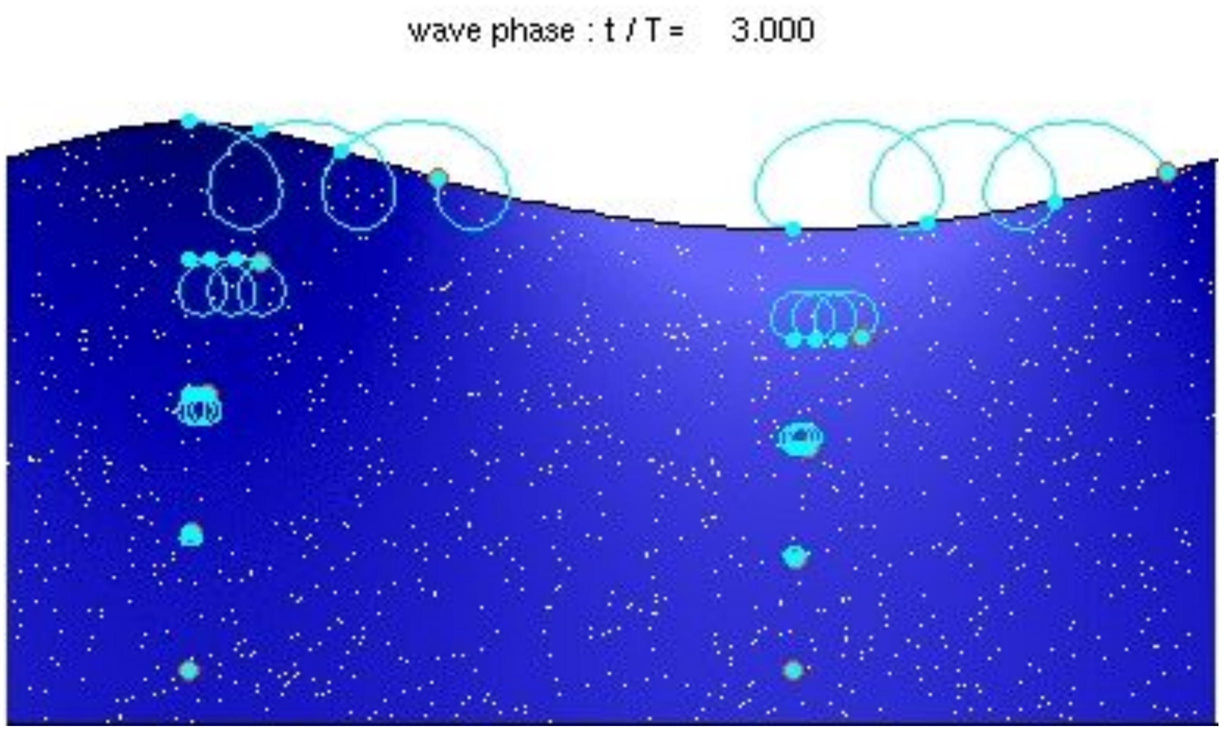
$$v_e = \frac{\pi\sqrt{2}}{\rho(f+w)D_e} e^{\frac{\pi}{D_e} z} * \tau_e * \sin\left(\frac{\pi}{4} + \frac{\pi}{D_e} z\right)$$



τ_e = Effective Wind Stress
D_e = Ekman depth
f = planetary vorticity
w = local vorticity

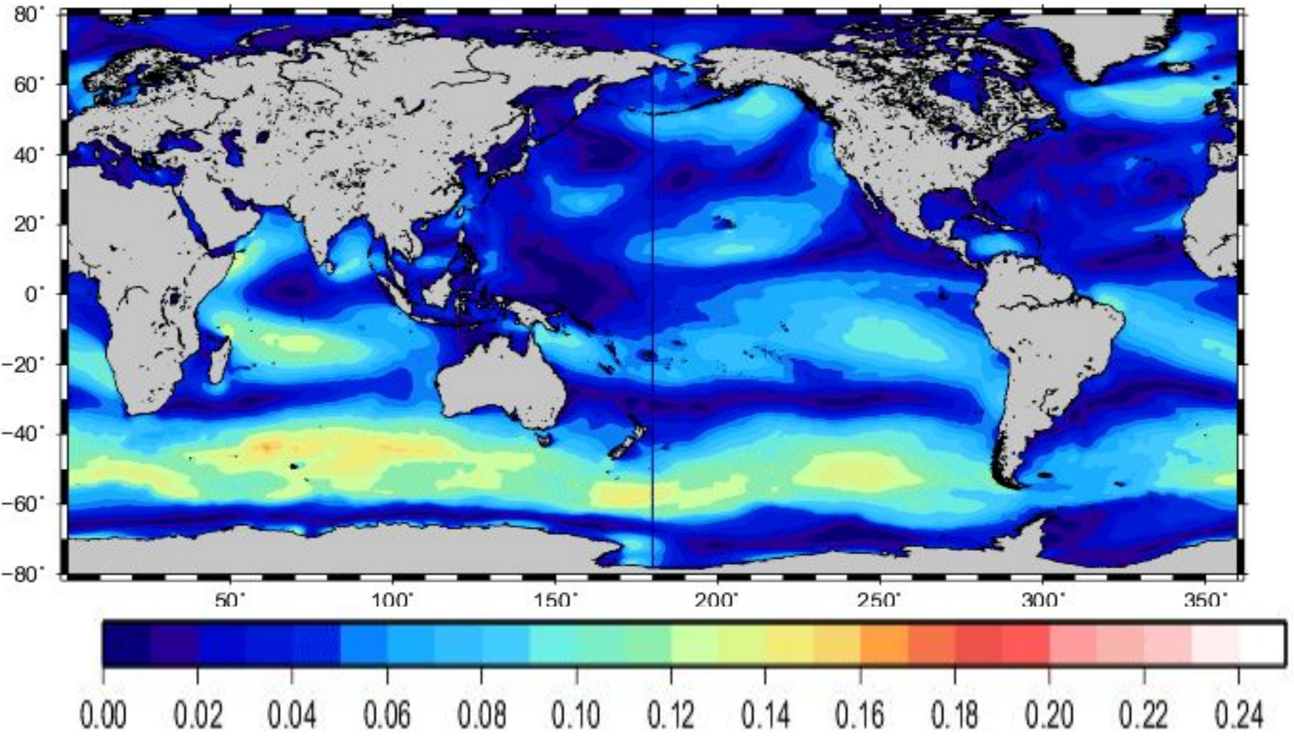
$$2\omega = \partial_x vgeost - \partial_y ugeost$$

Stokes drifts



Water particles that make up the waves do not travel in a straight line, but rather in orbital motions (Stokes, 1847). Their movement is enhanced at the top of the orbit and slowed slightly at the bottom. and thus have an additional movement in the direction of wave propagation.

Mean Stokes drift over one month (September 2009) calculated from the WW3 model



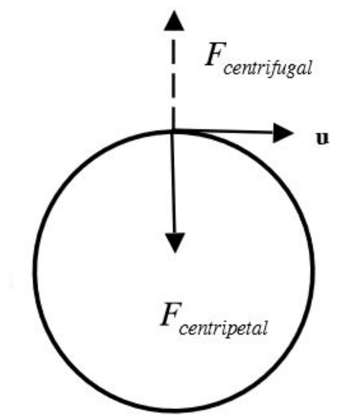
$$\overline{u}_S \approx \omega\, k\, a^2\, \mathrm{e}^{2kz} = \frac{4\pi^2\, a^2}{\lambda\, T}\, \mathrm{e}^{4\pi\, z/\lambda}.$$

a is the wave amplitude, k is the wave number:
k = 2π / λ,
ω is the angular frequency: ω = 2π / T ,
z is the vertical coordinate, with positive z pointing out of the fluid layer,
λ is the wave length and T is the wave period

=2m in 10 m/s wind speed

Inertial oscillations

When wind and wave forces that have set upper ocean motions cease to strongly act, water will not rest immediately. Energy imparted by the wind and waves takes time to fully dissipate. The Coriolis force will then continue to apply as a **centripetal force**, leading to **rotational flows**, referred as **inertial currents**. The period of rotation will vary with the local Coriolis parameter f (e.g. latitude dependent).



