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Regionalizing the impacts of wind and wave-induced currents on surface ocean dynamics: a long-term variability analysis in the Mediterranean Sea

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Key Points:

- Dynamically coherent regions in the Mediterranean are defined based on the Ekman, Stokes and geostrophic components variability.
- Ageostrophy dominates surface circulation short-term variability and exceed geostrophy over Northwest Mediterranean Sea in winter.
- Variations in the kinetic energy correlate well with the main Mediterranean climate modes in regions dominated by Ekman and Stokes.

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Abstract

Effects of wind and waves on the surface dynamics of the Mediterranean Sea are assessed using a modified Ekman model including a Stokes-Coriolis force in the momentum equation. Using 25 years of observations, we documented intermittent but recurrent episodes during which Ekman and Stokes currents substantially modulate the total mesoscale dynamics by two non-exclusive mechanisms: (i) by providing a vigorous input of momentum (e.g. where regional winds are stronger) and/or (ii) by opposing forces to the main direction of the geostrophic component. To properly characterize the occurrence and variability of these dynamical regimes we perform an objective classification combining self-organizing maps (SOM) and wavelet coherence analyses. It allows proposing a new regional classification of the Mediterranean Sea based on the respective contributions of wind, wave and geostrophic components to the total mesoscale surface dynamics. We found that the effects of wind and waves are more prominent in the northwestern Mediterranean, while the southwestern and eastern basins are mainly dominated by the geostrophic component. The resulting temporal variability patterns show a strong seasonal signal and cycles of 5 - 6 years in the total kinetic energy arising from both geostrophic and ageostrophic components. Moreover, the whole basin, specially the regions characterized by strong wind- and wave- induced currents, shows a characteristic period of variability at 5 years. That can be related with climate modes of variability. Regional trends in the geostrophic and ageostrophic currents shows an intensification of $0.058 \pm 1.43 \cdot 10^{-5}$ cm/s per year.

Plain Language Summary

The ocean dynamics plays a decisive role in the global balance of essential variables, such as heat, CO₂ or primary production, as well as in the dispersion of pollutants. However, the physical processes that control the mesoscale dynamics and its variability in the surface of the Mediterranean sea is not fully understood. Therefore, we have analyzed the regional contribution of the geostrophic and the wind and waves induced currents using a classification method based on a machine learning algorithm. We find that the effect of wind and waves is stronger over regions of the northwestern Mediterranean, while the southwestern and eastern basin is mainly dominated by geostrophy. We observe that regions where wind and wave dominate the dynamics co-vary with the main Mediterranean climate modes of variability. The geostrophic currents show an intensification with a clear shift in 2002, which suggests that this positive trend could be a part of a large decadal oscillation.

1 Introduction

Ocean currents are of crucial importance for the transport of physical, chemical and biological variables across the world oceans. They are the main responsible for the horizontal redistribution of energy, salt and heat, playing an important role in the climate system (Covey & Barron, 1988). In particular, the sea surface is a key transitional layer where most biological and biogeochemical activities concentrate and tightly interact with vigorous physical features (e.g. Hernández-Carrasco et al., 2014) ultimately affecting marine biodiversity patterns (e.g. Villarino et al., 2018) and atmosphere-ocean coupled processes (e.g. Bronselaer & Zanna, 2020). Hence, a precise knowledge of the circulation in the upper oceanic boundary layer and of its variability is key to many issues of broad scientific and practical importance, ranging from ecosystem and fisheries management (e.g. Dubois et al., 2016; Futch & Allen, 2019), the tracking of marine pollution including microplastic (e.g. Van Sebille et al., 2015) to marine safety such as search and rescue operations (e.g. Sayol et al., 2014).

Oceanic circulation results from movements of fluid in response to internal forces (pressure gradients and Coriolis forces) and external forces (gravity forces and frictional

67 forces, such as wind stress and waves at the surface and drag at the bottom and lateral
68 boundary layers). At the ocean surface, total currents result from several energy inputs
69 from diverse sources occurring at multiple scales. In particular, wind and waves inter-
70 act with the ocean general circulation, giving rise to a highly variable multi-scale envi-
71 ronment. During the last decade or so, mesoscale surface currents have traditionally been
72 interpreted as dominated by the geostrophy. This simplifying assumption, together with
73 the advances in satellite altimetry, have led the oceanographic community to estimate
74 surface horizontal currents from the balance between the pressure gradient and the Cori-
75 olis forces. However, although geostrophy provides a reasonable view of the low frequency/large-
76 scale motion of the ocean, it has limitations. As such, previous studies aimed at express-
77 ing total currents as a sum of both geostrophic and Ekman components (Sudre et al.,
78 2013; Rio et al., 2014). Despite relative improvements, our description of the upper oceanic
79 layer dynamics is still incomplete as it is also necessary to account for the high frequency
80 and ageostrophic motions caused by both wind- and wave-driven currents. Indeed, there
81 is growing evidence that the mesoscale ageostrophic flow plays an important role in the
82 transport and mixing processes, affecting the distribution patterns of transported ma-
83 terials (Dobler et al., 2019) such as, the fate of marine debris (Onink et al., 2019). More-
84 over, Fraser et al. (2018) have shown that wave-induced currents enhance ocean connec-
85 tivity around Antarctica, potentially affecting the local ecosystems.

86 Although great advances have been made in the last decades for measuring geostro-
87 phy at meso and larger scales or wind stress over the ocean surface, such as satellite scat-
88 terometers like QuikSCAT or ASCAT (Bourassa et al., 2019), wave and wind-wave com-
89 bined measurements are still limited to specific sites (mooring, stations and buoys) or
90 interpolated from radar radiometers (Ardhuin et al., 2018). However, the availability of
91 global forecasting systems both for wave and surface winds, allows the inclusion of these
92 high frequency velocities in recently developed models of the ocean circulation, by merg-
93 ing the different sources to obtain improved velocity products (Breivik et al., 2016; Onink
94 et al., 2019).

95 The wind-driven currents at the sea surface were initially studied by Ekman's sem-
96 inal work (Ekman, 1905). He proposed that the momentum balance between the tur-
97 bulence stress caused by the wind and Coriolis force can be modeled as a classical dif-
98 fusion problem but with a kinematic viscosity. Besides, gravity waves have an associated
99 current, the Stokes velocity resulting from the non-linearity of the wave orbital veloc-
100 ities (Stokes, 1847). From the Eulerian standpoint, the Stokes-drift-induced-current com-
101 ponent acts as an additive term that interacts with the mean ageostrophic current, ap-
102 pearing in the momentum equations as an external force such as, a vortex force or as the
103 Coriolis-Stokes force (McWilliams & Restrepo, 1999; Polton et al., 2005). The low and
104 high frequency velocities can be of the same order of magnitude depending on the in-
105 tensity of the local wind and wave fields (Polton et al., 2005; Breivik et al., 2016; Fraser
106 et al., 2018).

107 Despite substantial efforts in studying the effects of wind and waves on surface cur-
108 rents around the world (Kaiser, 1994; Polton et al., 2005; Ardhuin et al., 2009; Hui &
109 Xu, 2016; Onink et al., 2019), our knowledge of these ageostrophic currents and of their
110 impacts on the upper layer dynamics of the Mediterranean Sea is still poor. The Mediter-
111 ranean Sea is a semi-enclosed basin with large spatial and seasonal variability of both
112 winds and wave fields, making it an excellent laboratory to study the effects of the in-
113 teraction of the wind and wave induced currents in the general circulation. Sayol et al.
114 (2016) studied the energy and mass fluxes generated by wind-wave interactions in the
115 western part of the Mediterranean Sea and showed, that the induced surface transport
116 has a seasonal character, peaking during winter seasons. Recently, Morales-Márquez et
117 al. (2020) showed that this variability is largely controlled by large-scale climatic pat-
118 terns. The atmospheric circulation over the Mediterranean Sea can be indeed charac-
119 terized by specific modes of variability related to atmospheric teleconnections (Wallace

120 & Gutzler, 1981). The main climatic patterns influencing the Mediterranean dynamics
 121 are the North Atlantic Oscillation (NAO), the East Atlantic pattern (EA), the Scandi-
 122 navia pattern (SCAND) and the East Atlantic/Western Russia (EA/WR) (Barnston &
 123 Livezey, 1987; Morales-Márquez et al., 2020).

124 In this paper, we first derive analytical expressions to estimate the total oceanic
 125 surface currents as a sum of a geostrophic term and another ageostrophic one, taking into
 126 account wind and waves forcings. We then apply our expressions to altimetric and re-
 127 analyses datasets in order to compute surface currents over the whole Mediterranean Sea
 128 for the last 25 years. It allows investigating the relative contributions, which vary in space
 129 and time, of both geostrophic and ageostrophic components to the total kinetic energy.
 130 In order to identify the regions where the Ekman- and Stokes-induced flows affect sub-
 131 stantially the upper ocean dynamics, we perform an objective regionalization of the Mediter-
 132 ranean Sea. Homogeneous dynamical regions are unveiled using a machine-learning al-
 133 gorithm applied to an artificial neural network. Previous studies have proposed diverse
 134 objective regionalizations of the Mediterranean Sea (Ayata et al., 2018), using different
 135 statistical techniques, and based on different oceanic variables, e.g. climatological av-
 136 erages of temperature, salinity, nutrients concentrations (Reygondeau et al., 2017), trans-
 137 port properties of surface waters (Rossi et al., 2014) or phytoplankton variability (d’Ortenzio
 138 & d’Alcalà, 2009; Nieblas et al., 2014). By doing so, we analyze the regional variability
 139 of the dynamical impacts of both winds and waves on the surface circulation in the Mediter-
 140 ranean Sea. In each homogeneous dynamical region, we further extract the dominant tem-
 141 poral scales and study their relationships with the main climatic modes to assess the in-
 142 terannual variability of the currents field.

143 2 Sea Surface Currents

144 Total current at the sea surface (\mathbf{U}_t) can be expressed as the sum of the geostro-
 145 phy, $\mathbf{U}_g = u_g + iv_g$, and an ageostrophic velocity, $\mathbf{U}_a = u_a + iv_a$ which is associated
 146 with the wind and non linear wave-induced momentum along their direction of propa-
 147 gation:

$$\mathbf{U}_t = \mathbf{U}_g + \mathbf{U}_a. \quad (1)$$

148 2.1 Geostrophic currents

149 Considering a steady and Boussinesq flow, the geostrophic term can be obtained
 150 from the equilibrium between Coriolis and pressure gradient forces in the momentum equa-
 151 tion:

$$if\mathbf{U}_g = -\frac{1}{\rho_w}\nabla P, \quad (2)$$

152 where $\nabla = \frac{\partial}{\partial x} + i\frac{\partial}{\partial y}$ and P is the pressure. Using the hydrostatic balance in homoge-
 153 neous ocean, an expression of the geostrophic velocities can readily be obtained from the
 154 Sea Surface Height (SSH) as:

$$u_g = -\frac{g}{f}\frac{\partial(\text{SSH})}{\partial y}, \quad v_g = \frac{g}{f}\frac{\partial(\text{SSH})}{\partial x}, \quad (3)$$

155 where g is the acceleration of gravity and $f = 2\Omega\sin\phi$ is the Coriolis parameter with
 156 Ω the angular Earth velocity and ϕ the latitude.

