→ ESA ADVANCED OCEAN SYNERGY TRAINING COURSE 2019

4–8 November 2019 | Center of Mediterranean Architecture | Chania, Greece

Ocean Surface Currents from Space

Marie-Helene Rio

ESA-ESRIN, Frascati, Italy

NO direct measurement of ocean surface currents from space BUT…

ERS, ENVISAT,CRYOSAT, SENTINEL-3 TOPEX, JASON

Sea Surface Height

ENVISAT, Sentinel-1

SAR doppler velocity

GOCE, GRACE

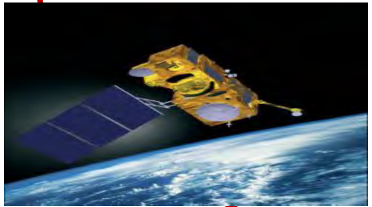
Geoid

Metop, QuickScat

Wind speed

Ocean Surface Currents

ENVISAT, Sentinel-3, MSG, Metop



Sea Surface Temperature

SMOS, Aquarius

Sea Surface Salinity

ENVISAT, Sentinel-2, Sentinel-3

Ocean Color

Sensor	Measured variable	Method	Surface current component retrieved	Spatio-temporal resolution
Altimeter + Gravimeter	Sea level above reference ellipsoid Geoid above reference ellipsoid	Optimally interpolated gridded field + Geostrophic approximation	Geostrophic current	100-400km 10-30 days
Scatterometer	Wind	Ekman model	Ekman current	25 km 12 hours
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Infrared Radiometer	SST	Optical flow, MCC	Total surface currents	Polar orbiting 10 km 1 day Geostationnary 10 km hourly
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SAR	Range Doppler Anomaly Shift	CDOP sea state component of Doppler shift	Radial component of total current minus wind drift (included in CMOD)	Snashots 10 km 3 days

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The geostrophic circulation

$E<10^{-3}$ $R_0<10^{-3}$ and $w<<u,v$

Away from the boundary layers and away from the equator, over large (> 50-100 km) spatial and long (>2-10 days) temporal scales ocean is to the first order in geostrophic balance.

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

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

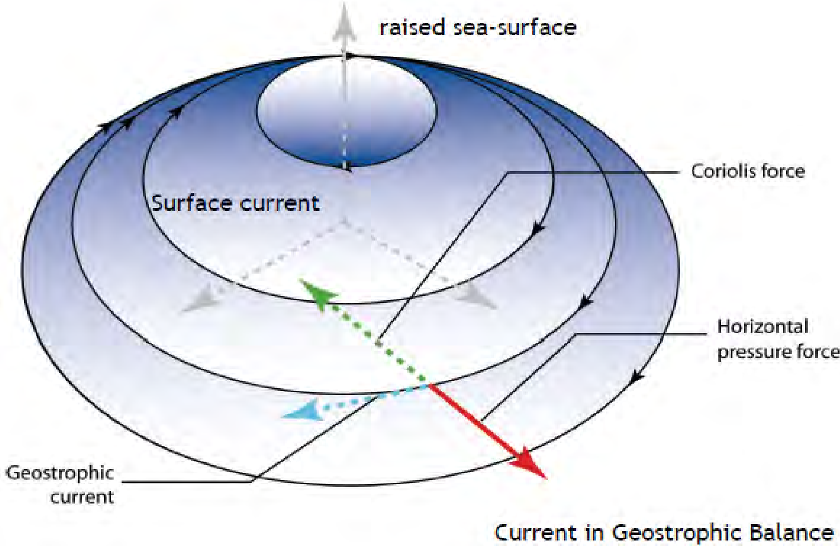
$$0 = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu$$

+ *Hydrostatic equation*

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

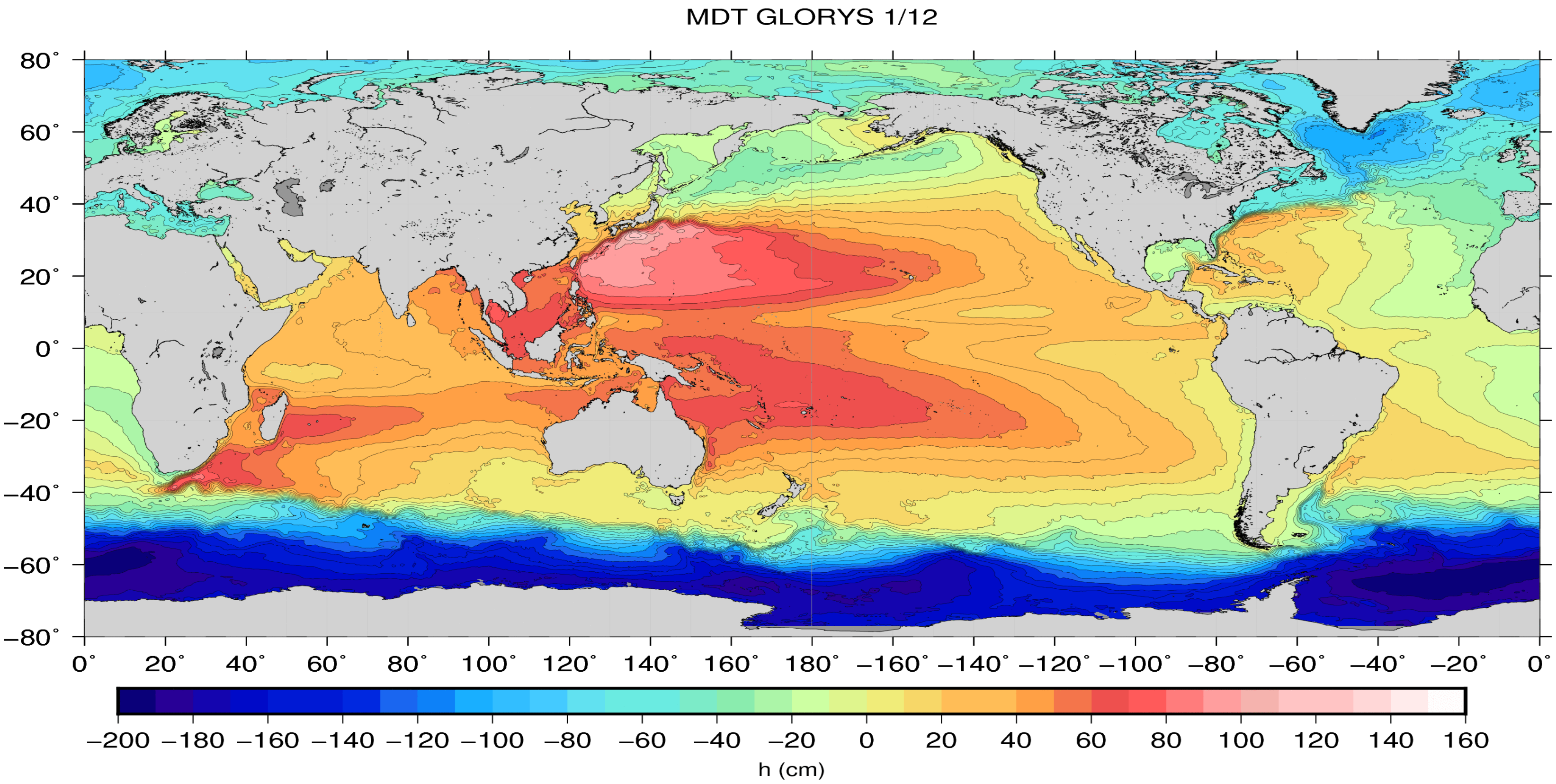
$$u_{geo} = -\frac{g}{f}\frac{\partial h}{\partial y}$$

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

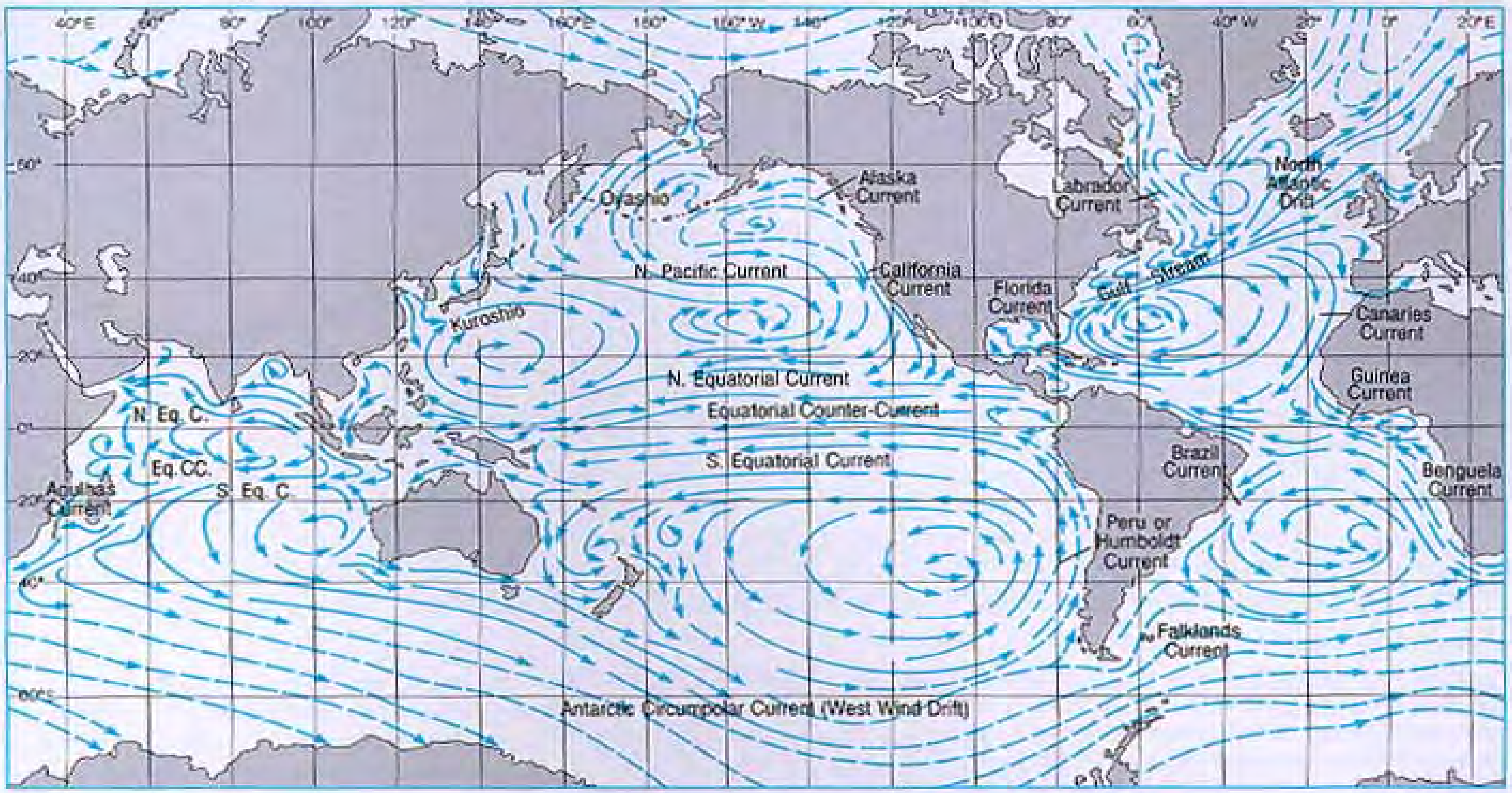


The ocean surface velocity field (u,v) can be readily obtained from the gradients of h, the sea level above the geoid h.

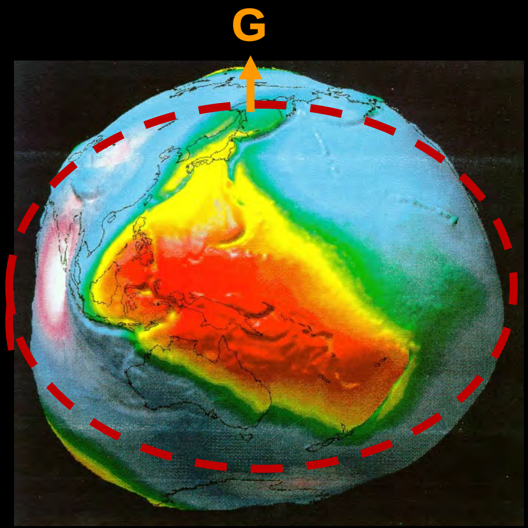
Mean Dynamic Topography from a high resolution (1/12°) ocean numerical model



The General Ocean Circulation



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



E : Reference Ellipsoid

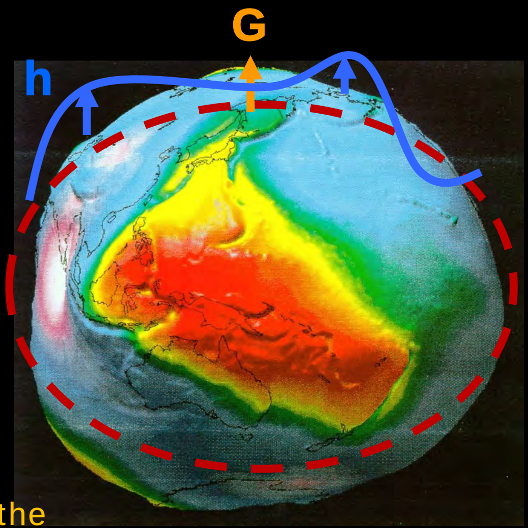
Equipotential of the gravity field

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

Gravity forces generating tides

Thermal forcing

Variations of the Atmospheric pressure



Wind effects

Hydrological Cycle

Coriolis Force due to the Earth Rotation

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**

Estimating ocean surface currents from space:

Altimetry + Gravimetry

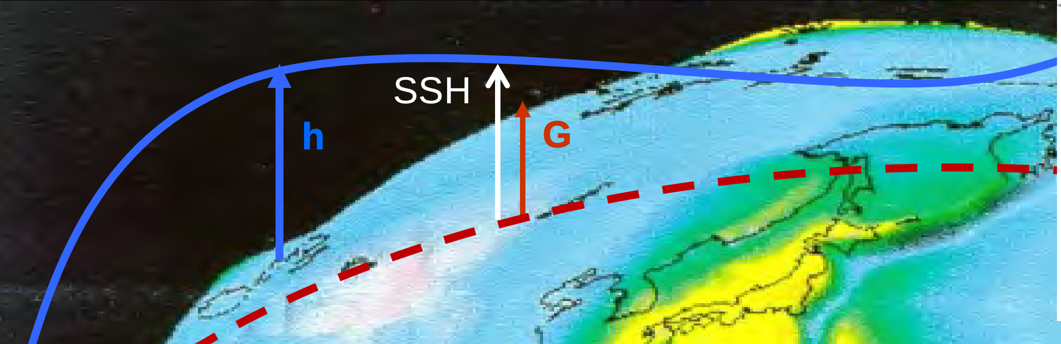
Some very simple equations

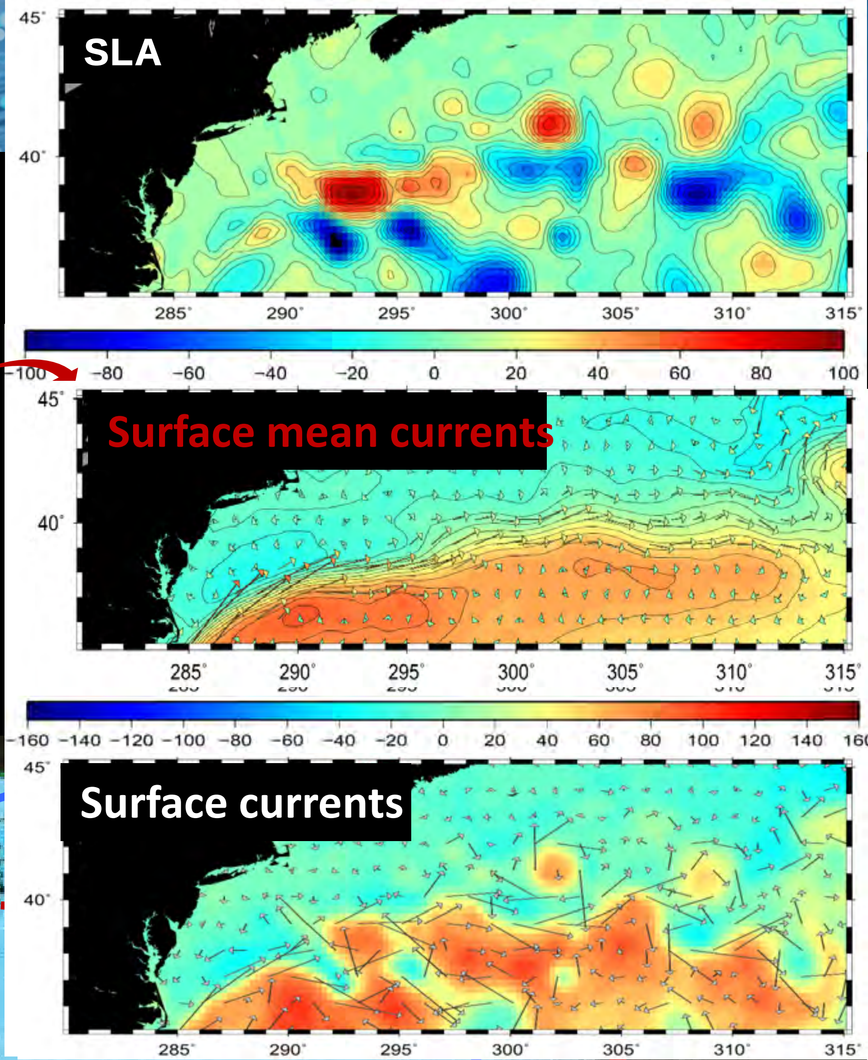
SSH=h+G ⟷ h=SSH-G

MSSH=MDT+G ⟷ MDT=MSSH-G

SSH-MSSH=h-MDT=SLA

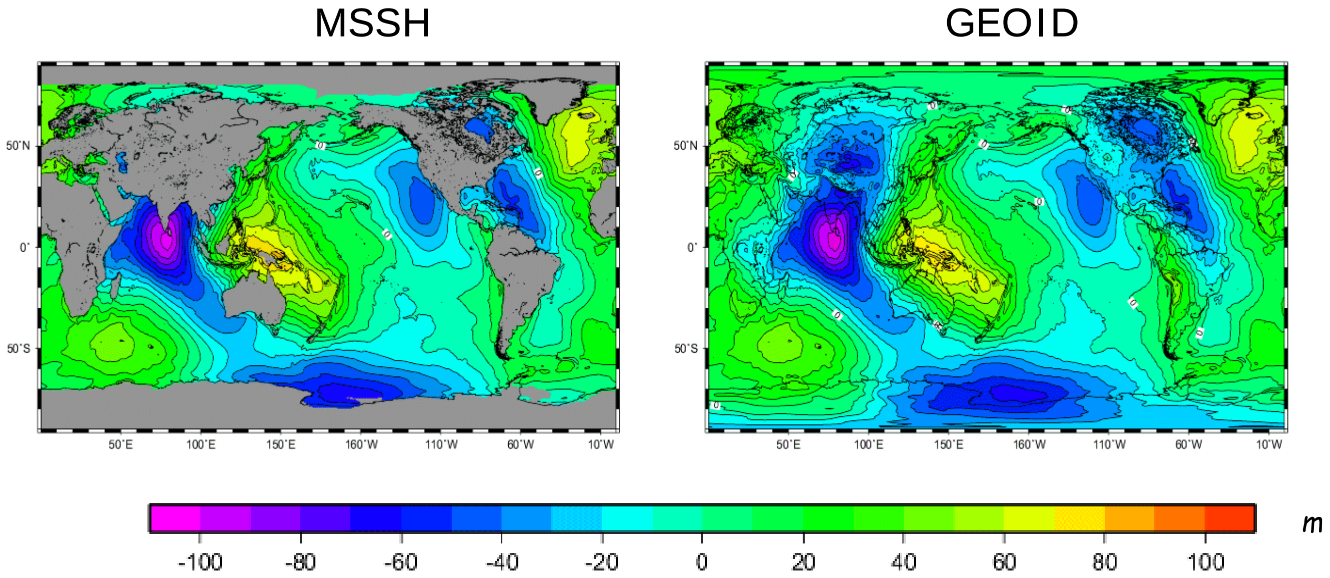
h=SLA+MDT





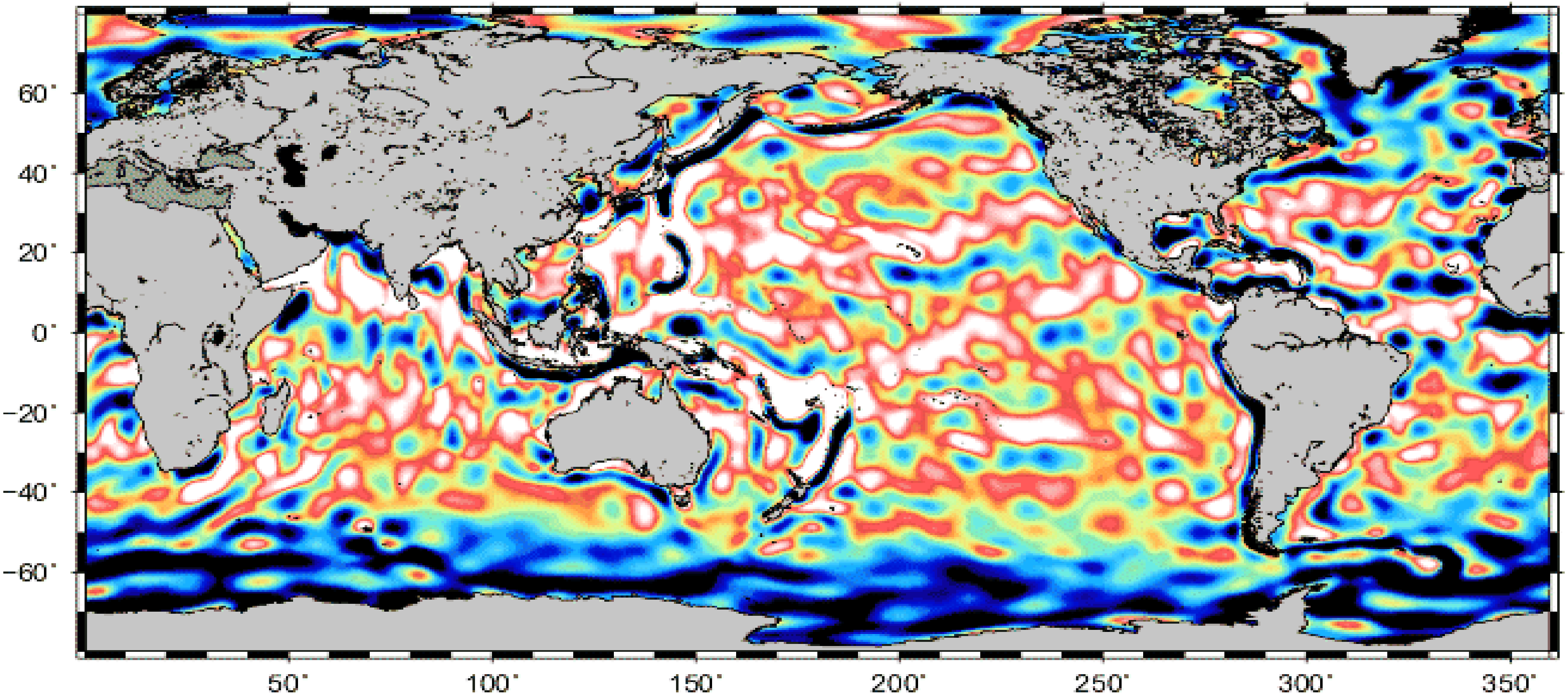
The MDT from altimetry and gravity data

MDT=MSSH –GEOID



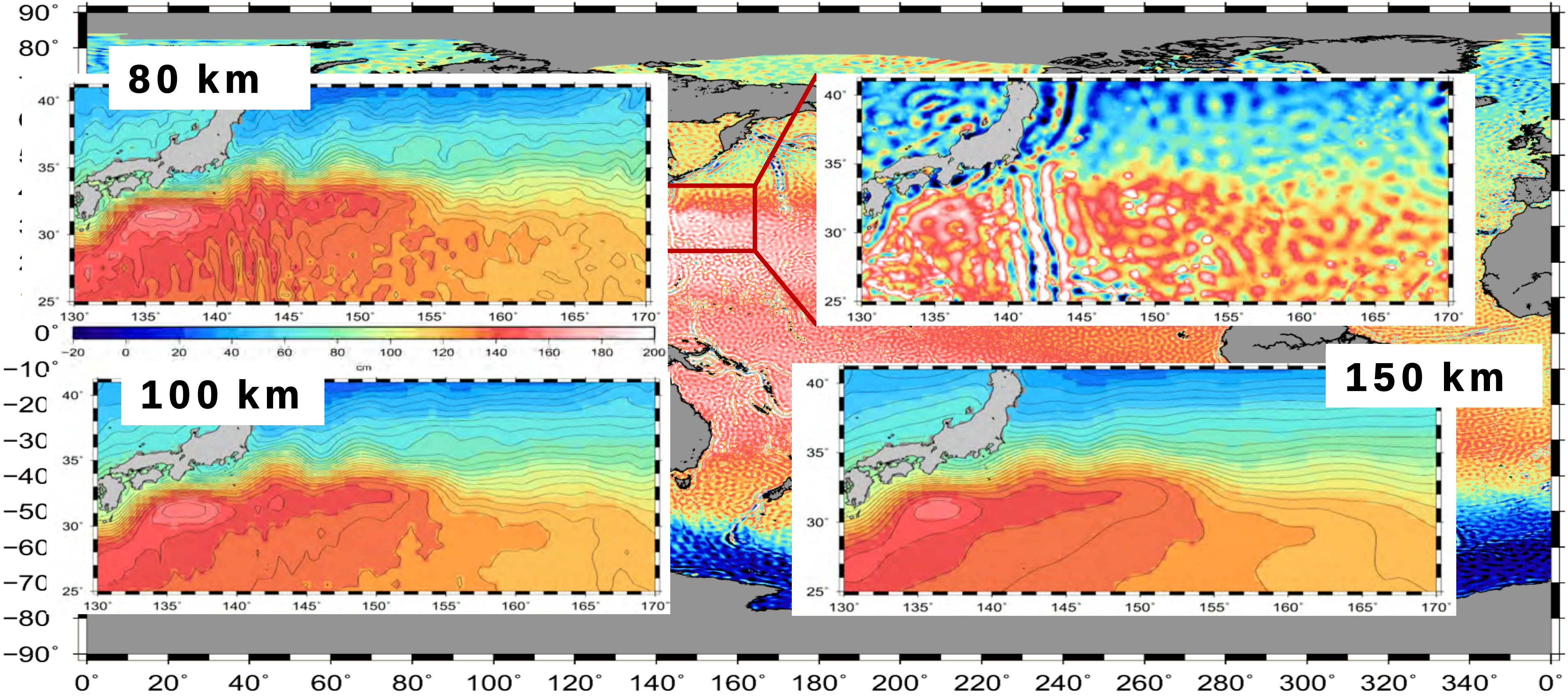
The MDT from altimetry and gravity data

1995: MDT=MSS OSU95 – CHAMP RAW DIFFERENCE



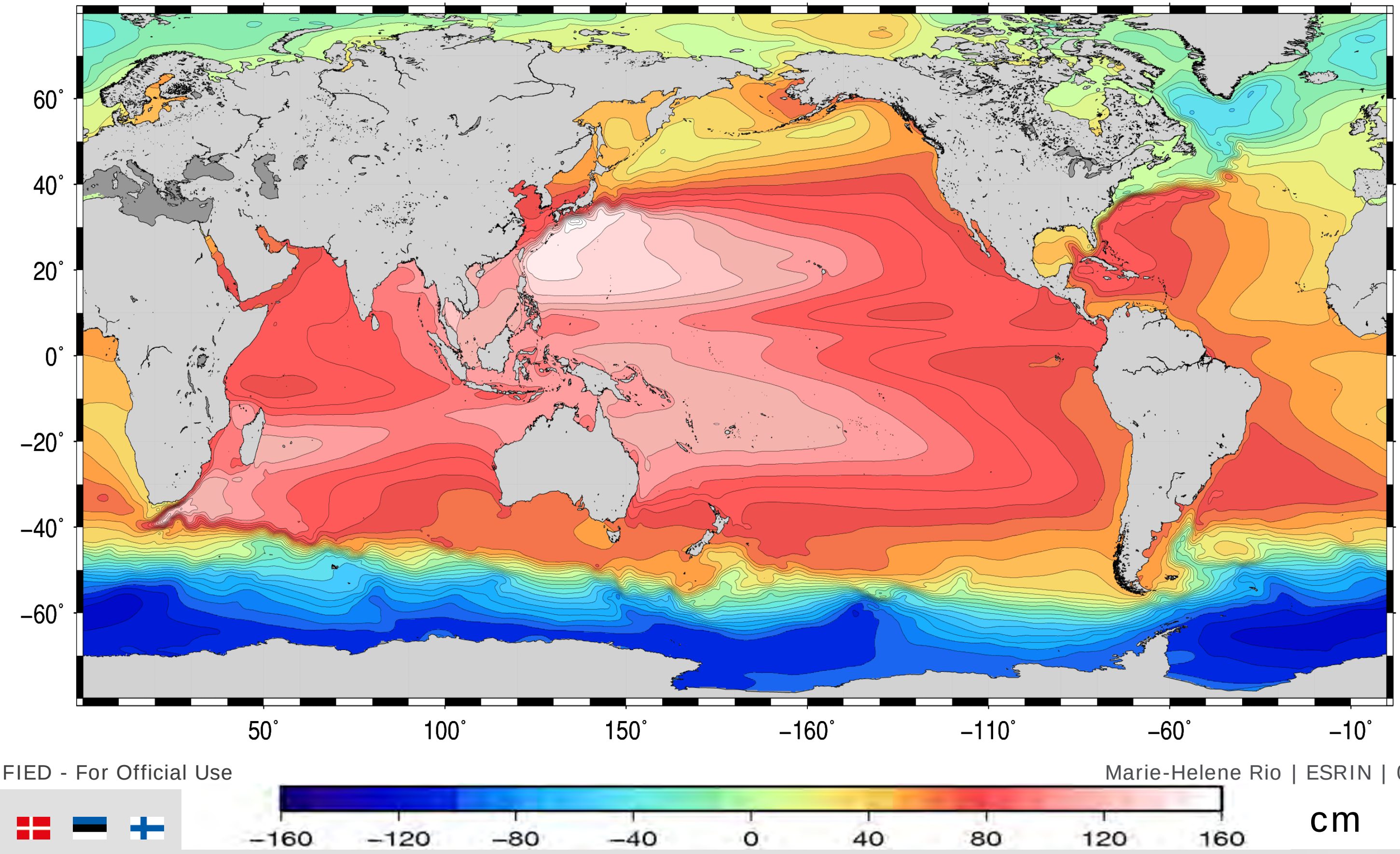
The MDT from altimetry and gravity data

Today: MDT=MSS CNES-CLS – GOCE   RAW DIFFERENCE



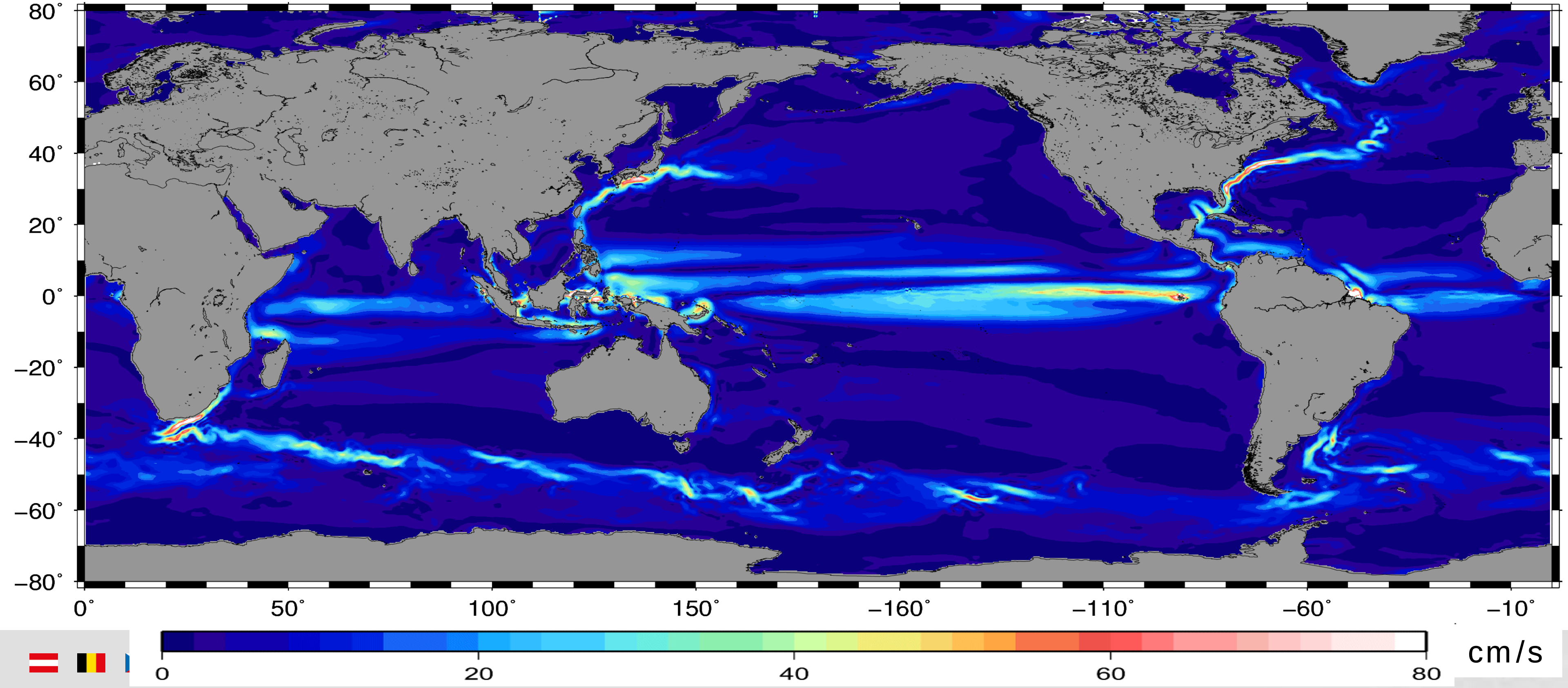
MDT=MSSH-Geoid

MDT=MSS CNES-CLS15 – GOCO05S   Optimally filtered

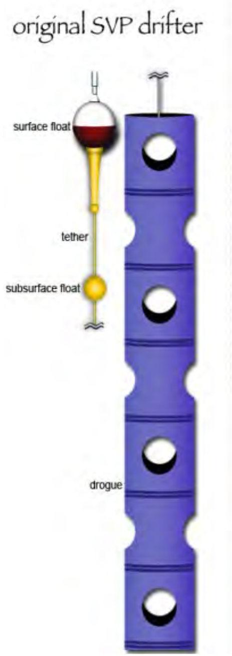


Mean geostrophic currents speed GOCO05S, 2015 : Resolution 100 km

GOCE (Gravity field and steady-state Ocean Circulation Experiment) mission, launched in 2009

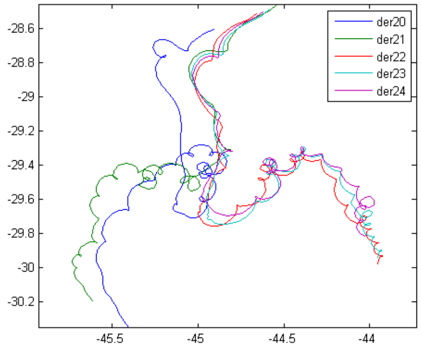


In-Situ measurements: drifting buoys

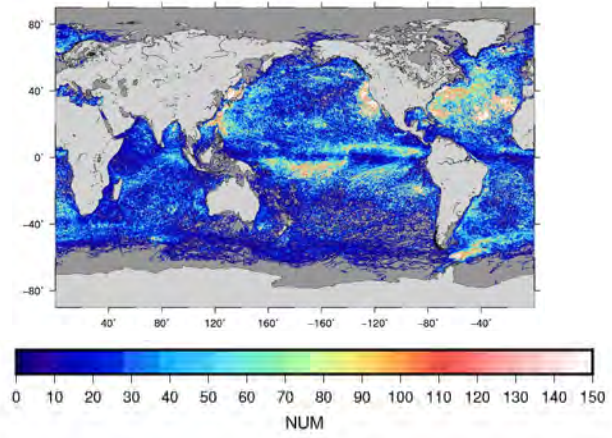


SVP (Surface Velocity Program) type

- Buoy position localized by Argos/Iridium
- Have been designed to minimize the direct wind slippage (less than 0.7 cm/s in 10 m/s winds)
- Holey-Sock drogue centered at 15 m depth -> advected by 15m depth currents
- Drogue loss detection sensor
- After quality control and position processing, regularly sampled velocities are estimated along the buoy trajectory.
- Time sampling: 1 hour, 6 hours
- Life time: ~400 days