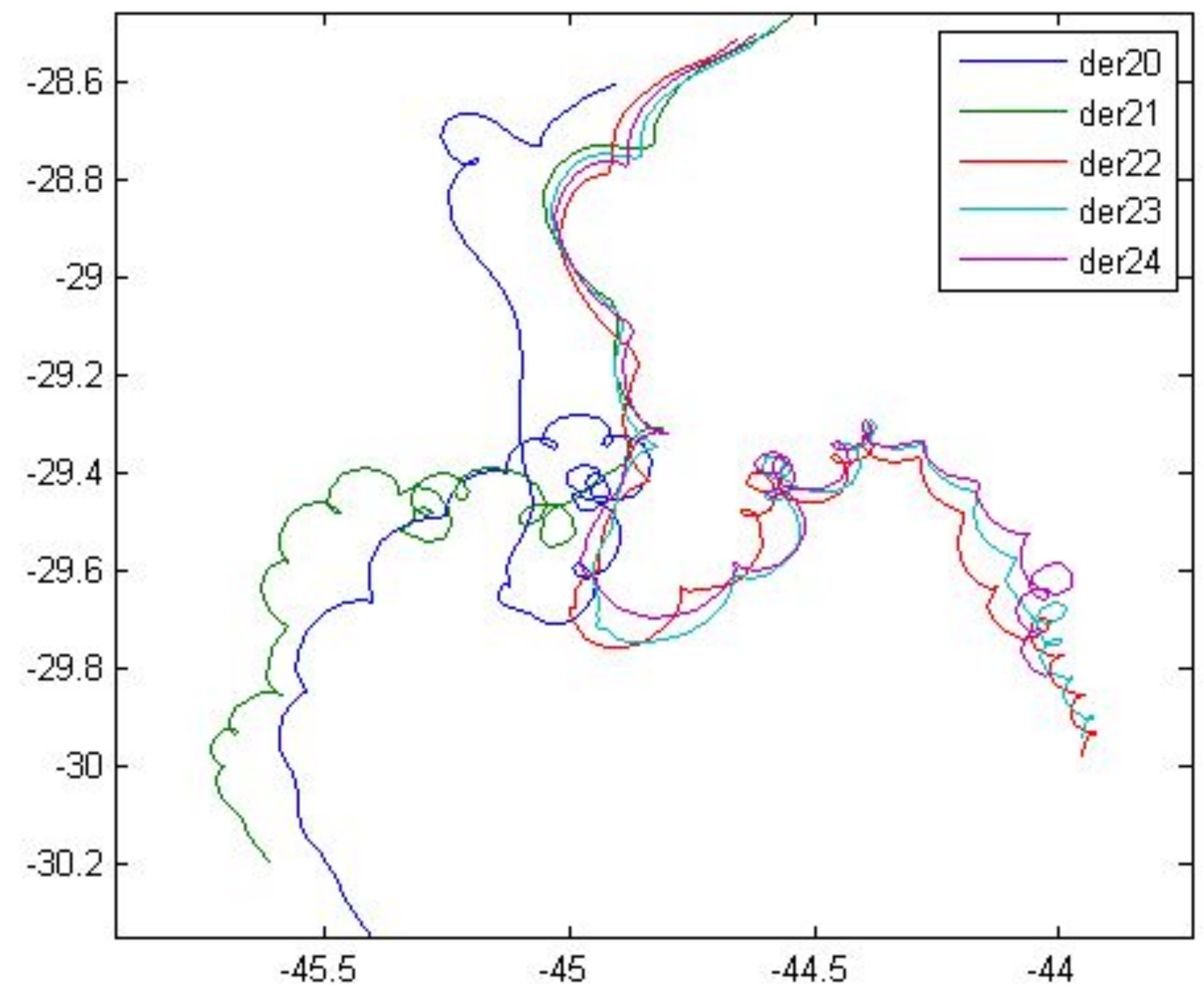
⇨ $u = U * \sin(ft)$

$v = U * \cos(ft)$

Circular oscillations with Period=2π/f
(f is the inertial frequency)

Inertial Period P depends on latitude:
10°N = 69 hours
30°N=24 hours
45°N=16.9 hours

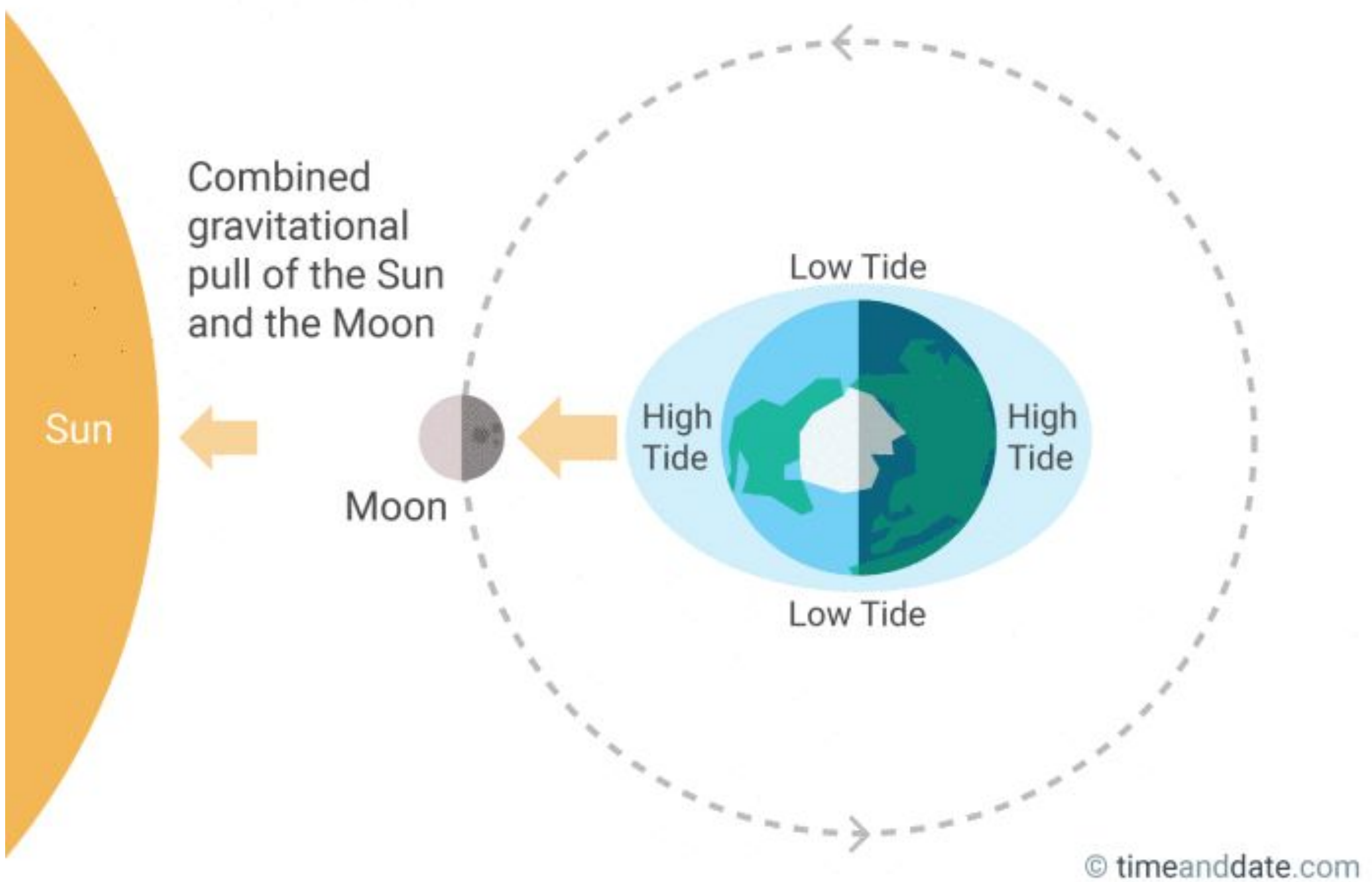
Radius of oscillations: r=U/f



Example of inertial oscillations offshore Brazil

Tidal currents

The ocean tide: Periodic variations of the ocean sea level due to the actions of celestial bodies in rotation around the Earth. More than 400 tidal waves with periods ranging from less than half a day to years.

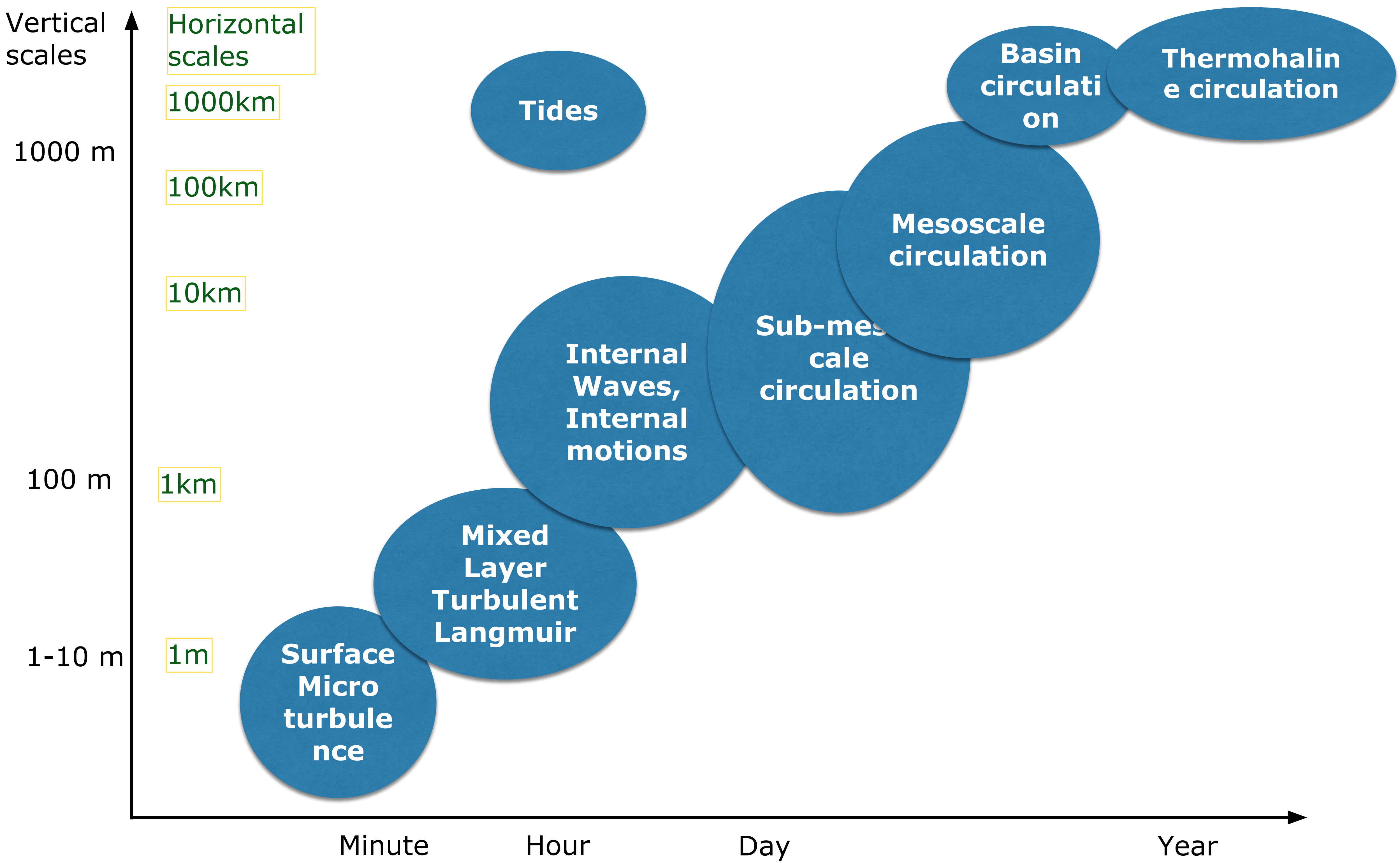


The vertical motion of the tides near the shore causes the water to move horizontally, creating currents.

