Skin Sea Surface Temperature schemes in coupled ocean-atmosphere modeling: the impact of chlorophyll-interactive e-folding depth.

Vincenzo de Toma[2], Daniele Ciani[1], Yassmin Hesham Essa[1,3,4], Chunxue Yang[1], Vincenzo Artale[1], Andrea Pisano[1], Davide Cavaliere[1], Rosalia Santoleri[1], and Andrea Storto[1]

[1] CNR-ISMAR, Consiglio Nazionale delle Ricerche, Istituto di Scienze Marine, via Fosso del Cavaliere 100, 00133 Rome, Italy.
[2] CNR-ISMAR, Consiglio Nazionale delle Ricerche, Istituto di Scienze Marine, Calata Porta Di Massa - Porto Di Napoli 80, 80133 Naples, Italy.
[3] GUF-IAU, Goethe University Frankfurt, Institut fuer Atmosphaere und Umwelt, Frankfurt, Germany
[4] ARC-CLAC, Agricultural Research Center, Central Laboratory for Agricultural Climate, Giza, Egypt

Correspondence to: Vincenzo de Toma (vincenzo.detoma@cnr.it)

Abstract. In this paper, we explore different prognostic methods to account for skin sea surface temperature diurnal variations in a coupled ocean-atmosphere regional model of the Mediterranean Sea. Our aim is to characterize the sensitivity of the considered methods with respect to the underlying assumption of how the solar radiation shapes the warm layer of the ocean. All existing methods truncate solar transmission coefficient at a ~~constant~~ warm layer reference depth which is constant in space and time; instead, we develop a new scheme where this latter is estimated from a chlorophyll dataset as the e-folding depth of solar transmission, therefore allowing it to vary in space and time depending on seawater's transparency conditions. ~~This allows spatial and temporal variations of the warm layer extent to depend on seawater transparency.~~ Comparison against satellite data shows that our new scheme, compared to the one already implemented within the ocean model, improves the spatially averaged diurnal signal, especially during winter, and the seasonally averaged one in spring, and autumn, while showing ~~with a~~ monthly, basin-wide averaged bias ~~on monthly scales year-round~~ smaller than 0.1 K year-round. In April, when most of the drifters' measurements are available, the new scheme mitigates the bias during nighttime, keeping it positive but smaller than 0.12 K during the rest of the monthly-averaged day. The new scheme implemented within the ocean model improves the old one by about 0.1 K, particularly during June. All the methods considered here showed differences with respect to objectively analyzed profiles confined between 0.5 K during winter and 1 K in summer for both the eastern and the western Mediterranean regions, especially over the uppermost 60 m. More in detail, the new scheme reduces the RMSE on the top 15 m in the central Mediterranean for summertime months, compared to the one already implemented one within the ocean model. Overall, the surface net total heat flux shows that the use of a skin SST parametrization brings the budget about 1.5 W/m^2 closer to zero on an annual basis, despite all simulations showing an annual net heat loss from the ocean to the atmosphere. Our "chlorophyll-interactive" method proved to be an effective enhancement of existing methods, its strength relying on an improved physical consistency with the solar extinction implemented in the ocean component.

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1 Introduction

Air-sea fluxes govern the energy exchange at the ocean-atmosphere interface. A reliable representation of the Sea Surface Temperature (SST) diurnal cycle, i.e. the typical SST oscillation/excursion between night and day mainly due to solar heating, is crucial to accurately estimate air-sea heat fluxes (Kawai and Wada, 2007, Soloviev and Lukas, 2013), whose direct measurement is very difficult. Indeed, diurnal warming events can often exceed 5 K depending on weather conditions (Soloviev and Lukas, 1997) and geographical location, typically at tropical and mid-latitudes but also occasionally at high latitudes (Karagali and Høyer, 2013). Large diurnal warming events can lead to changes in air-sea heat flux locally reaching up to $60 W/m^2$ (Fairall et al., 1996, Ward, 2006, Kawai and Wada, 2007, Marullo et al., 2010, Marullo et al., 2016) on a variety of scales, ranging from the short regional ocean weather ones to large seasonal or long-term ones.

