

# Causes of the regional variability in observed sea level, sea surface temperature and ocean colour over the period 1993-2011

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## Abstract

We analyse the regional variability in observed sea surface height (SSH), sea surface temperature (SST) and ocean colour (OC) from the ESA Climate Change Initiative (CCI) datasets over the period 1993-2011. The analysis focuses on the signature of the ocean large-scale climate fluctuations driven by the atmospheric forcing and do not address the mesoscale variability. We use the ECCO version 4 ocean reanalysis to unravel the role of ocean transport and surface buoyancy fluxes in the observed SSH, SST and OC variability. We show that the SSH regional variability is dominated by the steric effect (except at high latitude) and is mainly shaped by ocean heat transport divergences with some contributions from the surface heat fluxes forcing that can be significant regionally (confirming earlier results). This is in contrast with the SST regional variability, which is the result of the compensation of surface heat fluxes by ocean heat transport in the mixed layer and arises from small departures around this background balance. Bringing together the results of SSH and SST analyses, we show that SSH and SST bear some common variability. This is because both SSH and SST variability show significant contributions from the surface heat fluxes forcing. It is evidenced by the high correlation between SST and buoyancy forced SSH almost everywhere in the ocean except at high latitude. OC, which is determined by phytoplankton biomass, is governed by the availability of light and nutrients that essentially depend on climate fluctuations. For this reason OC show significant correlation with SST and SSH. We show that the correlation with SST display the same pattern as the correlation with SSH with a negative correlation in the tropics and subtropics and a positive correlation at high latitude. We discuss the reasons for this pattern.

## 1. Introduction

Oceans have been routinely monitored from space for more than 30 years now. In 1978, NASA launched Seasat, TIROS-N and Nimbus-7, the first three satellites dedicated to ocean observations. Seasat carried five complementary sensors and provided the first estimates from space of Sea Surface Height (SSH), surface wind stress, Sea Surface Temperature (SST), surface wave field and polar ice extent. Unfortunately these estimates covered only a short period because Seasat failed after 105 days in space. TIROS-N lasted more than 2 years in orbit and produced the first really useful maps of SST. The Nimbus-7 carried the Coastal Zone Color Scanner (CZCS), which remained operational for seven and a half years, from October 1978 until June 1986. CZCS was the first satellite sensor specifically developed to study ocean colour properties. Since then, about a dozen ocean-observing satellite missions have been launched by several space agencies to monitor continuously more than 10 essential ocean variables including SSH, ocean currents, tides, wave height (with radar altimeters),

54 wind speed and direction (with microwave scatterometers), SST (with infra-red and microwave  
55 radiometers), sea surface salinity (with microwave imaging radiometers with aperture synthesis),  
56 ocean colour (with multispectral imagers) and ocean bottom pressure (with space gravimetry).

57  
58 In >30 years of measurements, satellite missions have revolutionised our understanding of the oceans.  
59 By providing a global mapping repeated with high temporal resolution, they have revealed the intense  
60 spatio-temporal variability, which characterizes the cycles of the oceans. This new picture challenged  
61 earlier views based on previous sparse measurements collected from ships and buoys. It spurred  
62 oceanographers to make considerable progress in the understanding of the role of the ocean in the  
63 physical (e.g. climate and weather), chemical (e.g. carbon cycle), and biological (e.g. primary  
64 production) processes of the Earth.

65  
66 In this paper we are interested in the role played by the ocean in the climate system. We revisit the 30-  
67 year-long satellite record of ocean observations and summarize what we have learned from it about the  
68 ocean variability at climatic time scales (i.e. interannual to multidecadal time scales). We take  
69 advantage of the satellite archive, which gives an almost global view of the oceans, to explore the  
70 regional ocean variability.

71  
72 Among all ocean variables remotely sensed from space, SSH, SST and Ocean Colour (OC) have long  
73 enough continuous records (~20 years or longer) to address the interannual to multidecadal variability  
74 in the ocean. For this reason we focus on these three variables. We also analyse the hydrographic data  
75 (temperature and salinity) obtained from in-situ measurements because these data are covering a large  
76 part of the ocean since the 1970s (Rhein et al. 2013, Abraham et al. 2013) and provide highly valuable  
77 information to interpret the signal showed by the other variables. Our objectives are to 1) describe the  
78 dominant spatio-temporal patterns of nonseasonal SSH, SST and OC variability, 2) discuss the current  
79 state of knowledge regarding the physical mechanisms responsible for these patterns and 3) explain  
80 the covariance at climatic time scales between these spatio-temporal patterns. The paper is outlined as  
81 follows: section 2 recalls the physical background governing the SSH, SST and OC variability and  
82 explains how this variability is linked to the atmospheric forcing and the ocean circulation. Section 3  
83 shows the spatio-temporal patterns of SSH, SST and OC obtained from the satellite archive and uses  
84 the outputs of an ocean reanalysis to unravel the role of the atmospheric forcing and the ocean  
85 circulation in these patterns. In section 3, we also take the opportunity to show how the patterns in  
86 SSH, SST and OC relate to each other and we discuss to which extent the information in SSH, SST  
87 and OC provided by satellites enables to monitor the mechanisms responsible for the ocean variability  
88 at climatic time scales. Section 4 summarizes the paper, reviews the future issues and concludes.

## 94 **2. Physical background**

### 95 **2.1. SSH variations deduced from the pressure budget of the water column**

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98 SSHs indicate the level of the top of ocean water columns relative to a defined reference, which is  
99 constant with time; we use in general the mean sea surface as a reference. SSH variations indicate  
100 variations in the volume of water columns and are governed by two processes: the changes in mass  
101 and the changes in density of those water columns. Both processes can be tracked through vertical  
102 pressure changes in the pressure budget. Mathematically the pressure budget under the hydrostatic and  
103 Boussinesq approximations can be written as follows (Gill and Niller, 1973):

$$104 \quad \eta = \frac{1}{\rho_0 g} (p_b - p_a) + \eta_{st} \quad (1)$$

105  
106

107 where  $\eta$  is the SSH,  $p_a$  is the sea surface atmospheric pressure,  $p_b$  is the ocean bottom pressure,  $\rho_0$  is  
 108 a constant density,  $g$  is the acceleration due to gravity and  $\eta_{st}$  is the steric sea level.  $\eta_{st}$  is given by  
 109  $\eta_{st} = \frac{1}{\rho_0} \int_{-H}^0 \rho dz$  in which  $H$  is the ocean bottom depth and  $\rho$  the water density deviation from  $\rho_0$ .

110  
 111 The pressure budget breaks down SSH in 3 signals: 1) a signal which depends on the atmospheric  
 112 pressure (i.e. the inverted barometer response, Wunsch and Stammer 1997), 2) a signal which depends  
 113 on bottom pressure and gravity, and 3) a signal which depends on the water density along the water  
 114 column. The first signal is isostatic (i.e. dynamically irrelevant) on climate time scales and thus will  
 115 not be analysed here. The second signal represents the mass component in sea level. It arises  
 116 essentially from changes in bottom pressure caused by the water mass redistribution by the ocean  
 117 circulation in response to the atmospheric forcing. This signal is the dominant component in sea level  
 118 changes at high frequencies (for periods  $<1$  months, Forget and Ponte 2015). It is much smaller at  
 119 interannual and higher time scales but it remains sizeable in particular at high latitudes (see section 3).  
 120 The mass signal in sea level can arise also from changes in the gravity field of the Earth (i.e. changes  
 121 in local  $g$ ). These changes can be due to on-going ice loss in ice sheets for example or to the current  
 122 solid earth response to ice loss in ice sheets that occurred during the last deglaciation (Tamiseia 2011).  
 123 However, this component in sea level change is significant only at time scales longer than  
 124 multidecadal time scales and will not be considered here. The third sea level signal in the pressure  
 125 budget represents the steric component in sea level. This signal arises from changes in the sea-water  
 126 density due to changes in the water column temperature (thermosteric sea level) or salinity (halosteric  
 127 sea level). These changes in temperature and salinity are caused by buoyancy forcing at the surface or  
 128 redistribution of heat and freshwater within the ocean by the ocean circulation in response to the  
 129 atmospheric forcing. The redistribution of heat and freshwater can occur through advection or  
 130 diffusion (including diapycnal, isopycnal and convective mixing) such that steric sea level can be  
 131 written as follows:

$$132 \eta_{st} = \eta_T + \eta_S = \eta_T^a + \eta_T^d + \eta_T^{BF} + \eta_S^a + \eta_S^d + \eta_S^{BF} \quad (2)$$

133 where subscripts  $T$  and  $S$  refer to thermosteric and halosteric sea level and  $a$ ,  $d$  and  $BF$  refer to  
 134 advection, diffusion and buoyancy forcing (a more detailed explanation of the  $\eta_{st}$  budget can be found  
 135 in Piecuch and Ponte 2011). Thermosteric effects ( $\eta_T$ ) dominate steric sea level variability over most  
 136 of the ocean (see section 3 and Köhl 2014, Forget and Ponte 2015). At interannual and longer time  
 137 scales, heat advection in the oceans ( $\eta_T^a$ ) plays the leading role in thermosteric sea level variability  
 138 nearly everywhere, suggesting that the heat redistribution in the ocean is nearly adiabatic. However, in  
 139 many locations at all latitudes, the air-sea heat flux ( $\eta_T^{BF}$ ) role is sizeable and can reach the same order  
 140 of magnitude as the advection term. This is unlike the heat diffusion term which is negligible almost  
 141 everywhere (except locally in the Arctic and along the Antarctica margin close to deep water  
 142 formation regions, Forget and Ponte 2015) and thus will be neglected here.

143  
 144 The interannual and longer time-scale variability in steric sea level is predominantly the result of  
 145 internal reorganisation of water masses in the ocean forced by anomalies in surface wind stress  
 146 (Stammer et al. 2013). When this contribution is removed, the remaining variability is explained by  
 147 buoyancy forcing anomalies (i.e., surface heat and freshwater exchanges) and the intrinsic oceanic  
 148 variability spontaneously generated by the ocean circulation under the repeated seasonal atmospheric  
 149 forcing (Penduff et al. 2011, Serazin et al. 2015). It is essentially surface heat fluxes anomalies, which  
 150 explain most of this residual variability (see section 3). The surface heat fluxes can be written as  
 151 follows:

$$152 \eta_T^{BF} \approx \varepsilon \frac{Q_{net}}{\rho_{ml} C_p} = \varepsilon \frac{Q_{sh} + Q_{lh} + Q_{sw} + Q_{lw}}{\rho_{ml} C_p} \quad (3)$$

153 where  $\varepsilon$  is the thermal expansion coefficient,  $\rho_{ml}$  is the density of the upper ocean mixed layer and  $C_p$   
 154 is its heat capacity. Surface heat fluxes are caused by the turbulent energy fluxes (sum of the sensible  
 155 heat flux  $-Q_{sh}$  and latent heat flux  $-Q_{lh}$ ), which are broadly proportional to the wind speed, the air-

160 sea temperature and humidity differences. Radiative fluxes (sum of the downward solar radiative flux  
161  $-Q_{sw}$  - and the longwave radiative flux  $-Q_{lw}$  -) are functions of air temperature, humidity, and  
162 cloudiness.