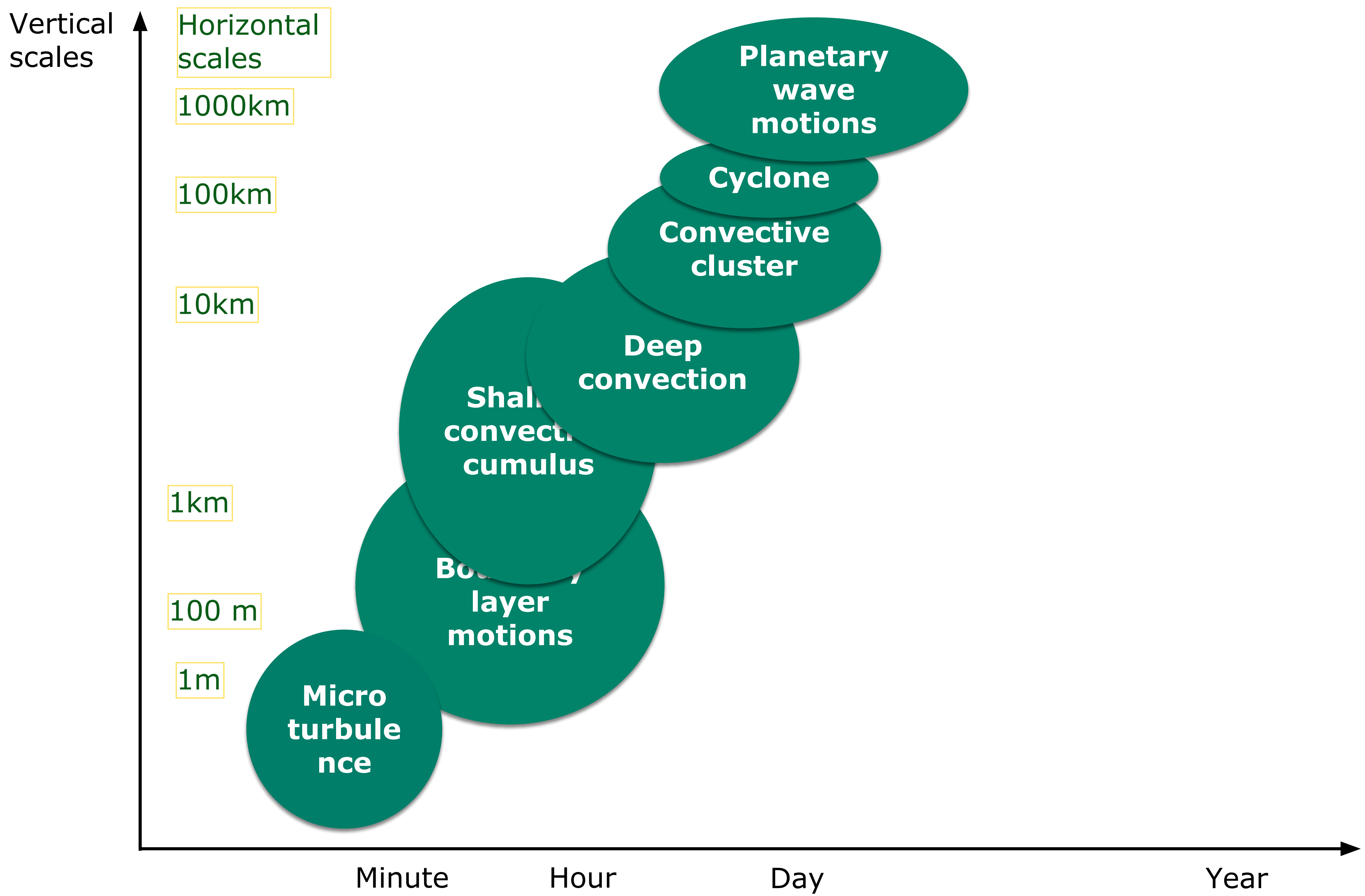
Current intensity depends on :

- the different phases of the moon. When the moon is at full or new phases, tidal current velocities are strong. When the moon is at first or third quarter phases, tidal current velocities are weak.
- the relative positions of the moon and Earth. When the moon and Earth are positioned nearest to each other, the currents are stronger than average . When the moon and Earth are at their farthest distance from each other, the currents are weaker .
- the shape of bays and estuaries also can magnify the intensity of tides and the currents they produce

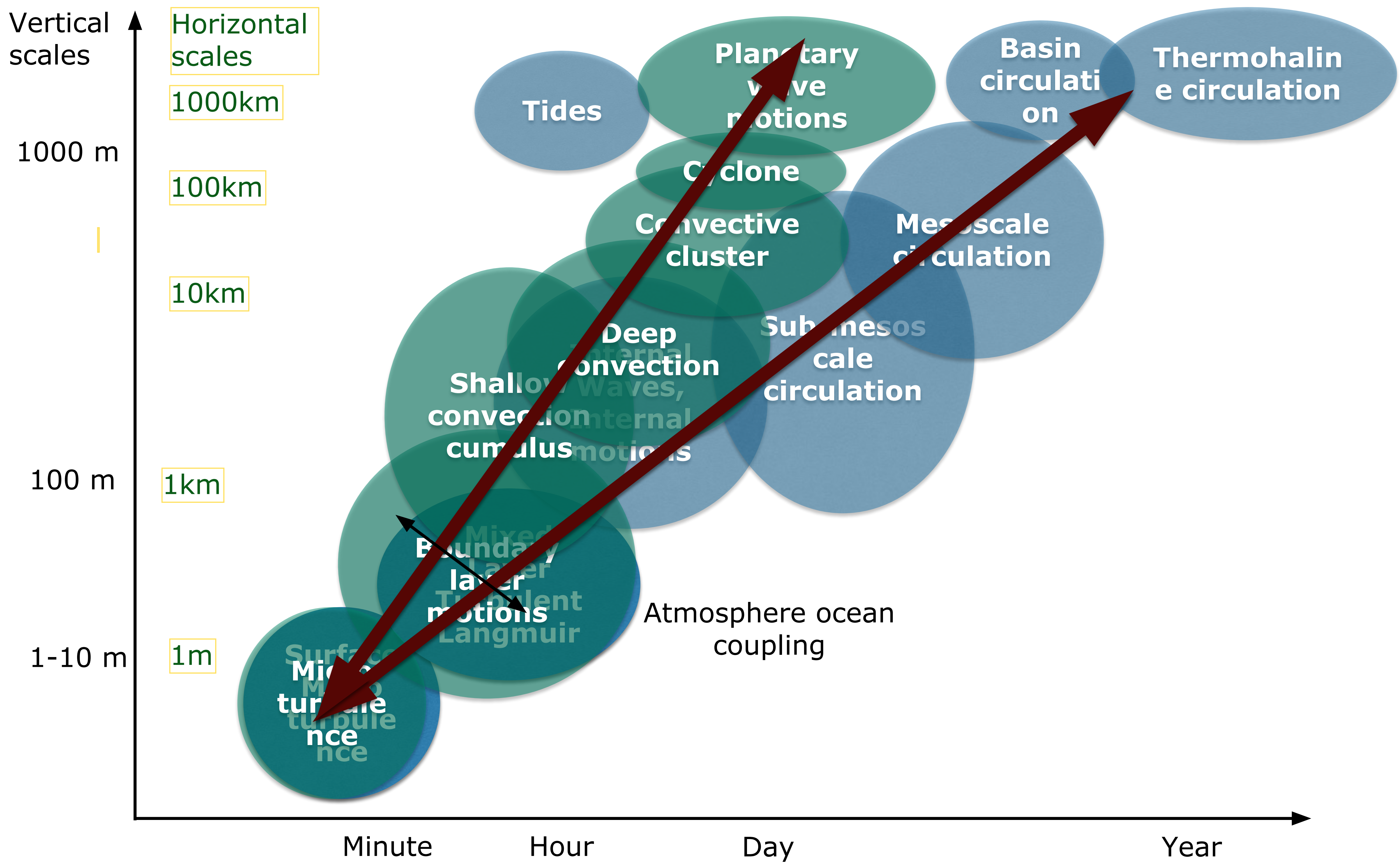
Ocean space-time scales



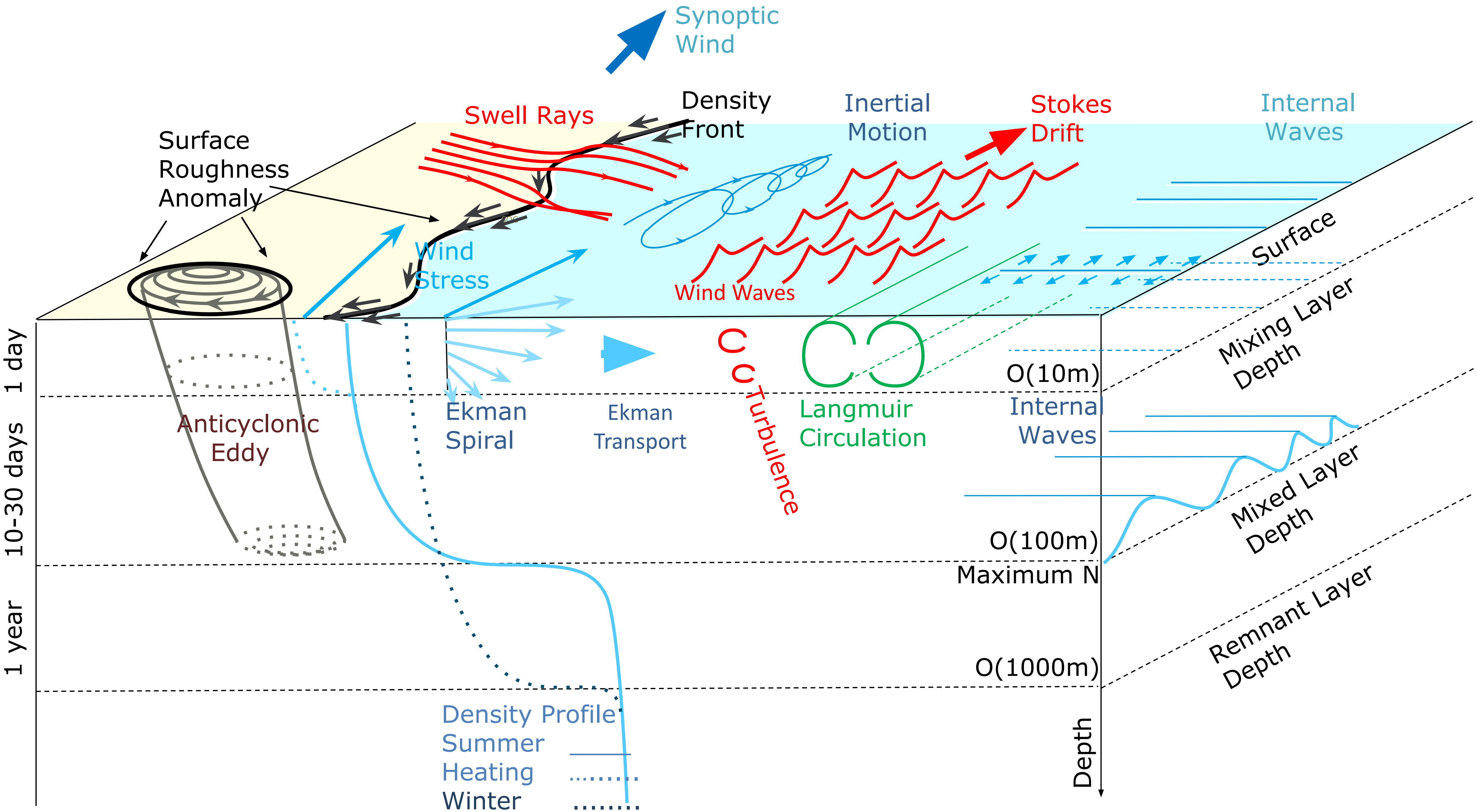
Atmosphere space-time scales



Ocean and Atmosphere space-time scales



Ocean circulation



Ocean circulation monitoring

In-situ measurements

- Temperature/salinity profiles -> steric height h_s

  XBT, CTD, gliders, Argo floats

- Drifting buoys: Lagrangian measurement of the total ocean current at a given depth
- ADCP
- Current meters