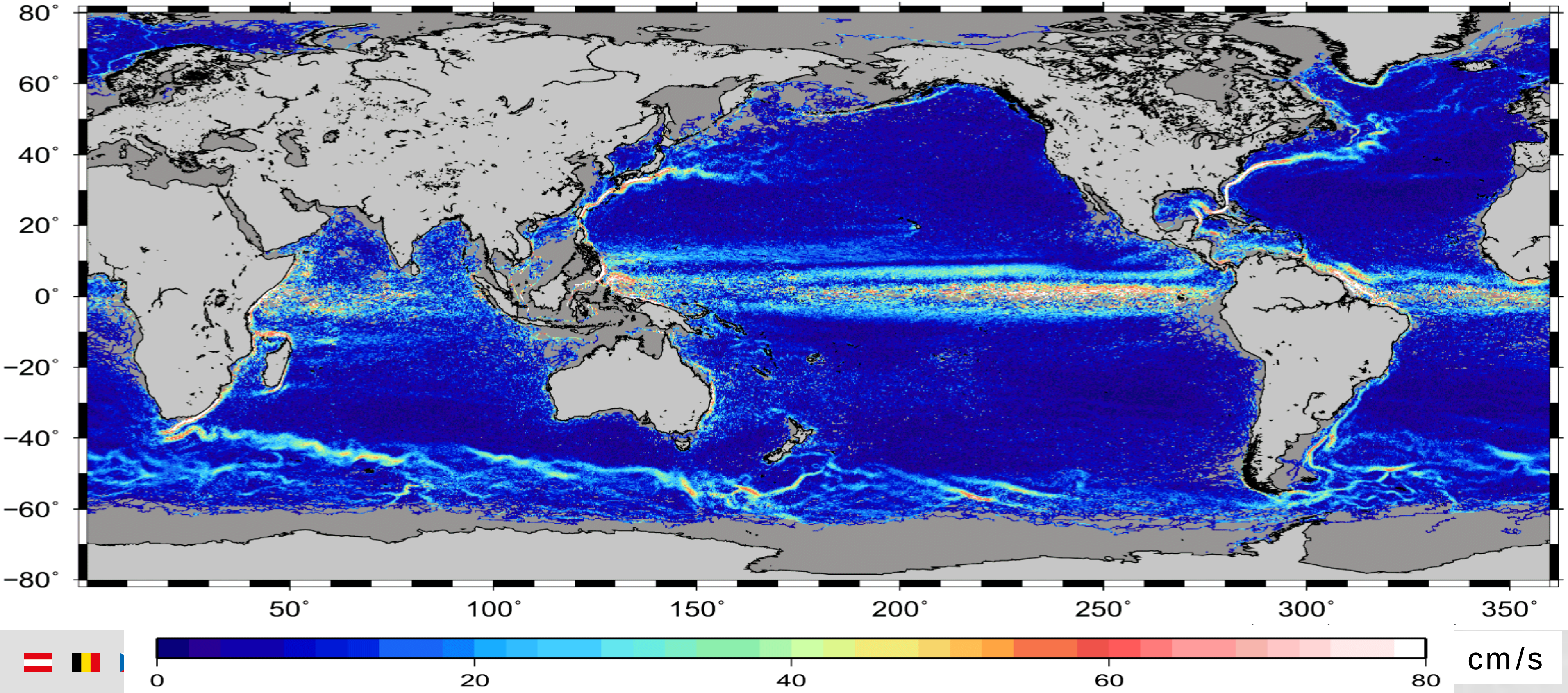


Number of obs (1993-2016)



$$U_{buoy} = U_{geost} + U_{ekman} + U_{tides} + U_{inertial} + U_{stokes} + U_{ageost\_hf}$$

Mean geostrophic currents speed From in-situ measurements



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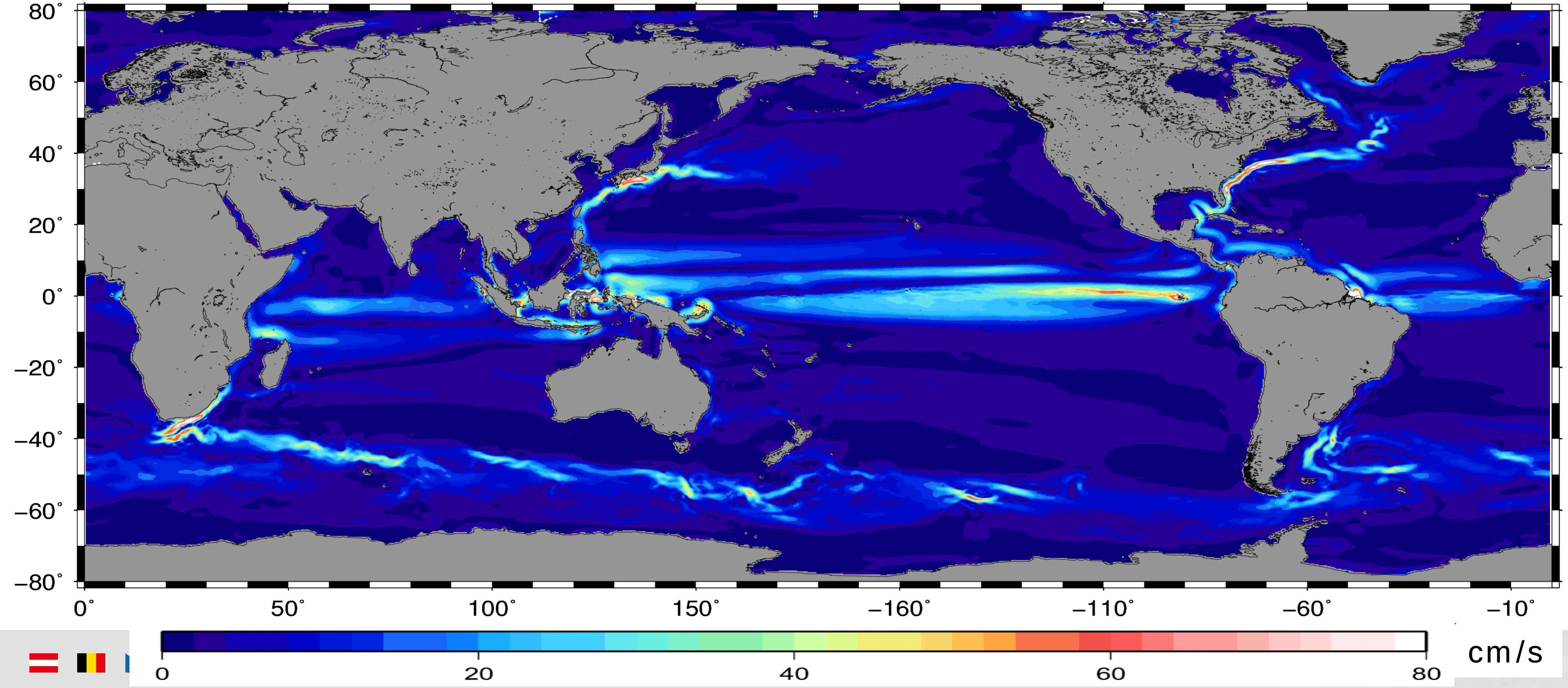
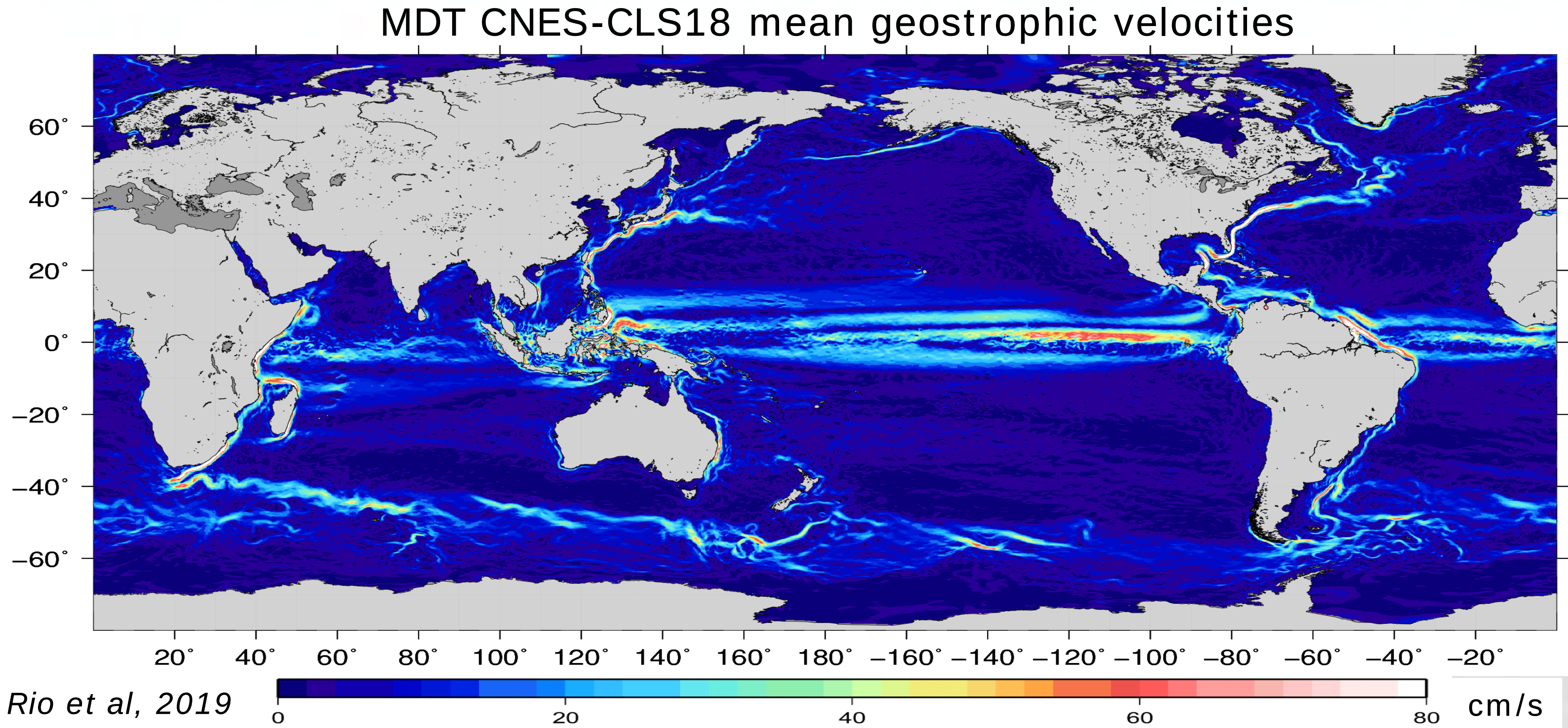
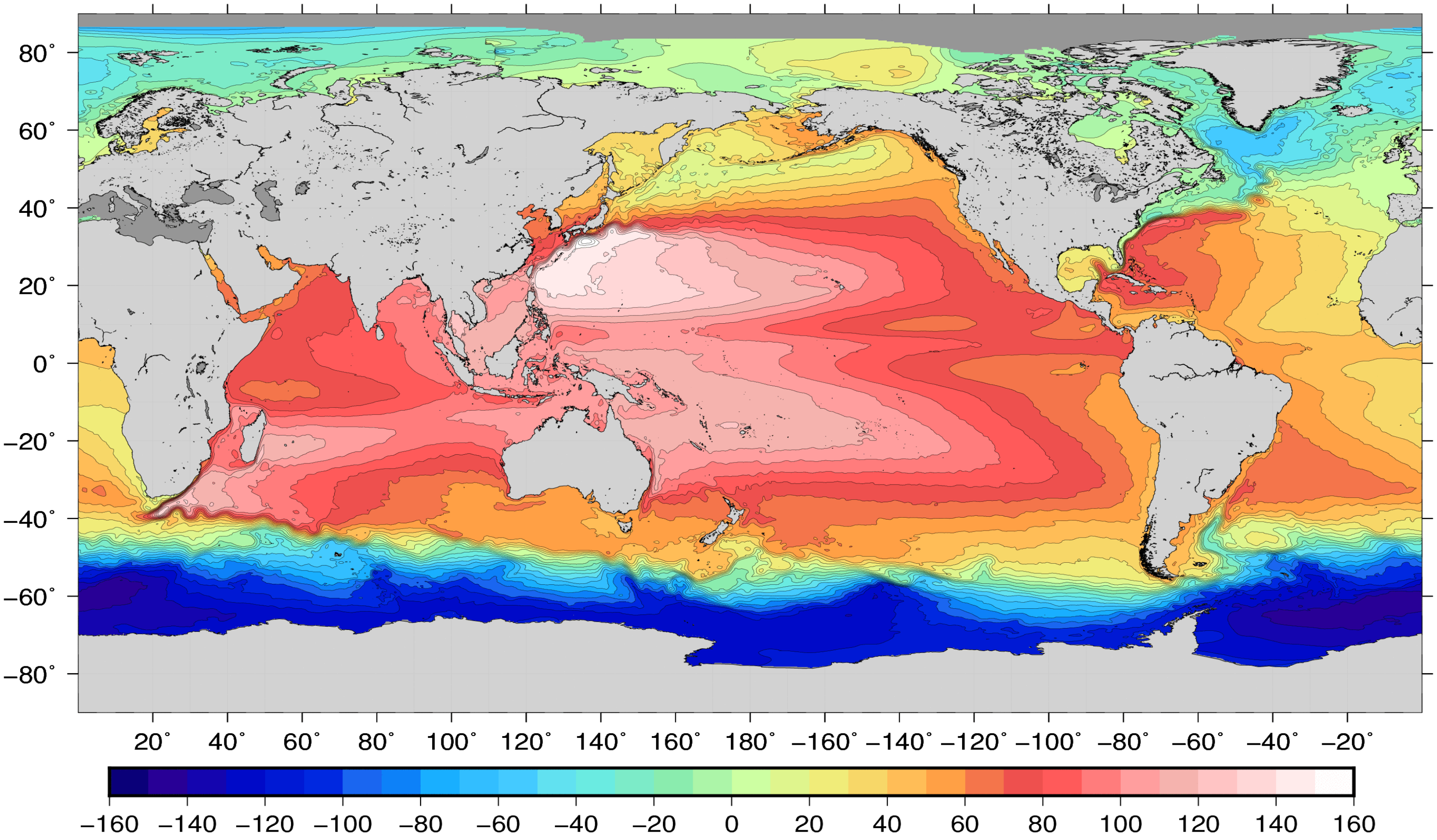