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## 165 2.2. SST variations deduced from the heat budget of the upper-ocean mixed layer

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167 SST from space closely reflects upper-ocean mixed layer temperatures, being within 0.2°C of upper-  
168 ocean mixed layer temperatures, other than where dynamic processes drive near-surface temperature  
169 gradients (Grotsky et al. 2008). It is governed by the processes regulating the exchange of energy at  
170 the sea surface and at the bottom of the mixed layer. These processes are both atmospheric (wind  
171 stress and buoyancy fluxes) and oceanic (heat transport by currents, vertical mixing and boundary  
172 layer depth influence). Mathematically, the heat budget can be written as follows (Deser et al., 2010)

173

$$174 \quad \frac{\partial T}{\partial t} = \frac{Q_{net,ml}}{\rho C_p h} + (\overrightarrow{U_{geo}} + \overrightarrow{U_{ek}}) \cdot \nabla T + \frac{W_e - W_{ek}}{h} \cdot (T - T_b) \quad (4)$$

175

176 where  $T$  is the SST,  $h$  is the mixed layer depth,  $\overrightarrow{U_{geo}}$  is the geostrophic current velocity,  $\overrightarrow{U_{ek}}$  is the  
177 Ekman current velocity,  $W_e$  is the vertical entrainment rate,  $W_{ek}$  is the Ekman pumping velocity and  
178  $T_b$  is the temperature at depth that is entrained in the mixed layer. Ekman and geostrophic currents  
179 contribute to the heat budget of the mixed layer through horizontal advection while entrainment  
180 velocity and Ekman pumping change the SST through vertical advection.  $Q_{net,ml}$  is the net surface  
181 energy flux which enters the mixed layer. It is the sum of the turbulent energy fluxes, the component  
182 of downward solar radiative flux, which is absorbed within the mixed layer and the longwave radiative  
183 flux. In general  $Q_{net,ml} \approx Q_{net}$  and both terms depend on the same variables: wind speed, air-sea  
184 temperature and humidity difference, cloudiness. However in regions of clear water and shallow  
185 mixed layer they can be different because a non-negligible portion of the downward solar radiative  
186 flux penetrates below the mixed layer (see section 3).

187

188 Thus, both SSH (through the steric sea level  $\eta_{st}$ ) and SST have a dependence on the surface heat  
189 fluxes  $\frac{Q_{net}}{\rho_{ml} C_p}$ . This common dependence is expected to give rise to correlated variability at interannual  
190 and longer time scales, in particular in response to the atmospheric forcing. In contrast SST has a  
191 different dependency on ocean circulation than SSH: SST depends on the circulation of the upper  
192 mixed layer while SSH depends on the circulation of the ocean from the surface down to the bottom  
193 (see for example Figure 1d in Piecuch and Ponte, 2011). This will result in different responses of the  
194 SST and SSH to the oceanic circulation and variability at interannual and longer time scales that are  
195 not correlated. In section 3, we will explore with the use of an ocean reanalysis the complex relation  
196 which exists between SSH and SST variability through their common response to surface heat fluxes.

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## 201 2.3 Bio-optics of ocean-colour

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203 In the global ocean, phytoplankton biomass is essentially governed by the availability of light  
204 and nutrients (locally temperature and concentration of predators -i.e. zooplankton- also can play a  
205 role). A common approach to estimate phytoplankton biomass is to measure the concentration of  
206 chlorophyll (the main pigment in phytoplankton cells), because of the central role this pigment plays  
207 in photosynthesis, because it is produced uniquely by plants, and because it is easily measured, and  
208 can be estimated from satellite ocean-colour observations.

209

210 Phytoplankton cells are viable in the upper layer of the ocean, where sufficient light is available and  
211 where recycled nutrients are available, and additional nutrients can be brought up from the deep  
212 oceans through upwelling, wind mixing, advection and other physical processes. The solar irradiance

213 penetrating the ocean will be attenuated with depth due to the optical properties of pure seawater itself,  
214 and also due to the presence of particles, in particular phytoplankton cells and the chlorophyll pigment  
215 they contain.

216  
217 Ocean colour is determined by the spectral variance of reflectance, defined as the ratio of upwelling  
218 irradiance (radiant flux per unit surface area,  $\text{W m}^{-2}$ ) at the surface of the ocean to the downwelling  
219 irradiance at the same depth. In satellite applications, it is also customary to use the term remote-  
220 sensing reflectance for the ratio of upwelling radiance (radiant flux per unit surface area and unit  
221 steradian,  $\text{W m}^{-2}\text{sr}^{-1}$ ) to downwelling irradiance at the surface. Both irradiance reflectance and remote-  
222 sensing reflectance, which vary with the wavelength of light considered, are functions of absorption  
223 ( $a$ ) and back-scattering ( $b_b$ ) coefficients of light, which in turn, are affected by the absorption and  
224 back-scattering properties of phytoplankton, which vary with the concentration of phytoplankton in  
225 the water, and also with the type of phytoplankton present. Though other substances, such as detritus  
226 and coloured dissolved organic material contribute to the variability in ocean colour, it is often  
227 assumed, as a first approximation, that phytoplankton may be treated as the single, independent  
228 variable that determines ocean colour in the open ocean. Remote-sensing reflectance at wavelength  $\lambda$   
229 can be written as:

$$230$$
$$231 \quad R_{rs}(\lambda) = f(a(\lambda), b_b(\lambda)) \quad (5)$$
$$232$$

233 where  $f$  is a function that increases with back-scattering  $b_b$  and decreases with absorption  $a$ . The  
234 function  $f$  also incorporates the effect of the angular structure in the light field on  $R_{rs}$ . As noted  
235 earlier, the absorption and back-scattering coefficients are both functions of the concentration of  
236 phytoplankton in the water, typically measured as the biomass  $B$  in chlorophyll units:

$$237$$
$$238 \quad a(\lambda) = a_w(\lambda) + a_B(\lambda) + a_Y(\lambda) + a_X(\lambda) \quad (6)$$
$$239$$

240 and

$$241$$
$$242 \quad b_b(\lambda) = b_{bw}(\lambda) + b_{bB}(\lambda) + b_{bX}(\lambda) \quad (7)$$
$$243$$

244 Where the subscripts  $w$ ,  $B$ ,  $Y$  and  $X$  stand for pure seawater, chlorophyll concentration, concentration  
245 of coloured dissolved organic matter (sometimes called yellow substance) and particles in suspension  
246 other than phytoplankton, respectively. In the visible domain of the electromagnetic spectrum,  
247 reflectance is a small part of the solar flux that reaches just below the surface, of the order of 5%. The  
248 rest penetrates into the ocean. The same optical properties that determine reflectance at the sea surface  
249 also dictate the rate of light penetration in the ocean. Inside the water-column, the decrease in  
250 irradiance level with depth can be described using an exponential function as

$$251$$
$$252 \quad I(z, \lambda) = I_0(\lambda)e^{-K(\lambda)z} \quad (8)$$
$$253$$

254 where  $I(z, \lambda)$  is irradiance at depth  $z$  and wavelength  $\lambda$ ,  $I_0(\lambda)$  is incident irradiance (just below the  
255 surface), and  $K(\lambda)$  is the diffuse vertical attenuation coefficient in  $\text{m}^{-1}$  (Kirk, 1994). The attenuation  
256 coefficient can be expressed as

$$257$$
$$258 \quad K(\lambda) = g(a(\lambda), b_b(\lambda)) \quad (9)$$
$$259$$

260 where  $g$  is an increasing function of both  $a$  and  $b_b$ .

261  
262 Because  $K$  increases with chlorophyll, phytoplankton-rich waters display a high attenuation  
263 coefficient, and the irradiance will not penetrate as deep as in low chlorophyll waters. The photic  
264 depth (defined as the depth at which irradiance reaches 1% of  $I_0$ ) is deeper in clear waters and  
265 shallower in waters characterized by a high chlorophyll biomass (e.g., Edwards et al., 2001).

266

267 The energy absorbed at a particular depth yields a local temperature increase with time  $t$  given by  
268 (e.g., Lewis et al., 1983)

269  
270 
$$\frac{\partial T}{\partial t} = \frac{1}{\rho_0 c_p} \frac{\partial I}{\partial z} \quad (10)$$

271  
272 where  $T(z)$  is the temperature change (in °K) of the water due to the heating,  $c_p$  is the specific heat  
273 capacity (in J kg<sup>-1</sup> K<sup>-1</sup>) and  $\rho_0$  is the density of the water (kg m<sup>-3</sup>). This equation does not account for  
274 the decrease in downwelling light with depth due to back scattering at depth  $z$  (which does not  
275 contribute to local heating), or for heat gains by attenuation of upwelling irradiance at that depth  
276 (Edwards et al., 2001; Zaneveld et al., 1981).

277  
278 Since  $I$  tends to zero below the photic depth, the heating will be confined to a layer near the surface in  
279 chlorophyll-rich waters, whereas in low-chlorophyll waters, the heat energy will penetrate farther  
280 down the water column. In other words, in an oceanic region where surface chlorophyll concentration  
281 is high, the upper ocean layer will be warmer (and the deeper layer will be cooler) compared with a  
282 region where chlorophyll is absent (e.g., Sathyendranath et al., 1991; Edwards et al., 2001; Wu et al.,  
283 2007; Zhai et al., 2011).

284  
285 These considerations highlight the relationship between SST and ocean colour: the optical properties  
286 have a modulating influence on the distribution of solar heating with depth, and hence on SST. Within  
287 the mixed layer, it would be reasonable to assume that the chlorophyll concentration would be  
288 uniform, and equal to the value at the surface determined from ocean-colour remote sensing. Where  
289 the photic depth is significantly deeper than the mixed layer, the vertical structure in chlorophyll  
290 concentration would have to be taken into account (Lewis et al. 1983).