Eulerian measurement of the current (no stokes drift)

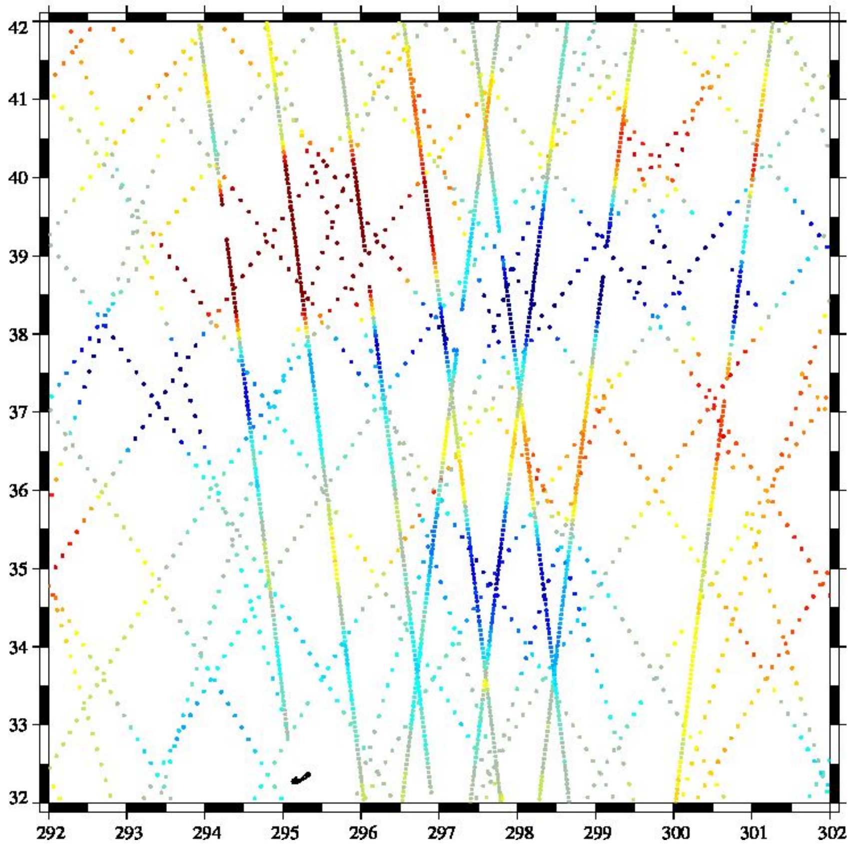
- Coastal HF radar

Remote sensing measurements

- Altimetry for mesoscale and large scale geostrophic currents
- Synthetic Aperture Radar (SAR) Doppler
- Scatterometers (+ altimetry + in-situ drifters) to derive Ekman currents
- Radiometers/Spectrometers for higher resolution currents

Geostrophic currents from altimetry

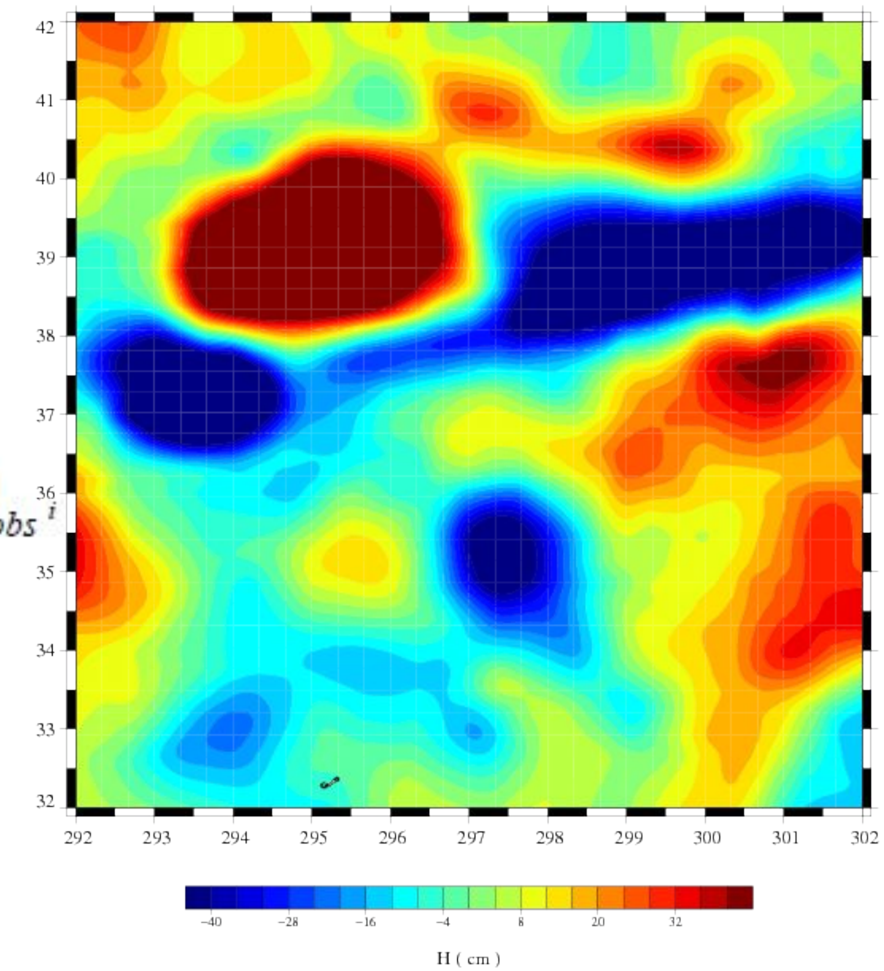
Altimetric anomalies along the tracks from 4 different satellites in the Gulfstream.



Objective Analysis

$$\theta_{est}(\mathrm{x}) = \sum_{i=1}^{n}\sum_{j=1}^{n} A_{ij}^{-1} C_{xj} \Phi_{obs^i}$$

Altimeter anomaly map

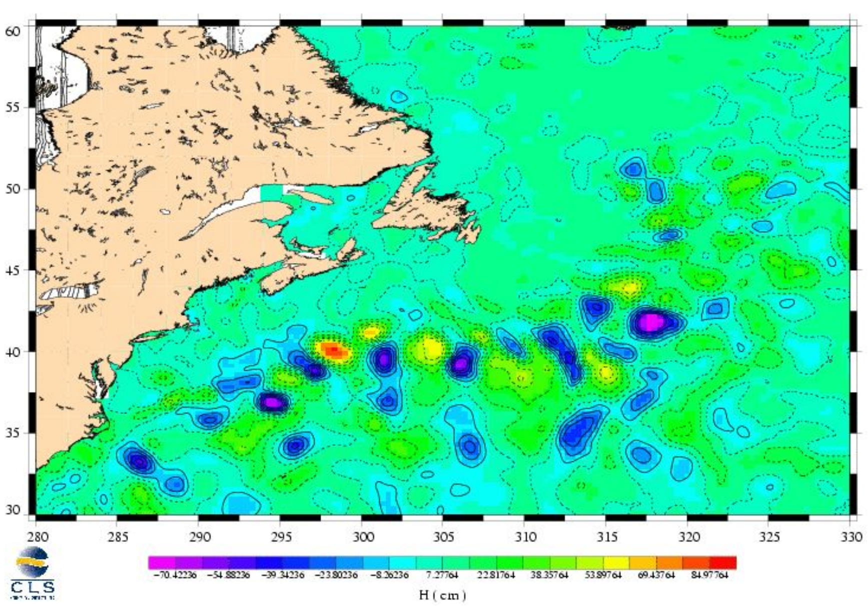


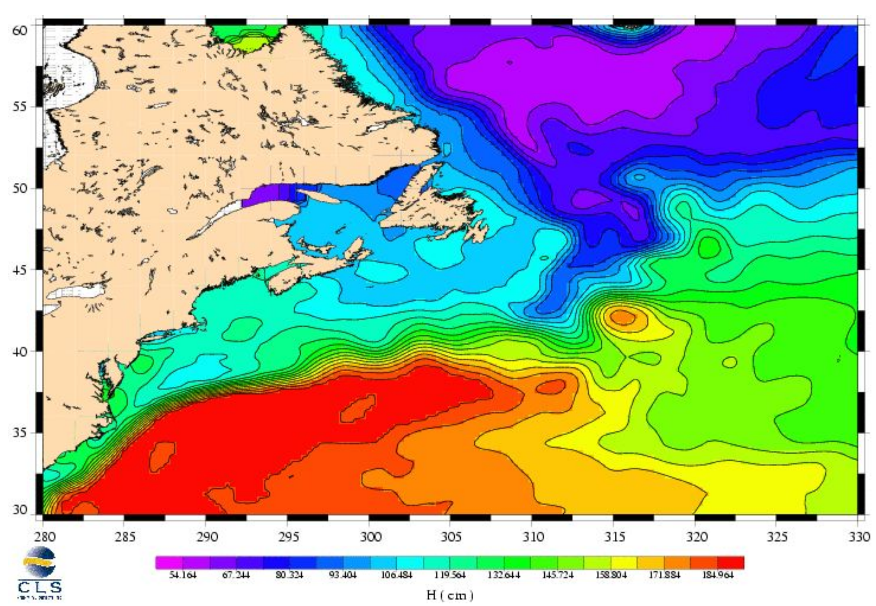
$$A_{ij} = \left\langle \Phi_{obs^i} \Phi_{obs^j} \right\rangle = \left\langle \Phi_i \Phi_j \right\rangle + \left\langle \varepsilon_i \varepsilon_j \right\rangle \qquad C_{xi} = \left\langle \theta(x)\, \Phi_{obs^i} \right\rangle = \left\langle \theta(x)\, \Phi_i \right\rangle$$