157 2.2 Ageostrophic currents: wind and wave driven components

158 The wind- and wave-induced ageostrophic currents in the upper boundary layer are
 159 obtained from the horizontal Ekman-wave induced momentum equation for a steady and

160 Boussinesq flow (Lewis & Belcher, 2004; Huang, 1979; Polton et al., 2005):

$$if\mathbf{U}_a = \frac{\partial}{\partial z} \left(A_z \frac{\partial \mathbf{U}_a}{\partial z} \right) - if\mathbf{U}_s, \quad (4)$$

161 where $\mathbf{U}_a = u_a + iv_a$ denotes the horizontal ageostrophic velocity in complex notation,
 162 $\mathbf{U}_s = u_s + iv_s$ is the wave-induced Stokes velocity, resulting $if\mathbf{U}_s$ the term from the
 163 Coriolis-Stokes force (rotation acting on the Stokes drift), and A_z is the vertical eddy
 164 viscosity of sea water. Previous works (Huang, 1979; Polton et al., 2005) have shown that
 165 the flow is significantly modified by the Coriolis–Stokes force not only at the near-surface
 166 layer but throughout the entire Ekman layer. We assume that the vertical viscosity is
 167 constant and equal to $A_z = 1.0710^{-2} \text{m}^2 \text{s}^{-1}$ (McWilliams et al., 1997). While other ap-
 168 proaches considered a vertical parametrization of A_z (Wenegrat & McPhaden, 2016; Polton
 169 et al., 2005), we use a constant value since: (i) it would only affect the estimation at the
 170 surface boundary condition and, (ii) the wave-induced circulation changes are indepen-
 171 dent of the vertical mixing parametrizations when the typical depth scale of the waves
 172 effect is smaller than the typical Ekman layer.

173 Assuming a monochromatic wave field propagating in deep water with a wavenum-
 174 ber $\mathbf{k} = (k_x, k_y)$, the Stokes drift velocity, $\mathbf{U}_s = U_s \hat{\mathbf{k}}$, is related to the wave as (Phillips,
 175 1966):

$$U_s = a^2 \omega k e^{2kz}, \quad (5)$$

176 being a the wave amplitude, $\omega = \sqrt{gk}$ the wave frequency at deep waters, $k = |\mathbf{k}|$ and
 177 the wave number unit vector:

$$\hat{\mathbf{k}} = \cos(\theta_w) + i \sin(\theta_w), \quad (6)$$

178 with θ_w the mean direction of propagation waves, which is not necessarily parallel to the
 179 wind stress. We compute the Stokes including the sea and swell components of the wave.
 180 Thus, the combined effect of wind and waves are not only provided in the generation area
 181 but also while waves propagate across their swell.

182 Both boundary conditions required by the second-order ordinary differential equa-
 183 tion (Eq. 4) are given at the free surface and at the vanishing boundary as:

$$A_z \frac{\partial \mathbf{U}_a}{\partial z} = \frac{1}{\rho_w} \left(\boldsymbol{\tau} + \frac{\partial \mathbf{S}}{\partial \mathbf{X}} \right), \quad \text{at } z = 0, \quad (7)$$

$$\mathbf{U}_a \rightarrow 0, \quad \text{as } z \rightarrow -\infty, \quad (8)$$

185 where ρ_w the sea water density and $\boldsymbol{\tau}$ is the wind stress at the sea surface, $\boldsymbol{\tau} = \rho_a C_D u_{10} \mathbf{u}_{10}$,
 186 where ρ_a is the air density (1.2 kg/m^3), u_{10} is the 10-m wind speed and C_D is the neu-
 187 tral drag coefficient taken as, $C_D = (2.7/u_{10} + 0.142 + 0.0764u_{10})/1000$ following Large
 188 et al. (1994). S_{ij} are the components of the radiation stress provided at the surface by:

$$\frac{\partial \mathbf{S}}{\partial \mathbf{X}} = \left(\frac{\partial S_{xx}}{\partial x} + \frac{\partial S_{yx}}{\partial y} \right) + i \left(\frac{\partial S_{xy}}{\partial x} + \frac{\partial S_{yy}}{\partial y} \right),$$

$$S_{xx} = \frac{E}{2} \cos^2 \theta_w, \quad S_{xy} = S_{yx} = \frac{E}{2} \sin \theta_w \cos \theta_w, \quad S_{yy} = \frac{E}{2} \sin^2 \theta_w,$$

189 with $E = \rho g a^2 / 2$.

190 The steady-state solution of Eq. 4 subjected to boundary conditions (Eq. 7-8) is:

$$\mathbf{U}_a(z) = \frac{\boldsymbol{\tau}}{\rho_w A_z m} e^{mz} + \frac{\frac{\partial \mathbf{S}}{\partial \mathbf{X}}}{\rho_w A_z m} e^{mz} + \frac{m^2 \mathbf{U}_{s0}}{4k^2 - m^2} e^{2kz} - \frac{2km \mathbf{U}_{s0}}{4k^2 - m^2} e^{mz}, \quad (9)$$

191 with $\mathbf{U}_{s0} = \mathbf{U}_s(z=0)$, $m = \sqrt{if/A_z} = (1+i)\lambda$ and $\lambda = \sqrt{f/(2A_z)}$. The characteris-
 192 tic depth of the Ekman layer is defined as $\delta_e = 1/m$ and the characteristic Stokes depth
 193 scale as $\delta_s = 1/2k$.

194 To clarify the importance Coriolis-Stokes interaction, Eq.9 is rewritten as,

$$\mathbf{U}_a(z) = \mathbf{U}_E(z) + \mathbf{U}_{\tau_w}(z) + \mathbf{U}_S(z) + \mathbf{U}_{ES}(z). \quad (10)$$

195 Each term constituting Eq. 10 corresponds to the different components of the ageostrophic
 196 velocity. $\mathbf{U}_E(z)$ represents the classical Ekman component. $\mathbf{U}_{\tau_w}(z)$ accounts for the
 197 surface current induced by the wave radiation stress, which will not be analyzed sepa-
 198 rately in the following sections because its value is small compared to the other compo-
 199 nents, $\mathbf{U}_S(z)$ is the Stokes component, that decreases over the Stokes depth scale, be-
 200 ing much shallower than the Ekman layer ($\delta_s \ll \delta_e$). The latter component is corre-
 201 lated with the dynamical response to the Coriolis–Stokes force, being different than the
 202 Lagrangian Stokes drift U_s given by Eq. 5. The last term, $\mathbf{U}_{ES}(z)$ is the Ekman-Stokes
 203 component that accounts for the non-linear interaction between wind and waves acting
 204 over the entire Ekman layer (Polton et al., 2005).

205 Here, \mathbf{U}_a is integrated over 1 meter depth since the mean Stokes layer depth is gen-
 206 erally smaller than 2m in the Mediterranean Sea (Sayol et al., 2016). Fig. 1 displays an
 207 example of the vertical distribution of the Stokes (red arrows), Ekman (blue arrows) and
 208 ageostrophic (black arrows) velocity components along the water column in the Mediter-
 209 ranean Sea. It clearly shows that the Stokes component has a smaller influence (only a
 210 few meters near sea surface), while the Ekman component has a significant effect at deeper
 211 depths. Also, it is important to note how the Stokes component modify the vertical dis-
 212 tribution of the ageostrophic velocities in spite of its relative small value as compared
 213 to the Ekman velocity (see the difference between blue and black arrows).

214 The velocity fields obtained with this formulation have been validated with the drifters-
 215 database provided by the National Institute of Oceanography and Experimental Geo-
 216 physics Institute (OGS) of Italy. This data encompasses drifters in the Mediterranean
 217 Sea from 1986 to 2016. The averaged separation distance between the real and the vir-
 218 tual drifter trajectories averaged over all the hourly initializations is smaller (up to 15
 219 km smaller after 72 hours of integration, not shown) when the virtual drifter trajectory
 220 is computed using the total velocity field as compared to the geostrophic velocities. Be-
 221 sides, we found that the variance of the difference between the virtual and the real drifters
 222 is also smaller when we advect the drifters in the total velocity field, showing that er-
 223 rors in \mathbf{U}_T are significantly reduced as compared to \mathbf{U}_g .

224 Further modifications of the Ekman model have been considered in the last years,
 225 such as including an additional ageostrophic component caused by the geostrophic stress
 226 (McWilliams et al., 2015; Wenegrat & McPhaden, 2016). The global comparison between
 227 the different ageostrophic components performed in Wenegrat and McPhaden (2016) shows
 228 that this term has a higher influence at low latitudes while the Coriolis-Stokes stress dom-
 229 inates at higher latitudes, as in the case of the Mediterranean Sea. Additionally, in the
 230 very surface layers of the ocean this component (geostrophic stress) has not a significant
 231 value on the effective stress profile, and a relative value respect to the surface wind stress
 232 smaller than 3% in spring and summer and smaller than 8% in winter and autumn over
 233 the Mediterranean Sea, as shown in Figure 9 of Wenegrat and McPhaden (2016). They
 234 also found that the contribution of Coriolis-Stokes stress is 5 times greater than the geostrophic
 235 stress in the Mediterranean Sea, with a seasonal variability significantly larger in the Coriolis-
 236 Ekman induced currents. This small contribution suggests that the results obtained in
 237 this study will not be substantially modified if the geostrophic stress is considered in the
 238 Mediterranean basin.

239 3 Data

240 3.1 Wave and atmospheric data

241 Gridded wave and sea surface wind data can be obtained from remote sensing equipped
 242 with scatterometer (Bourassa et al., 2019) and from model outputs. However, while satel-

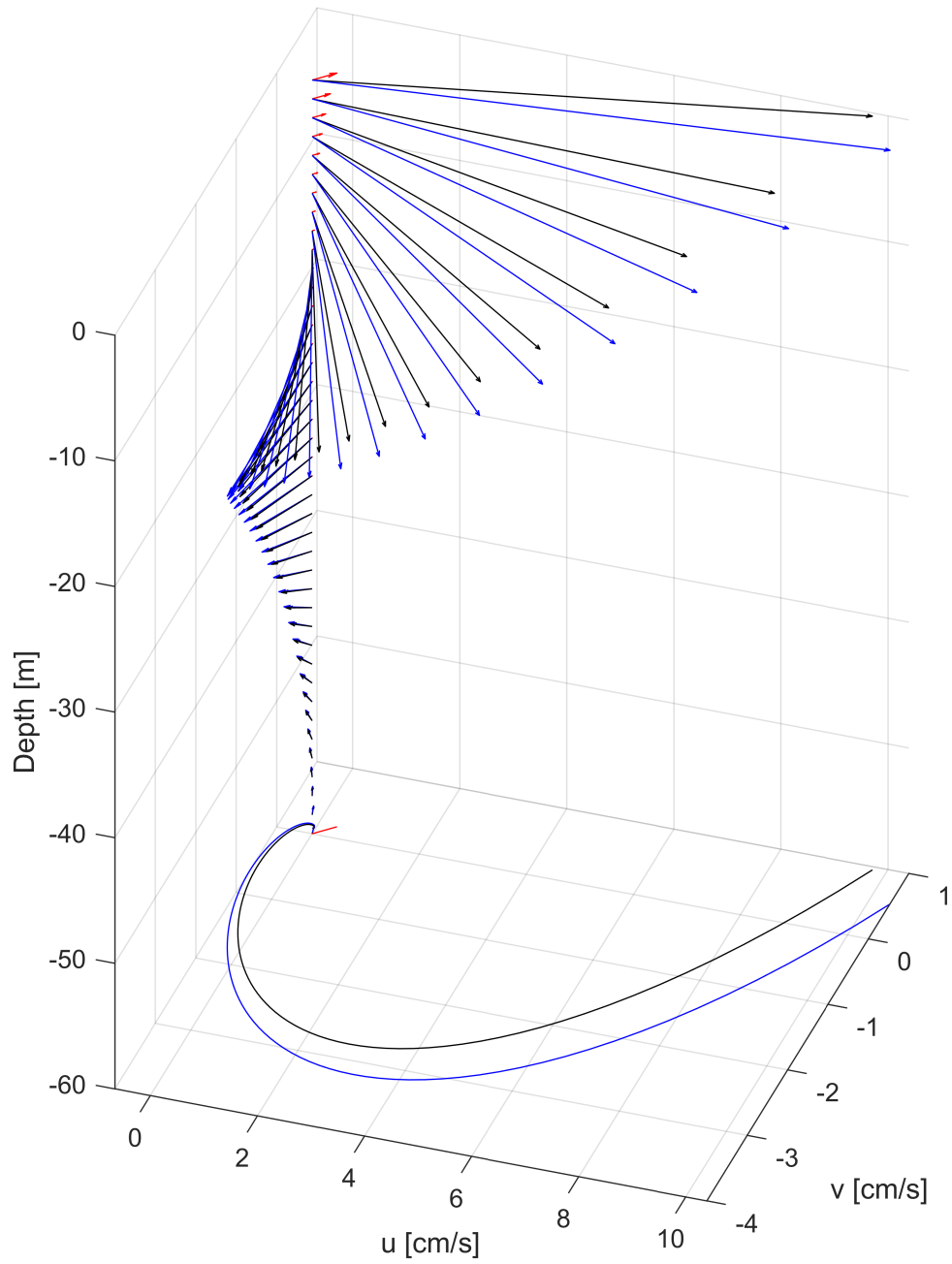


Figure 1. Ekman (blue line), Stokes (red line) and ageostrophic (black line) velocity profiles at spatial point 6°E, 38.5°N corresponding to the 5th of February 2014 at 6 : 00 UTC.