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### 293 **3. Spatio-temporal patterns in SSH, SST and OC and their relation**

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295 In this section we analyse the spatio-temporal patterns of SSH, SST and OC obtained from the satellite  
296 archive. To estimate these patterns we use the ESA Climate Change Initiative (CCI) project datasets  
297 because they have been developed to be the most homogeneous and stable satellite records at  
298 interannual to decadal time scales as possible (Ablain et al. 2015, Merchant et al. 2015,  
299 Sathyendranath et al. 2016). Among SSH, SST and OC CCI records, the SSH and SST records cover  
300 the same period: 1993-2014. In the following we focus our analysis on this period 1993-2014 because  
301 it is the longest period covered by the CCI datasets. In addition to satellite datasets we use an update  
302 of the analysis of ocean subsurface temperature and salinity by Ishii and Kimoto (2009) to estimate the  
303 steric effect in SSH variability. This analysis is based on temperature and salinity data from the World  
304 Ocean Database and Atlas, the Global Temperature-Salinity in the tropical Pacific from IRD, the  
305 Centennial in situ Observation Based Estimates sea surface temperature and Argo profiling floats data.  
306 This in-situ measurements record has become almost global since 2006 for the upper 2000 m (Argo  
307 array). Before 2006 the historical in situ measurements are very sparse (time, space and depth), and  
308 they are sparser for salinity than for temperature profiles. In particular a large fraction of the  
309 deep/abyssal ocean (below 700m depth before 2006 and below 2000m depth after 2006) still lacks in  
310 in-situ measurements. We also make use of an ocean reanalysis to unravel the role of the atmospheric  
311 forcing and the ocean circulation in the patterns of SSH, SST and OC. The reanalysis used here is the  
312 ECCO (Estimating the Circulation and Climate of the Ocean)-Production Version 4 Release 1 solution  
313 (Forget et al. 2015; Forget and Ponte 2015), hereafter referred to simply as ECCO. This product  
314 represents a model solution that has been constrained to observations (e.g., satellite altimetry, Argo  
315 floats, historical hydrography, etc.) based on the method of Lagrange multipliers. The optimization is  
316 achieved by making iterative adjustments to the initial conditions, boundary conditions, and internal  
317 model parameters (consult Wunsch and Heimbach, 2007, for more details on the general procedure).  
318 This solution covers the period 1992-2011. The ocean model setup is global, including the Arctic, and  
319 is fully coupled to an interactive sea ice and snow model. The spatial grid has a nominal 1-degree  
320 horizontal resolution, telescoping to 1/3-degree in the tropics and effectively 40 km in the Arctic, and

321 uses 50 vertical levels. Initial-guess bulk-formula surface forcing is taken from the Interim European  
322 Centre for Medium-Range Weather Forecasts Reanalysis (ERA-Interim) of Dee et al. (2011) and  
323 iteratively adjusted as outlined above. The model also uses parameterization schemes to incorporate  
324 the effects of geostrophic eddies, vertical mixing, and salt plumes (see Forget et al., 2015, for more  
325 details on this solution). We have chosen ECCO v4 because it has been shown to fit well to  
326 observations of SSH (Forget and Ponte 2015), in situ temperature and salinity (Forget et al. 2015),  
327 SST (Buckley et al., 2014) as well as other ocean circulation and climate variables (e.g., Piecuch et al.  
328 2015).

329  
330 In this paper we focus on the climate fluctuations of the ocean and we analyse their signature on the  
331 SSH, SST and OC variables. In several regions, such as western boundary current regions, individual  
332 meso-scale eddies can generate substantial variability in ocean variables and mask the underlying  
333 climate fluctuations. To remove this noise from eddies we perform a spatio-temporal smoothing on the  
334 SST and SSH data (see section 3.3.2 for OC analysis) which filters out the smaller spatial scales and  
335 the shorter time scales consistently in observations and model. The spatio-temporal smoothing consists  
336 in applying first a 30-day boxcar window and then a spatial Gaussian filter which removes the  
337 variability on scales smaller than  $3^\circ$ . Like in Forget and Ponte (2015), we have chosen 30 day for the  
338 temporal smoothing scale because it corresponds to the longest repeat cycle amongst observation  
339 datasets and thus it enables to get full maps of each observation dataset before performing the spatial  
340 smoothing. For the spatial smoothing scale we have chosen  $3^\circ$  because it enables to separate properly  
341 the ocean mesoscale, which relates to the baroclinic Rossby radius of deformation that is of the order  
342 of hundreds of km, from the ocean large scale climate fluctuations driven by the atmospheric forcing,  
343 which are of the order of thousands of km (Forget and Ponte 2015).

344  
345 In the following sections, the spatio-temporal patterns of SSH, SST and OC at interannual and decadal  
346 time scales are analysed in terms of standard deviation and trend of the time series over the period  
347 1993-2011. For SSH and SST data, the linear trend over 1993-2011 is estimated simultaneously with  
348 the seasonal cycle from monthly time series. After removing the trend and the seasonal cycle, we  
349 apply a 13-month Hanning window to remove remaining intraseasonal signals and then we estimate  
350 the standard deviation on the residual time series.

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### 356 **3.1 sea level and how it relates to ocean mass, temperature and salinity**

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#### 358 *3.1.1 Regional variations*

360 SSH observations from the CCI sea-level dataset show considerable interannual variability (Fig.1a).  
361 This variability is maximum within the tropics. Temperature and salinity observations have a very  
362 sparse coverage in the Arctic and do not allow evaluation of the steric effect in this region. But for the  
363 rest of the ocean, they show that most of the interannual variability in sea level is dominated by the  
364 interannual variability in steric sea level (Fig.1b). This result is confirmed by independent data from  
365 space gravimetry over the period 2004-2014. Indeed the Gravity Recovery And Climate Experiment  
366 (GRACE) mission, which measures the time varying gravity field of the Earth, provides continuous  
367 estimates of the ocean bottom pressure since 2004. These estimates confirm that ocean bottom  
368 pressure interannual variability is smaller than the steric variability by an order of magnitude except at  
369 high latitudes and in shallow shelf seas ( $>60^\circ\text{N}$  and  $<55^\circ\text{S}$ , Piecuch et al., 2013; Ponte and Piecuch,  
370 2014).

371 At mid and low latitudes the steric sea level signal is essentially due to temperature changes (Fig 1c).  
372 Salinity changes play only a local role (Fig. 1d) but this role can be sizeable in several regions like the  
373 Eastern Indian ocean (Llovel and Lee 2015) or the North Atlantic (Wunsch et al., 2007; Köhl 2015;  
374 Forget and Ponte 2015).

375



**Fig.1 : (a) Standard deviation in SSH ( $\eta$ ) from the CCI Sea-level dataset. (b),(c),(d) Standard deviation in respectively steric height ( $\eta_{Steric}$ ), thermosteric height ( $\eta_T$ ) and halosteric height ( $\eta_S$ ) from an update of Ishii and Kimoto (2009). Heights are in mm. Note that thermosteric and halosteric data have large uncertainties in regions where the in-situ temperature and salinity observations are poor in the southern ocean before 2006.**

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The picture is similar for the trends in sea level as for the standard deviation. They are largely dominated by the steric effect (Fig2.a,b). The halosteric effect is much smaller than the thermosteric effect but it is sizeable in many regions and should not be neglected (Fig. 2.c,d). Interestingly, in the few regions where the halosteric signal is sizeable, like in the southern tropical Pacific, its effect tends to compensate the thermosteric effect. Such compensation suggests nearly adiabatic transport of the water masses in these regions (an example of such adiabatic transport is heaving of the water column) as suggested by previous authors (e.g., Wunsch et al. 2007, Durack et al. 2014).



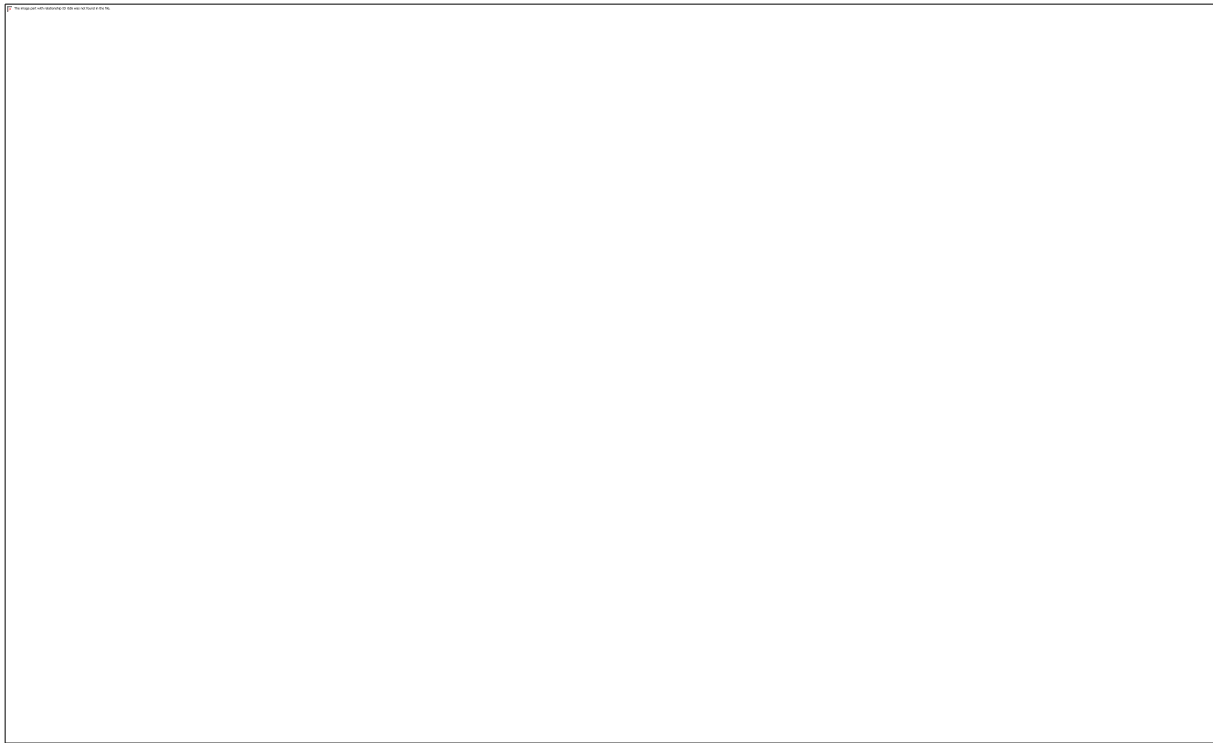


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**Fig.2 : (a) Trends over 1993-2011 in SSH ( $\eta$ ) from the CCI Sea-level dataset (Ablain et al. 2015). (b),(c),(d) Trends over 1993-2011 in respectively steric height ( $\eta_{steric}$ ), thermosteric height ( $\eta_T$ ) and halosteric height ( $\eta_S$ ) from an update of Ishii and Kimoto (2009). Trends are in mm/yr. Note that thermosteric and halosteric data have large uncertainties in regions where the in-situ temperature and salinity observations are poor in the southern ocean before 2006.**

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*3.1.2 the effects of wind stress and surface buoyancy fluxes*