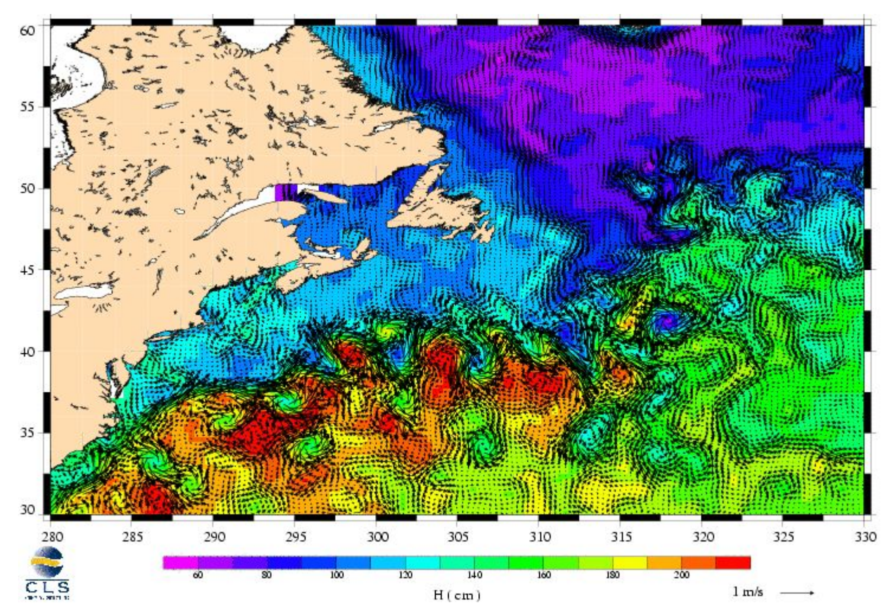
Geostrophic currents from altimetry

From Sea Level Anomaly to the Absolute Dynamic Topography

Altimetric Anomaly (SLA) + Mean Dynamic Topography = Absolute Dynamic Topography







$$u_g = -\frac{g}{f}\frac{\partial h}{\partial y}$$

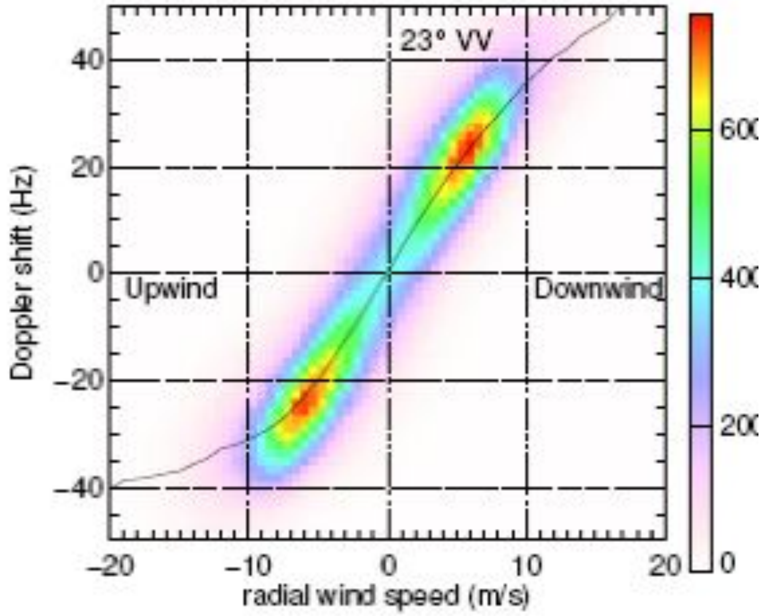
$$v_g = \frac{g}{f}\frac{\partial h}{\partial x}$$

(will be computed in TD)

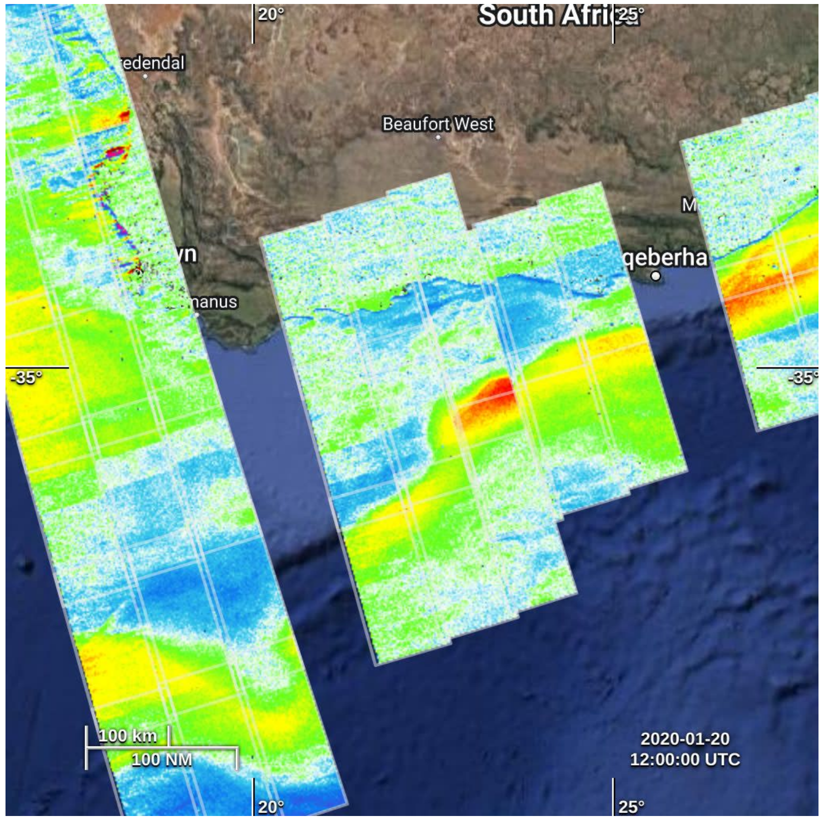
Geostrophic approximation

SAR Doppler

Chapron et al, 2005 ; Johannessen et al, 2008; Rouault et al, 2010 have demonstrated the strong value of using the Doppler shift measurements from the ENVISAT ASAR data for retrieving the radial component of the surface current.



A Doppler shift is measured between the Signal emitted by the instrument and the signal **backscattered** by the sea surface and **measured** by the SAR antenna. It is due to:



- The known movement of the satellite in orbit,
- A wave-state contribution highly correlated to wind speed which can be estimated using an empirical relationship between the range Doppler velocity and the near surface wind field, Mouche et al. (2012) with a C-band Doppler (CDOP) algorithm. These local wind contributions are mainly **from wave orbital motion, but also from Ekman and Stokes drift.**
- A **measure of the sea surface current**, that contains the contributions, projected onto the range direction, of the **total currents**.

Ekman/Stokes currents from wind

$$u_e = \pm \frac{\pi\sqrt{2}}{\rho(f+w)D_e} e^{\frac{\pi}{D_e}z} * \tau_e * \cos\left(\frac{\pi}{4} + \frac{\pi}{D_e}z\right)$$

$$v_e = \frac{\pi\sqrt{2}}{\rho(f+w)D_e} e^{\frac{\pi}{D_e}z} * \tau_e * \sin\left(\frac{\pi}{4} + \frac{\pi}{D_e}z\right)$$

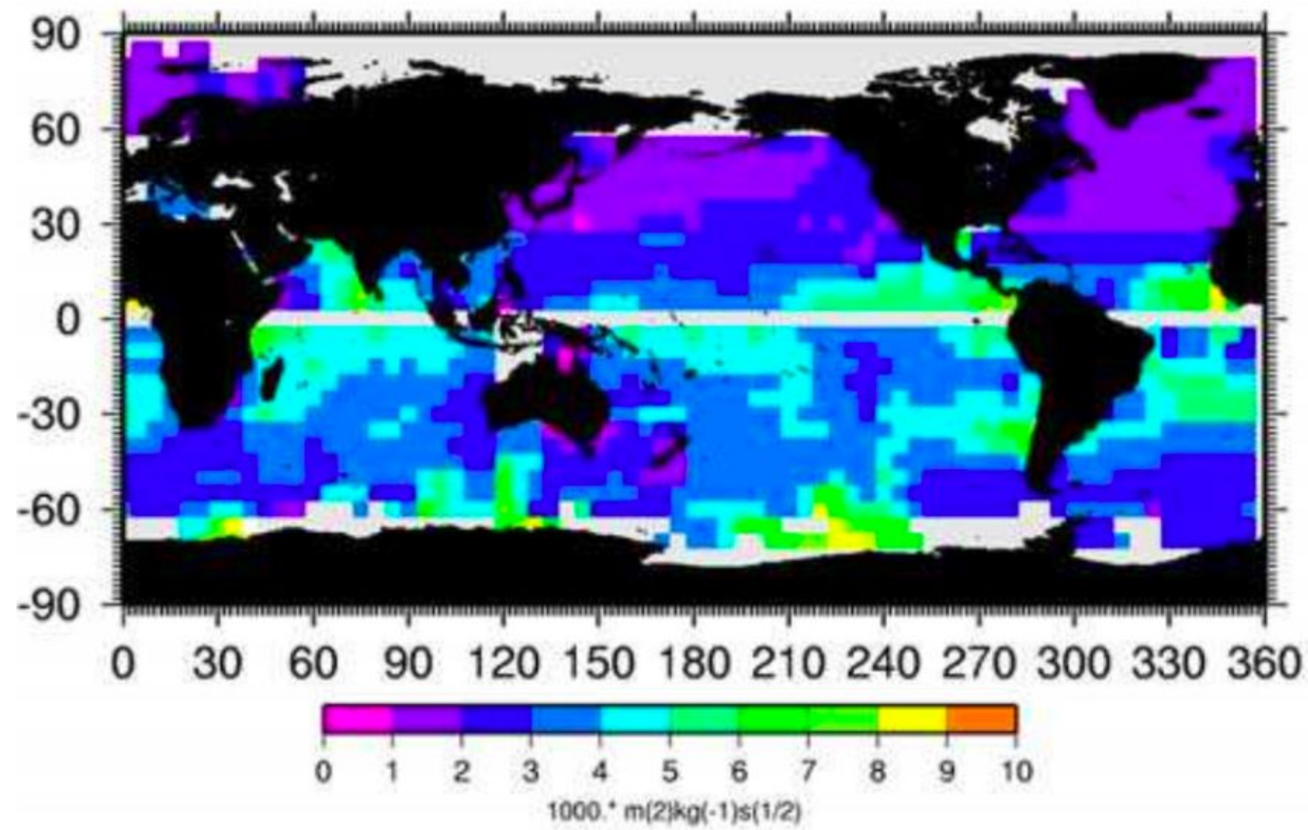
$$\vec{u}_e = \frac{b\vec{\tau}}{\sqrt{f}} e^{i\theta}$$