Mean geostrophic currents speed
From Altimetry+GOCE+in-situ measurements

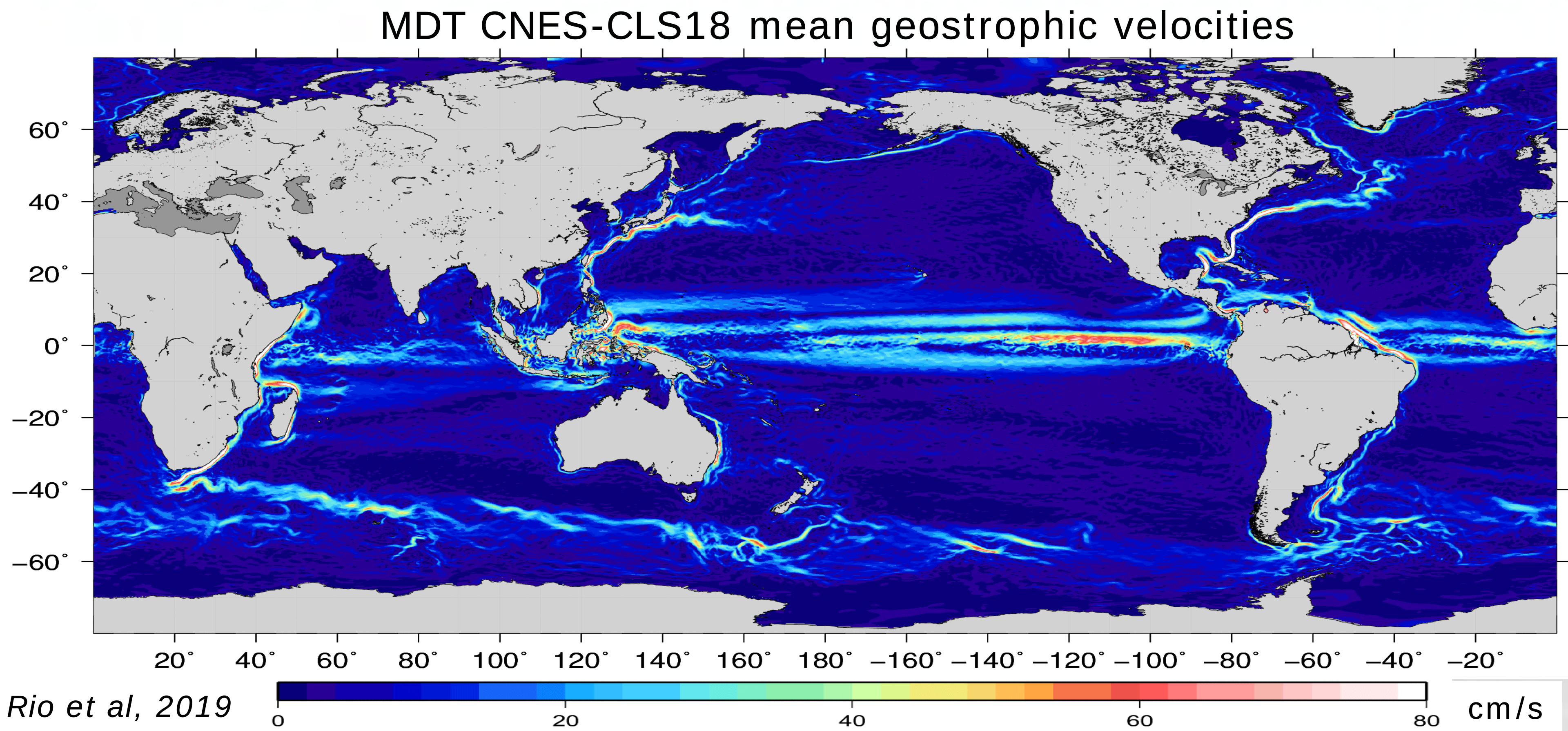


Rio et al, 2019

CNES-CLS18
Mean Dynamic Topography

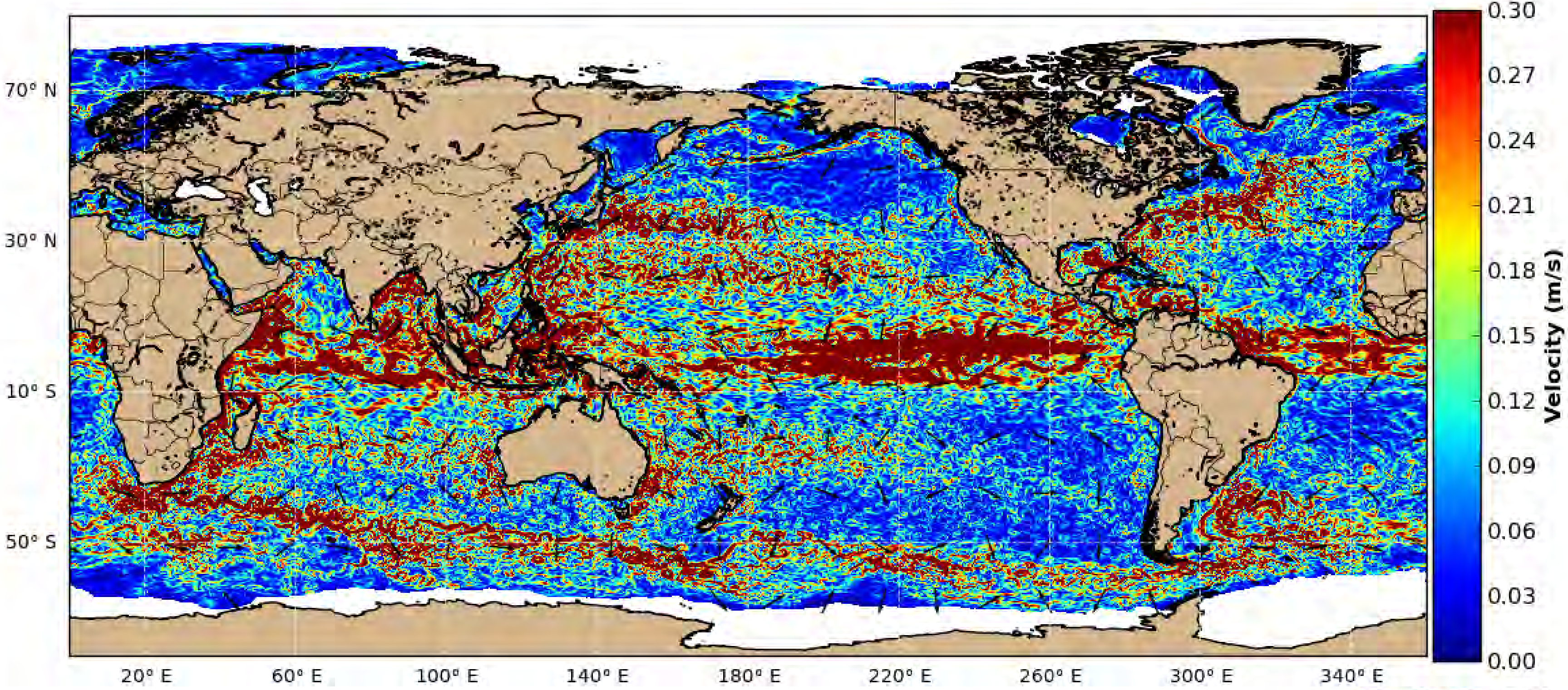


Mean geostrophic currents speed From Altimetry+GOCE+in-situ measurements

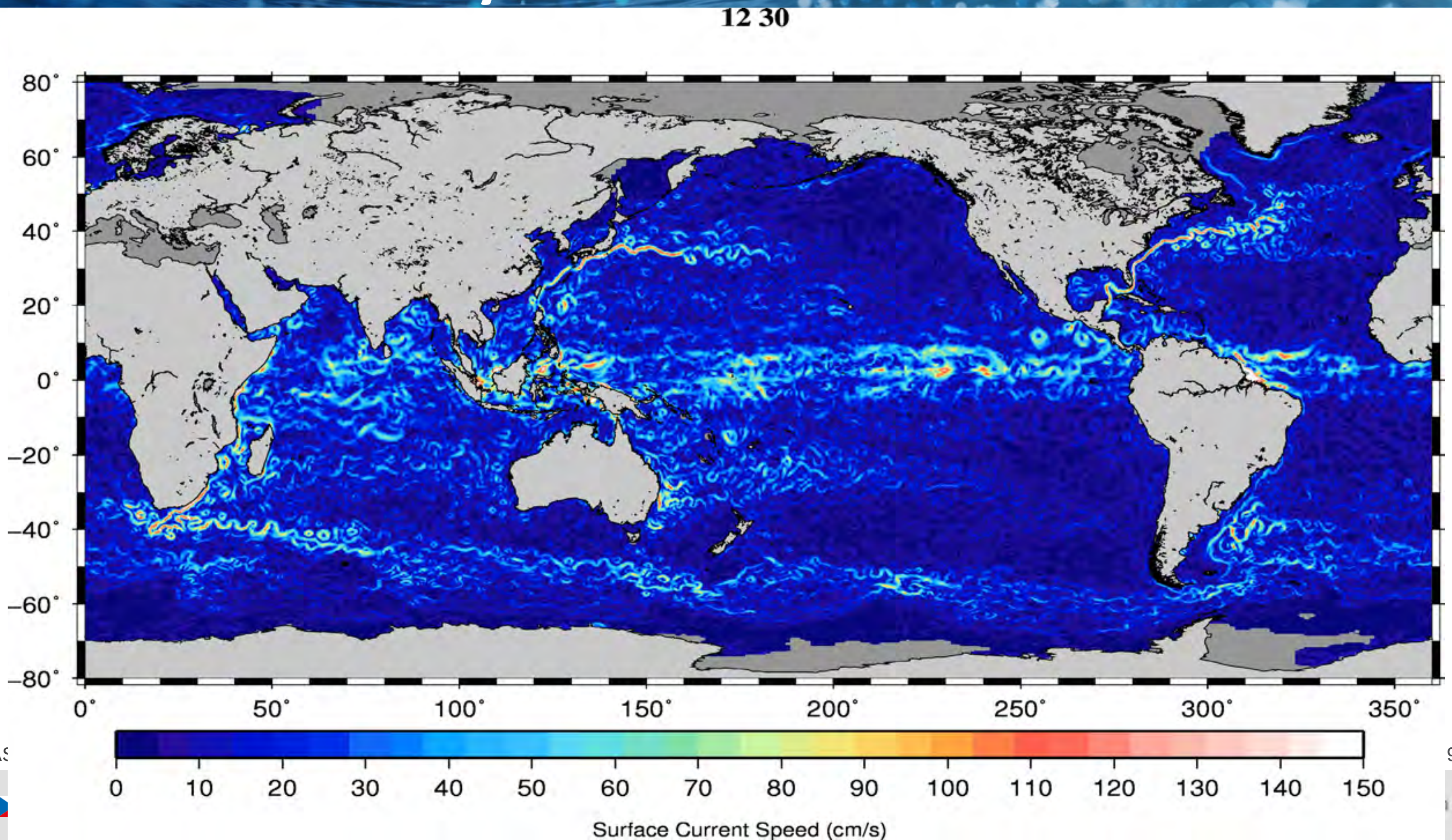


Rio et al, 2019

Geostrophic velocity anomaly speed



24 years of geostrophic currents speed From Altimetry+GOCE+in-situ measurements



Estimating ocean surface currents from space: Altimetry + Gravimetry

From 1993, 25 years of geostrophic currents from gravity+altimetry, (+ near real time and real time products). By far the most exploitable system for ocean currents retrieval.
However:
❖Only the **geostrophic component of the surface current** is obtained

Missing ageostrophic components include

- Ekman currents
- Stokes drift
- Inertial oscillations
- Tidal currents

For a detailed discussion on ocean surface currents components
See Globcurrent Technical Note #1
OTC19 ftp site
from_esa/ForStudents/Day2/Lecture4

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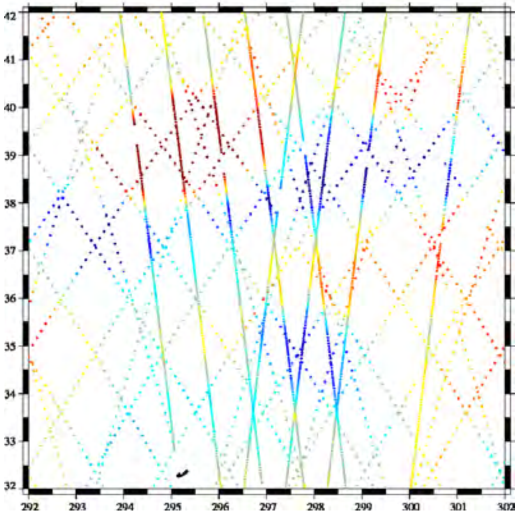
❖The **spatial and temporal** resolution depend on the altimeter constellation

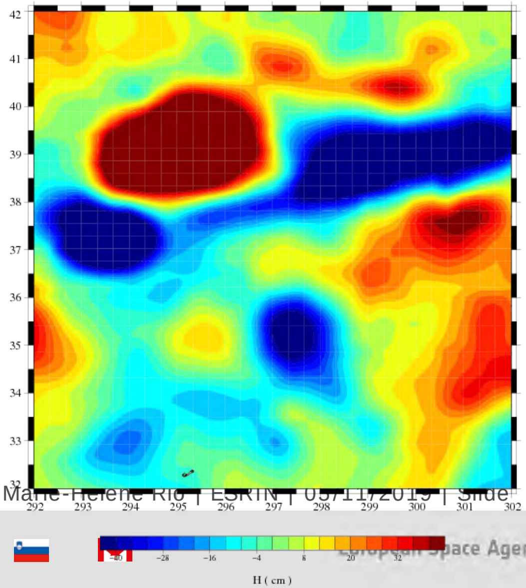
Altimetric anomalies along the tracks from 4 different satellites in the Gulfstream.



Objective Analysis

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

Altimeter anomaly map



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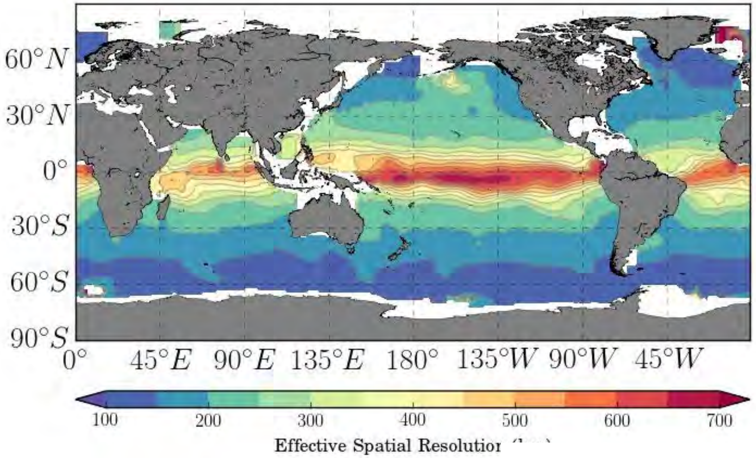
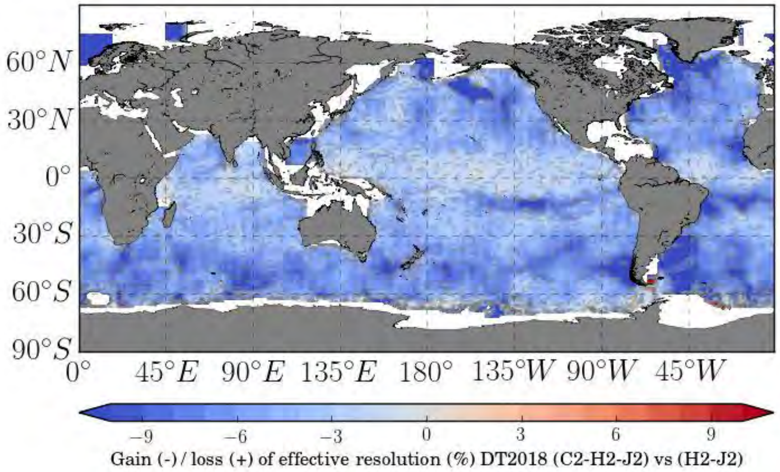
Effective spatial resolution
3 altimeters – monthly time scale



Resolution Gain from 2 satellites
to 3 satellites



Ballarotta et al, 2019

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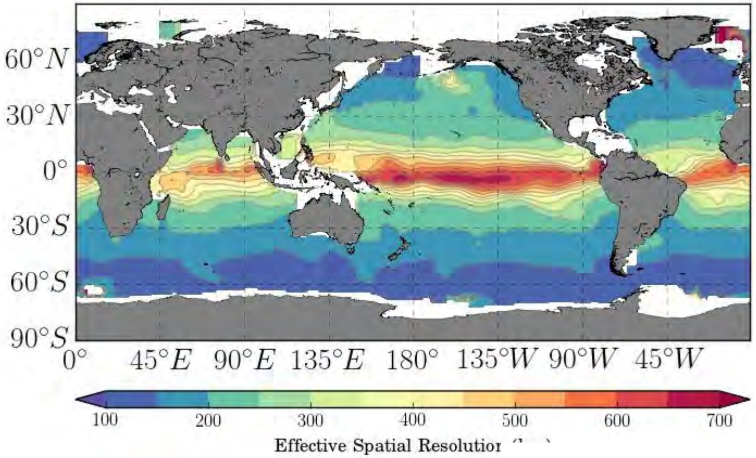
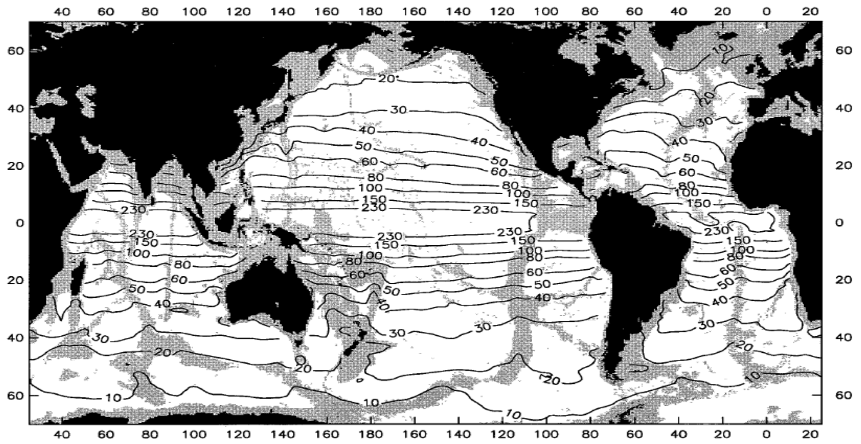


Ballarotta et al, 2019

First Baroclinic Rossby Radius of Deformation:



Chelton et al, 1998

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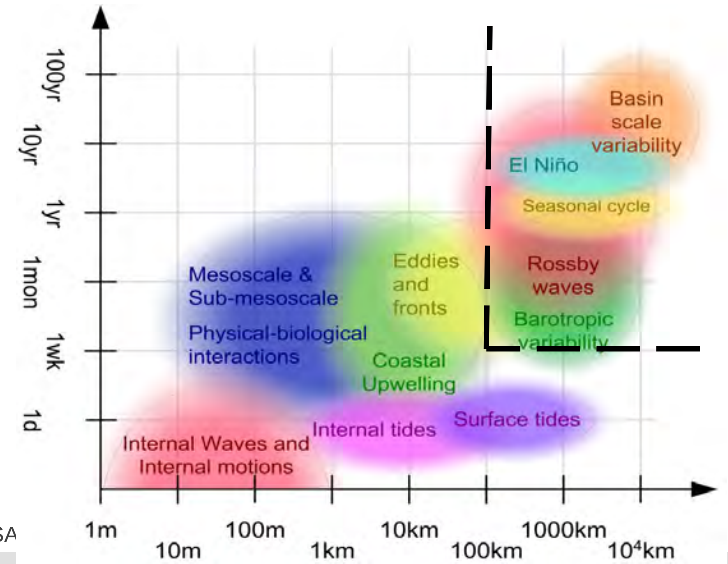

In order to go beyond the altimeter system limitations, **new sensors and new methodologies must be explored**

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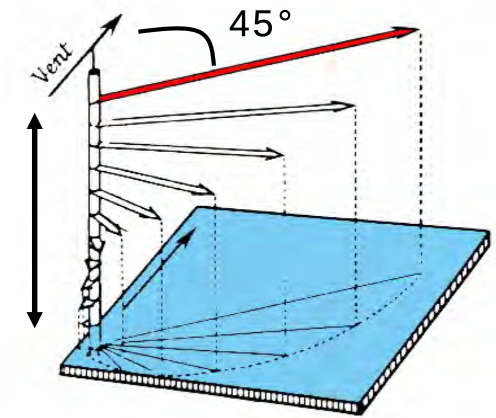
The Ekman currents

Wind-driven Ekman

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

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

β θ



Model

Rio et al, 2003, 2014

$$\vec{u}_e = \beta \vec{\tau} e^{i\theta}$$

τ_e = Effective Wind Stress

D_e = Ekman depth

f = planetary vorticity

w = local vorticity

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

The Ekman currents

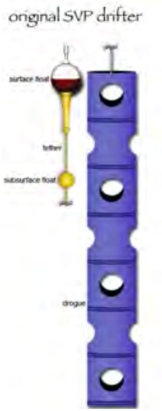
Model

Rio et al, 2003, 2014

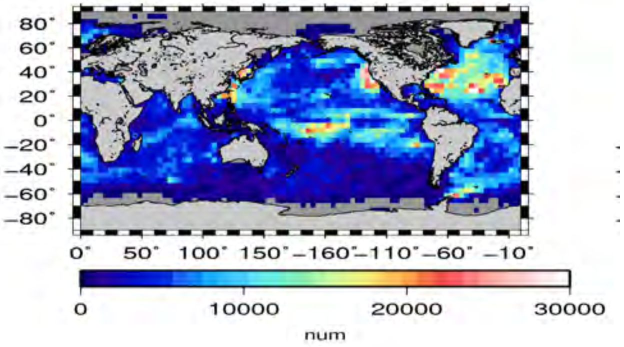
$$\vec{u}_e = \beta \vec{\tau} e^{i\theta}$$

$\vec{u}_{buoy} - \vec{u}_{alti}$

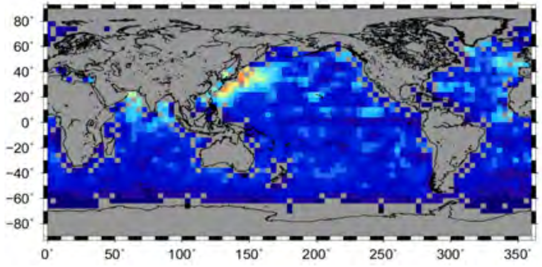
Wind stress (ERA-INTERIM from ECMWF)

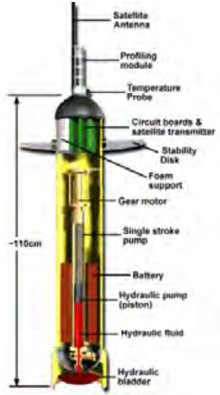


Number of SVP buoy velocities Drogue ATTACHED at 15m depth
Period: 1993-2014



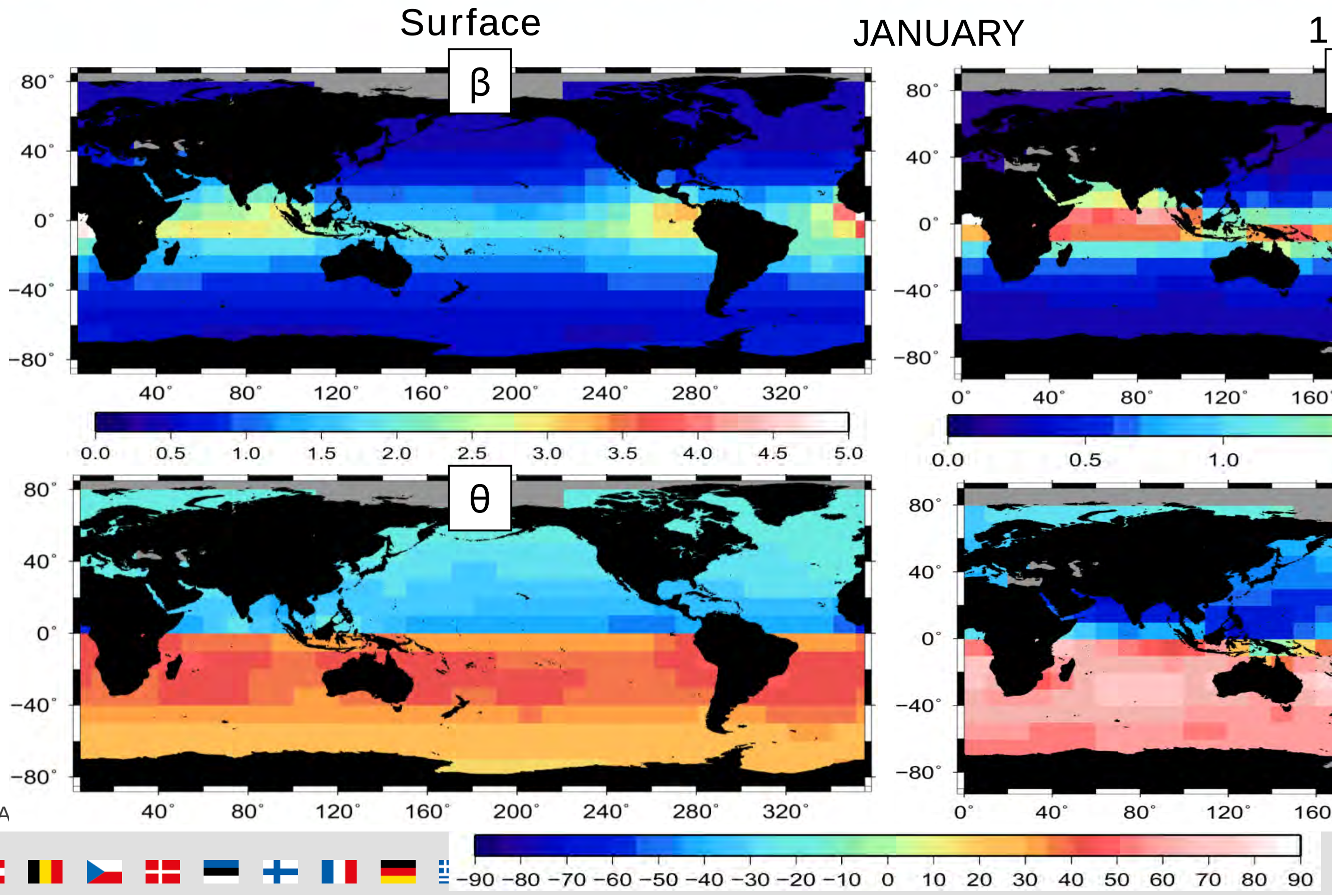
Number of Argo float surface velocities Period: 1997-2014