Therefore, there is a wide interest in the development of models to accurately reconstruct SST diurnal variations in order to improve the representation of air-sea energy exchanges, especially, but not solely, within the coupled ocean-atmosphere modeling framework (Penny et al., 2019).

The net energy flux across the air-sea interface results from four contributions: the net solar radiation; latent and sensible heat fluxes, and the net thermal radiation. The last three contributions depend on SST and have a direct impact in determining ocean heat uptake or dynamical processes such as deep-water formation (Chen and Houze Jr., 1997). Ideally, the most accurate flux estimate would imply the knowledge of the temperature right at the atmosphere-ocean separation interface. From an observational point of view, the skin SST is the temperature immediately adjacent to the ocean surface (~10-20 microns depth) that is measurable, typically from infrared radiometers, and thus a key parameter to understand heat flux exchange (Minnet et al., 2019). Indeed, following what is measurable by current sensors, the GHRSST-PP (i.e. the Global ocean data assimilation experiment High Resolution SST Pilot Project) introduced the distinction between skin, sub-skin, depth, and foundation SST (Donlon et al., 2007), which can be respectively regarded as successive, better-to-worse approximations to the ideal target, i.e. SST right at the interface, which is actually impossible to measure. However, in most of the widely used ocean models and configurations, the too-coarse vertical resolution does not allow to direct modeling skin SST (the first model layer being only around 0.5 - 1 meter thick, e. g. the ocean model NEMO – see the sketch in figure 1). Therefore, one must use schemes to reconstruct skin SST variations. Sadly, the only thing one can be sure about is that in general no model will be able to perfectly reproduce skin SST diurnal variations, and there are different ways to approach this challenging problem, each one still with its own limitations (see Kawai and Wada, 2007 and references therein). Simplified ~~models~~ approaches widely employed in ocean and atmosphere state-of-the-art models parameterize the skin SST dynamics via the distinction of two main effects: the cool skin and the warm layer. Due to its interactions with the atmosphere, the temperature right at the interface of separation ~~ocean surface~~ is supposed to be almost anywhere and anytime ~~cooler~~ lower than ~~the the ones below~~ the temperature of the waters infinitesimally close to it, resulting in the ocean being covered with a thin cool skin layer~~.~~ ~~One~~one of the very first and simpler models

75 assumes this cool skin temperature difference as proportional to the ratio between heat fluxes and kinematic
76 stress (Saunders, 1967), via the Saunders' constant.

77 The cool skin effect is very important in obtaining accurate estimates of the latent and sensible heat flux,
78 especially because its consideration modifies specific humidity at the ocean surface, which is one of the factors
79 in the bulk formula. Indeed, latent and sensible heat fluxes are defined as the heat transfer across the
80 ocean/atmosphere interface due to turbulent air motions (the former including the one resulting from
81 condensation or evaporation). For example, a recent study in the South China Sea showed that during nighttime
82 the cool skin temperature difference is around 1 K, and there's currently a large uncertainty in the Saunders'
83 constant (Zhang et al., 2021~~0~~). A warm layer (in which diurnal warming effectively takes place) develops below
84 this cool skin, and its extent reaches a depth at which the penetration of solar radiation can be neglected (usually
85 fixed to 3m by most ~~of~~ existing parameterizations – see section 3.3 for more details). Diurnal warm layer
86 anomalies (which can sometimes exceed 3K) can potentially impact both the atmosphere and ocean mean state
87 on a variety of spatial (ranging from regional, basin-wide to global ones) and temporal scales (relevant for
88 weather or seasonal forecast to long-term climatic trends) (Donlon et al., 2007). The skin SST diurnal warming
89 amplitude increases under low surface winds (smaller than 2 m/s) and intense solar radiation (higher than
90 typical daily peaks, around 900 W/m^2) conditions, smaller in winter and at the poles than in summer and in
91 the tropics. The accuracy of skin SST models, and therefore their ability to reconstruct skin SST diurnal
92 variations is crucial especially in heat budget closure problems, which are still a subject of active debate
93 especially in climate change hot spot regions such as the Mediterranean domain (see Marullo et al., 2021 and
94 references therein). Skin SST schemes are also crucial for assimilating daytime SST data from satellite sensors
95 (Penny et al., 2019; Storto and Oddo, 2019, Jansen et al., 2019), with obvious impact on the accuracy of
96 numerical weather and ocean predictions; a correct account of skin SST diurnal variations in turn is crucial for
97 flux calculations, which is already a very delicate problem also from an instrumental point of view.