(will be computed in TD)

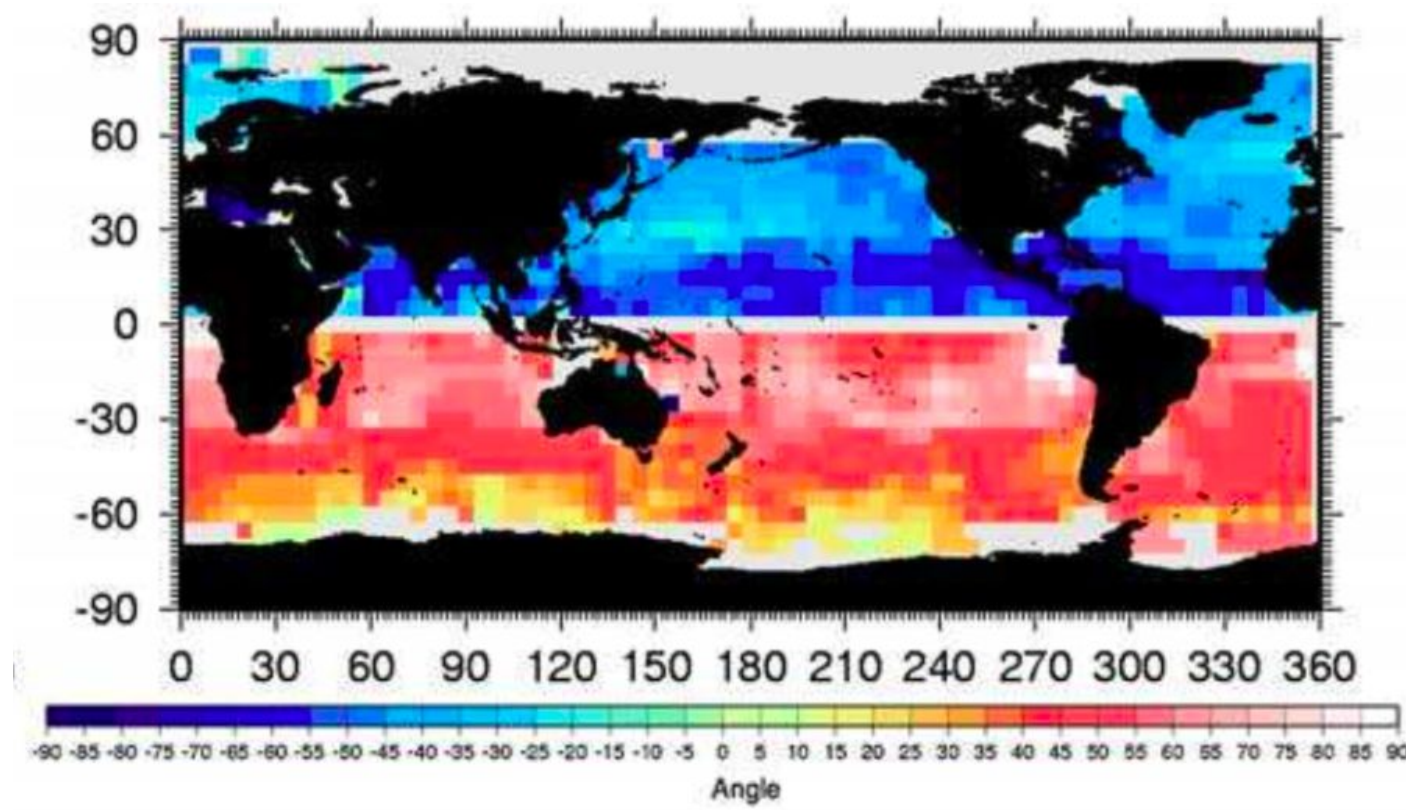
Where τ is the wind stress from model or observed wind, b the amplitude factor and θ the phase difference between the wind-driven currents and the wind stress.

b and θ parameters are determined by comparing altimetric geostrophic currents with in-situ surface velocity from drifters (see [Rio et al 2003]).

b coefficients in winter

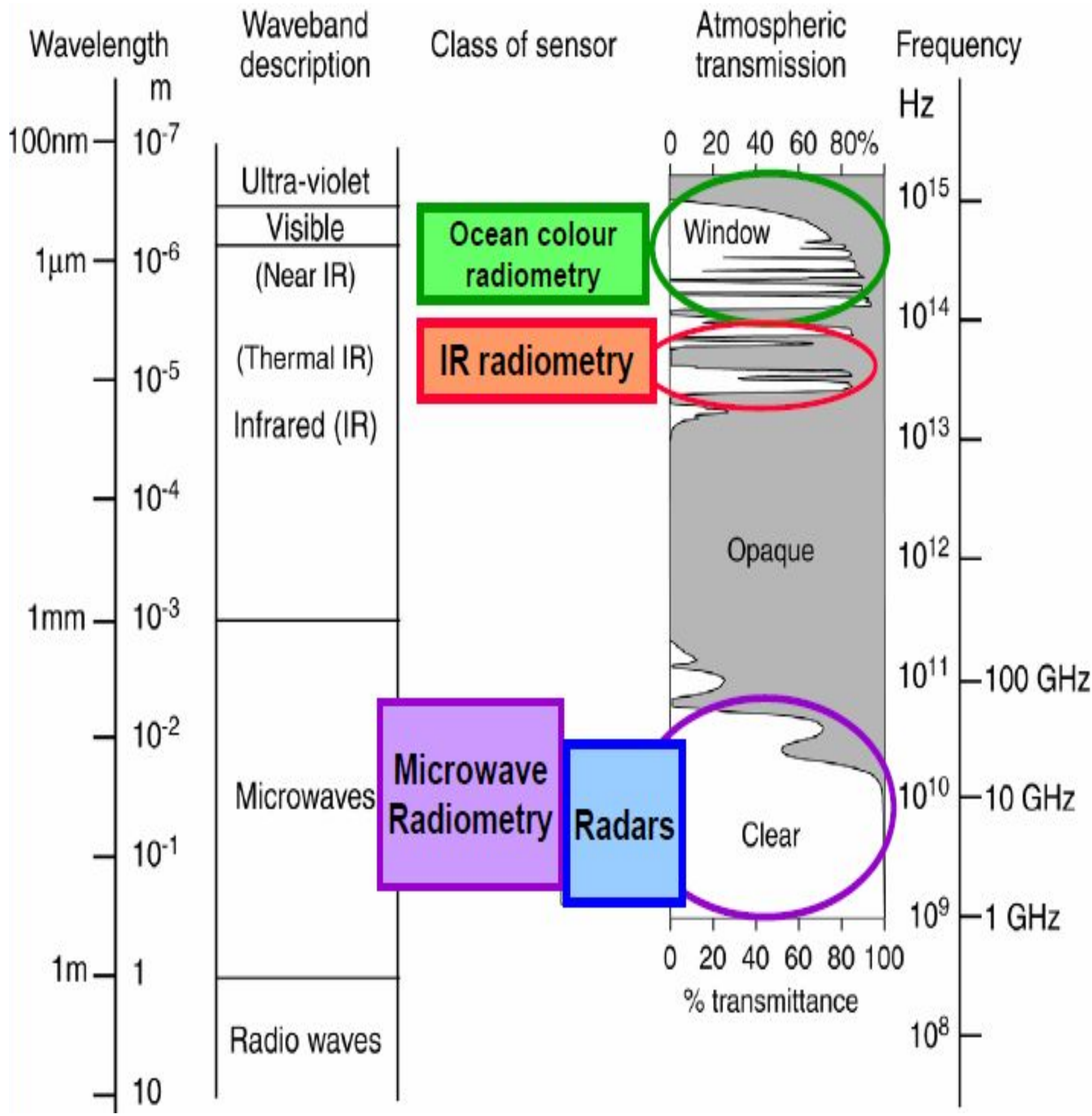


θ coefficients in winter



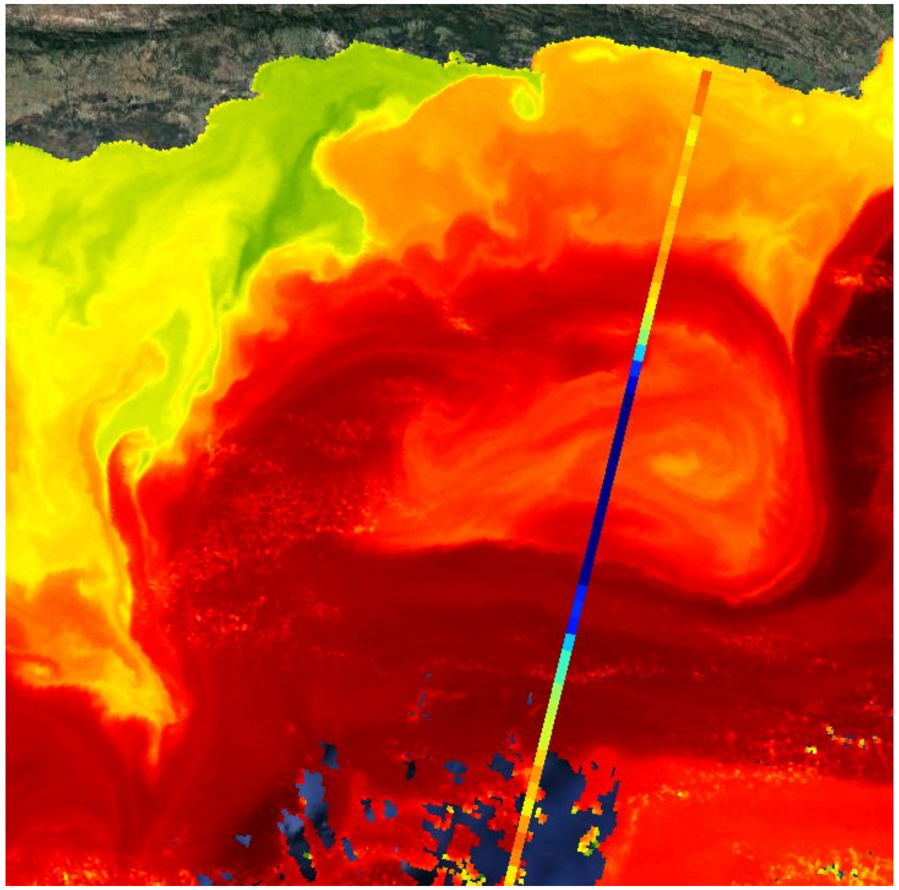
Synergy with indirect observations

- Radiometers
  - Ocean color: from the spectral signature of the sun radiation reflected at the surface of the ocean.
  - Temperature, Salinity: from the energy emitted by the surface of the ocean.