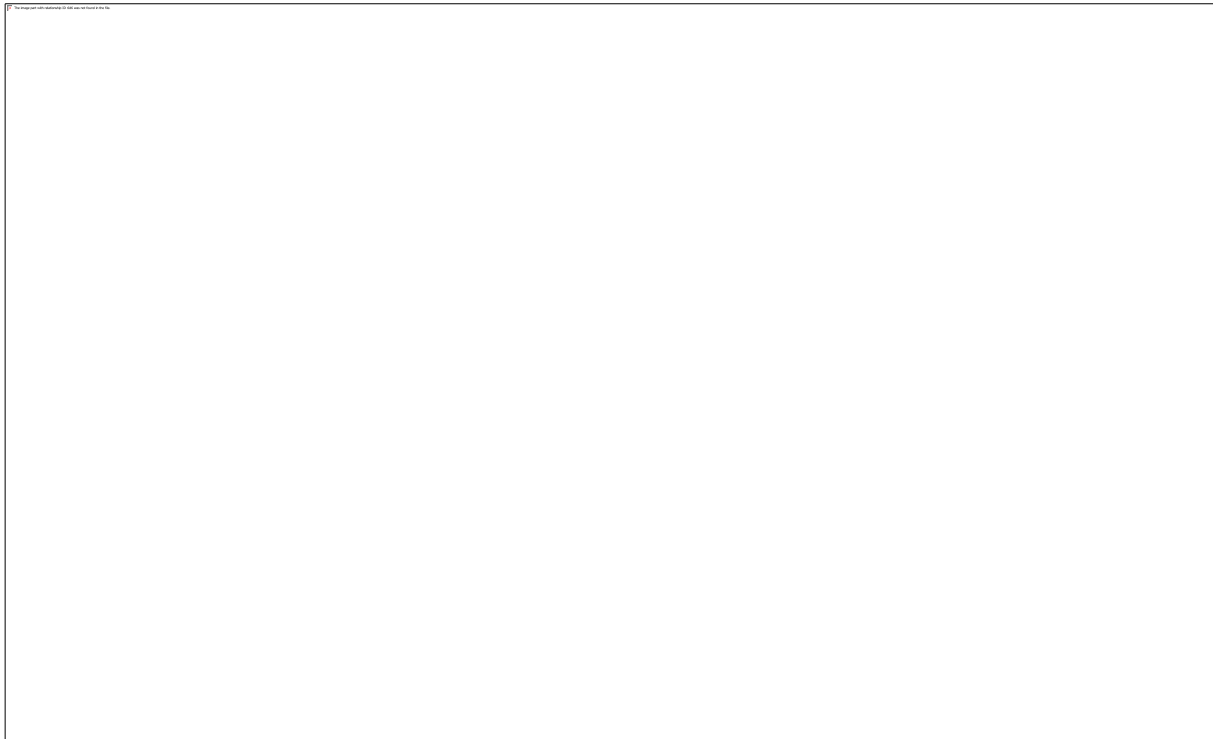
399 The ECCO model estimate shows similar patterns in SSH variability and trends as those obtained from  
400 the CCI observations (Fig. 3a, 4a). The ECCO estimate of the ocean bottom pressure (Fig.3c, 4c) and  
401 of the steric effect (Fig. 3b, 4b) corroborates the notion that SSH variability and trends are dominated  
402 by the steric effect in general. At high latitudes, in the polar oceans where altimetry data is not  
403 available, ECCO also provides SSH estimate. In these regions, the SSH variability and trends appear  
404 to be substantial and actually dominated by the mass signal (Fig. 3c, 4c). This mass signal in the  
405 Arctic and in the southern ocean has been confirmed by GRACE observations since 2004 (Purkey et  
406 al. 2014, Makowski et al. 2015). It is the result of the barotropic circulation caused by wind forcing  
407 (Frankcombe et al. 2013, Volkov and Landerer 2013, Volkov 2014, Peralta-Ferriz et al. 2014,  
408 Fukumori et al. 2015, Makowski et al. 2015). In the Arctic, ECCO shows that the steric effect is  
409 actually sizeable and should not be neglected in comparison to the mass signal. This steric effect has a  
410 significant halosteric component coming from the variability of the freshwater inputs in the Arctic and  
411 the sea ice (see Figure 4 in Köhl ,2014, which suggests that both mixed layer processes and heaving of  
412 isopycnals contribute to interannual halosteric variability in the Arctic). At low and mid latitude,  
413 ECCO confirms that the sea level variability and trends are almost entirely of thermosteric origin with  
414 some local halosteric effect, which tends to compensate the thermosteric effect as in observations.

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**Fig.3 : (a),(b),(c),(d) Standard deviation in respectively SSH ( $\eta$ ), thermosteric height ( $\eta_T$ ), halosteric height ( $\eta_S$ ) and bottom pressure ( $p_b$ ) from ECCO version4. Heights are in mm**

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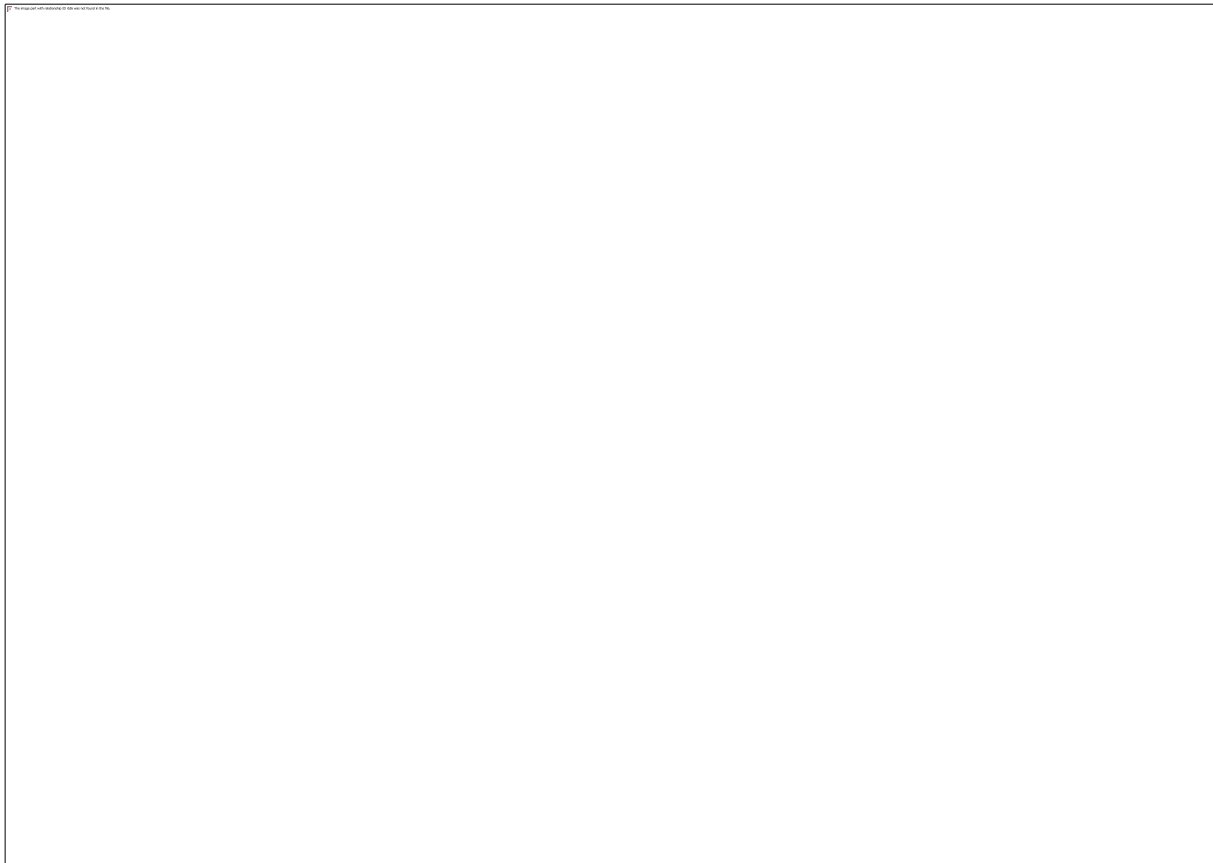


**Fig.4 : (a),(b),(c),(d) Trends over 1993-2011 in respectively SSH ( $\eta$ ), thermosteric height ( $\eta_T$ ), halosteric height ( $\eta_S$ ) and bottom pressure ( $p_b$ ) from ECCO version4. Trends are in mm/yr**

An advantage of reanalyses over observations is that they allow unambiguous identification of the anomalous forcings, which are responsible for the interannual variability and the trends in SSH. In this subsection, we use perturbation experiments run based on the ECCO model setup to distinguish the influences of wind stress and buoyancy exchanges on sea level. This set of experiments is

429 described more extensively by Forget and Ponte (2015). In more detail, each member of that set  
430 employs different surface boundary conditions. In each run, a common forcing component was  
431 prescribed, which comprised both fully variable surface buoyancy exchanges and a climatological  
432 mean seasonal cycle in wind stress. What differed between the simulations was that interannual  
433 changes in the wind stress were or were not turned off over the global ocean. The difference between  
434 these two experiments (shown in Figs. 5a and 6a) represents the oceanic response to interannual and  
435 decadal wind stress changes; the experiment without interannual or decadal wind stress changes (but  
436 retaining the common forcing component; shown in Figs. 5b and 6b) reflects the ocean's adjustment to  
437 interannual and decadal changes in buoyancy exchanges as well as any nonlinear intrinsic changes (cf.  
438 Penduff et al. 2011; Piecuch and Ponte 2012)

439  
440 The perturbation experiments reveal that SSH variability is essentially forced by surface wind stress  
441 anomalies (see Fig. 5.a,c and also Stammer et al. 2013, Forget and Ponte 2015). The remainder  
442 variability, which is forced by anomalous surface fluxes of buoyancy along with any nonlinear  
443 intrinsic variability (cf. Penduff et al. 2011), is in general significantly smaller except in some regions  
444 (Fig. 5.b,d) such as the region of the Antarctic Circumpolar Current (ACC), in the Kuroshio extension,  
445 in the Arctic, in the North Atlantic, in the North Pacific and in the tropical Pacific confirming earlier  
446 results from Thompson et Ladd (2004), Cabanes et al. 2006 and Piecuch and Ponte (2012, 2013).  
447



448  
449 **Fig5. Interannual variability in  $\eta$  due to wind stress (a) and to buoyancy forcing (b) from ECCO version 4.**  
450 **(See the text for the definition of  $\eta$  due to wind stress and to buoyancy forcing). Ratio of the interannual**  
451 **variability in  $\eta$  due to wind stress (c) and buoyancy forcing (d) over the interannual variability in total  $\eta$**   
452

453 In the case of SSH trends over 1993-2011, the perturbation experiments reveal that both wind stress  
454 forcing and buoyancy forcing play a leading role in SSH trends but in different regions. In the tropics  
455 it is the wind stress anomalies which are responsible for the SSH trends (Fig. 6.a). Indeed, the large  
456 positive pattern in the western tropical Pacific has been associated to a deepening of the thermocline in  
457 response to trade wind intensifications, likely linked to the negative phase of the pacific decadal  
458 oscillation in recent decades (Timmermann et al. 2010, Merrifield 2011, MacGregor et al. 2012, Qiu  
459 and Chen 2012, Meyssignac et al. 2012, Moon and Song 2013, Meehl et al. 2013, Palanisamy et al.