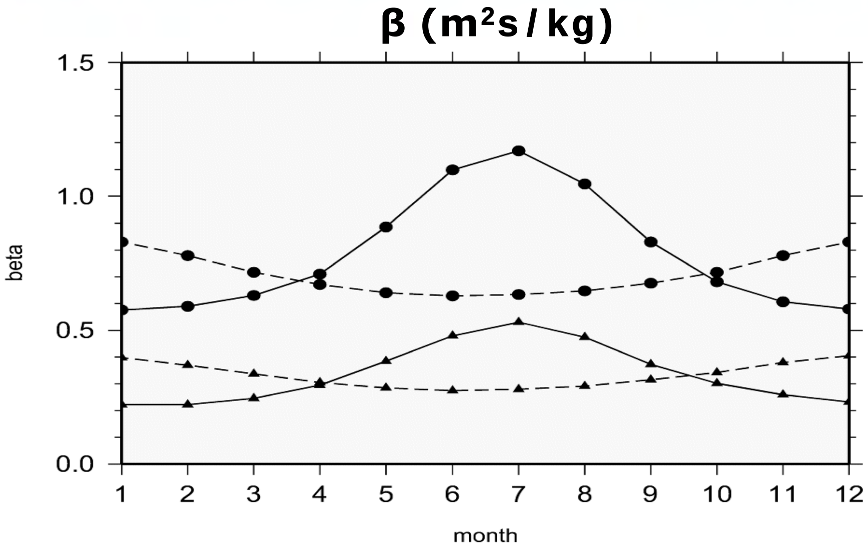


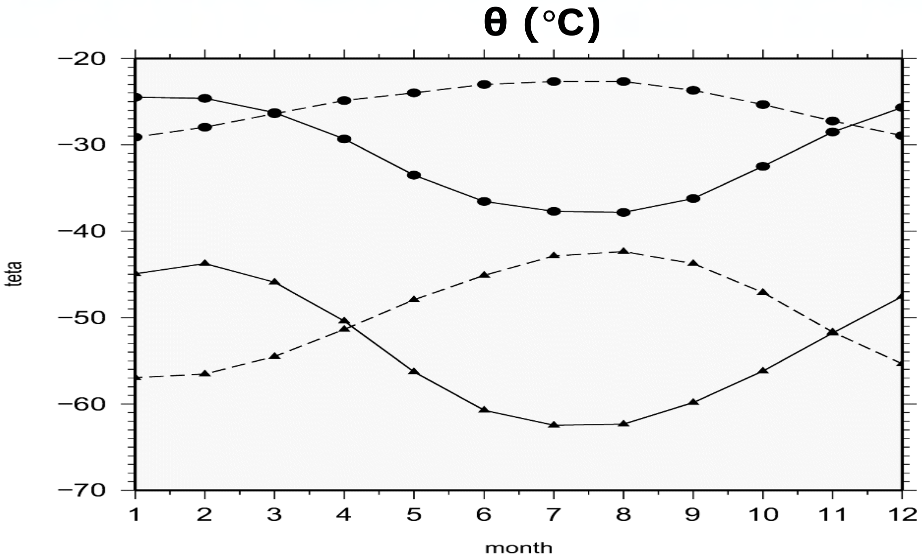
β and θ are estimated through least square fit by month and 4° boxes
At the surface using the Argo float surface velocity dataset **from YoMAHA**
At 15m depth using SVP Drifting buoys **flagged as DROGUED by the SD-DAC**

The Ekman currents



The Ekman currents





Northern Hemisphere: solid line
Southern Hemisphere: dashed line
Surface: circles
15m depth: triangles

In Summer stratification increases => De decreases

$$\beta = \frac{\pi\sqrt{2}}{\rho f D_E} e^{\frac{\pi}{D_E} z}$$ increases

$$|\theta| = \left(\frac{\pi}{4} + \frac{15}{D_e}\right)$$ increases

The Ekman currents

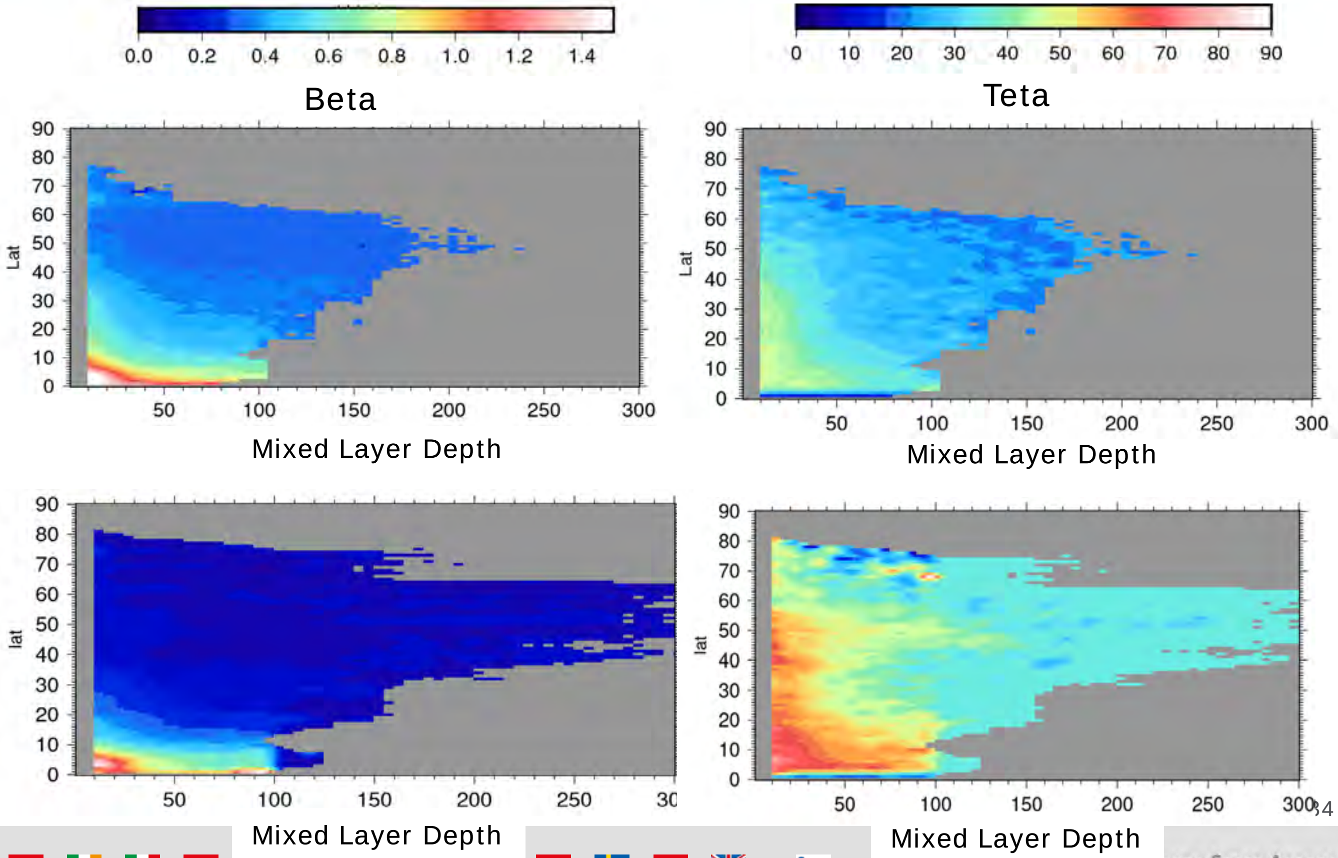
SURFACE

$$\vec{u}_w(z=0) = \beta_0 e^{i\theta_0} \vec{\tau}^{0.6}$$

15m depth

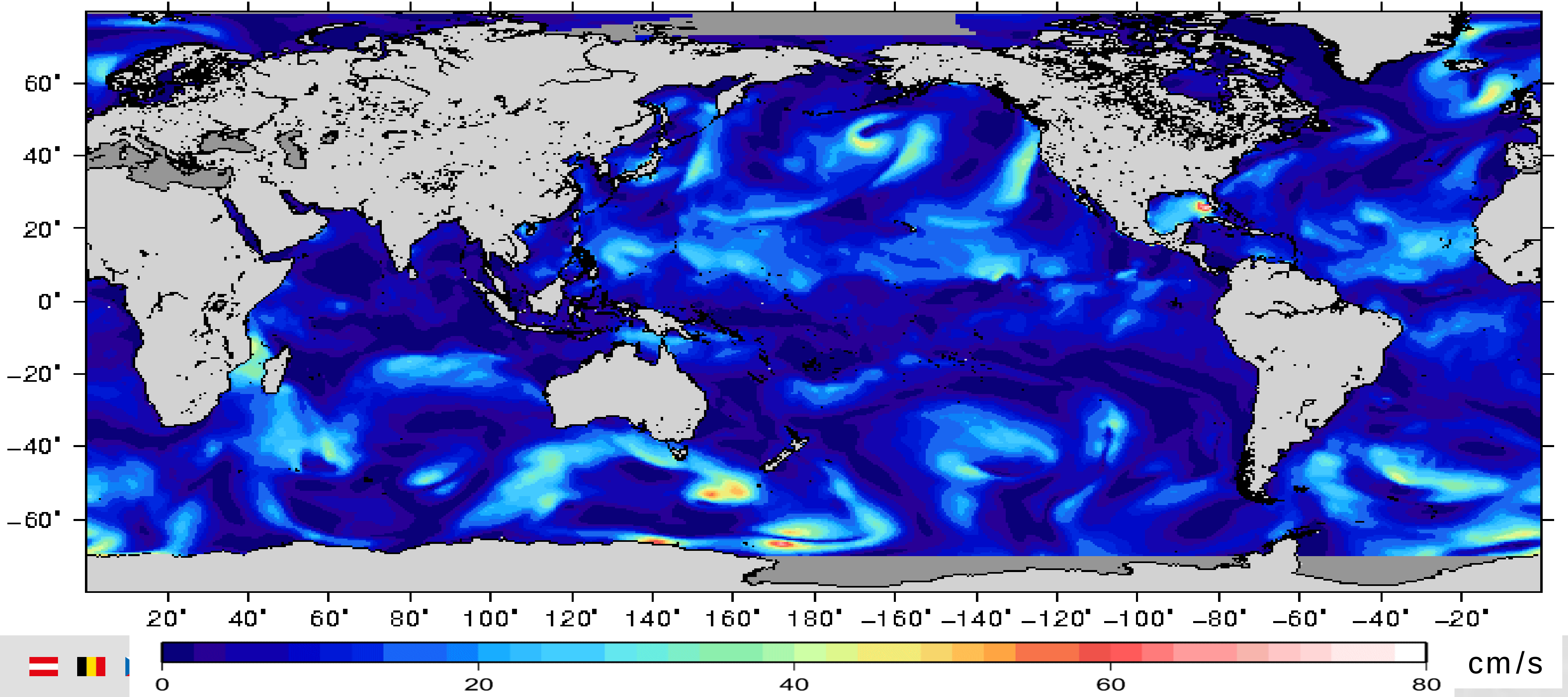
$$\vec{u}_w(z=15) = \beta_{15} e^{i\theta_{15}} \vec{\tau}^{0.7}$$

MLD from the weekly ARMOR3D T/S fields (Guinehut et al, 2012)