98 Within these prognostic schemes, seawater's transparency conditions (e.g., estimated using chlorophyll
99 concentration) have great implications in the way solar radiation is absorbed within the ocean's uppermost layer
100 (Morel and Antoine 1994). Ohlmann et al. 2000 quantified with the help of radiative transfer calculations effects
101 of physical and biological processes on solar radiation transmission, classifying as main factors chlorophyll
102 concentration, cloud cover and solar zenith angle. Ohlmann and Siegel 2000 and Lee et al. 2005 are further
103 examples of how radiative transfer models are used to develop solar transmission parameterization which is fit
104 to the sum of exponentials (the number of terms in the sum depending on the variable which has been
105 considered). To the best of our knowledge, these ideas have not been implemented nor tested within the
106 prognostic scheme for skin SST present in the ocean model NEMO, which just relies on chlorophyll-calibrated
107 coefficients though (Gentemann et al 2009).

108 Our main aim here is therefore to improve existing skin SST prognostic schemes, investigating the impact of
109 variable ~~accounting for~~ seawater's transparency conditions in modeling solar radiation extinction in the upper
110 ocean. The use of chlorophyll concentration as a proxy for seawater's transparency is not new. In fact, given

its covariance with Secchi disk depth (estimated from reflectance at various wavelength), it has been often applied by the ocean color community to study the dynamics of oligotrophic gyres (Leonelli et al., 2022 and references therein). The paper is structured as follows: after this introduction, we describe the data and coupled modeling system in section 2. The mathematical context in which we developed our new method, whose novelty stands in allowing the warm layer's extent to vary in space and time according to a chlorophyll-concentration climatology follows in section 3. In section 4 we present results, discussing them and drawing conclusions in section 5.

2   Data and Modeling System

We describe here the data and the coupled regional modeling system used in this study. Our description here is functional to the scope of this paper, and far from a complete depiction of each dataset. We refer readers to the documentation and relevant literature for detailed information on each dataset and model.

Operational MED DOISST within CMEMSWe describe here the data and the coupled regional modeling system used in this study. Our description here is functional to the scope of this paper, and far from a complete depiction of each dataset. We redirect the documentation and the appropriate literature describing each data and model in depth.

2.1   Operational MED DOISST within CMEMS Operational MED DOISST within CMEMS

The MEDiterranean Diurnal Optimally Interpolated Sea Surface Temperature (MED DOISST) product, operationally distributed and freely available within the Copernicus Marine Environmental Service (CMEMS) provides gap-free (L4) hourly mean maps of sub-skin SST at $1/16^o$ horizontal resolution over the Mediterranean domain, covering from 2019 to present. Sub-skin SST is defined as the temperature at the base of the cool skin layer, typically sensed by microwave radiometers, and representative of a depth of few millimeters from the ocean's surface (Minnet et al., 2019).

This product combines satellite data acquired from the Spinning Enhanced Visible and InfraRed Imager (SEVIRI) and model data from the Mediterranean Forecasting System (MedFS), respectively used as observations and first guess for an optimal interpolation, giving a L4 field representative of subskin SST (see Pisano et al., 2022 and references therein). In all diagnostics involving these data (and presented in the following sections), regions where the percentage of model datavalid SEVIRI measurements is higher lower than 50% have been masked out both in CMEMS MED DOISST and our experiments.

2.12.2 iQuam in-situ data

SST from drifter data were used for validation purposes and acquired from the iQuam (In situ SST Quality Monitor) archive (Xu and Ignatov, 2014). The iQuam provides high-quality and quality controlled (QC) in-situ SST data collected from various platforms, such as drifters, Argo Floats, ships, tropical and coastal moored

145 buoys. iQuam SST data are also provided along with quality level flags ranging from 0 to 5, with 5
146 corresponding to the highest quality level (Xu and Ignatov, 2014). For this study, SST with quality level equal
147 five were selected from drifters only, since they provide the temperature measurement closest to the surface
148 (compared to the other available instruments), ranging between 20-30cm (depending on the drifter type).