460 2015). In the extra tropics, the SSH trends are essentially driven by the buoyancy forcing (Fig.6.b)  
461 while the wind stress forcing contribution is smaller but remains significant in several regions (such as  
462 the North and South Pacific and the Arctic, see Fig. 6.a,b.). Note that the trends in SSH driven by the  
463 buoyancy forcing are positive all over the ocean reflecting the general warming of the ocean.  
464 Interestingly, they are maximum in the subtropical gyres and in the North Atlantic subpolar gyre.  
465



466 **Fig. 6 : Trends in  $\eta$  between 1993 and 2011 (in mm/year) from ECCO version 4. (a): contribution due to**  
467 **wind stress; (b): contribution due to buoyancy forcing.**  
468  
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### 470 **3.2 SST and how it relates to sea level and ocean temperature and salinity**

#### 471 *3.2.1 Regional variations*

472 Interannual variability in SST is presented in Fig. 7 as the standard deviation of de-seasonalised, de-  
473 trended monthly SST anomalies. Very low SST variability at high latitudes in areas of perennial sea  
474 ice reflects the damping of variability in SST by the freezing and melting of that sea-ice: in the SST  
475 data set, SST is set to the freezing temperature of sea water where the ocean is ice covered. Variability  
476 is greater in areas of seasonal sea ice, where variations in sea ice extent are reflected also in SST.  
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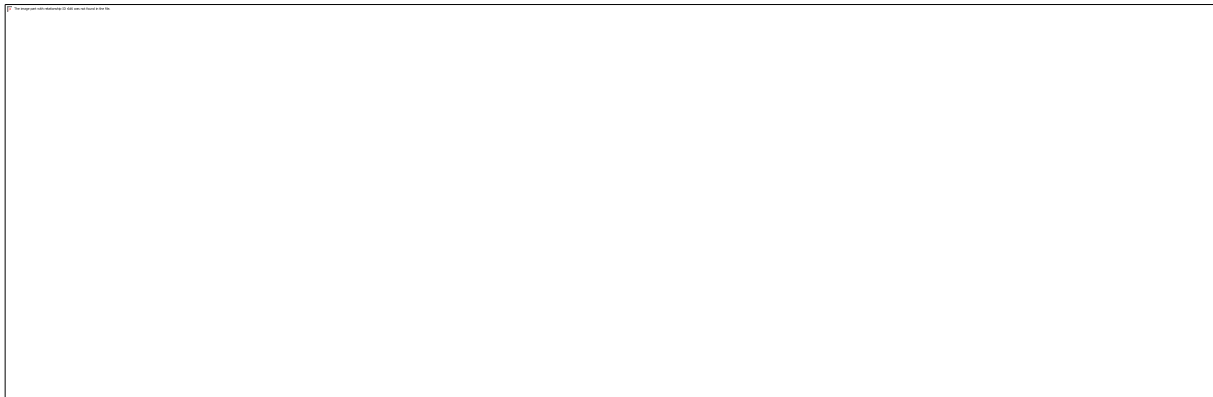
479 In the extra-tropics poleward of sea ice, the variability in monthly SST is seen to be relatively high,  
480 typically of order 1 K. Much of the large-scale development of SST anomalies in the extra-tropics is  
481 driven by large-scale reorganization of atmospheric circulation anomalies. Atmospheric variability  
482 with timescales longer than  $\sim 10$  days is effective at driving SST anomalies that reflect the temperature  
483 of the upper mixed layer, because of the large thermal inertia of the upper-ocean mixed layer  
484 (Frankignoul & Hasselmann 1977, Deser et al. 2003). (Note that however SST can respond strongly to  
485 higher frequency atmospheric variability when the depth of the response is shallow, as in diurnal  
486 variability.) The large-scale nature of atmospheric teleconnection patterns is imprinted upon the SST  
487 anomaly field essentially via the surface energy fluxes and Ekman currents (e.g., Cayan 1992,  
488 Marshall et al. 2001, Visbeck et al. 2003). Spatial variations in mixed-layer depth (which themselves  
489 reflect the recent history of wind stress) modify the effective thermal inertial, and also play a role in  
490 determining the magnitude of the SST anomalies.  
491

492 Across much of the tropics, variability in monthly SST is low,  $< 0.5$  K. The clear exception is the  
493 equatorial Pacific ocean, where SST variability  $> 1.5$  K is present eastwards of  $180^\circ\text{W}$ , associated with  
494 the El Nino Southern Oscillation (ENSO). ENSO is a coupled mode of variability, in which large-  
495 scale atmospheric circulation anomalies develop in close interaction with the SST variability, in  
496 contrast to extratropical anomalies (e.g. Deser et al. 2010).  
497

498 Local processes such as upwelling, entrainment, and lateral advection also contribute to SST  
499 variability. For example, vertical advection plays a prominent role along the coastal and equatorial  
500 upwelling zones, with variability being wind-driven. Horizontal advection is important along the  
501

502 western-boundary current regions (e.g., the Gulf Stream and Kuroshio Current). Oceanic processes  
503 also play an indirect role in SST variability by affecting the depth of the upper-ocean mixed layer.  
504

505 The trends in SST over the period are relatively uniform around 0.3 K decade<sup>-1</sup> across much of the  
506 Atlantic and Indian Oceans. The North Atlantic, including the Barents Sea, the sub-Greenland gyre  
507 and Labrador Sea, shows a warming trend exceeding 0.5 K decade<sup>-1</sup>. Across the Pacific Ocean, the  
508 SST trend largely reflects the change in the tendency of the Pacific Decadal Oscillation over the period  
509 (Fig.7a), as previously noted with respect to SSH.  
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513 **Fig7. Variability and trends in SST from the CCI dataset (Merchant et al., 2015) a) Standard deviation in K. b)**  
514 **Linear trend in over 1993-2011 in K per decade.**  
515

### 516 3.2.2 The effect of surface heat fluxes and ocean transport 517

518 The ECCO model estimate shows general similar patterns in SST variability as those obtained from  
519 the CCI observations (Fig. 8a) except at mid latitude and in the southern ocean where it tends to  
520 underestimate the SST variability notably in eddy active regions (in the Kuroshio and Gulf stream  
521 extensions, the Malvinas current and the Agulhas current). It shows also general similar patterns in  
522 SST trends (Fig. 9a) except in the North Atlantic (from the Gulf Stream extension to the Barents sea)  
523 and in the eastern part of the north Pacific where trends are underestimated by a few tenth of  
524 °K/decade and in the north east Indian ocean and China sea where trends are overestimated by a few  
525 tenth of °K/decade.

526 As for SSH, we use ECCO output to infer the role of surface heat fluxes and ocean transport  
527 divergences in SST variations. We integrate vertically the upper-ocean temperature budget (equation 7  
528 in Piecuch and Ponte, 2012) over a constant climatological mixed layer depth computed for each grid  
529 cell. The role of surface heat fluxes on SST is diagnosed by calculating the changes in SST due to the  
530 surface heat forcing term  $\frac{Q_{net,ml}}{\rho C_p h}$ . The ocean transport divergences effect is diagnosed by calculating

531 the changes in SST due to the sum of the advection and diffusive terms. This form of the upper-ocean  
532 temperature budget does not distinguish between Ekman and geostrophic transport contributions as in  
533 equation (4). However, it allows the same separation of SST in terms of surface heat forcing and ocean  
534 transport divergences as for steric sea level (section 3.1). In that sense, it allows to compare easily  
535 between upper ocean heat and steric sea level budgets, being able to infer whether surface heat fluxes  
536 or ocean transport divergences are more or less important for upper ocean heat or steric sea level  
537 changes.

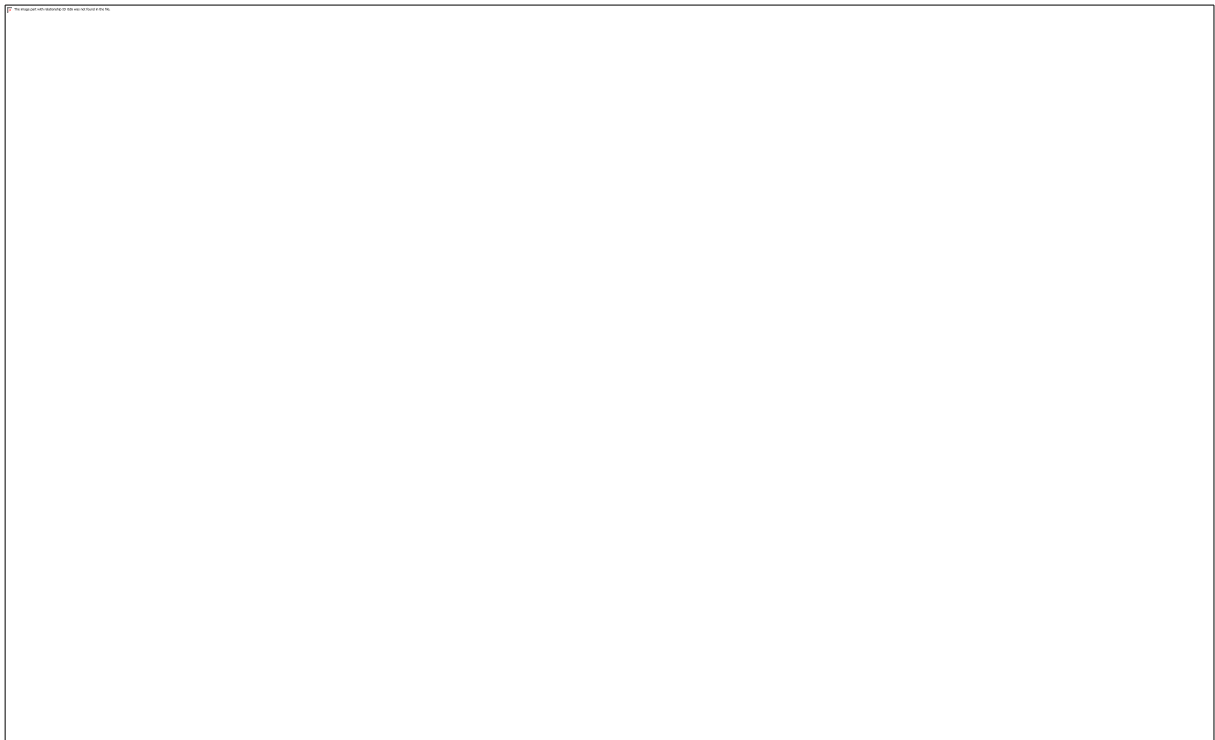
538 Both standard deviation patterns in SST due to surface heat flux and ocean transport divergences are  
539 very similar (see Fig S1a,b in the supplementary material). They show high variability in the tropics  
540 and upwelling regions and low variability at higher latitude with a minimum in the arctic ocean north  
541 of 80°N and in the southern ocean. The amplitudes of both standard deviation maps are also  
542 significantly higher than the standard deviation of the SST map almost everywhere: this is a result of  
543 integrating the ocean temperature budget in time. The effect of surface heat fluxes on SST is opposite  
544 to the effect of ocean transport divergences almost everywhere and they tend to cancel out (see Fig  
545 S1a,b). However the cancellation of surface heat fluxes effect with ocean transport divergences effect