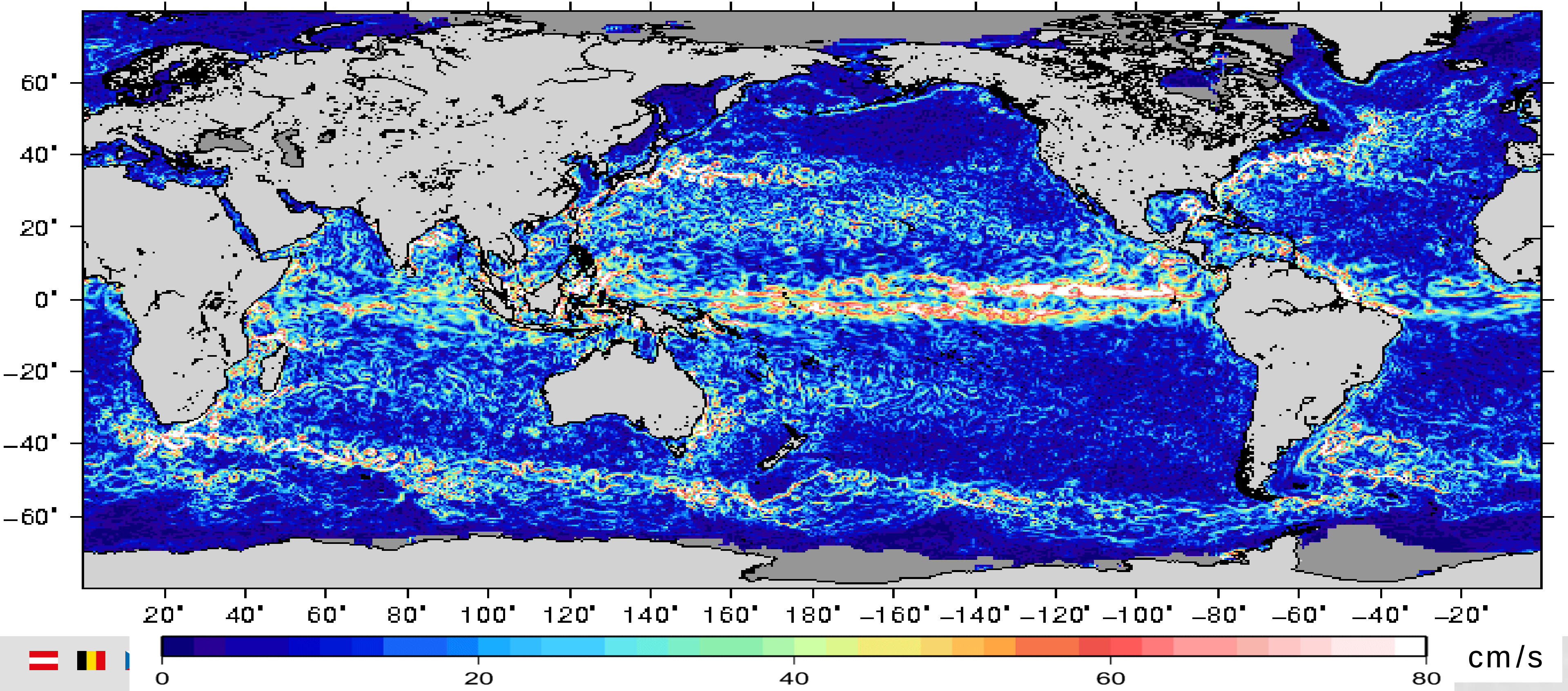


Rio et al, 2019

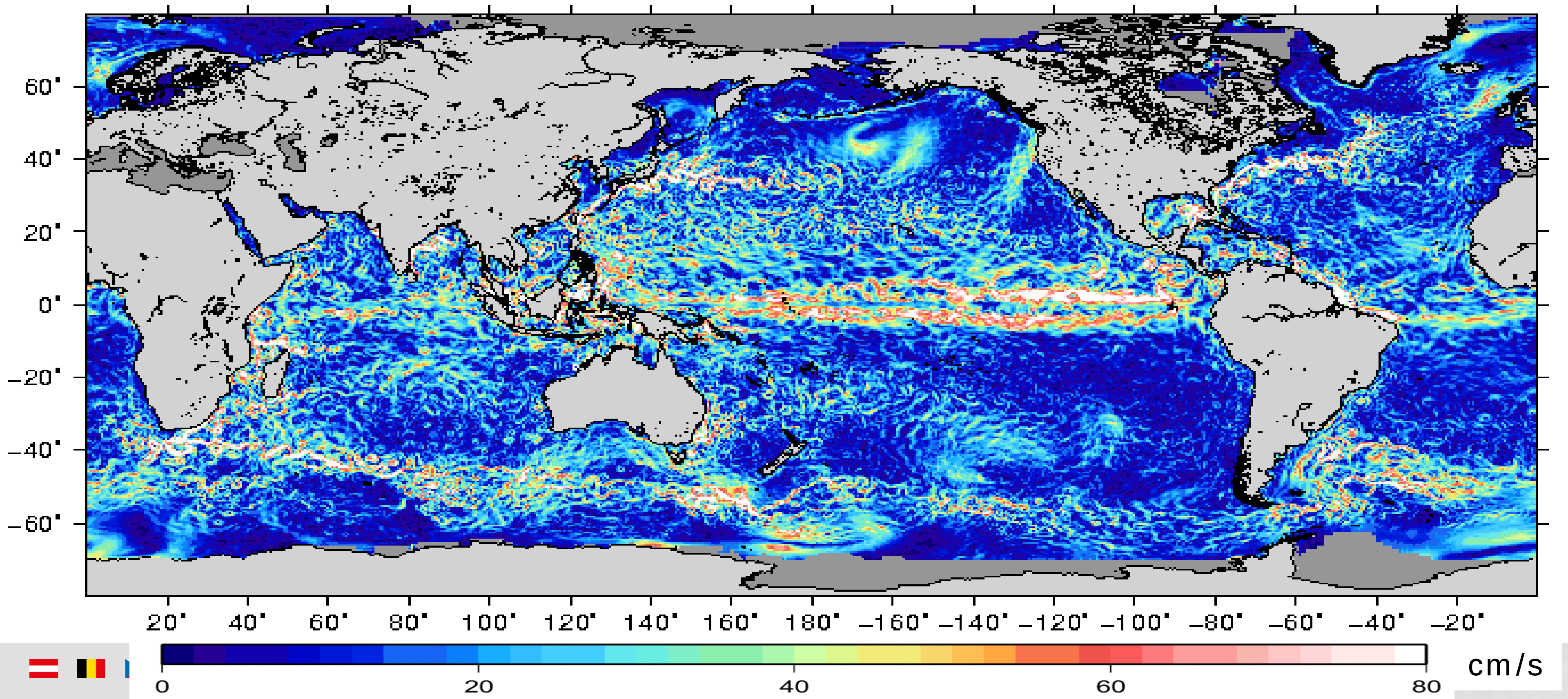
The Ekman current May, 5th 2016



The Geostrophic current May, 5th 2016



Ocean Surface Current from Gravity+Altimetry+Wind



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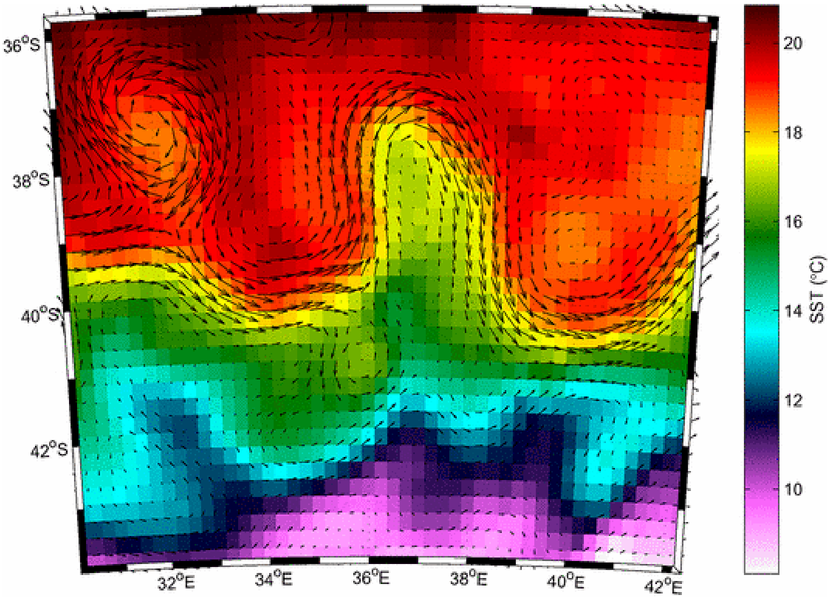
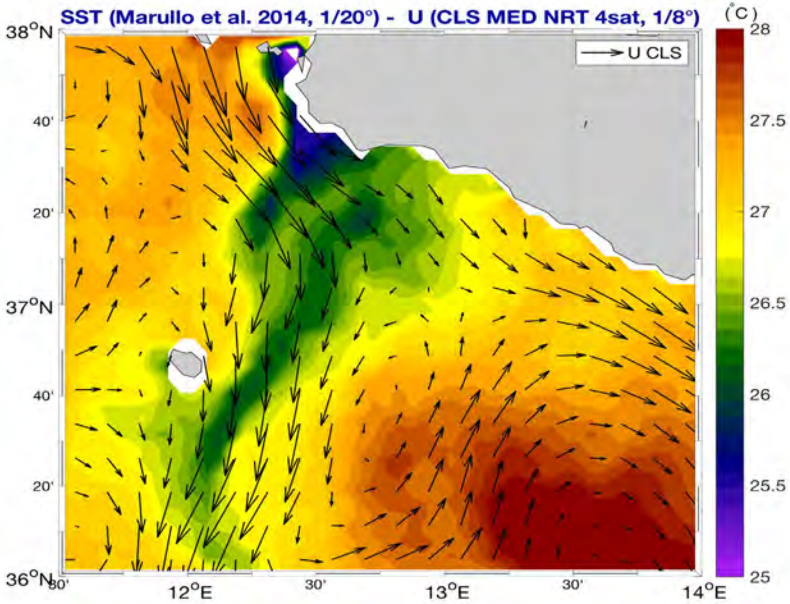
Scatterometer	Wind	Ekman model	Ekman current	25 km 12 hours
Microwave Radiometer	SST	Optical flow, MCC	Total surface currents	25 km 1 day
		E-SQG	Geostrophic surface currents	
Infrared Radiometer	SST	Optical flow, MCC	Total surface currents	Polar orbiting 10 km 1 day Geostationnary 10 km hourly
		E-SQG	Geostrophic surface currents	
L-Band radiometer	SSS	Optical flow, MCC	Total surface currents	100km 3-10 days
		E-SQG	Geostrophic surface currents	
Spectrometer	Ocean color	Optical flow, MCC	Total surface currents	Polar orbiting 10 km 1 day Geostationnary 10 km hourly
SAR	Range Doppler Anomaly Shift	CDOP sea state component of Doppler shift	Radial component of total current minus wind drift (included in CMOD)	Snashots 10 km 3 days

Deriving surface currents from tracer information

January, 1st 2004
Microwave SST product
Altimeter geostrophic velocities



August, 6th 2013
Microwave + Infrared (SEVIRI) 1/20° SST product
Altimeter geostrophic velocities



Effective Surface Quasi Geostrophy (E-SQG) Method

Under favourable environmental conditions, the streamfunction ψ 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)

$$\frac{\partial}{\partial y}\psi_{sqg} = -u$$
$$\frac{\partial}{\partial x}\psi_{sqg} = v$$

Currents

$$\widehat{\psi_{sqg}}(\vec{k}, z) = \frac{f}{N_{eff}\rho_0 k}\widehat{\rho'_s}(\vec{k}).e^{\left(\frac{N_{eff}kz}{f_0}\right)} \Rightarrow \rho'_S = -\alpha \cdot T_S' - \beta S_S' = -\alpha' T_S'$$

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.

$$\hat{\psi}(\vec{k}) = F_T(k)\hat{T}_s(\vec{k}),$$

$F_T(k)$ is the transfer fonction is to be determined using independent observations.

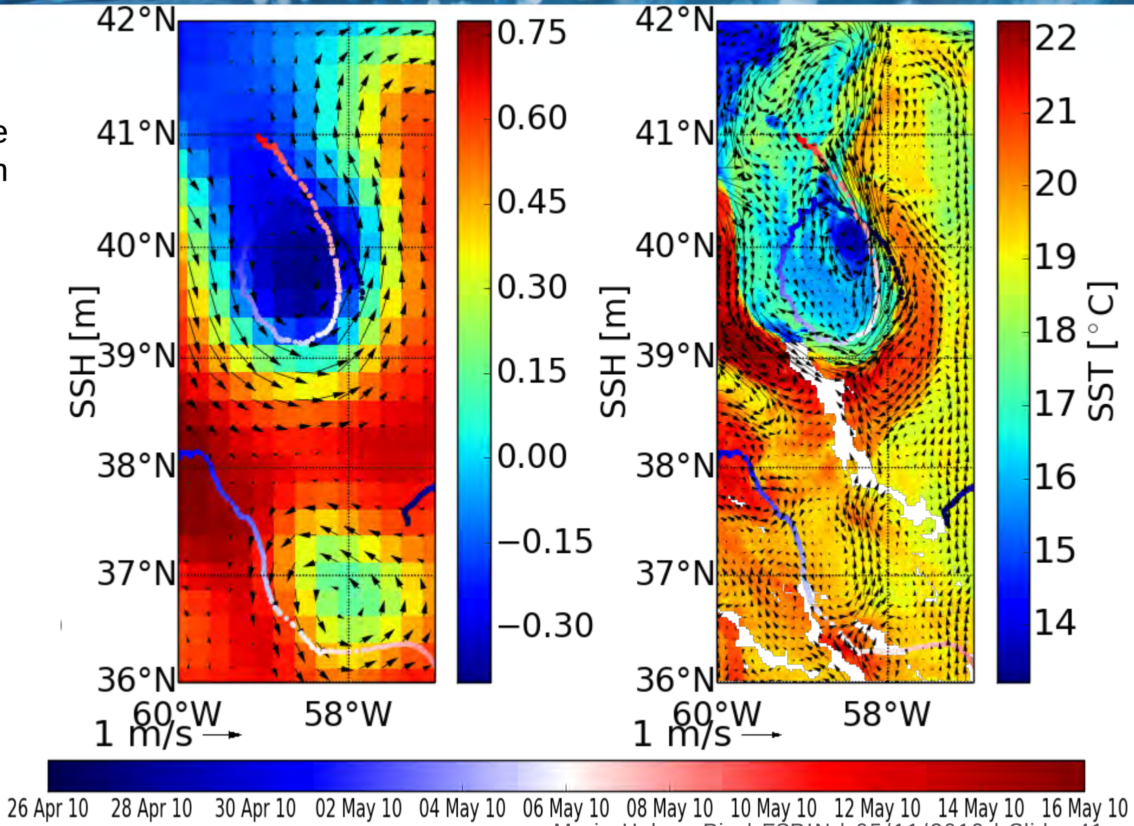
Isern-Fontanet et al. [2014]: the geostrophic streamfunction at the ocean surface is proportional to the SSH (η).

$$\psi(\mathbf{x}) = \frac{g}{f_0}\eta(\mathbf{x}).$$

So that the transfer fonction can write:

$$F_T(k) = \frac{g}{f_0}\frac{\langle|\hat{\eta}|\rangle_k}{\langle|\hat{T}_s|\rangle_k},$$

Combination of the phase of SST measurements and the amplitude of SSH measurements.



Correlation between the reconstructed stream-function and altimeter streamfunction:

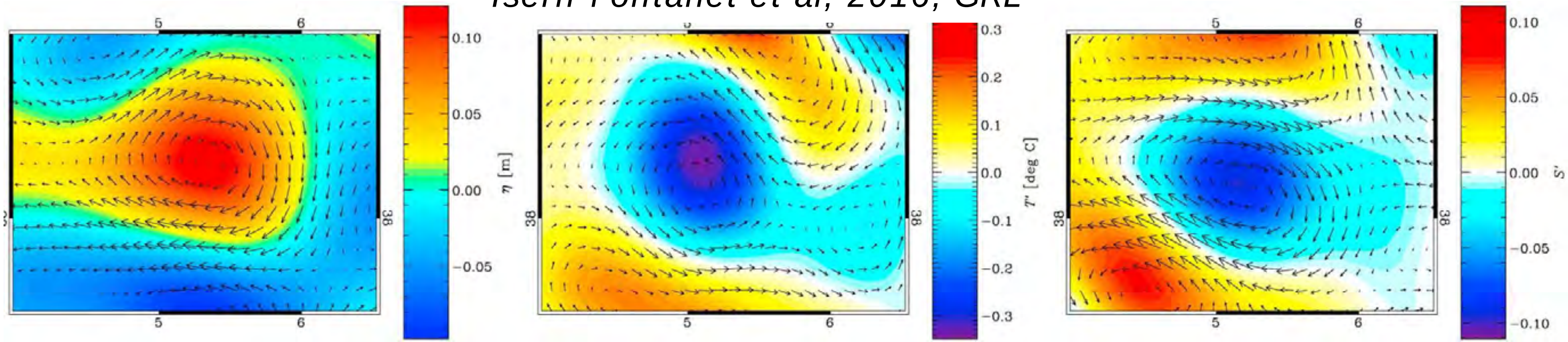
Limitations

▪Limited to **the retrieval of mesoscale (30-300km),** not the large scale currents

▪The SQG method is valid in baroclinic instabilities areas, and **strong gradients areas** (ACC, Gulfstream, Kuroshio…)

▪In addition, the validity of the SQG approximation is limited to cases when the SST is a good proxy of the density anomaly at the base of the mixed layer.

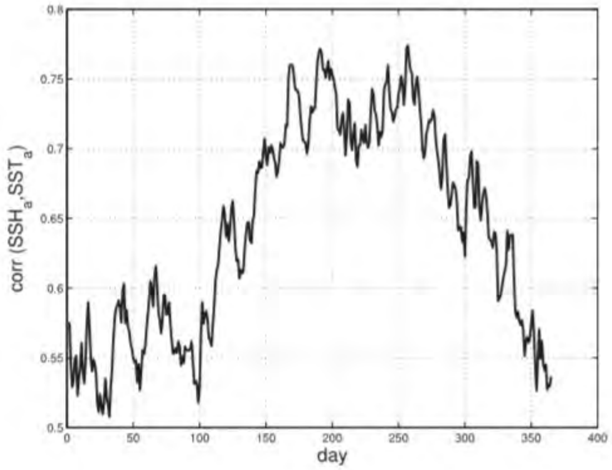
Isern-Fontanet et al, 2016, GRL



(left) SSH anomaly with the geostrophic velocities overplotted, (middle) SST anomaly with the velocities derived from SST, and (right) SSS anomaly with the velocities derived from SSS corresponding to 12 May 2011.

Limitations

- The coldest SST anomalies are reported to efficiently 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



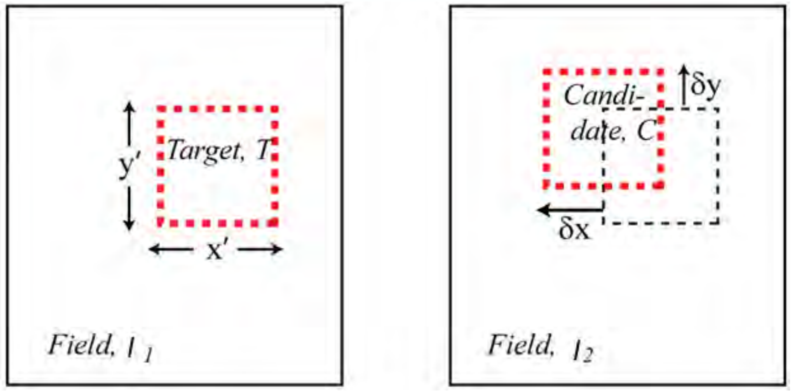
Time series of the global correlation between SSH and SST anomaly fields in 2004 in the Agulhas return current

Legoff et al, 2016

The Maximum Cross Correlation (MCC) method

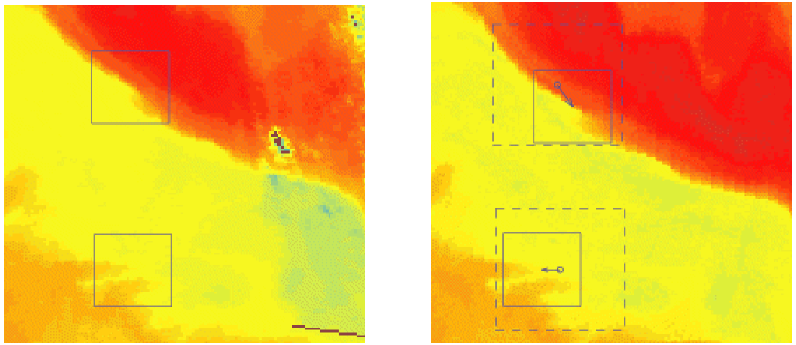
The method consists in calculating the displacement of small regions of patterns from one image to another.
We look for the pair δx_{max} δy_{max} that achieves the maximum cross-correlation between a target region in the first image and a candidate region in the second image

$$[u\ v] = [\delta x_{max}\ \delta y_{max}\ ] / \Delta t$$



First Image

Second Image



This can be applied on any image: a brightness temperature, a water-leaving radiance or a derived product such as sea surface temperature (SST) or chlorophyll concentration

Emery et al, 1986

The Maximum Cross Correlation (MCC) method

Application on GOCI (Geostationnary Ocean Color Imager) Ocean Color images in the Tsushima Strait

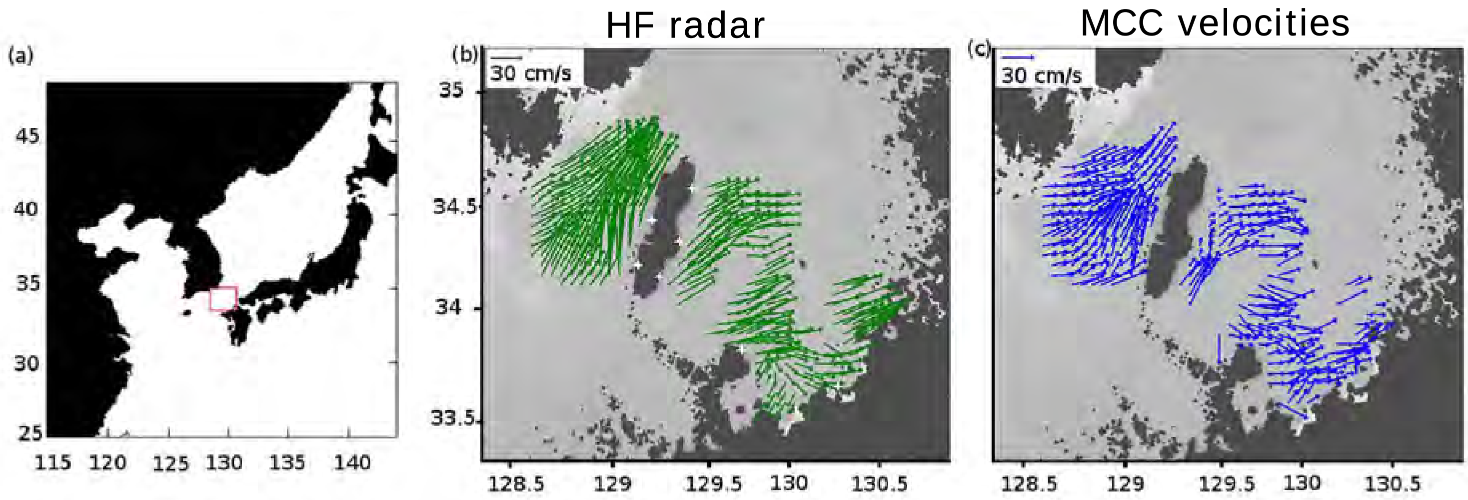


Figure 2. (a) Total region covered by the GOCI sensor with the red box indicating the Tsushima Strait area used in this study; (b) arrows indicate mean velocities from HF radar and locations of the 7 HF radar stations are shown as white dots; c) arrows indicate mean velocities from MCC methodology. Data are from 26 March 2012 and all MCC image pairs have been used. Velocities are the mean over corresponding time periods and locations. Radar-derived velocities are only shown where there are MCC velocities.

Warren et al, 2016

Limitations

- **Cloud-cover and isothermal/isochromatic ocean surface conditions drastically limit the spatial and temporal velocity coverage provided by the MCC method**. Clouds block the ocean surface in both thermal and ocean color imagery, and there are no features for the MCC method to track in isothermal/isochromatic regions.
- Accurate spatial alignment and coregistration of the imagery used in feature tracking is required. Consequently, the technique has been more **often used in coastal regions**, where landmarks are available to renavigate the satellite data.
- MCC techniques work well for intervals between images of 6-24 hrs, but are not so reliable for longer gaps due to evolution of the features, including rotation and shear.

« Optical flow » methods: inversion of a tracer conservation equation

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

Challenge: only **along-gradient velocity** information can be retrieved from the tracer distribution at subsequent times in **strong gradients areas**.