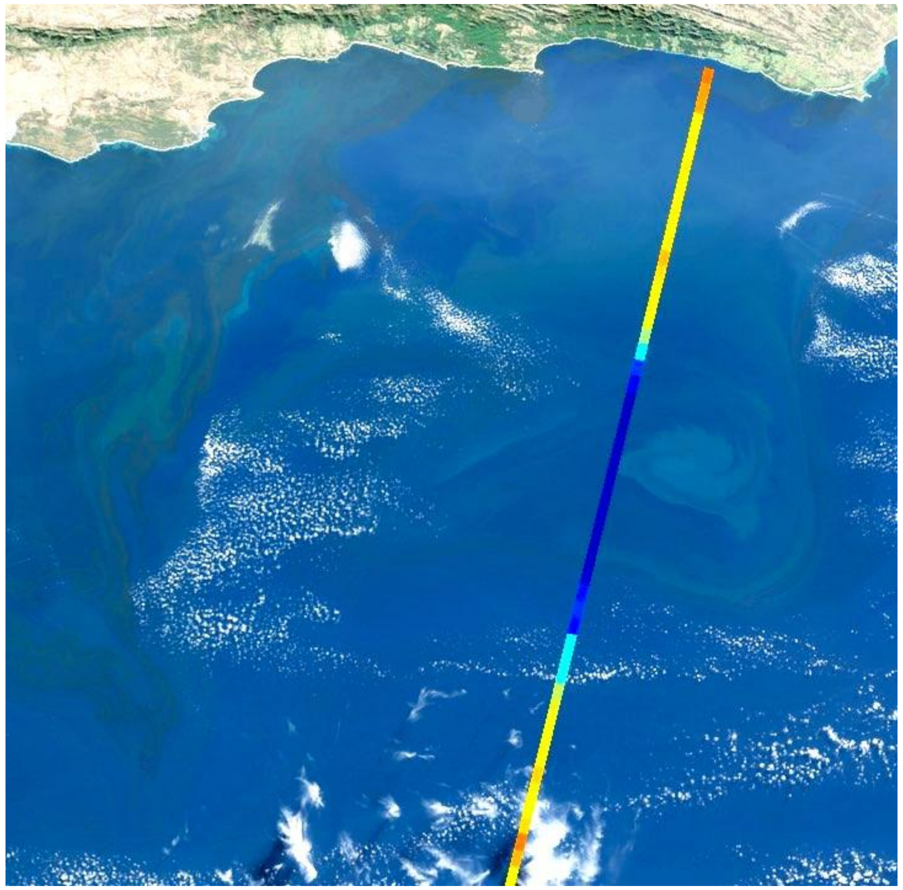


Synergy with indirect observations

- Ocean Color, Temperature and Salinity can be considered as passive tracers advected by the ocean currents.
- There are different methods to benefit from the synergy tracer-currents to improve our assessment of the upper ocean motions:
  - ➡ the analysis of successive images informs on the ocean current field ***Maximum Cross Correlation (MCC), optical flow***
  - ➡ Under favorable environmental conditions, the streamfunction ψ from which geostrophic velocities are derived, can be calculated from surface density values from which SST may be considered as a proxy ***e-SQG approximation***



Synergy from Sentinel3 sensors:
SSH from SRAL,
Ocean Color from OLCI,
and SST from SLSTR.



Effective Surface Quasi-Geostrophy

Under favourable environmental conditions, the streamfunction h from which geostrophic velocities are derived, can be calculated from surface density values (Lapeyre et al, 2006; Klein et al., 2008):

Inversion of the Quasi Geostrophic Potentiel Vorticity conservation equation in the horizontal Fourier transform domain (valid for space scales of 10-200km)

$$\begin{cases} \dfrac{\partial}{\partial y} h_{sqg} = -u \\ \dfrac{\partial}{\partial x} h_{sqg} = v \end{cases} \Leftarrow \hat{h}_{sqg}(\vec{k}, z) = \frac{f}{N_{eff}\rho_0 k} \hat{\rho}'_s(\vec{k}) \cdot e^{\left(\frac{N_{eff} k z}{f_0}\right)} \Rightarrow \rho'_s = -\alpha \cdot T_s' - \beta S_s' = -\alpha' T_s'$$

Currents Streamfunction Surface Density Anomalies **SST anomaly**

N_{eff} is the effective Brunt-Vaisala frequency (constant stratification assumed)
$\alpha' N_{eff}^{-1}$ is a free parameter that needs to be set up to account both interior PV and the partial compensation of salinity and temperature.

Limitations:

- The coldest SST anomalies are reported to trace the lowest SSH anomalies for all seasons, while the warmest SST anomalies solely match the largest SSH anomalies during winter.

- SST-derived SSH reconstruction using the surface quasi geostrophic approximation should take into account stratification effects, especially during summer

Optical Flow: tracer conservation

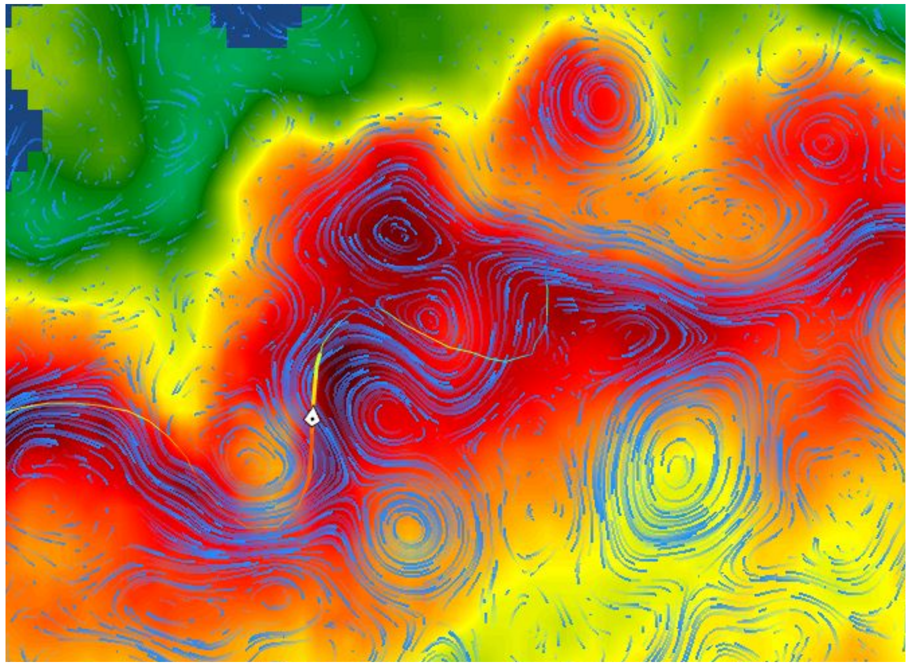
Require the velocity field (u,v) to obey the tracer concentration c evolution equation and inverse it for the velocity vector:

$$\frac{\partial c}{\partial t} + u\frac{\partial c}{\partial x} + v\frac{\partial c}{\partial y} = F(x, y, t)$$

c represents the concentration of any tracer as Sea Surface Temperature, Sea Surface Salinity, Chl-a concentration,
F(x,y,t) represents the source and sink terms

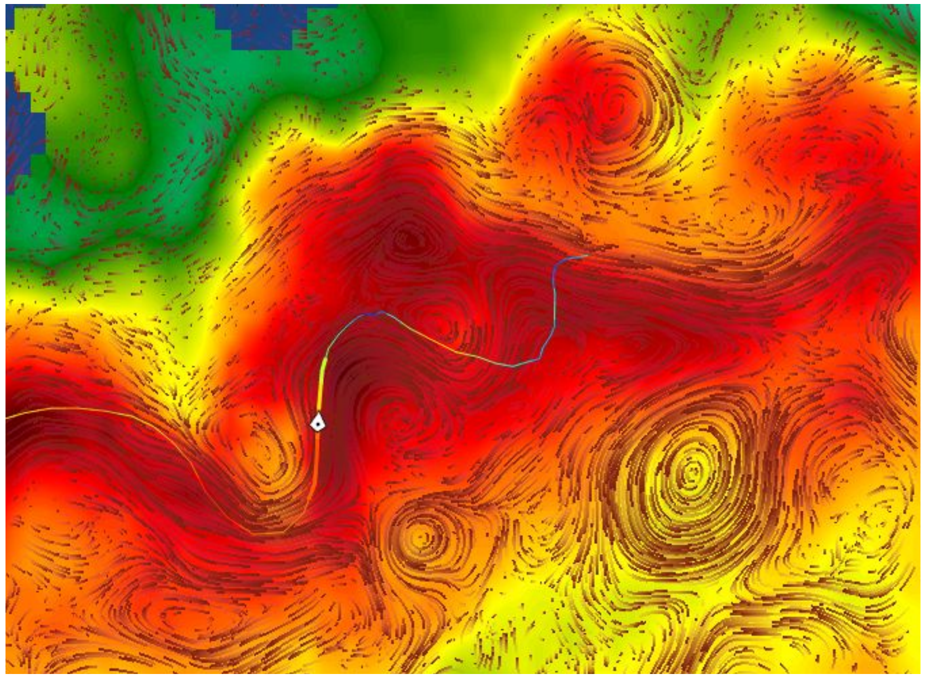
Limitation:

- only along-gradient velocity information can be retrieved from the tracer distribution at subsequent times in strong gradients areas.
- Strong uncertainties on external fluxes