546 is not exact and gives rise to SST variability. Fig. 8b and c show the standard deviation of the residual  
547 in SST due to surface heat fluxes and ocean transport respectively after removing the cancellation part  
548 (see also Fukumori and Wang 2013). Figure 8b shows that the interannual variability in SST due to  
549 the surface heat flux forcing is larger almost everywhere except in the tropical Pacific where ocean  
550 transport divergences play a similar role as surface heat fluxes in the local SST variability.  
551 The same compensation process between surface heat fluxes and ocean transport divergences effects  
552 on SST occurs at longer time scales. The trend pattern in SST due to surface heat flux and ocean  
553 transport are very similar and tend to cancel out each other (see Fig. S2a,b in the supplementary  
554 material). Trends in SST due to surface heat fluxes are very large in the tropics and very low in  
555 boundary current regions and at high latitude (Fig. S2a) while trends in SST due to ocean transport  
556 divergences are opposite in the same regions reflecting that the ocean gains most of the heat in the  
557 tropics and transports it to higher latitudes where it is released to the atmosphere or buried in the  
558 deeper layers of the ocean. However the effect of surface heat fluxes and ocean transport divergences  
559 on SST trends do not fully cancel out everywhere. Fig. 9b,c shows the residual in SST trends due to  
560 excess surface heat fluxes or ocean transport divergences respectively after removing the cancellation  
561 part. Figure 9b shows that surface heat fluxes dominate over ocean transport divergences and explain  
562 the SST trends in the tropical Indian ocean, in the western tropical Pacific ocean, in the North Pacific  
563 and North Atlantic subtropical gyres. It also dominates in the North Atlantic subpolar gyre and on the  
564 southern edge of the South Pacific subtropical gyre. In all upwelling regions (California current,  
565 Humboldt current and to a lesser extent in the Benguela current) it is the ocean transport divergences  
566 effect which dominates and explain the SST trends. In the eastern tropical Pacific, in the Indian and  
567 South Pacific subtropical gyres and in the North Atlantic subpolar gyre, it also dominates.  
568 Interestingly enough, the dominance of the ocean transport divergences over the surface heat fluxes,  
569 which explain the negative trends in SST in the California current, the Humboldt current and the eastern  
570 tropical Pacific, is consistent with the increase of cold deep water entrainment in these regions in  
571 response to increasing trade winds associated with the decreasing PDO over 1993-2011.  
572 The case of the western tropical Pacific region where the SST trends are dominated by surface heat  
573 fluxes warrants further discussion. At first sight, this result seems inconsistent with Fig. 6, which  
574 indicates positive SSH trends in this same region due to wind forcing (see section 3.1.2). But there is a  
575 possible interpretation. The time-mean SST budget in the western tropical Pacific Ocean is a balance  
576 between strong increase due to surface heat fluxes and strong decrease due to ocean transport  
577 divergences (see Fig. S2). Over the study period, well-reported-on wind stress changes act to reduce  
578 the magnitude of that background ocean transport divergence contribution, making it less negative  
579 (and resulting in the steric sea level rise in this area). This results in their being an excess of surface  
580 heat forcing contribution (or deficit of ocean transport divergence contribution), which leads to the  
581 result in Fig. 9 that the western tropical Pacific SST trends are attributed to surface heat flux forcing.  
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**Fig8. a) Standard deviation in SST from ECCO. b) standard deviation in SST due to the surface heat fluxes (after removing the cancellation part with ocean transport divergences) c) standard deviation in SST due to the ocean transport divergences (after removing the cancellation part with surface heat fluxes)**



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**Fig9. Same as Fig.8 but for trends**

### *3.2.3 comparison with SSH variability*

At interannual time scales SSH and SST show a fairly large common variability in the tropical band and in upwelling regions. Fig 10a shows the correlation map of SST and SSH detrended interannual time series computed from the CCI observational datasets. The correlation is higher than 0.7 in most

597 of the tropics, in the California current, the Humbolt current and in the Canari current. In the Bengela  
598 current the correlation is not significant but it is potentially because this narrow coastal current is not  
599 well sampled by satellite altimetry and its variability in SSH is not captured properly by the CCI  
600 product (Note that the correlation is significant and high in the Bengela current in the ECCO estimate  
601 which support this hypothesis –see Fig.10b-). In the extra tropics the correlation is in general non-  
602 significant except south east off Greenland and in some local eddies in the Antarctic circumpolar  
603 current, in the Kuroshio extension and in the Gulf stream. The ECCO estimate confirms this global  
604 picture and shows a similar pattern in the correlation map for SST and SSH as in the CCI correlation  
605 map (see Fig. 10b). The one difference between the ECCO correlation map and the CCI correlation  
606 map is that the ECCO estimate shows actually significant correlations in large regions of the extra-  
607 tropics but these correlations are low and below 0.6 in general.

608 To get insights on the cause of the common variability in SSH and SST we correlate the detrended  
609 interannual time series of SST with the wind driven SSH response (see Fig.10c) and with the  
610 buoyancy forced SSH response (see Fig.10d). Figures 10c and d show that in general, SST is more  
611 closely related to the buoyancy-forced sea level response than it is to either the total (wind+buoyancy)  
612 sea level or the wind driven sea level. In particular, for buoyancy-driven sea level, the average  
613 correlation coefficient between sea level and sea surface temperature is 0.55 and the correlation  
614 coefficient is significant over 68% of the global ocean. These same numbers for the wind-driven sea  
615 level are 0.25 and 29%, respectively. And these same numbers are 0.49 and 63% for the fully forced  
616 (wind+buoyancy driven) sea level, respectively. The reason for the high correlation between SST and  
617 buoyancy forced SSH is that the SST variability is dominated in most regions of the ocean by the SST  
618 response to surface heat fluxes (see section 3.2.2 and Fig. 8b) and that in many regions also surface  
619 heat fluxes effects dominates over freshwater fluxes effects in the buoyancy forced sea level response.  
620 As a result, surface heat fluxes appears as the main cause for the common variability in SSH and SST.  
621 In a few regions this statement does not hold. In regions where the freshwater effect on the buoyancy  
622 forcing is dominant over the heat flux effect, like in the Pacific warm pool or in the Arctic region, the  
623 correlation of buoyancy forced sea level with SST is not significant (see Fig.10c). In the tropics and in  
624 the upwelling regions, the wind stress is also a cause of sizeable common variability in SST and SSH  
625 along with the surface heat flux (see Fig 10d). This is because wind stress variability in the tropics  
626 generate at the same time a zonal pressure gradient and a zonal tilt in the thermocline, which make  
627 respectively sea level and SST vary in phase in this region.

628 The picture is different for trends in SSH and SST. The relationship between SST and buoyancy  
629 forced sea level is somewhat less clear for trends than it is for interannual to decadal time scales. For  
630 example, the correlation coefficient between the spatial patterns in SST (Fig 9a) and SSH (Fig. 2.b) is  
631 0.32, between SST and wind driven SSH (Fig.6a) is 0.15, and between SST and buoyancy forced sea  
632 level (Fig.6b) is 0.28. This means that at long time scales, while SST is more correlated with  
633 buoyancy-driven sea level than it is with wind-driven sea level, SST is even more correlated (but still  
634 modestly so) with the total sea level. This indicates that probably the relationship between SST and  
635 buoyancy-forced sea level is quite complex in general, and likely depends critically on time scale,  
636 among other factors. One striking feature when looking at figures 9a and 6b is that, out of the tropics  
637 the SST pattern seems quite similar to the buoyancy forced sea level except around Greenland. Around  
638 Greenland, sea level trends are more subdued while surface temperature trends are more pronounced.  
639 This decoupling is potentially due at least in part to freshwater fluxes and halosteric sea level changes  
640 in this region. Indeed, models suggest that such salinity effects on sea level are important in this  
641 region, and act to compensate and offset sea level changes due to temperature effects (i.e.,  
642 thermosteric height, Köhl , 2014). In the eastern tropical Pacific the negative SST trends are in phase  
643 with the negative SSH trends which are essentially caused by wind stress. This is consistent with the  
644 increasing trade winds associated to the decreasing PDO since the late 70s. Another interesting region  
645 is the Arctic (and to a lesser extent the south of the southern ocean) where the SST pattern does not  
646 correlate with any sea level pattern. The reason is probably that mass (bottom pressure) trends  
647 probably play an important role in sea level trends as well as the halosteric effects and hence make the  
648 SSH trends independent from the SST trends.