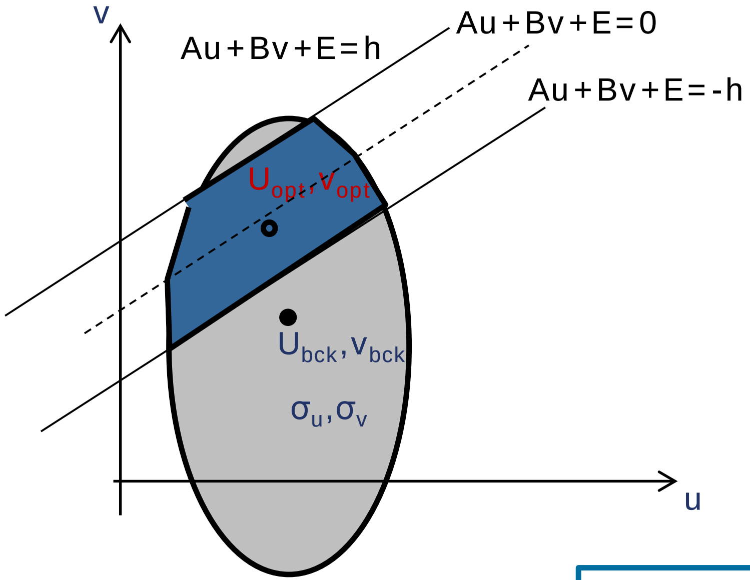
Synergy : The method is used on successive SST images using the altimeter geostrophic velocities as background so as to obtain an optimized 'blended' velocity (u_{opt}, v_{opt}). *Rio et al, 2016; Rio and Santoleri, 2018; Ciani et al, 2019*

Optimal SSH / SST combination

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

$$A = \frac{\partial SST}{\partial x} \qquad B = \frac{\partial SST}{\partial y} \qquad E = \frac{\partial SST}{\partial t} - F_{bck}$$

$|F-F_{bck}| < h$



Spatial SST variations

$$q = \sqrt{\left(\sigma_u^2 \sin^2\varphi + \sigma_v^2 \cos^2\varphi\right)} \qquad \varphi = \operatorname{Arc\,tan}\left(-\frac{A}{B}\right)$$

$$\alpha = \frac{Au_{bck} + Bv_{bck} + E - h}{\sqrt{A^2+B^2}} \qquad \beta = \frac{Au_{bck} + Bv_{bck} + E + h}{\sqrt{A^2+B^2}}$$

$$d = \frac{\left|Au_{bck} + Bv_{bck} + E\right|}{\sqrt{A^2+B^2}} \qquad p = \frac{\sin\varphi\cos\varphi\left(\sigma_v^2 - \sigma_u^2\right)}{q^2}$$

$$u_0 = \frac{F(\min(\beta,q)) - F(\max(\alpha,-q))}{G(\min(\beta,q)) - G(\max(\alpha,-q))} \qquad v_0 = pu_0$$

$$F(x) = -\frac{2\left(q^2-x^2\right)^{3/2}}{3}$$

$$G(x) = x\left(q^2-x^2\right)^{1/2} + q^2\sin^{-1}(x/q)$$

Temporal SST variations

Forcing term estimates

Forcing term error

Background velocity error

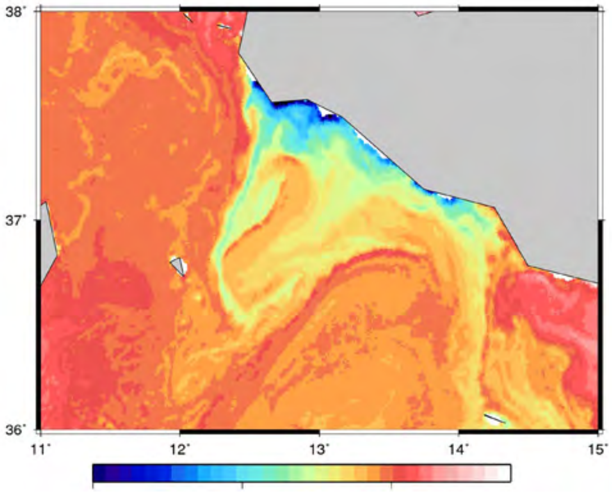
$$u_{opt} = u_{bck} + u_0\sin\varphi + v_0\cos\varphi \qquad v_{opt} = v_{bck} - u_0\cos\varphi + v_0\sin\varphi$$

Piterbarg et al, 2009
Rio et al, 2016;
Rio and Santoleri, 2018

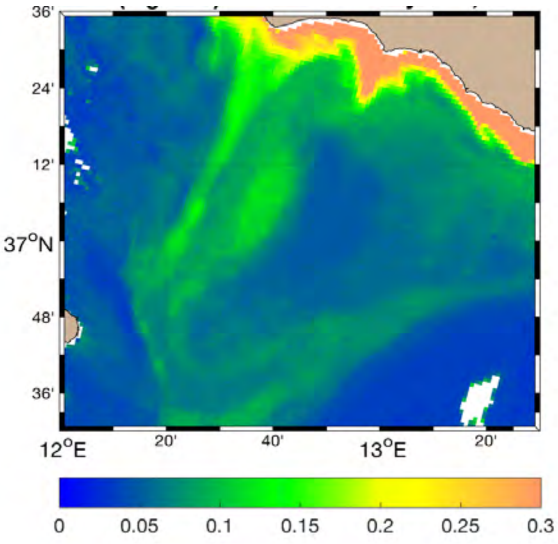
Deriving surface currents from tracer information

Sentinel-3 data on July, 28th 2016

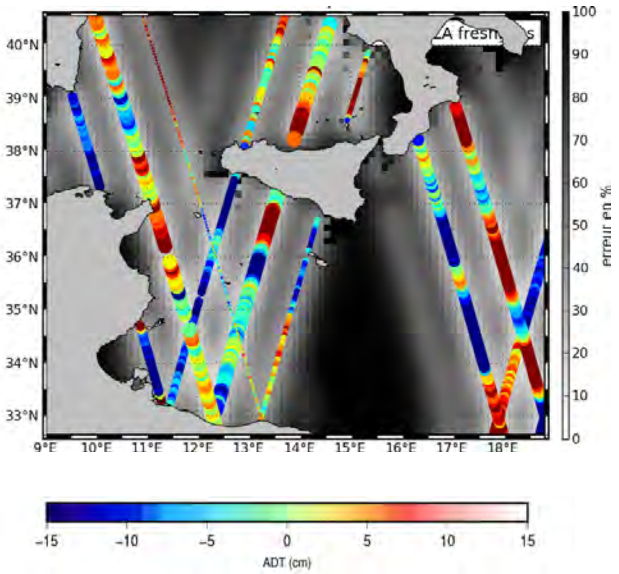
S-3 Sea Surface Temperature



S-3 Chl-a concentration

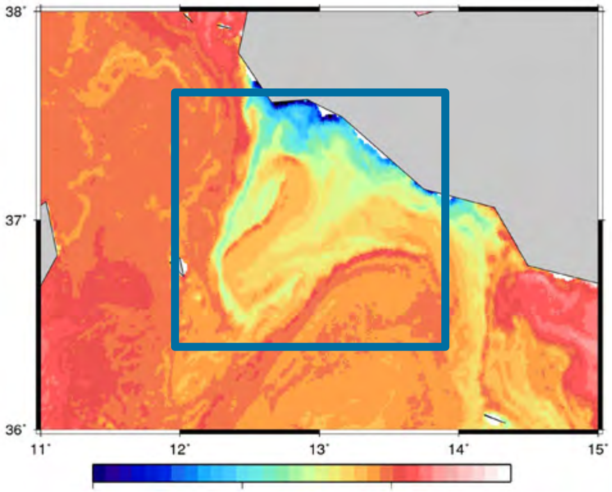


S-3 along-track Sea Level Anomalies (+- 5 days)