Table 1. Statistical comparison between the ERA-Interim and ERA5 dataset.

Variable	R^2	SCI	RB
SWH	0.96	0.20	-0.04
U_{10}	0.90	0.42	-0.01
V_{10}	0.89	0.44	-0.05

lites collect indirect observations of wind and waves (Ardhuin et al., 2018), data are acquired along tracks, generating maps with an effective resolution of approximately 40-50km and one week. Since the wave field changes at high frequency, that is for periods spanning a few hours, remote-sensed winds are not the most suitable dataset in order to study the wave effect on surface circulation. Concurrently, there exist nowadays consistent and global database about the wave field, also providing high-resolution wind velocities, that are generated by model reanalyses. Such model reanalyses have been extensively validated with different *in-situ* observations (Berrisford et al., 2011) and have already been used to study transport in the ocean (Breivik et al., 2016).

Surface waves and 10-m above the sea surface wind velocities are provided by the ERA-Interim reanalysis (Dee et al., 2011). Wave fields are obtained using the WAM wave model with the assimilation of available global measurements of ERS1 wave height data (Janssen et al., 1997). These reanalysis data are provided by local GRIB code of the European Centre for Medium-Range Weather Forecasts (ECMWF) covering the period between 1979 and 2019 with a temporal resolution of 6 hours and a spatial resolution of 0.125° both in latitude and longitude in the Mediterranean Sea (Fig. 2). For a detailed description of these products the reader is referred to Berrisford et al. (2011).

While new wind and wave reanalysis are currently available in the Mediterranean Sea, we use ERA Interim since it provides wind and wave fields at a spatial resolution consistent with the geostrophic velocity field; i.e. 0.125° in the Mediterranean Sea against 0.25° of winds provided by ERA5, and even coarser for the wave field (0.5° against 0.125° of ERA-Interim). Furthermore, we have compared both products and we have found a high correlation and low values of Scatter index and Relative bias between both reanalysis data, as shown in Table 1 (Roelvink et al., 2009).

The leading climatic modes of variability in the Mediterranean Sea, NAO, EA, EA/WR and SCAND have been downloaded from the NOAA Climate Prediction Centre (<https://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml>; last access on: 27 February 2020). NAO is usually defined as the sea level pressure difference between the Iceland Low and the Azores High (Hurrell et al., 2003). The EA index consists of a north-south dipole of anomaly over the North Atlantic, with a strong multidecadal variability. The EA/WR is represented with four main anomaly centers; positive phase is associated with positive wave height anomalies located over Europe and negative wave height anomalies over the central North Atlantic. Finally the SCAND pattern is composed with a primary circulation center over Scandinavia, with weaker centers of opposite sign over western Europe. Climate indices are constructed through a rotated principal component analysis of the monthly mean standardized 500-mb height anomalies in the Northern Hemisphere, ensuring the independence between modes at a monthly scale due to orthogonality (Barnston & Livezey, 1987).

3.2 Geostrophic velocity field

Geostrophic currents are derived from the Sea Level Anomaly (SLA) provided by the Copernicus Marine Environment Monitoring Service (CMEMS) through the product *Mediterranean Sea Gridded L4 Sea Surface Heights and derived variables reprocessed*

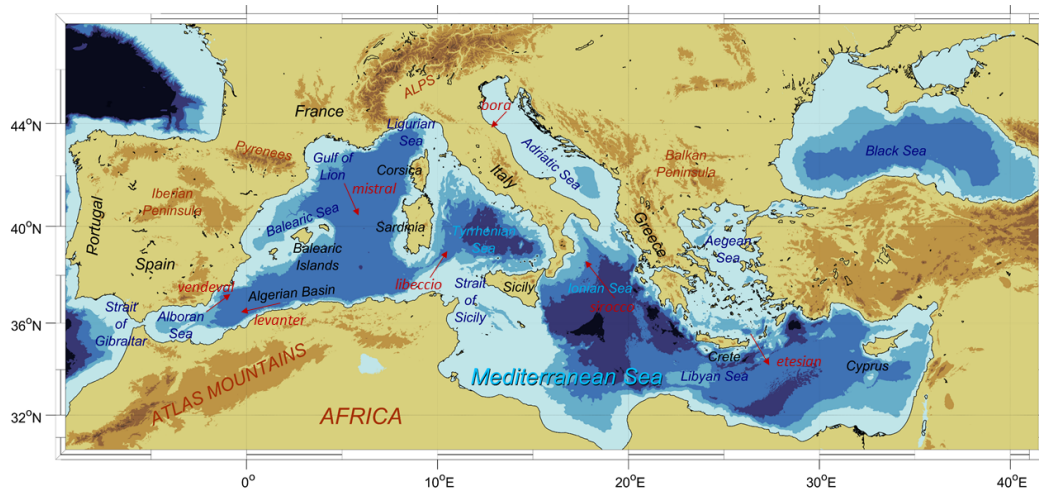


Figure 2. Topography of the Mediterranean basin and naming convention of the main geographical locations used in the paper.

(1993-ongoing) (https://resources.marine.copernicus.eu/?option=com_csw&view=details&product_id=SEALEVEL_MED_PHY_L4_REP_OBSERVATIONS_008_051; last access on: 7 February 2019). This product merges the different altimeter missions available (Jason-3, Sentinel-3A, Haiyang-2A, Saral/AltiKa, Cryosat-2, Jason-2, Jason-1, TOPEX/Poseidon, ENVISAT, GFO, ERS1/2). SLA data are homogenized by the DUACS multimission altimeter data processing system in order to generate gridded L4 absolute geostrophic velocities and optimal reprocessed products for long-term analysis, including the robust estimation of regional mean sea levels trends (Pujol et al., 2016). This data set has a daily temporal resolution and is provided over a regular mesh of 0.125° over the entire Mediterranean Sea.

Velocity fields \mathbf{U}_g and \mathbf{U}_a are computed every 6 hours for 25 years from 1993 to 2018. For the geostrophic component, daily data are linearly interpolated to 6-hourly time step, while for the ageostrophic component, each of the terms are computed for each model output.

4 Statistical Methods

4.1 Self Organizing Maps

Self-Organizing Maps (SOMs) is a statistical method using unsupervised learning neuronal network which is especially suited to extract patterns in large datasets (Kohonen, 1982). SOM is a nonlinear mapping tool that reduces the high-dimensional feature space of the input data to a lower dimensional (usually 2D) network of units called neurons. Through the machine learning algorithm, SOMs are able to compress the information contained in large and complex dataset into a single set of patterns. Similar neurons are mapped adjacent on the network, since SOM preserves topology. This helps to improve the visualization of the patterns, being one of the advantages of this technique.

SOM learning process algorithm inserts the input velocity fields into a neural network which is modified along an iterative procedure. Each neuron is represented by a

weight vector containing as many components as the dimension of the input sample data. At each iteration, the neuron whose weight vector is closest (as measured by minimum Eulerian distance) to input data vector is retrofitted together with its topological neighbors towards the input sample according to a neighborhood relationship specified with a given mathematical function. At the end of the training process, SOM approximates the probability density function of the input data associating each neuron with a reference pattern.

The SOM technique is able to be applied both in the spatial and temporal domains. Since we are interested in classifying the regions in the Mediterranean Sea according to the temporal variability of each of the velocity components, we implement SOM analysis in the time domain. The input data set is constituted not only by the total velocity time-series (\mathbf{U}_T) at each grid point, but also by coupling the geostrophic (\mathbf{U}_g), Ekman (\mathbf{U}_E) and Stokes (\mathbf{U}_S) velocities at the same grid point; as such, it allows analyzing the simultaneous variations of these terms. The resulting time-series are normalized before starting the learning process. At its completion, each neuron will correspond to a specific velocity temporal pattern for \mathbf{U}_T , \mathbf{U}_g , \mathbf{U}_E and \mathbf{U}_S . Then, the time-series of the velocity components at each grid point are classified in accordance with the SOM temporal patterns, providing a map of different sub-regions characterized with a particular temporal variability. To compromise the levels of the regionalization and its interpretability, we retain 6 neurons (2x3 SOM) for the temporal analysis. Preliminary tests using larger numbers of neurons returned more detailed temporal patterns for numerous sub-regions which are, however, difficult to clearly distinguish by their dynamical behaviors (see the supplementary material Fig. A1, Fig. A2 and Hernández-Carrasco and Orfila (2018)). We use a hexagonal map lattice in order to have equidistant neighbors and do not introduce artificial anisotropy. We opted for a linear mode for the initialization, a batch algorithm for the training process, and an 'ep' type of neighborhood function since this parameter configuration produces lower quantitative and topological errors and a minimize computational cost (Liu et al., 2006).

4.2 Wavelet power spectral method

Wavelet transform of a time-series x_n ($W^X(s)$) performs a time-frequency domain decomposition of the time-series by varying the wavelet scale s and by estimating its spectral characteristics as a function of time (Torrence & Compo, 1998). Wavelet is able to extract local-frequency information from a temporal signal in order to extract the dominant modes of variability and detect changes over time (Torrence & Compo, 1998). Wavelet uses a Fourier transform approach on a sliding temporal window returning frequencies at each time step, therefore being well suited for identifying periodic phenomena with changing spectra (Kaiser, 1994). This tool facilitates the study of time-series that contain non-stationary power at many different frequencies (Daubechies, 1990), as is the case here. We used a Morlet wavelet transform, which is a plane wave of wavevector ω_0 modulated by a Gaussian of unit width with an adimensional frequency $\omega_0=6$ (i.e. it contains 6 complete cycles of the temporal scale that is being analyzed). This wavelet base function is adequate to be localized in both time and frequency spaces and therefore to properly assess changes in the wavelet amplitude over time (Torrence & Compo, 1998). To distinguish the signal from the underlying noise, a threshold above the 95% confidence interval of a red-noise spectrum was used. The ability of wavelets to extract significant frequencies in localized time periods provides a powerful tool to characterize the patterns resulting from the previously-described SOMs analysis in the time domain.

4.3 Combined SOM-Wavelet coherence analysis

To assess the response of the sub-regions identified by the SOMs to large-scale forcing, we use an approach based on the Wavelet Coherence Analysis (WCA) between two time-series (Grinsted et al., 2004). WCA characterizes cross-correlations by identifying

the main frequencies, phase differences and time periods over which the relationships between the variability of the currents components (geostrophy, Ekman and Stokes) and the main relevant large-scale forcing (e.g. NAO, EA, EA/WR and SCAND indices) are tight in each region. To do so, we first analyze the variability in both frequency and time of each velocity components characteristic time and the time series of the climate indices, using the continuous wavelet transform.