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651 **Fig10. Correlation coefficient between SST and SSH for CCI (a), between SST and total SSH for ECCO (b), between**  
652 **SST and the wind forcing driven SSH (c), between SST and the buoyancy forced SSH (d). Only positive correlations**  
653 **are shown because actually no significant negative correlation between SST and SSH were found.**  
654

### 655 **3.3 Ocean colour and how it relates to SST and SSH**

#### 656 *3.3.1 General principles*

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658 To investigate how ocean colour relates to SST and SSH, we must first understand the  
659 dominant processes driving the biophysical interactions, which include the constraints on  
660 phytoplankton dynamics imposed by the physical environment, and also the feedback mechanisms by  
661 which phytoplankton could modify their environment.  
662

663  
664 In the open-ocean waters of the tropics and subtropics, light is plentiful all-year round and  
665 phytoplankton biomass increases when nutrients become available in the mixed layer through various  
666 vertical processes, such as wind mixing and upwelling. The transport of cold waters from depths to the  
667 surface, carrying nutrients with them, reduces SST. In such cases, a decrease in SST can be considered  
668 a proxy for nutrient supply. These regions are therefore characterised by a negative correlation  
669 between SST and chlorophyll concentration.  
670

671 In contrast, in high-latitudes, light availability is the dominant driver: phytoplankton biomass shows a  
672 clear seasonality (tied to the annual solar cycle) and surface nutrients are generally replenished by  
673 deep-mixing in winter. In these regions, the seasonal changes in light at the sea surface are  
674 accompanied by seasonal warming of the waters, such that an increase in SST may be taken here to be  
675 a proxy for increase in available light.  
676

677 Sea-level variations can reflect alterations in ocean circulation and ocean density, occurring in  
678 response to changes in wind forcing, ocean warming, and changes in water and ice mass exchange  
679 between the land and the oceans (Church et al., 2013). Some of these processes form zones of  
680 convergence and divergence in the ocean, which are characterised by enhanced water-column  
681 stratification and upwelling respectively. In the tropics and subtropics, zonal wind-stress patterns  
682 cause convergence zones and increase sea level (Palanisamy et al., 2015). Such zones are also

683 characterised by deep thermoclines and low nutrient availability, which are unfavourable for  
684 phytoplankton production; and a negative correlation between SSH and chlorophyll concentration can  
685 be observed (Turk et al., 2001). In high-latitudes, oceanic convergence zones also show an increase in  
686 SSH, but the conditions may still favour phytoplankton production due to enhanced upper ocean light  
687 availability and supply of nutrients from winter mixing. In these regions, positive correlations can be  
688 found between SSH and chlorophyll (e.g., Racault et al. 2016, Wilson and Cole, 2005, Brewin et al.,  
689 2014).

690  
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### 692 *3.3.2 Correlation analyses based on CCI datasets*

693 Ocean-colour observations from the OC-CCI dataset show high inter-annual variability  
694 (Fig.11a) at high-latitudes and in coastal upwelling highly productive regions, whereas low variability  
695 is observed in the oligotrophic gyres (where phytoplankton production is minimum). Trends based on  
696 monthly chlorophyll anomalies are shown in Fig.11b. Correlation coefficient calculated using monthly  
697 means and monthly anomalies over the period 1998-2010 between chlorophyll (OC-CCI) and sea level  
698 (SL-CCI), and between chlorophyll (OC-CCI) and SST (SST-CCI) are presented in Fig.12. Although  
699 the correlation between chlorophyll and SSH tend to be weaker when compared with that between  
700 chlorophyll and SST, they display similar regional patterns throughout most of the global oceans. The  
701 tropics and subtropics typically show negative correlations (i.e., chlorophyll is low when SST and  
702 SSH are high), whereas the high-latitudes typically show positive correlations (i.e., chlorophyll is high  
703 when SST and sea level are high), as expected from the rationale presented above. In high latitudes,  
704 positive correlations between chlorophyll and SST are weaker when monthly anomalies  
705 (representative of inter-annual variability) are used compared with monthly means (when the  
706 seasonality has not been removed), but they show similar patterns (Fig.12). The correlation patterns  
707 displayed in the tropics, subtropics and high-latitudes are consistent with previous studies, which have  
708 been carried out for different time periods and different satellite sensors (e.g., Wilson and Cole, 2005,  
709 Brewin et al., 2012, Siegel et al., 2012, Brewin et al., 2014). This consistency suggests that the  
710 patterns are independent of the time period selected for the analysis, and of differences in chlorophyll-  
711 retrieval and atmospheric-correction algorithms used (Brewin et al., 2014; note also that OC-CCI  
712 product includes data from three sensors – SeaWiFS, MODIS and MERIS – which have been  
713 processed using SeaDAS (Fu et al., 1998) and POLYMER atmospheric correction algorithms  
714 (Steinmetz et al., 2011), and show a significant increase in data coverage (Sathyendranath &  
715 Krasemann, 2014, Racault et al., 2015, Sathyendranath et al., 2016).

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### 717 *3.3.3 Regional variations*

718 Other considerations have to be invoked to explain variations superimposed on these large-  
719 scale, latitudinal patterns. Some of these regions are examined more closely below.

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#### 721 *High-latitude HNLC regions*

722 In certain regions that are known as High Nutrient-Low Chlorophyll (HNLC) regions (Boyd et al.,  
723 2007), some trace nutrients (notably iron) may remain low, even when other nutrients such as  
724 nitrogen, phosphorus and silica are available, limiting phytoplankton production. In HNLC regions in  
725 high latitudes (including parts of the Pacific and the Southern Ocean), the correlation between  
726 chlorophyll and SST may become weak or negative, indicating that phytoplankton growth is  
727 predominantly limited by nutrient availability rather than light availability. Based on observed  
728 relationships between chlorophyll and SST and chlorophyll and SSH, Le Quéré et al. (2002) estimated  
729 that in the Southern Ocean between 40°S and 65°S, 44% of the total primary production was  
730 predominantly light limited, and 10% was driven by changes in stratification and nitrogen supply,  
731 which in turn was associated with changes in SST. The SST-related changes were higher in the south  
732 Atlantic part of the Southern Ocean, accounting for 30% of changes in the total primary production.  
733 To explain the remaining variability in biological productivity, the authors also considered changes in  
734 atmospheric dust deposition. However, in this region, atmospheric dust supply tends to remain low,  
735 originating from small dust sources in Argentina, Australia, and South Africa (Prospero et al., 2002)  
736 and a weak positive correlation between dust deposition and chlorophyll was found only in the west  
737 Pacific region of the Southern Ocean (Le Quéré et al., 2002). Their study demonstrates that simple

738 assumptions about light limitation and nutrient (nitrogen and iron) limitation may not be sufficient to  
739 explain fully the observed variations in chlorophyll distribution.

740

741 *Continental shelves and coastal regions*

742 In continental shelves and coastal regions, additional physical processes may also control the supply of  
743 nutrient in the euphotic zone, including coastal upwelling, Ekman pumping and river outflow. In the  
744 eastern subtropical North Atlantic, between 10°N and 25°N, the negative correlation found between  
745 chlorophyll and SSH, has been shown to result from an inverse relationship between SSH and  
746 nutricline depth (Pastor et al., 2014). In this region, the changes in nutricline depth and nutrient supply  
747 to the surface are governed primarily by changes in Ekman pumping (driving vertical advection  
748 processes), rather than by changes in stratification.

749

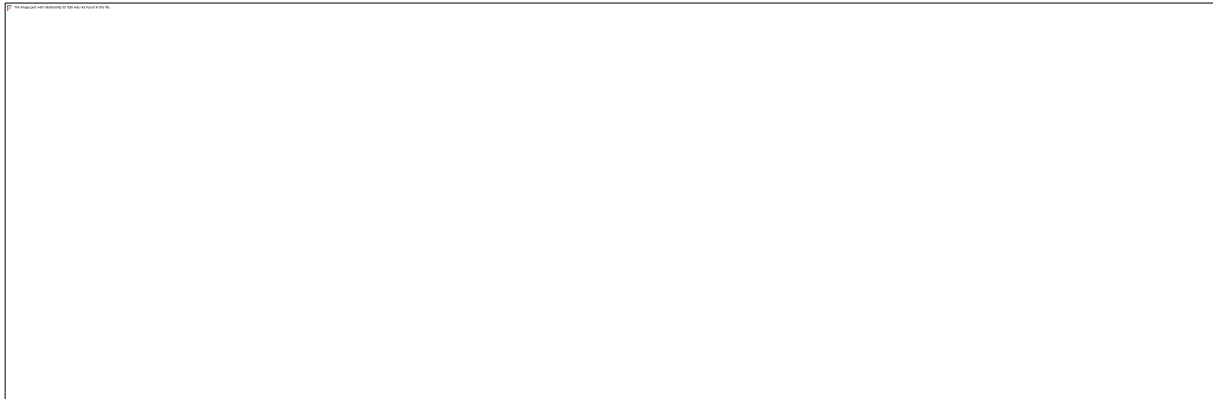
750 Atmospheric dust supply from the North African and Asian deserts, as well as coastal river discharge  
751 can also provide significant sources of nutrient supply for phytoplankton growth and modify the  
752 general latitudinal relationships between chlorophyll and SSH and between chlorophyll and SST. For  
753 instance, in the vicinity of the Amazon and Orinoco river plumes, positive relationships between  
754 chlorophyll and SST and chlorophyll and SSH are observed (Fig.12a,b). The chlorophyll  
755 concentration and the concentration of coloured-dissolved organic matter are higher in these river  
756 plumes than in surrounding waters (Smith and Demaster, 1996; Hu et al., 2004). The location and  
757 direction of these river plumes are influenced by anticyclonic eddies, characteristic of higher SST and  
758 SSH, which are capable of transporting the plumes hundreds of miles offshore (Johns et al., 1990;  
759 Corredor et al., 2004).

760

761 *Effect of eddies*

762 A prominent feature is shown off Western Australia where there is a clear positive relationship  
763 between chlorophyll and SSH (Fig.12a,c). In this region, anticyclonic eddies are thought to entrain  
764 shelf waters with regionally-high chlorophyll concentration (Pearce and Griffiths, 1991; Moore et al.,  
765 2007). These anticyclonic eddies then propagate westward and higher chlorophyll values are  
766 maintained through sub-mesoscale injection of nutrients into the anticyclonic-eddy boundary (Moore  
767 et al., 2007).