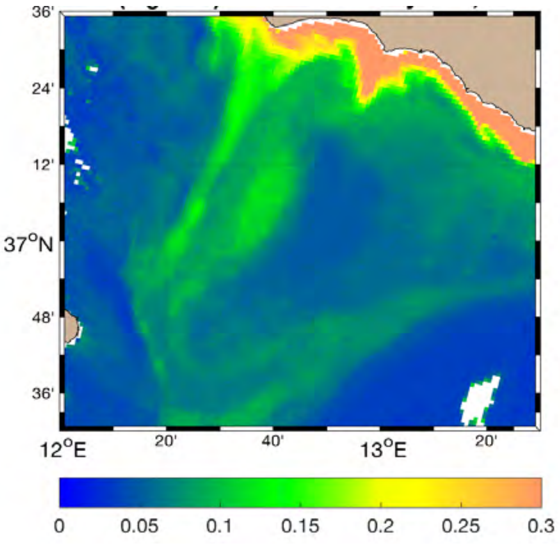


Sentinel-3 data on July, 28th 2016

S-3 Sea Surface Temperature

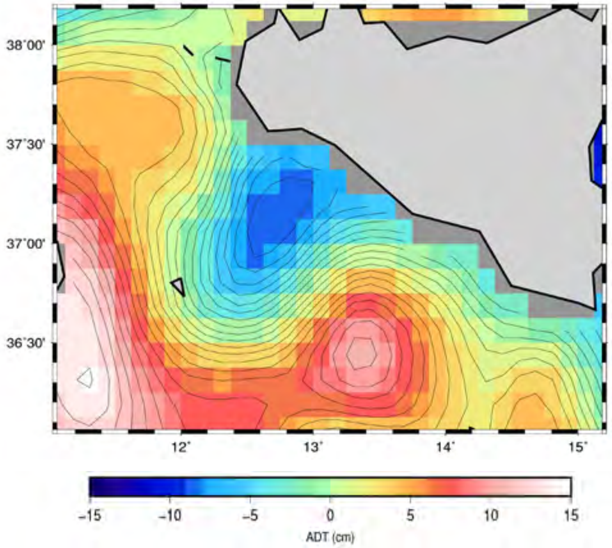


S-3 Chl-a concentration

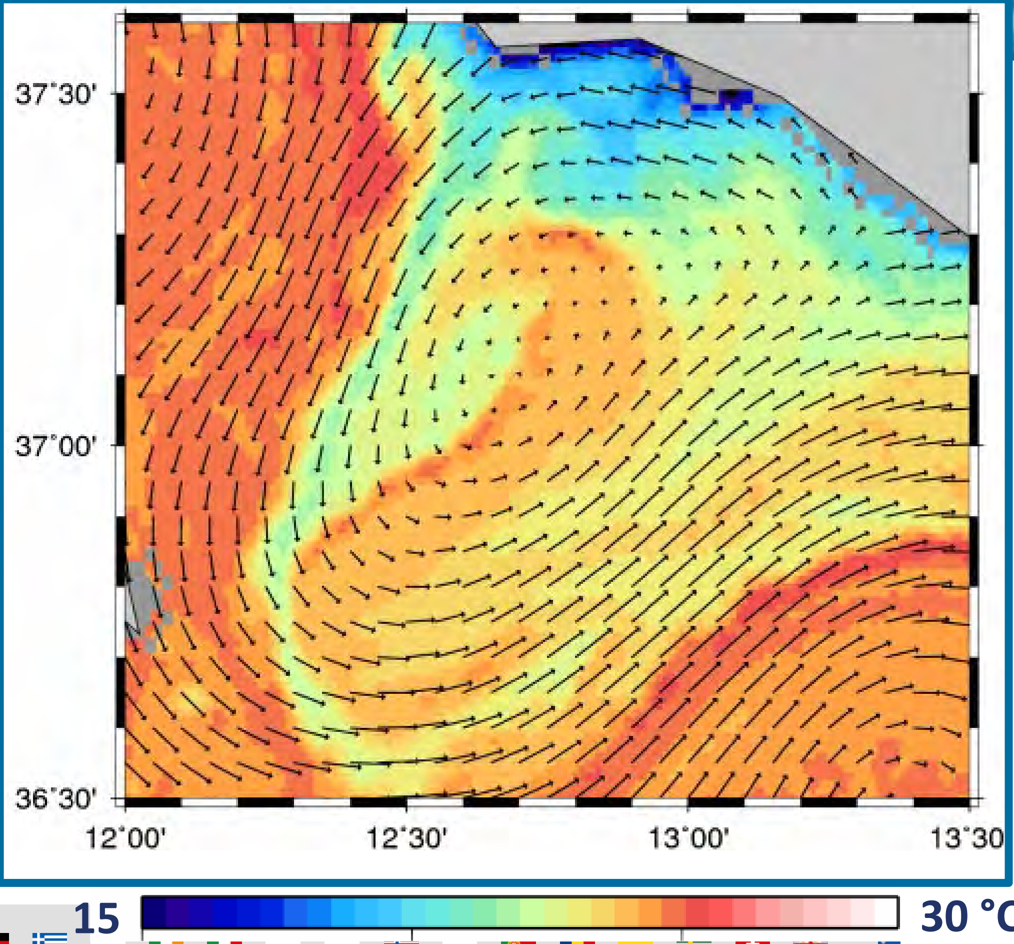


$mg.m^{-3}$

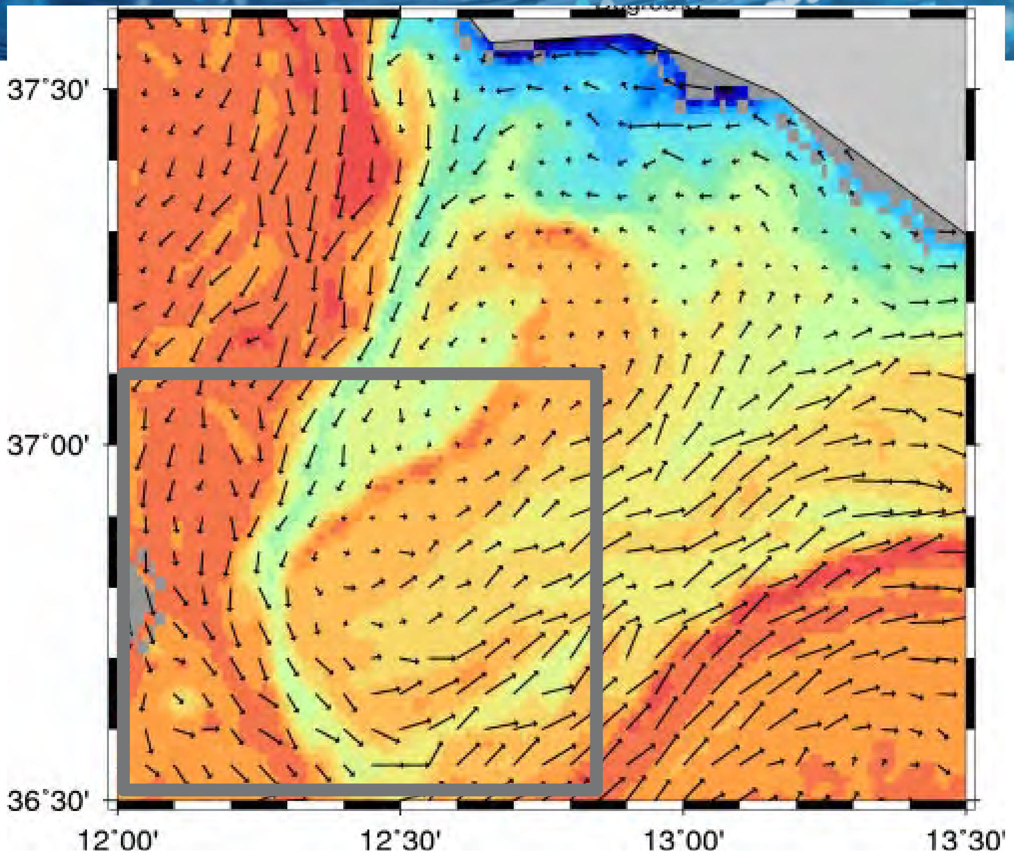
S-3 Absolute Dynamic Topography



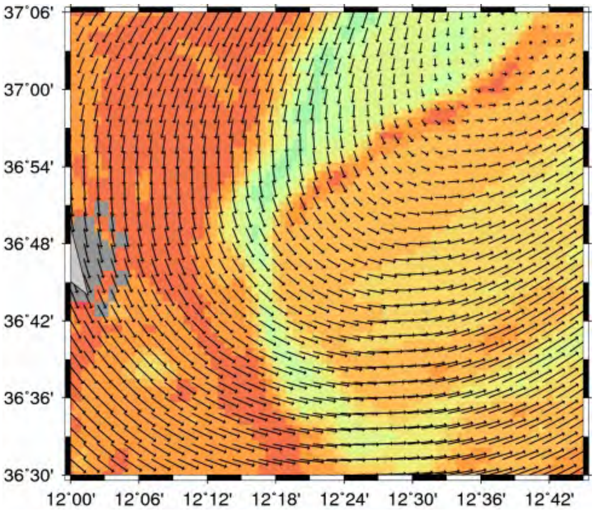
Sentinel-3 data on July, 28th 2016



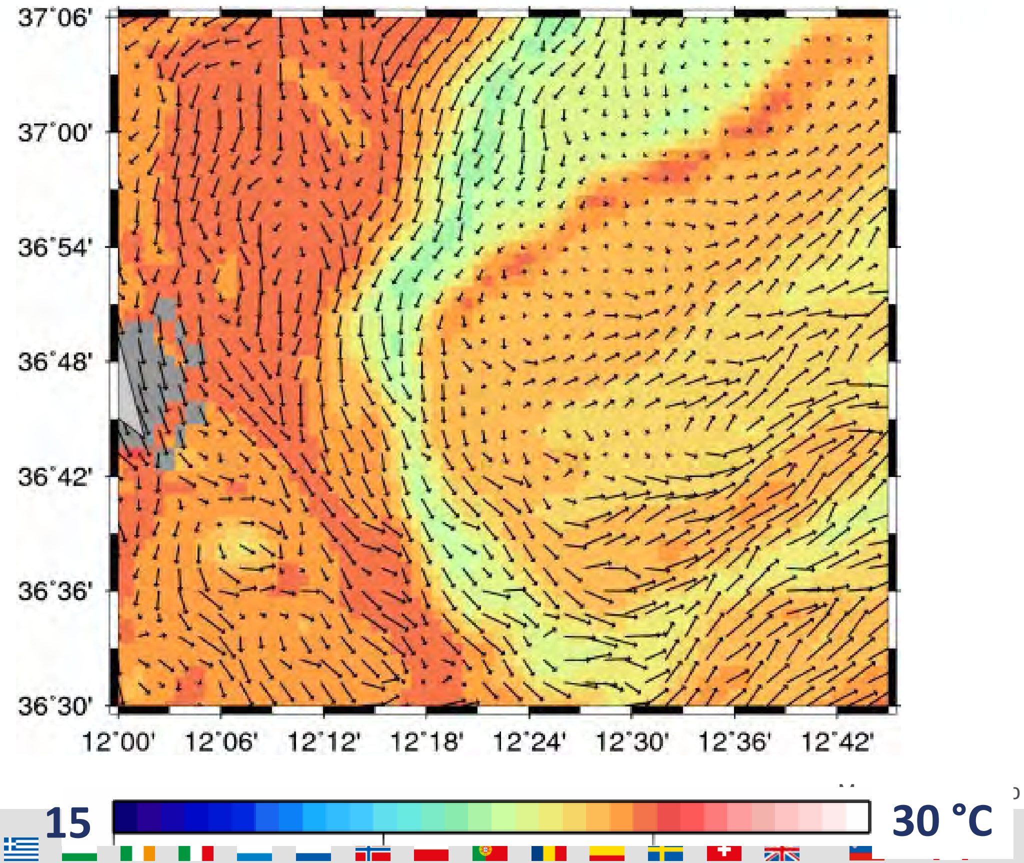
Sentinel-3 data on July, 28th 2016



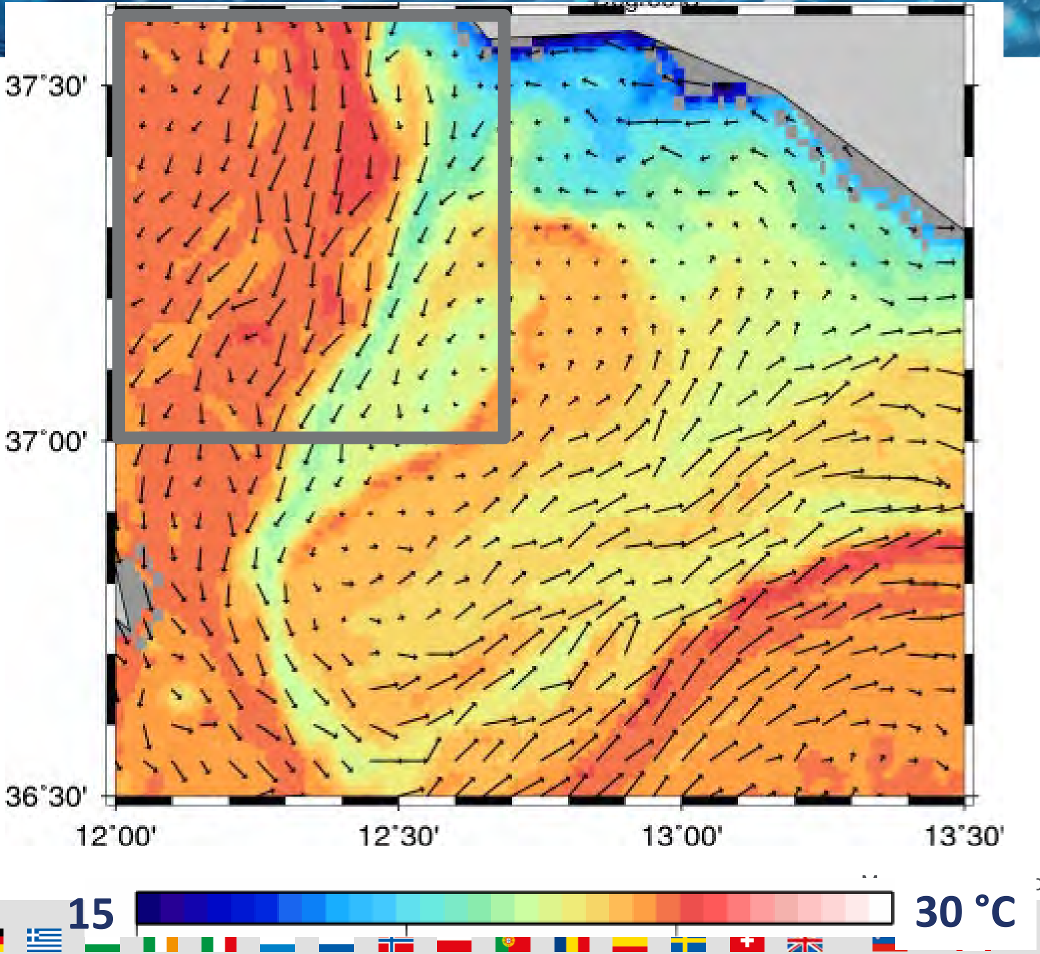
Sentinel-3 data on July, 28th 2016



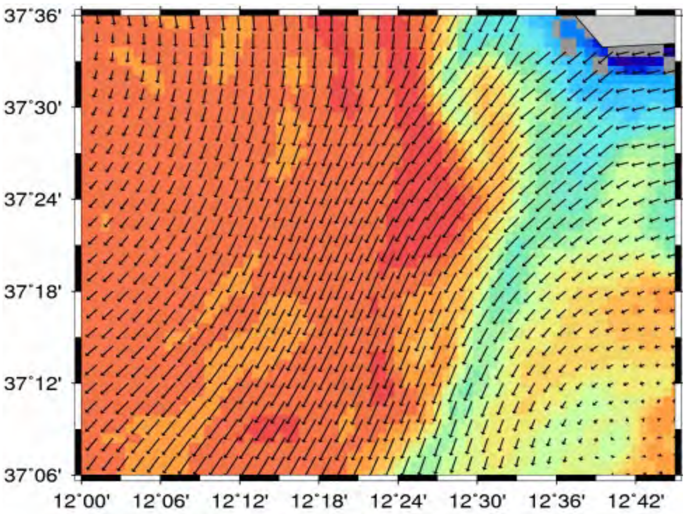
Sentinel-3 data on July, 28th 2016



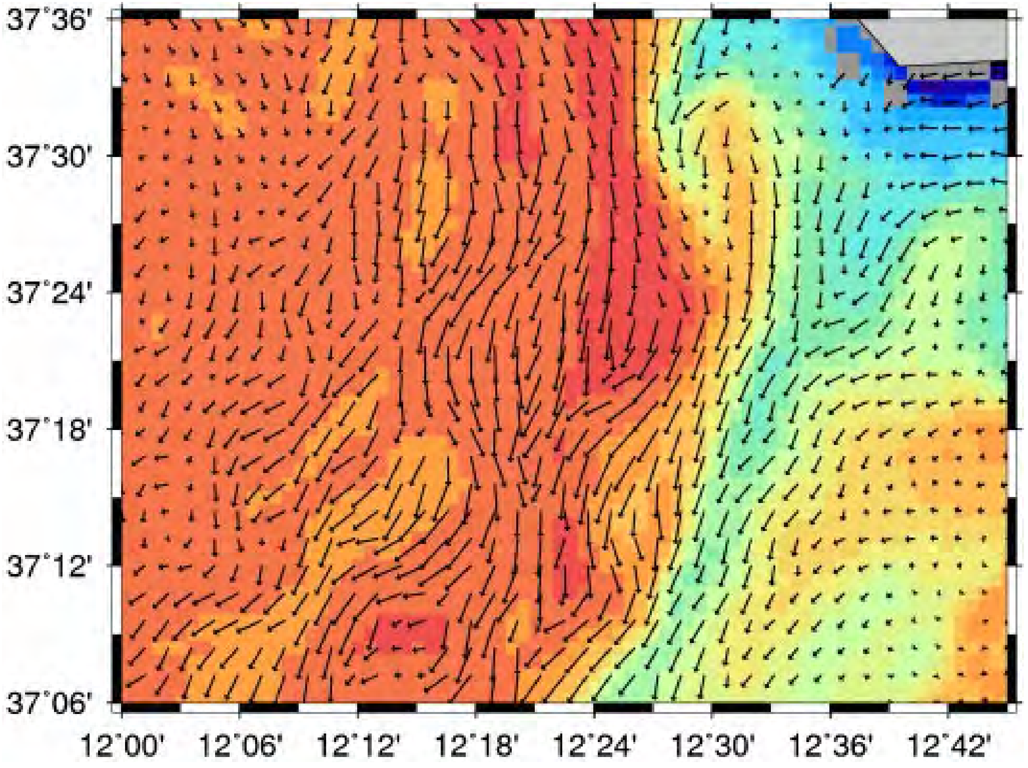
Sentinel-3 data on July, 28th 2016



Sentinel-3 data on July, 28th 2016



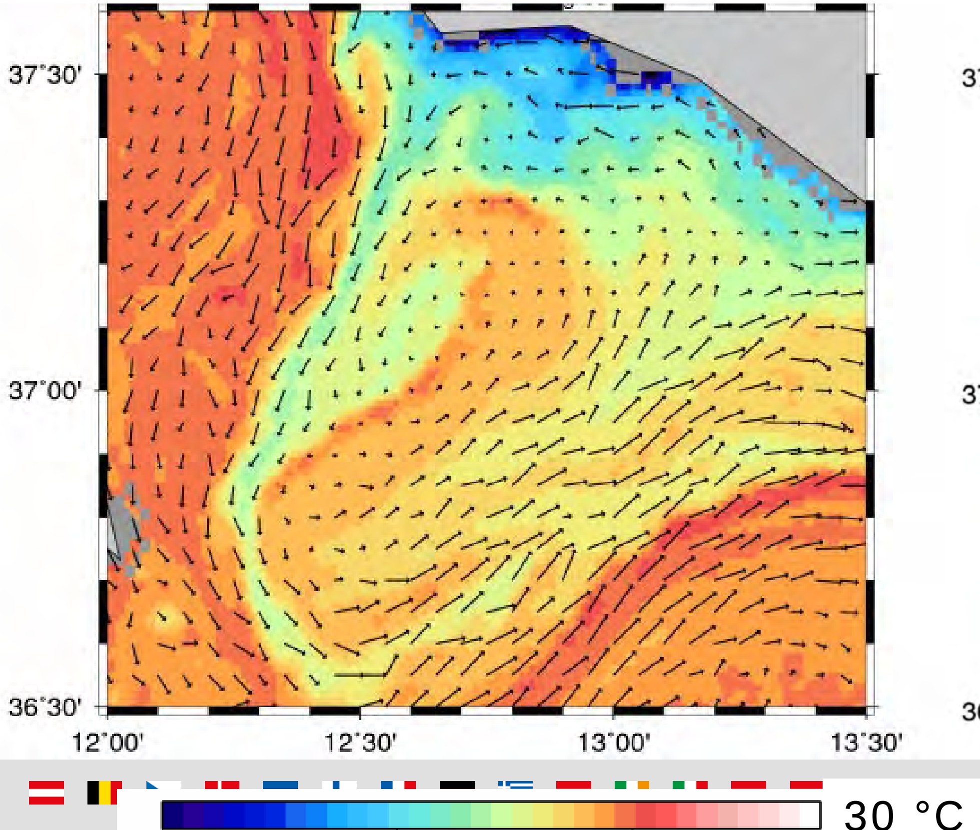
Sentinel-3 data on July, 28th 2016



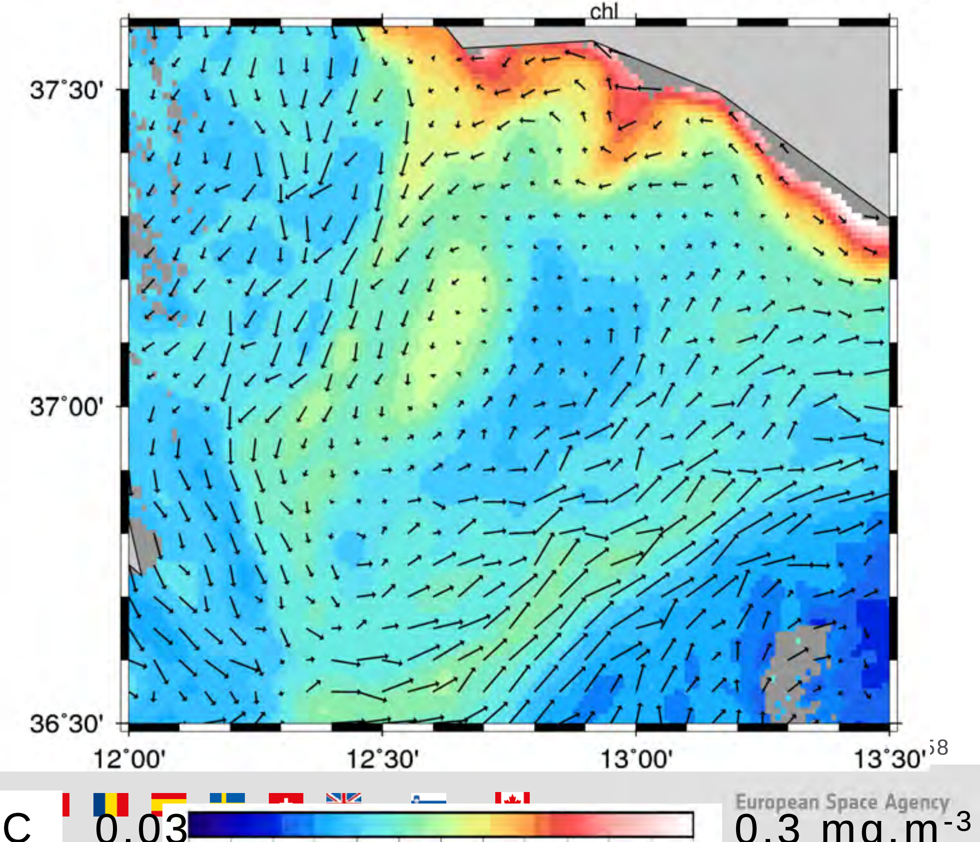
15
30 °C

Toward SSH/OC merging?

Optimal S3 SSH/SST velocities



Optimal S3 SSH/OC velocities

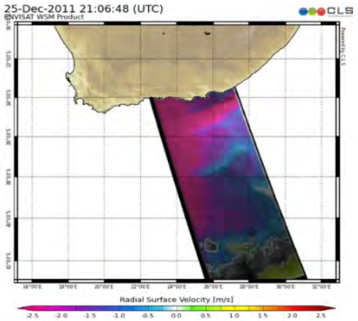


Sensor	Measured variable	Method	Surface current component retrieved	Spatio-temporal resolution
Altimeter + Gravimeter	Sea level above reference ellipsoid Geoid above reference ellipsoid	Optimal interpolated gridded field + Geostrophic approximation	Geostrophic current	100-400km 10-30 days
Scatterometer	Wind	Ekman model	Ekman current	25 km 12 hours
Microwave Radiometer	SST	Optical flow, MCC	Total surface currents	25 km 1 day
		E-SQG	Geostrophic surface currents	
Infrared Radiometer	SST	Optical flow, MCC	Total surface currents	Polar orbiting 10 km 1 day Geostationnary 10 km hourly
		E-SQG	Geostrophic surface currents	
L-Band radiometer	SSS	Optical flow, MCC	Total surface currents	100km 3-10 days
		E-SQG	Geostrophic surface currents	
Spectrometer	Ocean color	Optical flow, MCC	Total surface currents	Polar orbiting 10 km 1 day Geostationnary 10 km hourly
SAR	Range Doppler Anomaly Shift	CDOP sea state component of Doppler shift	Radial component of total current minus wind drift (included in CMOD)	Snashots 10 km 3 days

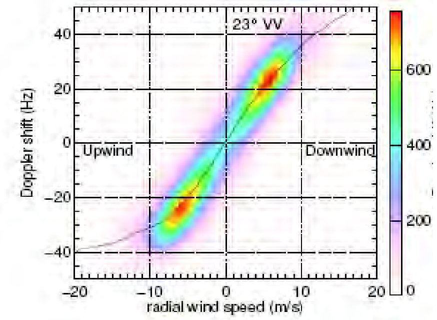
Surface velocities from SAR measurements

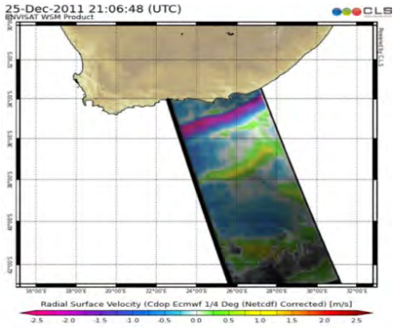
Chapron et al, 2005 ; Johannessen et al, 2008; Rouault et al, 2010

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:



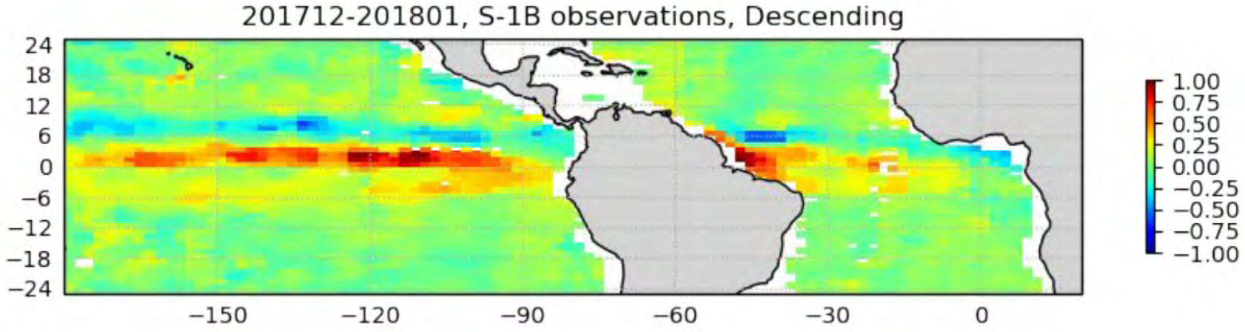
- The known movement of the satellite in orbit,
- A wave-state contribution highly correlated to wind speed which can be estimated using an empirical relationship between the range Doppler velocity and the near surface wind field, Mouche et al. (2012) with a C-band Doppler (CDOP) algorithm. These local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

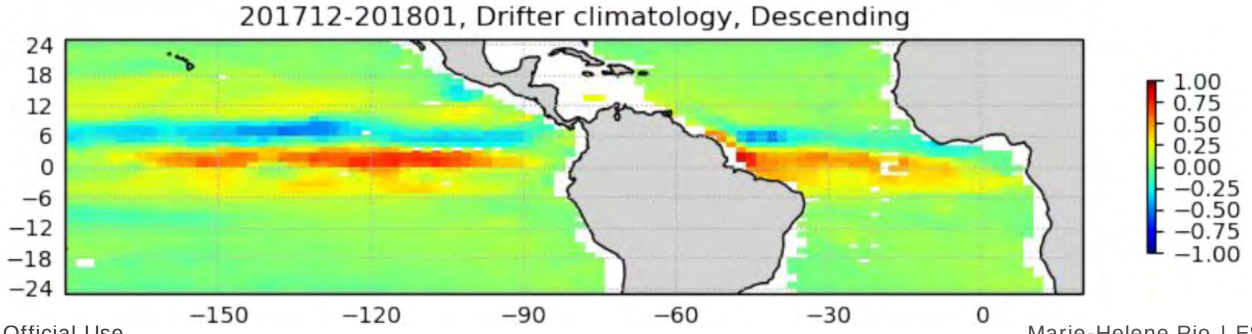




-a **measure of the sea surface current**, with 10km pixel size that contains the contributions, **projected onto the range direction**, of the **geostrophic currents, the tidal currents, the inertial oscillations.**

Descending tracks





Collard et al

Surface velocities from SAR measurements

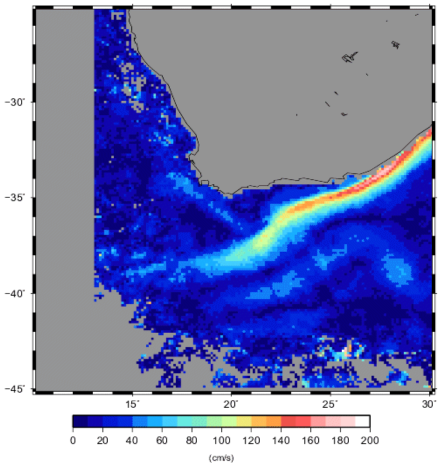
Synergy: SAR Doppler shift provides range velocities only. Altimeter velocity or SST front direction information can be used to recover the two components velocity.

$$\overset{*}{V}_a = \frac{V_a^{SAR}}{\cos(\beta_a)} \qquad \overset{*}{V}_d = \frac{V_d^{SAR}}{\cos(\beta_d)}$$

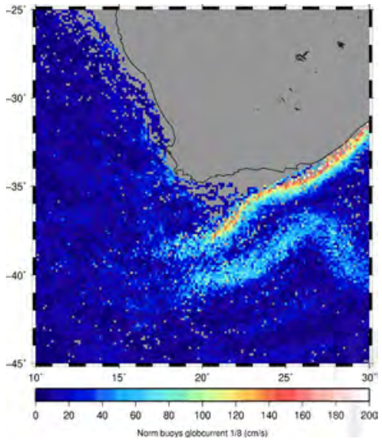
V_d^{SAR} V_a^{SAR} SAR-derived range velocities in ascending and descending passes

β_d β_a angle between the SAR range direction and the altimeter-derived current direction for ascending and descending passes

Mean SAR velocities



Mean drifter velocities



Bias U= -0.09m/s RMSU= 0.16 m/s
Bias V= -0.05 m/s RMS V=0.16m/s

Instantaneous SAR velocities versus instantaneous drifter velocities:

Bias U= -0.14m/s RMSU= 0.37 m/s Bias V= -0.12 m/s RMS V=0.28m/s

Conclusions

➢At the present time, **no direct measurement** of ocean surface currents from space

➢But, for the last two/three decades, a high number of space-borne sensors measuring different ocean variables from which ocean surface currents can be indirectly inferred:

SENSORS	VARIABLES	METHODS
Altimeter	altimetry	geostrophy
Radiometer	SST, SSS	MCC, e-SQG
Spectrometer	OC	Optical flow
SAR	rugosity	Doppler shift
Scatterometer	Wind	Ekman model

→ Ocean surface currents

➢**Each method has benefits and drawbacks** in term of physical content, spatial and temporal coverage, accuracy

➢For optimal exploitation of space data and best estimation of ocean surface currents, **synergy is needed** *(also with in-situ data!)*