Using the cross-Wavelet Transform (XWT), we determine the cyclic changes of the velocity components and their relationship with the climatic indices described above, in each of the sub-regions. The XWT of two time-series x_n and y_n indicates common power and relative phase in the frequency-time domain, given by $W^{XY}(s) = W^X(s)W^{Y*}(s)$, where $*$ represents the complex conjugate. $|W^{XY}(s)|$ is the cross-wavelet power and the complex argument $\arg(W^{XY}(s))$ is the relative phase between both time-series (shown in the Fig. 10 as arrows).

Finally the degree of coherence of the XWT at each time point is obtained by computing the coefficient R^2 given by the squared absolute value of the smoothed cross-wavelet spectrum, normalized by the product of the smoothed wavelet squared individual spectra, for each scale (Torrence & Compo, 1998; Grinsted et al., 2004), as:

$$R_n^2 = \frac{|S(s^{-1}W_n^{XY}(s))|^2}{S(s^{-1}|W_n^X(s)|^2)S(s^{-1}|W_n^Y(s)|^2)}, \quad (11)$$

whose values range from 0 (no correlation) to 1 (perfect correlation) and where S denotes the smoothing operator along the wavelet scale axis and along time. R^2 can be interpreted as a localized correlation coefficient in the frequency-time domain. It should be noted that while cross-wavelet analysis does not establish causative relationships, still allows identifying possible linkages between variables through the synchrony of their time-series.

Last but not least, wavelet coherent analysis is particularly suited to unveil regional relationships between global forcing (climate modes of variability) and the temporal velocity patterns obtained from the SOM given its ability to extract the frequencies and time periods when two time-series are correlated, wavelet coherent analysis is particularly suited to unveil regional relationships between global forcings (climate modes of variability) and the temporal velocity patterns obtained from the SOM.

5 Results and discussion

The overall picture of the mesoscale dynamics at the upper layer is mainly dominated by the geostrophic component for most space and time windows considered (not shown). However we found time periods where the ageostrophic velocities associated with wind and waves effects largely govern the main circulation over different regions of the Mediterranean Sea. As an example, Fig. 3 shows the total surface current and its respective components for the 19th of January 2005 at 12 : 00 UTC. It exemplifies a dynamical situation characterized by the net prevalence of Stokes and Ekman-induced velocities compared to the geostrophic component. At the geographical coordinate N38° E7°, 37', 30" (i.e. central location of the south-western Mediterranean basin), the maximum values of Stokes velocity reaches 15cm/s, being the Ekman velocity of 78cm/s which is largely exceeding the geostrophic velocity of 18cm/s. The contributions of \mathbf{U}_S and \mathbf{U}_E to the total velocity at this location for that particular date are 16.7% and 85.18%, respectively. As shown in Fig. 3, the spatial distributions of the ageostrophic velocities between the eastern and western basins clearly differ. While in the western Mediterranean, the total velocity is mainly governed by the Ekman component (i.e. intense winds blowing in the Gulf of Lion towards the center of the basin and modifying the Northern Current), the eastern Mediterranean basin is mainly governed by geostrophy (see Fig. 2 for the distinct hydrodynamical features).

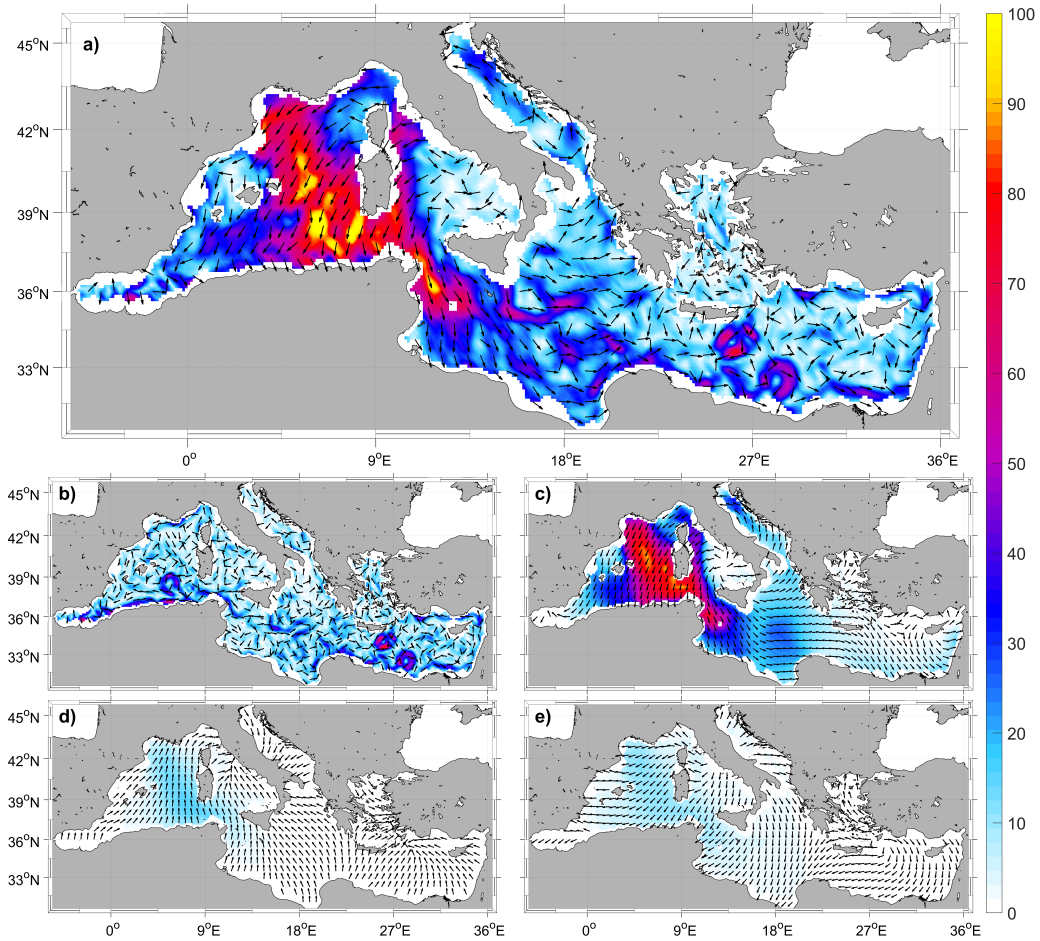


Figure 3. a) Total, b) Geostrophic, c) Ekman, d) Stokes and e) Ekman-Stokes velocity fields for January, 19th of 2005 at 12 : 00 UTC. The magnitudes (module, in cm/s) of each velocity component are displayed as background colors according to the color-scale. The black arrows represent the direction of the velocity fields. Only 1 of each 5 data points have been plotted for clarity.

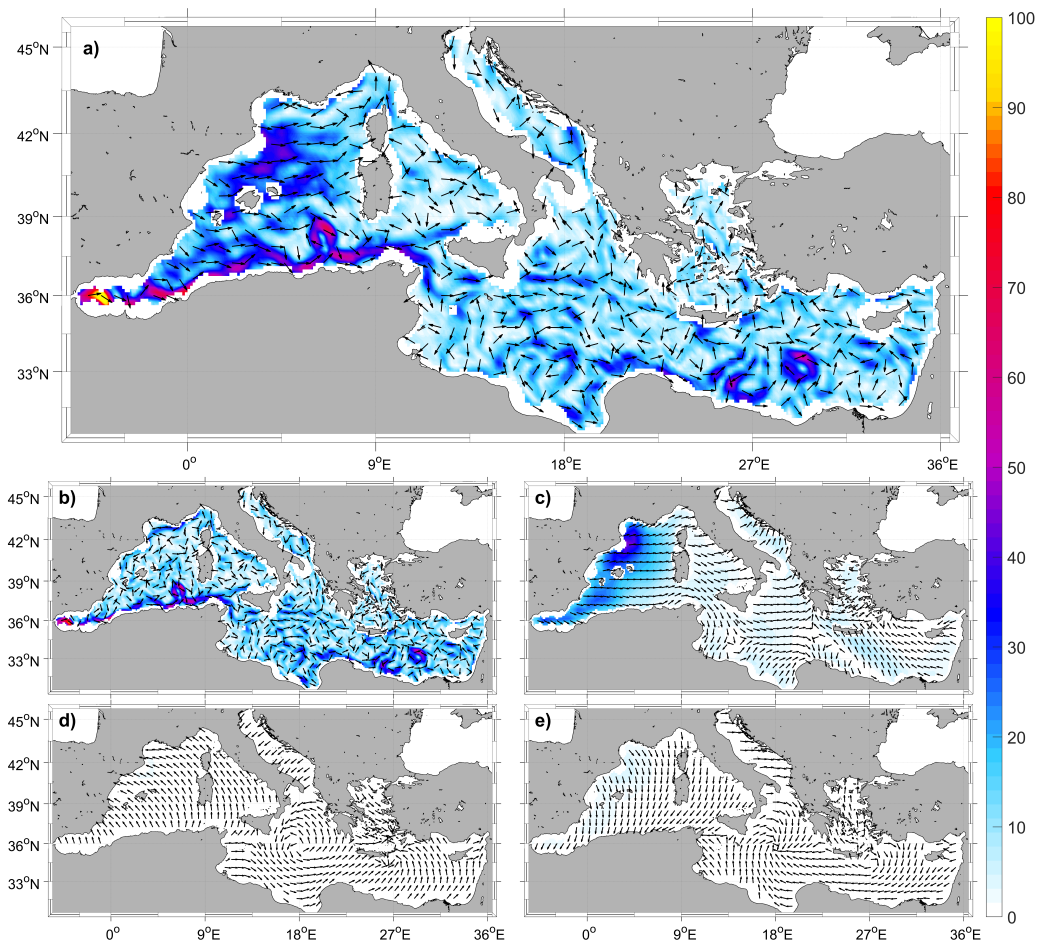


Figure 4. a) Total, b) Geostrophic, c) Ekman, d) Stokes and e) Ekman-Stokes velocity fields for February the 5th of 2014 at 6:00 UTC. The magnitudes (module, in cm/s) of each velocity component are displayed as background colors according to the color-scale. The black arrows represent the direction of the velocity fields. Only 1 of each 5 data points have been plotted for clarity.

410 The relevance of both Ekman and Stokes components on the total current is not
 411 only restricted to the situations where they reach maximum values, as shown by the pre-
 412 vious exemplary case (Fig. 3), since they can also have a noticeable impacts on the dy-
 413 namics with relatively small values. Indeed, the relative differences of direction between
 414 the wind stress and wave propagation on one hand, and the geostrophic component on
 415 the other hand, affect the total surface circulation. Fig. 4 displays an example correspond-
 416 ing to the 5th of February 2014 at 6 : 00 UTC where, even though the geostrophy rep-
 417 represents the main contribution on the total velocity, both Ekman and Stokes compo-
 418 nents suppress the Liguro-Provençal Current (Fig. 5). This suppressor effect of the Ekman
 419 component is not caused by its intensity, ($|\mathbf{U}_E|$ is similar to $|\mathbf{U}_g|$), but because its direction
 420 is opposite to the geostrophic current direction.