768



769

770

771 **Fig11. a) Standard deviation in chlorophyll concentration. b) Linear trend in chlorophyll concentration anomalies**  
772 **calculated over period 1998 and 2010. Only the linear regression coefficients, which are significant at the 95%**  
773 **confidence level are shown in color. Non-significant linear regression coefficients are in grey.**

774



775  
776 **Fig12. Relationships between ocean colour and SSH and between ocean colour and SST over the period 1998 to 2010.**  
777 **a) Correlation analysis between monthly means of chlorophyll concentration and SSH; b) Correlation analysis**  
778 **between monthly means of chlorophyll concentration and SST; c) Correlation analysis between monthly anomalies of**  
779 **chlorophyll concentration and SSH; and d) Correlation analysis between monthly anomalies of chlorophyll**  
780 **concentration and SST. Chlorophyll concentration data are from OC-CCI, SSH data are from SL-CCI and SST data**  
781 **are from SST-CCI. Only the correlation coefficients, which are significant at the 95% confidence level are shown.**  
782

### 783 **3.4 Bio-optical heating influence on SST, MLD, air-sea flux and how it may affect SSH**

784

785 The absorption of solar irradiance by phytoplankton cells (described in section 2.3) has been  
786 shown to influence considerably the heating rate of the upper layer of the world oceans. Using a  
787 mixed-layer model that incorporates ocean-colour observations, Sathyendranath et al. (1991)  
788 examined the effect of light attenuation on surface temperature by comparing model results forced by  
789 a pure-water case, and by a chlorophyll-dependent case. The authors estimated biologically-induced  
790 heating of surface temperature could reach a maximum of up to +4°C from August to September in the  
791 Arabian Sea, using satellite data for 1979. Subsequently, several investigators have used three-  
792 dimensional ocean models to study the influence of phytoplankton on SST and MLD at the regional  
793 and global scales (e.g., Nakamoto et al., 2000, Manizza et al., 2005, Wu et al., 2007, Zhai et al., 2011).  
794 As phytoplankton absorb a significant fraction of the incident solar radiation and increase the  
795 temperature in the surface layer, less radiation penetrates to greater depths. The increased temperature  
796 contrast between surface and deep waters can enhance water-column stability and reduce the depth of  
797 the mixed layer. For instance, Nakamoto et al. (2000) reported, in the Arabian Sea, a bio-optical  
798 heating of SST up to +0.6°C and mixed-layer depth decrease by 20m. During the Summer season, in  
799 the Labrador Sea, Wu et al. (2007) showed SST changes between -1.0 and +2.0°C, and mean MLD  
800 difference of 10m (i.e., ~ 20-50% shallower than it would be if phytoplankton were absent). Using a  
801 global ocean model, Manizza et al. (2005) observed temperature difference caused by the absorption  
802 of heat by phytoplankton between -0.2 and +0.6°C at mid and high-latitudes, with minimum values in  
803 winter and maximum values during the spring bloom. In their model results, the amplitude of the  
804 seasonal changes in the water-column thermal structure and the MLD increased from low to high-  
805 latitudes, with maximum MLD decrease by up to 20m observed at 60° in both hemispheres.  
806 Solar radiation and phytoplankton are not the only factor controlling SST and MLD in the oceans.  
807 Additional physical processes such as air-sea heat exchange, wind mixing, and horizontal advection

808 play important roles (see section 3.2). Zhai et al. (2011) demonstrated increased heat loss from the  
809 ocean to the air associated with bio-optical heating of the upper ocean caused by the presence of  
810 phytoplankton in the Gulf of St. Lawrence. Atmospheric heat gain induced by the presence of  
811 phytoplankton has been further examined in coupled ocean-atmosphere models, which show an  
812 amplification of the seasonal cycle of temperature in the troposphere, modulating the tropical  
813 convection patterns and atmospheric circulation (Shell et al., 2003), and affecting large patterns of  
814 climate variability such as ENSO (Zhang, 2015).  
815 The strong influence of the absorption of solar irradiance by phytoplankton cells on the heating rate of  
816 the upper layer of the world oceans suggests that bio-optical heating should have an influence not only  
817 on SST but also on SSH. This hypothesis is further supported by the fact that SST and buoyancy  
818 forced SSH show large common variability, which is due to the surface heat forcing (see section  
819 3.2.3).

820  
821  
822

## 823 4. Conclusion

824

825 In this paper we have analysed the regional variability of observed SSH, SST and OC from the CCI  
826 datasets over the period 1993-2011. We have focused our analysis on the signature of the ocean large-  
827 scale climate fluctuations driven by the atmospheric forcing and we did not consider the mesoscale  
828 activity, which has been filtered out. We used the ECCO version 4 ocean reanalysis to unravel the role  
829 of ocean transport and surface buoyancy fluxes in the ocean climate fluctuations and analyse the  
830 associated signatures on the observed SSH, SST and OC. Our analyses corroborate the findings from  
831 past studies on the role of ocean transport and surface buoyancy fluxes in the SSH variations at  
832 interannual to decadal time scales (Piecuch and Ponte 2011, Piecuch et al. 2013, Fukumori and Wang  
833 2013, Forget and Ponte 2015). They also provide new insights on the role of ocean transport and  
834 surface buoyancy fluxes in the SST variations and on the relation between the SSH, SST and OC  
835 variations at interannual to decadal time scales.

836

837

838 In agreement with previous studies, we show that the variability and trends in observed SSH over the  
839 last two decades are largely dominated by the steric effect except in shallow shelf seas where the mass  
840 effect is of the same order of magnitude as the steric effect and at high latitudes ( $>60^{\circ}\text{N}$  and  $<55^{\circ}\text{S}$ )  
841 where it dominates over the steric effect. The steric sea level signal is essentially due to temperature  
842 changes. Salinity changes play only a local role but this role can be sizeable in several regions in  
843 particular in the North Atlantic, in the Arctic and in the Southern ocean. The ECCO reanalysis reveals  
844 that the observed steric sea level variability and trends are essentially forced by surface wind stress  
845 anomalies in particular in the tropics where the SSH variability and trends are the most intense over  
846 the last two decades. The buoyancy forcing plays also a sizeable role but of smaller amplitude and  
847 more uniformly distributed. In the extra tropics where the wind stress forcing effect on SSH is smaller,  
848 the buoyancy forcing effect becomes significant and explain a sizeable part of the SSH variability and  
849 trends. In general, on average over the ocean, the buoyancy forcing effect on SSH trends is positive  
850 which reflects the penetration of heat into the ocean and the global warming of the ocean (Gregory et  
851 al. 2001, Suzuki and Ishii 2011). This result confirms that different reasons explain the global sea level  
852 rise from ocean warming and the regional sea level rise. While the buoyancy fluxes only are  
853 responsible for the total heat that enters the ocean and the associated global mean sea level rise, both  
854 ocean transport divergences caused by wind stress anomalies and the non-uniform buoyancy forcing  
855 (essentially at mid to high latitude) are responsible for the regional distribution of the heat within the  
856 ocean and thus for the regional sea level departures around the global mean.

857

858 Concerning the SST variability and trends over the last two decades, we show that the effect of local  
859 surface heat fluxes and ocean transport divergences locally are large and opposite to each other almost  
860 everywhere in the ocean. This finding primarily reflects the nature of the local steady state SST budget  
861 (in our case, integrated in time), wherein local heat forcing is compensated for and entirely cancelled  
862 out by the action of ocean transport processes. However, the cancellation between local forcing and

863 ocean transports is not exact, and the residual tendencies are what give rise to the resulting SST  
864 variability and trends. This result gives a unique perspective on the observed SST changes that, to our  
865 knowledge, has not been strongly emphasized elsewhere: almost everywhere, the mixed layer heat  
866 content is approximately in a local steady state balance with the ocean transport divergences effect  
867 almost entirely compensating the heat fluxes effect; in this context, observed changes in SST are  
868 interpreted as relatively small departures from this background steady state. In terms of interannual  
869 variability in SST, the effect of the surface heat flux forcing is larger than the effect of the ocean  
870 transport almost everywhere except in the tropical Pacific where both play a similar role. In terms of  
871 trends in SST, the picture is not so clear: both effect play a dominant role but in different regions.  
872

873 Thus we have provided complementary descriptions of mechanisms responsible for SSH and SST  
874 variability and trends. On the one hand, our exploration of SSH was more dynamical, focusing on the  
875 relevant forcing mechanisms (namely winds or buoyancy), without pinpointing particular processes  
876 (e.g., circulation versus mixing). On the other hand, our investigation of SST was more kinematical,  
877 targeting the underlying physical processes (e.g., advection or diffusion), without identifying the  
878 responsible drivers in all cases (i.e., wind or buoyancy). These separate analyses put on display the  
879 capabilities of the ocean-reanalysis machinery, and how such frameworks can help us understand, in a  
880 very detailed fashion, the nature of oceanic variability and change. In any case, bringing together the  
881 results of our SSH and SST analyses, we can see that SSH and SST bear some common variability.  
882 The main reason is that both SSH and SST variability show significant contributions from the surface  
883 heat fluxes forcing. This is evidenced by the high correlation between SST and buoyancy forced SSH  
884 almost everywhere except in a few regions where the freshwater effect on buoyancy forced SSH is  
885 dominant over the heat flux effect, like in the Pacific warm pool or in the Arctic region. In the tropics  
886 and in the upwelling regions, the wind stress forcing is also a cause of sizeable common variability in  
887 SST and SSH along with the surface heat flux forcing because wind stress variability in the tropics  
888 generate at the same time a zonal pressure gradient and a zonal tilt in the thermocline, which make  
889 respectively sea level and SST vary in phase in this region. SSH and SST bear also some common  
890 features in their trend patterns over the last two decades. However the picture is more complicated  
891 than for the interannual to decadal variability. This indicates that probably the relationship between  
892 SST and buoyancy-forced sea level or wind forced sea level is quite complex in general, and likely  
893 depends critically on time scale, among other factors.  
894

895 OC is a variable that is fundamentally different from SST and SSH because it depends on biological  
896 processes, which are forced by underlying physical processes. The concentration of chlorophyll, the  
897 main photosynthetic pigment present in all phytoplankton cells, can be used as key measure of the  
898 phytoplankton population and is a central variable in models used to estimate primary production (the  
899 rate at which phytoplankton produces organic matter from dissolved inorganic CO<sub>2</sub>). The growth of  
900 phytoplankton is primarily governed by the availability of light and nutrients, which in turn depend  
901 mostly on climate forcing conditions. Thus, OC variability and trends bear a strong dependence on  
902 climate fluctuations and can show significant correlations with SSH and SST. Although the correlation  
903 between OC and SSH tend to be weaker than between OC and SST, they display similar regional  
904 patterns throughout most of the global oceans. The tropics and subtropics show negative correlations  
905 because in this region phytoplankton growth is driven by nutrient availability in the mixed layer. Here,  
906 increases in nutrient availability are mainly driven by vertical entrainment (deeper mixed-layers)  
907 which influence SST and SSH (cooler waters entrained into the mixed-layer). At high-latitudes,  
908 phytoplankton growth is primarily driven by light availability. At these latitudes, changes in light at  
909 the sea surface are accompanied by warming of the waters which causes an increase in SST and  
910 buoyancy forced sea level and explain the positive correlation between OC, SSH and SST.  
911

912 In previous studies it has also been shown that phytoplankton can have a feedback effect on climate  
913 fluctuations. The absorption of solar irradiance by phytoplankton pigments tends to increase the  
914 heating rate of surface layers, stabilize the water column and reduce the depth of the mixed layer. As a  
915 result it changes the distribution of the heat in the water column. The regional effect on SST has been  
916 reported by several studies but no apparent effects on SSH have been reported yet. Given the  
917 significant influence of the absorption of solar irradiance by phytoplankton on the heating rate of the

918 upper layer, and the importance of the surface heat fluxes in the SSH variability, such effects of bio-  
919 optical heating on the SSH variability are probable. However, given the complexity of the feedback  
920 mechanisms controlling phytoplankton growth, evaluating the influence of bio-optical heating on SSH  
921 is not a trivial task, and to our knowledge, it has not been undertaken yet. In future studies, a possible  
922 starting point could be to examine the impact of phytoplankton on the total heat content of the water-  
923 column, and the mechanisms of heat redistribution, which would impact SSH.

924

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926

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