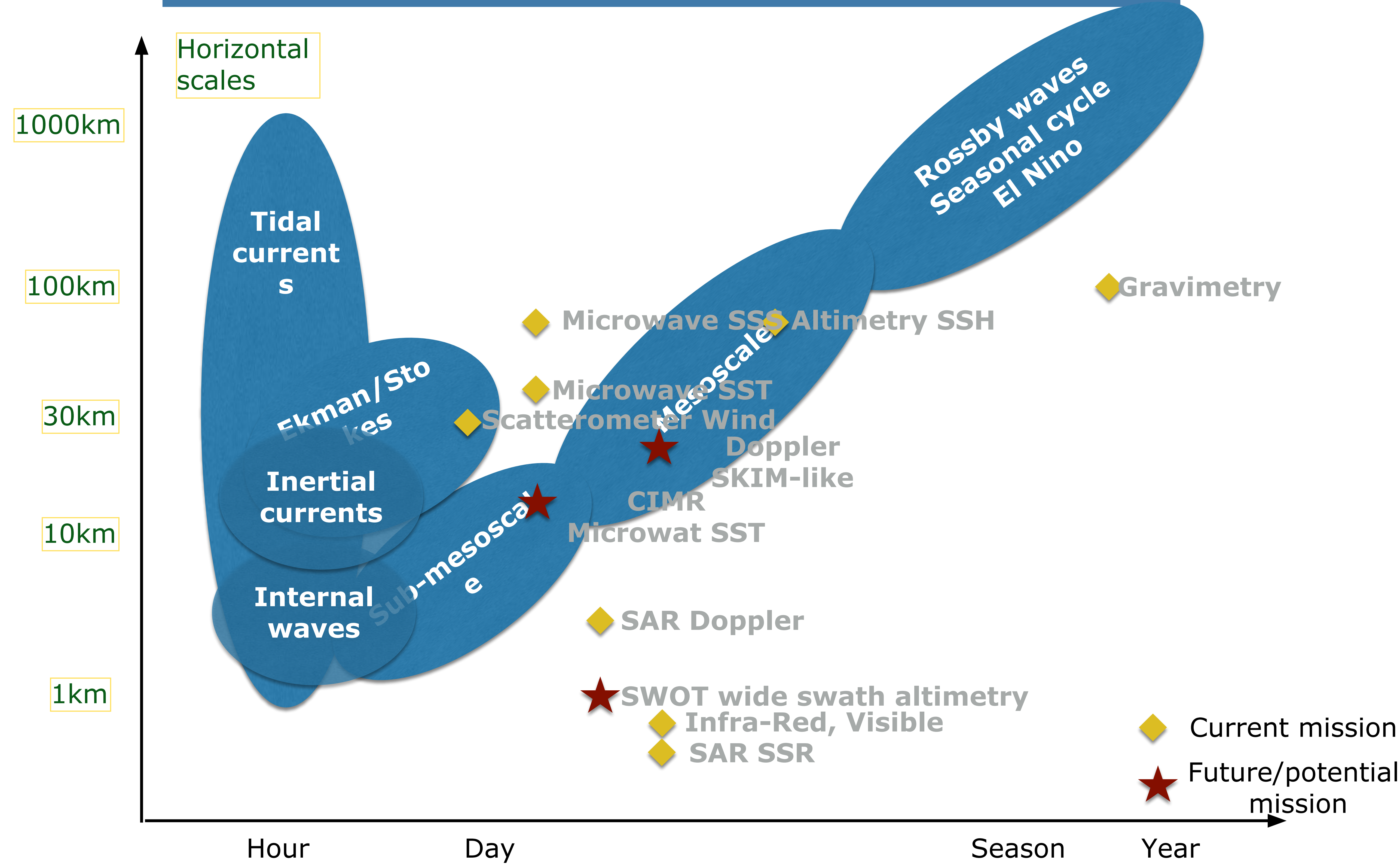
Altimetry derived geostrophic current over Microwave SST and surface drifter

Correct geostrophic current derived from altimetry using SST gradient



Corrected current over Microwave SST and surface drifter

Upper ocean circulation monitoring



3D Ocean circulation monitoring

Observing system	Coverage	Spatial resolution	Temporal resolution	Current component measured
Hydrological profiles (XBT, CTD, gliders, Argo floats	Surface and depth sparse	Vert. 1 m	20 minutes	Baroclinic component of the geostrophic current
Drifting buoys	Surface and depth sparse	Along-track	Surf: 1 hour/ 6 hours Deep: 1 day	Total current
ADCP	Surface and depth sparse	Vert. 10 m	1 hour	Total current
Current meters	Surface and depth sparse		30 minutes	Total current
HF radar	Surface costal	5-20 km	< 1 hour	Total (?) current
Altimeter	surface Global	100 km	10 days	Geostrophic current
SAR	Super sites	4-8 km	2-3 days	Radial component of total current minus wind drift
SST	MW: global IR: cloud sensitive	25 km 10 km	1 day	SQG: geostrophic MCC/OF: total / radial current in strong gradient areas

Any question?

Let’s move to the practicals to compute Gesotrophic and Ekman Currents from remote sensing observation

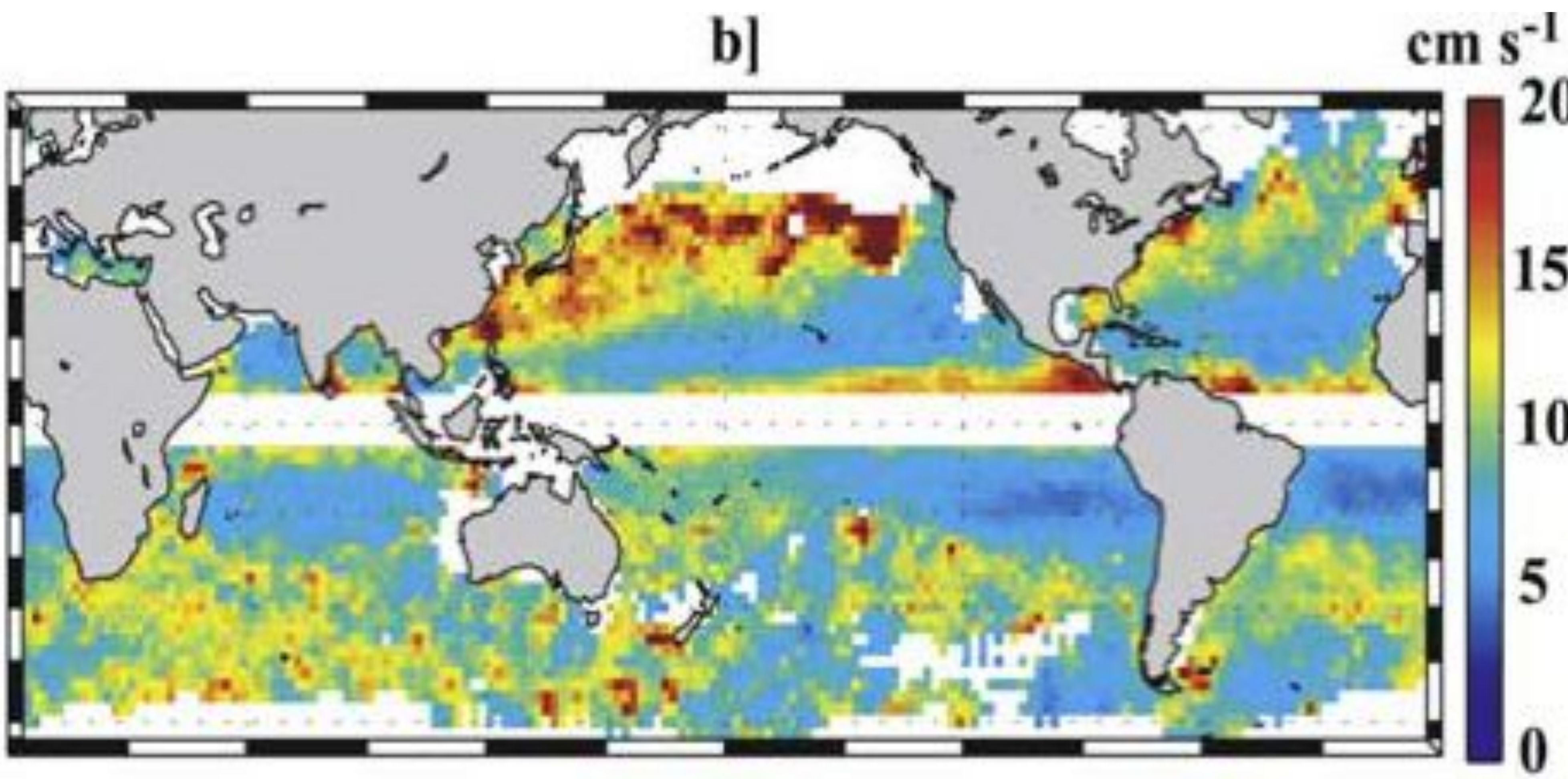
Inertial oscillations

As friction cannot be completely neglected, inertial oscillations in the real ocean decay in a few days. The amplitude of the inertial motion is proportional to the cumulative wind forcing term and inversely proportional to the water density and thickness of the mixed layer (Park et al, 2005).

Park et al (2005) :

- ✓ inertial amplitudes in the range 0-80 cm/sec with an average value of 13.7 cm/sec.
- ✓ inertial amplitude in the mid-latitude (30-45°N) band exceeds those in both the low (15-30°N) and high (45-60°N) latitude bands
- ✓ In three basins, the amplitude in summer is greater than that in winter by 15%-25%.

Inertial current amplitude from surface drifters. From Chaigneau et al (2008).



Future Doppler missions

Different concepts with a common strategy:

- Delayed Doppler effect is used to infer the sea motion in the satellite range direction
- Each scene is viewed from 2 or more azimuth angle to get motion vector

SEASTAR: Squinted Along-track interferometric SAR:
two-dimensional maps of total ocean surface current vectors and wind vectors at 1km resolution with unprecedented accuracy, supported by coincident directional swell spectra in coastal, shelves and polar seas.

DOPSCAT (EU)/ WaCM (Wind and Current Mission, US): : scatterometry with Doppler capability to provide simultaneous measurements of marine winds and surface currents. The mission seeks to monitor global surface ocean currents on a daily basis with a spatial resolution around 25km and errors better than 0.1-0.2 m/s

SKIM (Sea surface Kinematics Multiscale monitoring – EE9 candidate): Doppler-enabled rotating near-nadir Ka-band altimeter to measure total surface current vectors with an accuracy of 0.1 m/s for 40km / 10 days resolution together with the full directional wave spectrum.