421 It is worth noting that \mathbf{U}_{ES} ensures that the total velocity satisfies the wind stress
 422 boundary condition at the sea surface. Thus, it removes the sea surface stress caused by
 423 the Stokes component (\mathbf{U}_S) (Polton et al., 2005; Pearson, 2018). For this reason, \mathbf{U}_S and

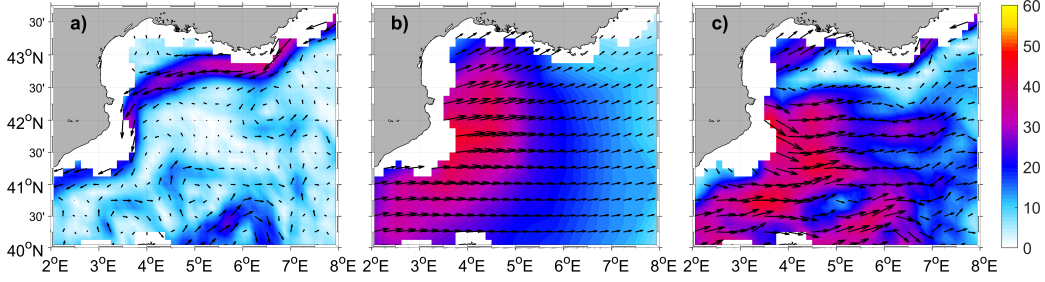


Figure 5. Zoom at the Liguro-Provençal current for a) Geostrophic, b) Ekman and c) Total velocity fields for February the 5th of 2014 at 6:00 UTC. The magnitudes (module, in cm/s) of each velocity component are displayed as background colors according to the color-scale. The black arrows represent the direction of the velocity fields.

424 \mathbf{U}_{ES} usually have opposite direction with the same order of magnitude, with a minor
 425 impacts on the total current. This is particularly appreciable when the Ekman layer is
 426 deeper than the Coriolis-Stokes depth ($\delta_s \ll \delta_e$), i.e. under short wave periods, where
 427 the effect on the current profile resembles the traditional pure Ekman solution (Polton
 428 et al., 2005).

429 These dynamical conditions associated with a large contribution of the wind and
 430 waves induced currents are not isolated cases since these ageostrophic circulation pat-
 431 terns occur frequently over different Mediterranean regions.

432 5.1 Regionalizing the impacts of wind and waves on the total surface 433 kinetic energy

434 To further characterize the regions and time periods for which the total surface dy-
 435 namics are governed by the Ekman and Stokes components, we perform a coupled SOMs
 436 analysis between the absolute value of \mathbf{U}_T , \mathbf{U}_g , \mathbf{U}_E and \mathbf{U}_S . Note that these magnitudes
 437 are closely related to the root-squared Kinetic Energy (henceforth referred as to KE) given
 438 by $KE=(u^2+v^2)^{1/2}$. We first apply the SOM algorithm to the 6-hour velocities for 2005,
 439 since this year presents maximum averaged values for the ageostrophic velocities and the
 440 areas influenced by each velocity component can be more clearly delimited.

441 The different temporal pattern extracted from the SOM analysis using a 2x3 neu-
 442 ral network in the time domain are shown in Fig. 6 for each of the velocity components.
 443 As expected, geostrophy dominates the low frequency variations while the Ekman and
 444 Stokes components modulate the high frequency signal of the total velocity, including
 445 the sub-daily variability (Onink et al., 2019). This high frequency signal shows the highly
 446 variable response of the upper layer dynamics to the rapidly evolving waves and wind
 447 forcing. In general, \mathbf{U}_g is of the same order of magnitude as \mathbf{U}_T , whereas \mathbf{U}_E is about
 448 half (or smaller) of \mathbf{U}_T 's intensity while \mathbf{U}_S is one order of magnitude smaller than \mathbf{U}_T .
 449 As observed in Fig. 6, due to the preservation of the topology, the SOM method organ-
 450 izes the patterns in the neural network according to the similarity in the intensity and
 451 variability of each velocity components. Patterns showing high contribution of geostrophy
 452 are located around the right top corner of the neuronal network (P2 and P3 in Fig.
 453 6), while patterns where the contribution of Ekman and Stokes velocities is large are found
 454 at the left-hand side of the neural network (P1 and P4). And between them, there are
 455 some intermediary patterns (P5 and P6). As revealed by some patterns, the wind and
 456 waves induced currents are more intense during winters, exceeding the value of the geostrophic

457 component in some patterns (i.e. P1, P4 and P5). This suggests a strong seasonal variability in the geostrophic signal which is further analyzed in section 5.2.1.

459 Fig. 7 shows the objective classification of the Mediterranean Sea in sub-regions based on the combined variability of total, geostrophic, Ekman and Stokes velocity components given by the temporal patterns described previously (Fig. 6). The region where the Ekman and Stokes components have the largest values (R1) corresponds to the temporal pattern P1. It identifies the northern and central sub-basins of the western Mediterranean as a region whose surface dynamics is largely affected by the wind and waves induced currents. It is indeed dominated by strong regional winds (i.e. 'mistral' and 'tramontane') blowing southward with the marine origin in the Gulf of Lion (Zecchetto & De Biasio, 2007; Obermann et al., 2018), where waves can be developed through the large fetch (Sayol et al., 2016; Morales-Márquez et al., 2020). This kind of winds although are stronger with longer duration and more frequent in winter, they also take place in summer (Soukissian et al., 2018). Surprisingly, we found in P1 the events with the larger values of U_T with velocities up to 40cm/s during the 19th of January, the 14th of February, the 11th of April and the 17th of December (although not easily appreciable in Fig. 6 since the original temporal pattern has been smoothed). Regarding the eastern and central parts of the basin, the influence of the Ekman and Stokes components is higher in the regions R4, R5 and R6, characterized by patterns P4, P5 and P6 (see green, yellow and orange regions in Fig. 7). These patterns can be associated with local winds such as, etesian and bora (Zecchetto & De Biasio, 2007), that although they do not have enough distance without any obstacle in order to the waves to be developed, they are able to cause a large Ekman velocity. Comparing the amplitude of Ekman and Stokes components over all the regions, we can observe that western basin is the region most impacted by wind and waves of the Mediterranean Sea, since there is a larger fetch.

482 Regions where the dynamics is mainly modulated by the geostrophy (low frequency signal) are characterized by P2 and P3 (Fig. 6) and shown by R2 and R3 in Fig. 7. They identify the well-known geostrophic circulation features in the Mediterranean Sea including the Alboran gyres, Levantine gyres and the detachment of eddies from the Algerian current through baroclinic instability (R2). Indeed, the Algerian current, which flows along the northern African shelf and then crosses the Strait of Sicily towards the southern Ionian Sea, is clearly identified by R3. It is also remarkable how the main Mediterranean gyres are well characterized within the same region (R2), showing a similar variability in the total kinetic energy of this geostrophic features between the western and eastern basins. These temporal patterns also identify the Liguro-Provençal current that is interrupted in the Gulf of Lions due to the effect of the Ekman and Stokes components (R1). Pattern P3 shows an increase of U_T during August and September, likely due to the importance of the geostrophic component (in contrast to the weakening of wind and waves). P4 characterizes the regions R4 (green areas in Fig. 7) associated with lower total kinetic energy (small values of U_T) and where the Ekman component is relative large, dominating the total velocity during winter season. This pattern identifies broad areas across the western and central parts of the Mediterranean Sea (Thyrrhenian, Adriatic, northern Ionian, Gulf of Gabes and Ebro shelf), as well as small regions around Cyprus (eastern basin). P5 and P6 are exclusive for the central and eastern Mediterranean, respectively; exhibiting intermediate values of Ekman and Stokes velocities, being higher the contribution of the geostrophy and the total kinetic energy in the eastern region (R6). It is worth mentioning that, the characteristic map of regions shown in Fig. 7 is in agreement with the main features of the surface dynamics in the Mediterranean Sea outlined in Millot (2005).

506 5.2 Regional assessment of the temporal variability

507 In this Section, we extend the analysis to the 25 years of data to assess the role of wind and waves at the inter-annual scale. Each velocity component is spatially averaged

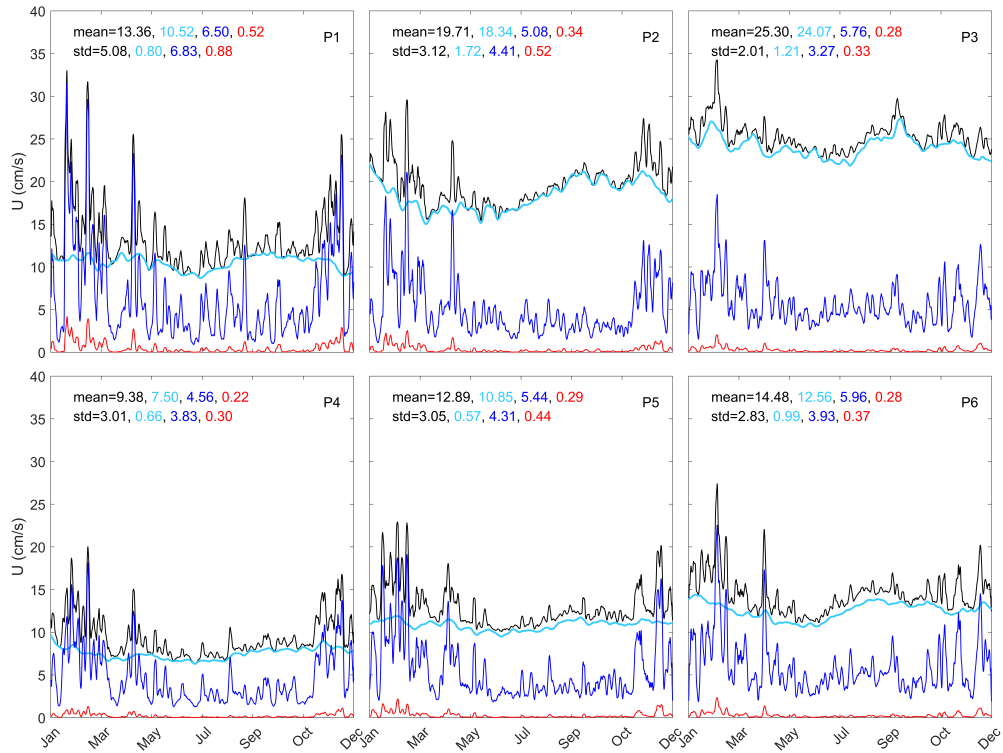


Figure 6. Temporal patterns of the absolute value of the total (black line), geostrophic (cyan line), Ekman (blue line) and Stokes (red line) velocity component fields extracted from the coupled SOMs technique for 2005. Patterns have been smoothed using a moving window of 3.5 days in order to facilitate comparison. The means and the standard deviations of each temporal pattern reported within each panel.

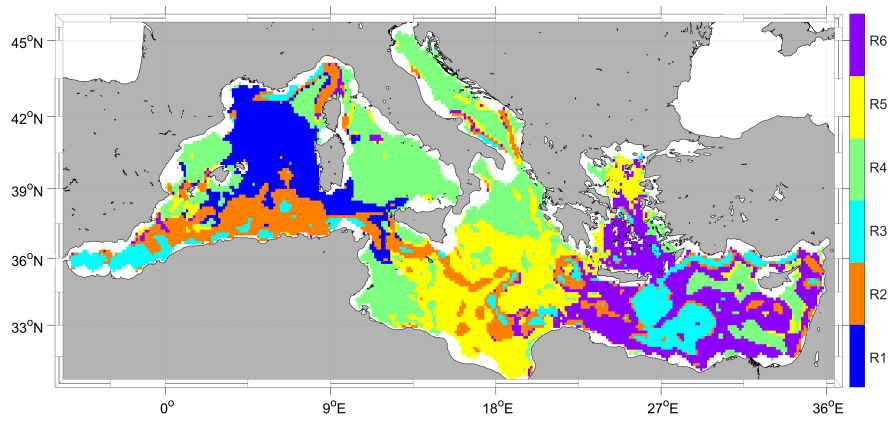


Figure 7. Regions unveiled from the SOM analysis according to the coupled variability of the absolute value of each velocity field component. R1 is dominated by the ageostrophic component; R2/R3, by the geostrophic one and R4/R5/R6 are intermediate patterns.

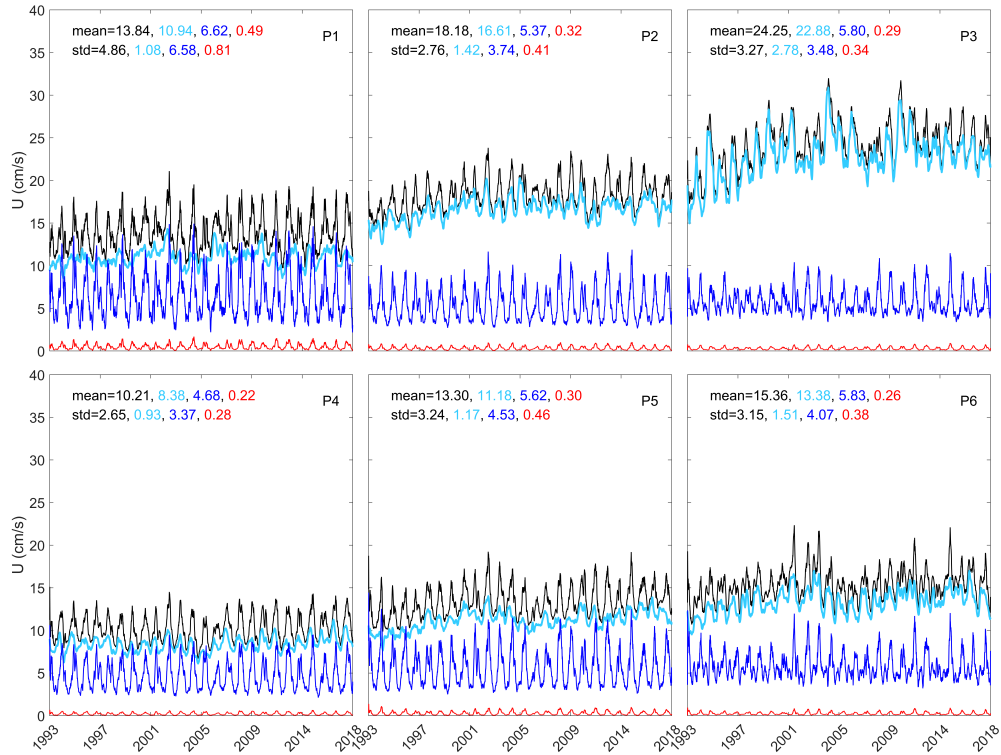


Figure 8. Mean total (black line), geostrophic (cyan line), Ekman (blue line) and Stokes (red line) velocity component module fields from 1993 to 2018 in the regions of the temporal SOMs of 2005.

509 every 6-h from 1993 to 2018 over each region identified by the previous SOM analysis
 510 (Fig. 7) to obtain the time-series reported in Fig. 8. The time series have been smoothed
 511 with a moving window of 45 days to improve readability. The different components exhibit
 512 similar variability than previously analysed for 2005, with geostrophy clearly dominating
 513 in patterns P2 and P3, and with the wind and wave induced velocities being prominent in
 514 pattern P1. The geostrophic component appears as the main contributor describing the large
 515 scale variability while Ekman and Stokes components incorporate the high frequency and a
 516 clear seasonal signal to the total velocity. Despite the fact that the values of Ekman and
 517 Stokes velocities are high during short time periods, they impact significantly on the total
 518 kinetic energy throughout the entire period analyzed. As seen in pattern P1, the Ekman
 519 component surpasses the geostrophy during winter. A similar situation occurs in R4, where the
 520 P4 presents smaller total kinetic energy with a large impact of the Ekman component in
 521 winter. In the central (R5) and eastern (R6) regions, the geostrophic velocities are larger
 522 than the Ekman and Stokes components except for a few occasional events, when the two latter
 523 are higher than the former. In general, the contribution of the Ekman component to the
 524 total velocity is larger in the central part (P5) than in the eastern one (P6). The effect
 525 of Ekman and Stokes components at the eastern part, P6, is particularly significant during
 526 2002, 2012 and 2015 winters (see Fig. 8).
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5.2.1 Short-term variability: annual and semiannual cycles

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An assessment of the temporal variability (i.e. dominant frequency bands as a function of time) of the different velocity components in each of the SOM regions identified in Fig. 7 is here performed applying a wavelet analysis to their corresponding temporal patterns (Fig. 9). All the regions show a strong seasonal signal (1 year characteristic period) for all the velocities except for the geostrophic component in R1. This strong intra-annual variability is mainly fueled by the ageostrophic components. While in regions R1, R2, R5 and R6 the annual geostrophic signal is interrupted, the Ekman and Stokes components contribute largely to the short term variability (annual cycle) of the total kinetic energy during these 25 years, as indicated by the marked seasonality of Ekman and Stokes components for the entire Mediterranean Sea (Fig. 8). It should be noted that \mathbf{U}_T also exhibits an important semi-annual cycle in R6, and in R3 to a lesser extent, during almost all 25 years except 1999. This characteristic period is also present in geostrophy but more discontinued than in the ageostrophic velocity. Note that the semi-annual signal in the geostrophic current in R1 from 2000 to 2008 is removed in the total velocity.

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5.2.2 Long-term variability: relation with climatic modes of variability

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Long-term oscillations are found in the total velocity with characteristic periods of around 2, 3 and 5 - 6 years over the whole basin. The long term variability on the total velocity field is modulated by the geostrophic component in all the regions. However, Ekman and Stokes components increase the spectrum power of these characteristic periods in some regions. In the western Mediterranean (R1), additional significant periods are identified around 2 and 3 years from 2010 to 2017, from 1999 to 2006 and from 2008 to 2013, respectively, as a result from the combination of the geostrophic and ageostrophic variability. As already suggested in Fig. 8, the Ekman component dominates the variability in this region during the 25 years period. In R2, there are significant signals with periods of 1.5 - 2 years and 2 - 4 years over 2013 - 2018 and 2001 - 2015, respectively, also due to the combined influences of the different velocity components. On the other hand, \mathbf{U}_g in R3 is practically the main contributor to the 1.5 - 2 years and 4 - 6 years cycles in the total velocity. Therefore \mathbf{U}_a has poor relevance in explaining the long-term variability in this region. Regions R4 and R5 present a 4 - 6 year-period well defined and a 1.5 years period in some specific years (see Fig. 9, R4 and R5). In R5, the annual signal is intermittent in \mathbf{U}_g being present during the 25 years in \mathbf{U}_a . Finally, \mathbf{U}_T in R6 registers cycles of 1 - 2.5 years and 1.5 years during 1997 - 2003 and 2013 - 2018, respectively. Periods ranging 5 to 6 years coincide with the characteristic periods of the dominant climatic patterns of variability acting over the Mediterranean Sea (Morales-Márquez et al., 2020).

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In order to get insights about the regional influence of the modes of atmospheric variability on the upper layer dynamics in the Mediterranean Sea, we perform a wavelet coherence analysis between the NAO, EA, EA/WR and SCAND indices and the \mathbf{U}_T in the dynamical regions previously identified (Fig. 10). This method allows identifying the frequency bands within which time series of KE for each SOMs region and the large scale atmospheric forcings co-vary.

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NAO is correlated with the total velocity with signals of around 1 year during 2014 to 2018 in all the SOM regions (see Fig. 10, NAO). For periods spanning 5 - 7 years the total velocity signal is anticorrelated with the NAO in all regions except R1 where the negative correlation is around 2.5 years. Note that R1 corresponds to the region where wind and waves are most relevant for the modulation of the high frequency variability of the total currents. This is in agreement with the results obtained by Morales-Márquez et al. (2020) where a strongly significant anticorrelation between extreme waves and the

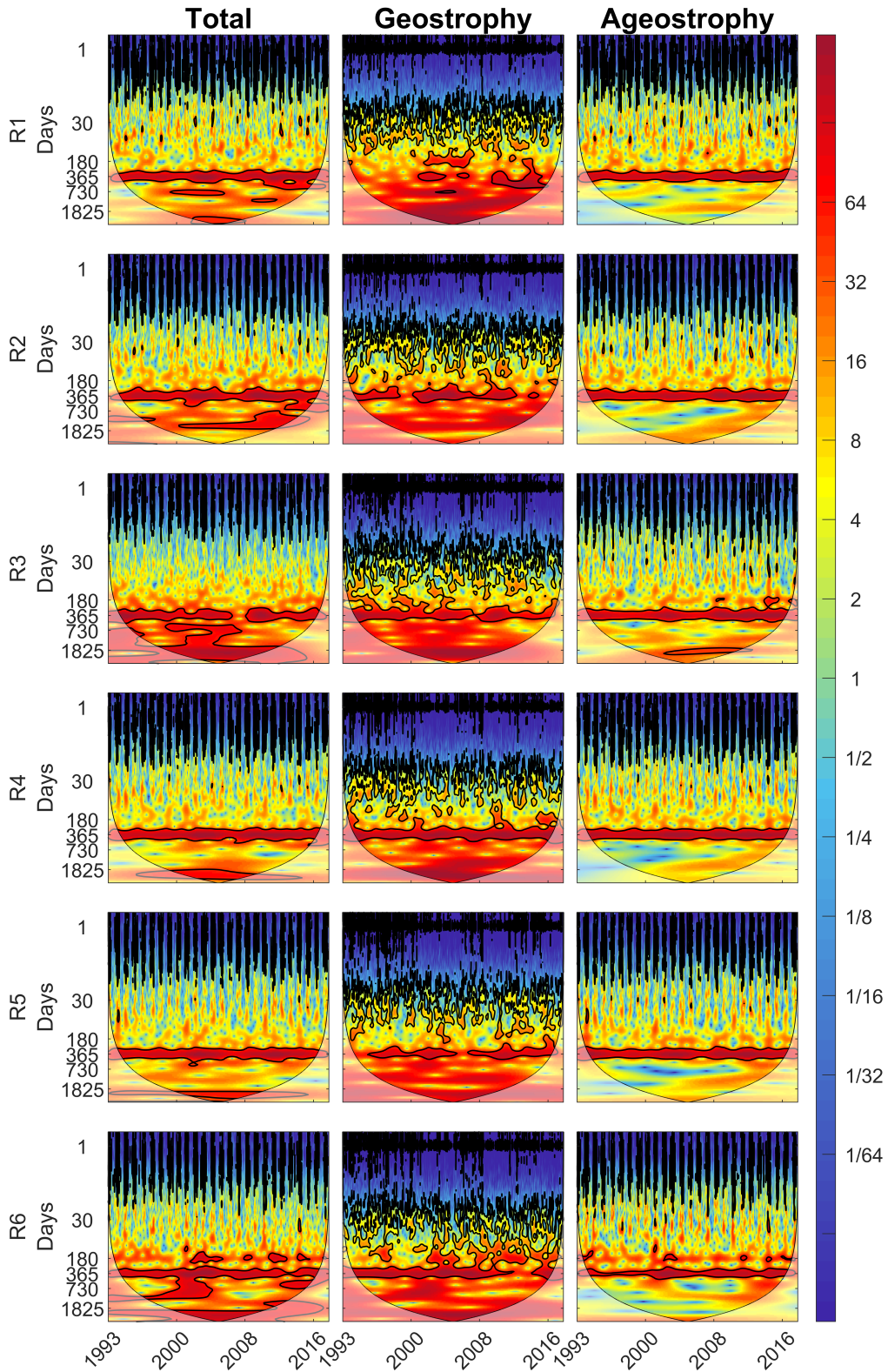


Figure 9. Wavelet power spectrum of the 6-hour time series of the spatially-averaged (over the SOM regions shown in Fig. 8) Total, Geostrophic and Ageostrophic velocity components from 1993 to 2018. Contours in black indicates the 95% significant levels. Lighter shades show the cone of influence (COI) where the edge effects may distort the Fourier analysis.

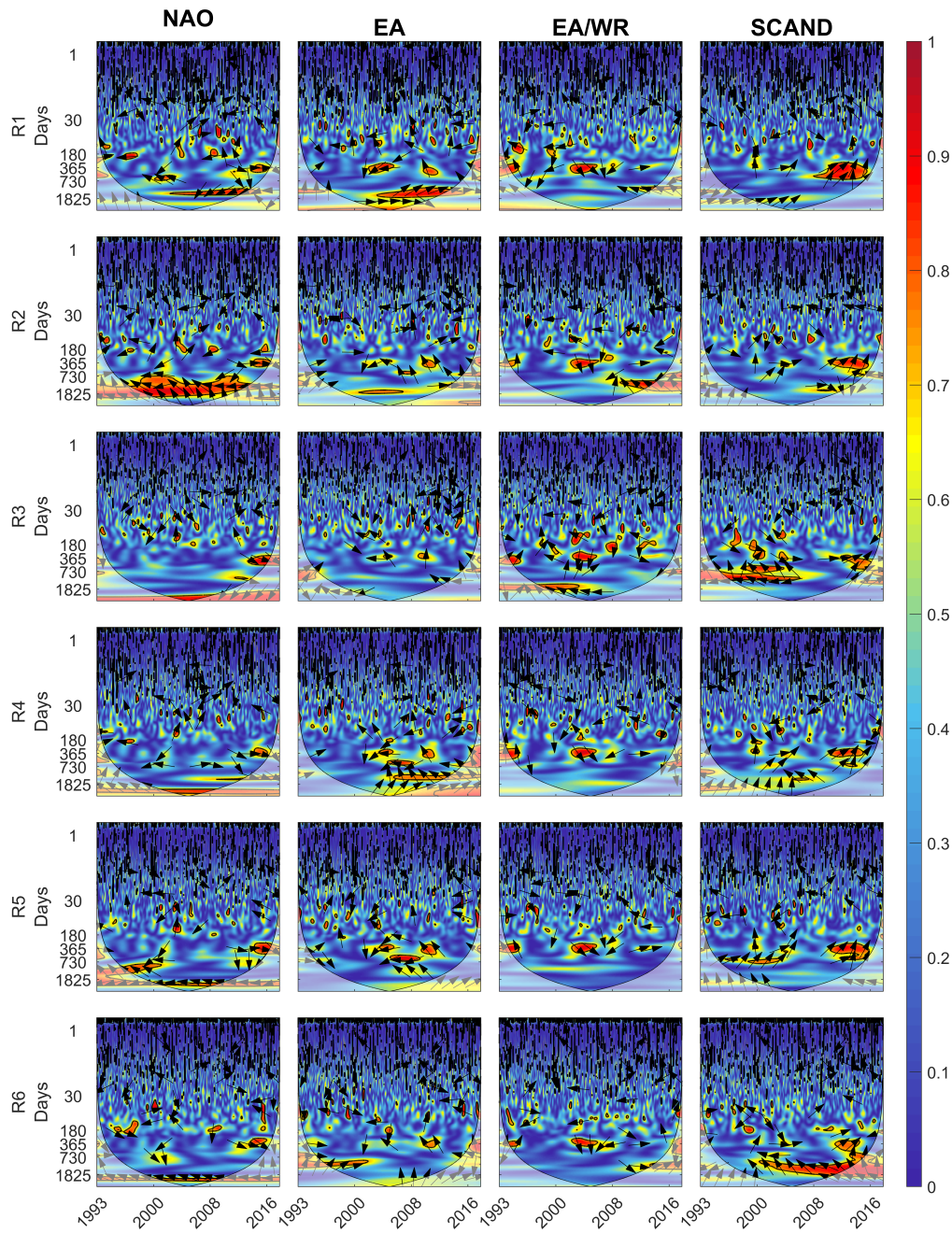


Figure 10. Wavelet coherence between the mean total velocity module for temporal 2005 SOMs patterns and the monthly values of NAO, EA, EA/WR and SCAND climatic indices from 1993 to 2018. The arrows determine the phase between both series. Arrows pointing to the right represent positive correlation (signals in phase) and when they point to the left, anti-correlation (signals in anti-phase). Contours indicate wavelet squared coherence.

579 NAO was obtained in the Mediterranean Sea. NAO has a significant influence in R2 with
 580 a negative correlation at 2- 5 years during the period of analysis, and in R5 with neg-
 581 ative correlation around 2 - 4 years from 1993 to 2002. In addition, NAO has an effect
 582 on the semiannual variability in all regions during 1996, 2003 and 2008 being less vis-
 583 ible in R3 and R5 (see Fig. 10, NAO).

584 The influence of EA on the variability of the total current in R1 and R4 is asso-
 585 ciated with 1.5 and 4 - 5 years signals after 2000 (see Fig 10, EA). In R1, correlation oc-
 586 curs between 2003 to 2016 with a 4 year-period and for the 25 years period around 7 years
 587 (see Fig. 10, R1, EA). Similar but less intense atmospheric influence is found in R5 and
 588 R6. In all regions a strong anticorrelation is shown around 1 year from 2002 to 2005 and
 589 from 2016 to 2018. EA also affects R4 and R5 with a 2 years signal between 2009 to 2012.
 590 Note that EA does not affect R3, that is where mesoscale surface dynamics is mainly con-
 591 trolled by the geostrophic component. R6 shows a positive correlation with the EA 3 year-
 592 period from 1993 to 2002.

593 The signature of the 1 year signal, associated with EA/WR, is clearly seen in all
 594 the Mediterranean surface dynamics between 2003 to 2005. The western Mediterranean
 595 (R1 and R2) shows an anticorrelation with EA/WR signals at 4 year-period from 2010
 596 to 2018, and around 5-year period in R3 from 1993 to 2007, in agreement with the re-
 597 lationships documented for extreme waves by Morales-Márquez et al. (2020).

598 The influence of SCAND index on total surface currents manifests itself with a pos-
 599 itive correlation at 1 - 2-year period for the whole basin after 2006. The impact of SCAND
 600 climate mode is more intense in Eastern Mediterranean as shown by the negative/positive
 601 correlation in the 1.5 - 3 years band between 1993 and 2006 in R3/R5 and by the strong
 602 negative correlation around 3 - 5 year-period during 2000 - 2018 in R6.

603 *5.2.3 Trends in the Kinetic Energy*

604 To analyze linear trends in the geostrophic and total velocity modules, the resid-
 605 ual of \mathbf{U}_T and \mathbf{U}_g are fitted by a linear regression in time at each spatial point (see Fig.
 606 11). We have verified these results with the Theil Sean estimator technique, obtaining
 607 similar trend patterns (not shown). The significance level is set at 90% with the Mann
 608 Kendall method and with a t -value adjusted of $N-2$ degrees of freedom (Pastor et al.,
 609 2018), being the number of points with no-significant values slightly larger using the Mann
 610 Kendall method. The estimated global Mediterranean trend of total speed is positive with
 611 a value of $0.058 \pm 1.43 \cdot 10^{-5}$ cm/s per year, being the geostrophic one higher with a value
 612 of $0.063 \pm 1.20 \cdot 10^{-5}$ cm/s per year (see Fig. 11). It suggests that surface velocities, and
 613 associated KE, are increasing over this 25 years period. While regions where the wind-
 614 and wave-induced velocities have the largest impacts (R1 and R4) do not exhibit clear
 615 and significant trends in the total velocity module (not shown), the geostrophic dom-
 616 inated regions (R3 and R2) show positive trends with a shift in 2003 ($0.59 \pm 8.15 \cdot 10^{-5}$
 617 and $0.37 \pm 6.56 \cdot 10^{-5}$ cm/s per year), see Fig. 12. These results are consistent with the
 618 KE increase presented in Ser-Giacomi et al. (2020), they explain this rise as a potential
 619 relation to an increment of a baroclinic instability since they show a decrease of the wind
 620 stress across the most of the western basin. While such mechanism could also explain
 621 the rising trend evidenced here, further analyses are needed to ascertain which mecha-
 622 nism is at play. Note however that the clear positive trend from 1993 to 2002 seems to
 623 slow down after 2003. It could indicate that this is not a proper trend but rather part
 624 of a longer oscillation or an artifact due to the inconsistency in the SLA dataset of 25
 625 years. However the altimeter product used in this study (see section 3) is the result of
 626 homogenization procedure among several altimeter satellite observations and is thus con-
 627 sidered suitable for trend analysis (Pujol et al., 2016). \mathbf{U}_T and \mathbf{U}_g present similar trends
 628 during the 25 years analyzed (see Fig. 11, a and b) with an increment in the eastern Mediter-
 629 ranean Sea and a decrease in the western basin. The global trend is generally positive

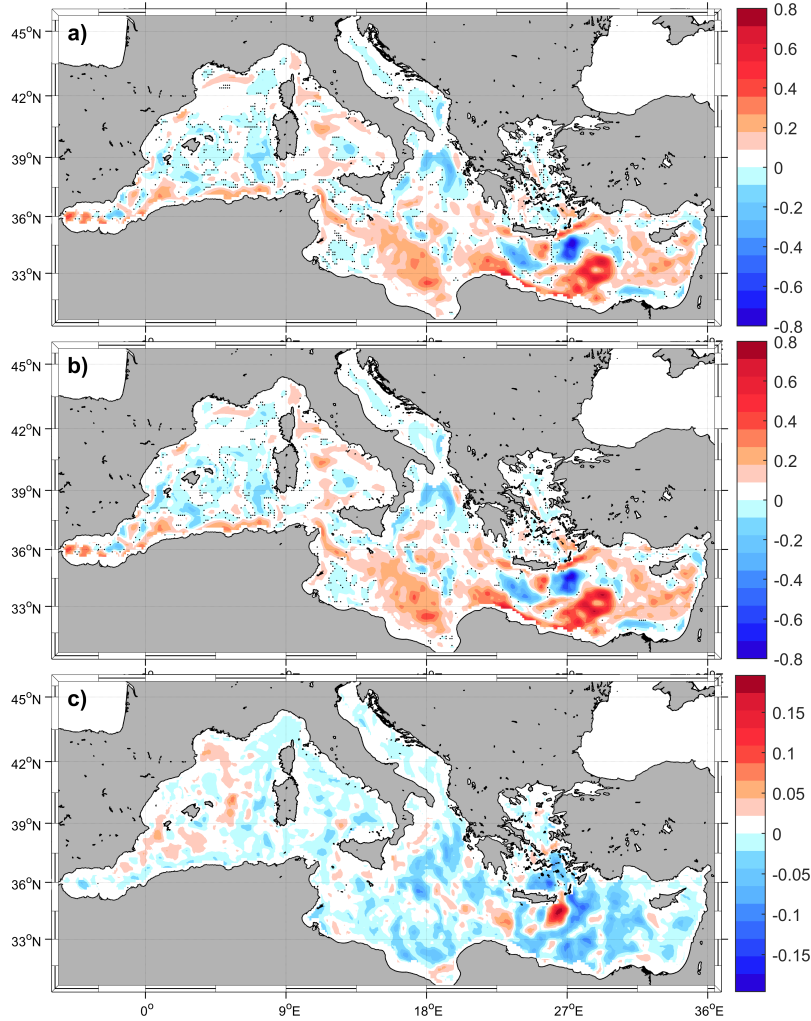


Figure 11. Trend of the a) Total velocity module and b) Geostrophic component module from 1993 to 2018 in cm/s per year. c) Difference between a) and b). No significant values at the 90% confidence interval are dotted (with Mann Kendall method).

630 in regions where the geostrophy is dominant, except in the Libyan Sea where both \mathbf{U}_T
 631 and \mathbf{U}_g tendencies are negative, in good agreement with Fig. 8. The maximum trend
 632 of $0.72 \pm 2.44 \cdot 10^{-5}$ cm/s per year for \mathbf{U}_T is found in the eastern part of the Mediterranean
 633 basin. In contrast, the minimum value in the Libyan Sea is $-0.81 \pm 2.42 \cdot 10^{-5}$ cm/s per
 634 year (Fig.11, a). The maximum and minimum trends for \mathbf{U}_g are found in the same re-
 635 gions with slightly smaller values, $0.73 \pm 2.20 \cdot 10^{-5}$ and $-0.77 \pm 2.85 \cdot 10^{-5}$ cm/s per year,
 636 (Fig.11, b). The ageostrophic input on the trend of the total velocity module is evalu-
 637 ated through the difference between both tendencies, \mathbf{U}_T and \mathbf{U}_g . Most values are close
 638 to zero in the whole Mediterranean (see Fig.11, c), except in the region with the min-
 639 imum trend of \mathbf{U}_T where the difference of trends is $\sim 0.2 \pm 9.89 \cdot 10^{-7}$ cm/s per year.
 640 In the western region, there are some areas with a small positive differences of trend of
 641 $0.05 \pm 6.33 \cdot 10^{-7}$ cm/s per year, corresponding to R1 of Fig. 7 and also to the regions
 642 of the main regional winds.

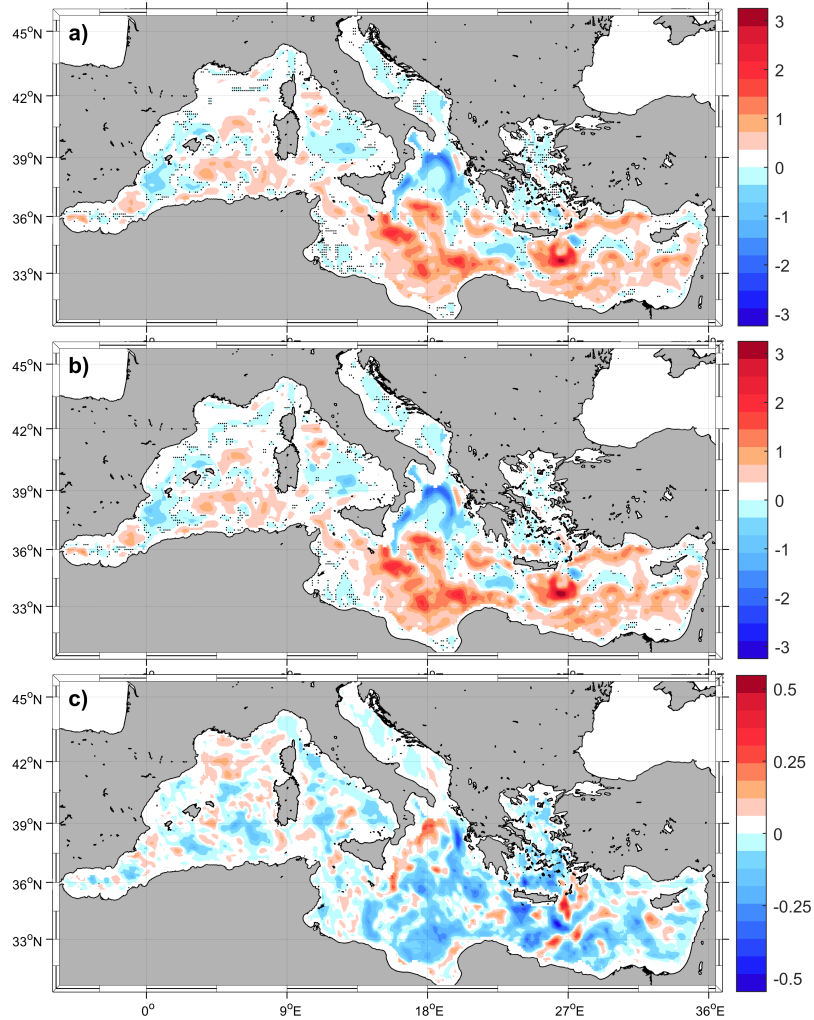


Figure 12. Trend of the a) Total velocity module and b) Geostrophic component module from 1993 to 2002 in cm/s per year. c) Difference between a) and b). No significant values at the 90% confidence interval are dotted (with Mann Kendall method).

643 Positive global trends of other oceanic variables have also been observed for the Mediter-
644 ranean. (Pujol & Larnicol, 2005) reported a trend in the root squared Eddy Kinetic En-
645 ergy of 0.7 cm/s per year, between 1993 and 2003, and (Pastor et al., 2018) showed a
646 linear trend for Sea Surface Temperature from 1982 to 2016 of 0.03 ± 0.003 °C per year.

647 6 Conclusions

648 This study analyzes the effect of Ekman and Stokes velocities on the total kinetic
649 energy in the upper layer of Mediterranean Sea. By solving the momentum equation (Eq.
650 4) we include the interaction between Ekman and Stokes drift on the geostrophic veloc-
651 ity. Total velocity is decomposed into different components: the geostrophic, Ekman, Stokes
652 and the interaction between Ekman and Stokes. The regional relevance of these differ-
653 ent components is evaluated through SOM decomposition, and their variability through
654 wavelet analysis.

655 Once the velocity components are obtained, a dynamical regionalization of the Mediter-
656 ranean Sea has been performed based on the local impacts of waves and wind on the to-
657 tal velocity variability. Ekman currents account for the short-term variability (seasonal,
658 semi-seasonal and smaller time scales) of the surface circulation, especially during win-
659 ter when the Ekman component occasionally exceeds geostrophy due to strong regional
660 winds. The regionalization shows that the effects of Ekman and Stokes are more marked
661 in the western than in the eastern Mediterranean basin. This is the result of the larger
662 fetch in the western basin, allowing the development of larger swells (Mao & Heron, 2008).
663 Regionalization of velocity components identifies two regions (associated with the main
664 Mediterranean gyres and the Algerian current) where the geostrophy modulates the to-
665 tal kinetic energy variability. These regions are characterized by a positive trend of the
666 module velocity of $0.14 \pm 2.15 \cdot 10^{-5}$ cm/s per year during the 25 years, with stronger in-
667 crements during 1993–2002. The dominant periods of the total currents in the entire
668 Mediterranean Sea, essentially dominated by geostrophy, are 1 and 5 - 6 years. In regions
669 where the inclusion of both Ekman and Stokes velocities returns a significantly differ-
670 ent flow field than the one obtained by geostrophic approximation, intermediate peri-
671 odicity values between 1 and 5 years are found. These signals of variability are related
672 with the principal climatic modes typical of the Mediterranean basin: the NAO, EA EA/WR
673 and SCAND patterns. NAO dominates, with a negative correlation, the large-scale, around
674 5 - 7 years in the whole basin except in the western Mediterranean, which was already
675 noticed by Morales-Márquez et al. (2020). Furthermore, NAO is correlated with the an-
676 nual variability during 2014-2018 and with the semiannual variability at the whole basin,
677 although these connections are weaker for geostrophy-dominated region. The EA index
678 has a positive large-scale correlation in the Mediterranean Sea (4 - 7 years), with the ex-
679 ception of the geostrophic modulated region. The long-term variability effect of EA/WR
680 on the currents velocity is negative and between periods of 4 to 5 years, in particular in
681 the Western Mediterranean. Finally, the SCAND mode of variability has a negative ef-
682 fect in periods of 3 -5 years in the eastern basin.

683 The methodology presented in this work can be used to better understand the phys-
684 ical, biological and chemical processes occurring at the upper layers of any ocean region
685 using only observations with a low computational cost. In the Mediterranean Sea, there
686 is a need for an improved wave model accounting for the feedback between the Ekman
687 currents on the wave field evolution, and how the Ekman layer evolves due to the non-
688 linear evolution of the wave field, which necessarily contemplates a non-steady Ekman
689 current, as shown in Shrira and Almelah (2020). Next step is devoted to extend this anal-
690 ysis to study transport properties from the Lagrangian point of view. Thus, several ap-
691 plications (e.g. floating debris, oil spill, Search and Rescue, jellyfish tracking, etc.) could
692 benefit from this approach to obtain reliable nowcast.

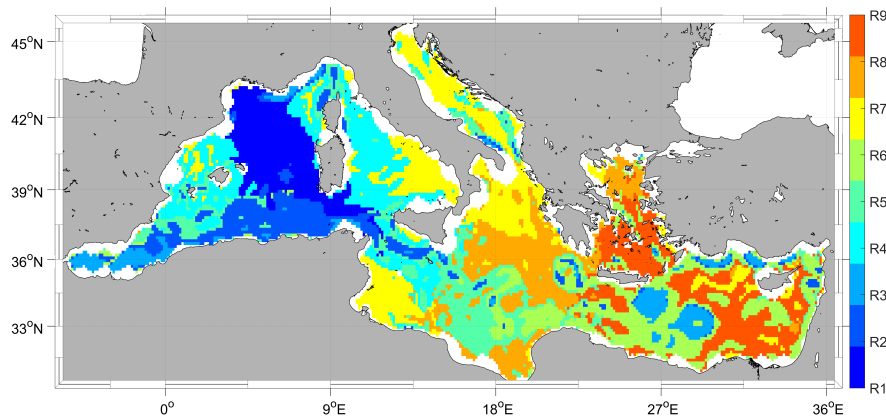


Figure A1. Temporal patterns of the absolute value of the total (black line), geostrophic (cyan line), Ekman (blue line) and Stokes (red line) velocity component fields extracted from the coupled SOMs technique for 2005 with 9 neurons.

Appendix A Supplementary material

Acknowledgments

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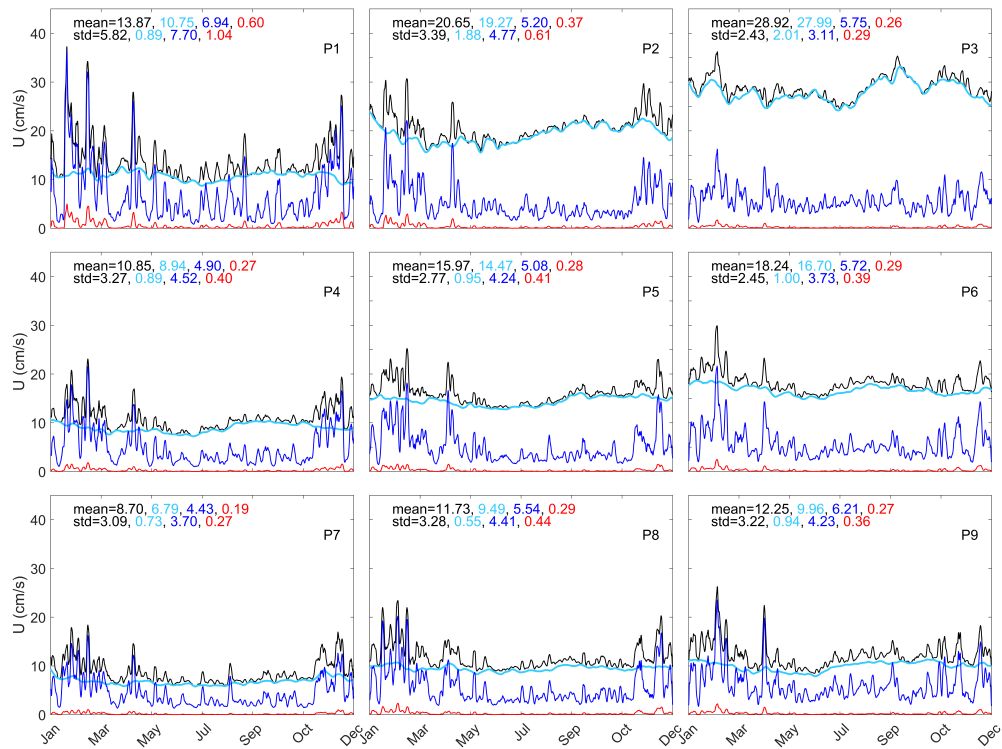


Figure A2. Regions unveiled from the SOM analysis, with 9 neurons, according to the coupled variability of the absolute value of each velocity field component.

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