149

150 Additionally, we interpolated model outputs on drifters' location in time and space. Table S1 ~~resumes~~ outlines
151 the number of available measurements for each given month and hour of the day. A total number of 555919
152 records were available after the quality flag and platform selection, with the month of April being the most
153 populated one, with 222996 measurements, and 10361 measurements at 9:00 am.

154

155 **2.2 2.3 EN4 objective analysis**

156 EN4, the quality controlled subsurface ocean temperature and salinity profiles and objective analyses, were
157 used to assess the impact on the temperature vertical profiles. To facilitate the comparison, we made use of the
158 objective analyses after bias corrections of Expendable Bathythermograph (XBT) calibrations (Gouretski and
159 Reseghetti, 2010, Gouretski and Cheng, 2020), which give a gridded version of the dataset on a 1-degree regular
160 grid. In the comparison, model outputs were interpolated on this grid.

161 **2.3 2.4 Mediterranean Chlorophyll concentration**

162 Chlorophyll data were used to estimate e-folding depths' seasonality (see Methods, Section 3). These data
163 are a daily interpolation at 0.3 km horizontal resolution over the Mediterranean domain, and result from a
164 merging between multiple sensors (MERIS - MEdium Resolution Imaging Spectrometer from ESA, SeaWiFS
165 - Sea-viewing Wide Field-of-view Sensor and MODIS - Moderate Resolution Imaging Spectroradiometer from
166 NASA, VIIRS - Visible Infrared Imager Radiometer Suite from NOAA, and most recently the Copernicus
167 Sentinel 3A OLCI - Ocean and Land Colour Instrument), as detailed in the product description (see Volpe et
168 al., 2019 and references therein for further details).

169

170 **2.4 2.5 ECMWF Atmospheric Reanalysis - ERA5**

171 We used heat fluxes (net solar radiation, latent and sensible heat fluxes, net thermal radiation) from ERA5
172 at 0.25° horizontal and hourly temporal resolution (Hersbach et al., 2020) as reference for comparing
173 performances across simulations with different skin SST schemes. Despite their possible biases in air-sea fluxes,
174 atmospheric reanalyses ~~at day~~ today are still widely thought to provide the best gap-free and dynamically
175 consistent reconstructions of the atmosphere system (Valdivieso et al., 2017, Storto et al., 2019).

176 **2.5 2.6 Mixed Layer Depth 1969-2013 Climatology**

177

178   Data from a mixed layer depth (MLD) climatology was used to test to what extent our modified scheme
179   correctly represents the seasonality of the mixed layer.
180   This monthly gridded climatology was produced using MBT, XBT, Profiling floats, Gliders, and ship-based
181   CTD (Conductivity, Temperature, Depth) data from different databases and carried out in the Mediterranean
182   Sea between 1969 and 2013. As for the model outputs, MLD is calculated with a $\Delta T = 0.1°C$ criterion relative
183   to 10m reference level on individual profiles (Houpert et al., 2015a, Houpert et al., 2015b).

184

185   **2.62.7 ISMAR Mediterranean Earth System Model (MESMAR)**

186   MESMAR is a newly developed coupled regional modeling framework for the Mediterranean region (Storto
187   et al., 2023). MESMAR includes the following components:

188   • the ocean model: NEMO v4.0.7, with horizontal resolution of about 7 *km*, 72 unevenly spaced vertical
189   levels (the first and the last levels being respectively about 0.5m and 200m thick) and a timestep of 7.5
190   minutes (NEMO System Team, 2019);

191   • the atmosphere model: WRF v4.3.3, with 41 vertical hybrid levels and horizontal resolution of about 15
192   *km*, covering the European branch of the international Coordinated Downscaling Experiment (EURO-
193   CORDEX) domain, and a timestep of 1 minute (Skamarock et al., 2019);

194   • an interactive runoff model: HD v5.0.1, with a timestep of 30 minutes and 1/12° degree horizontal
195   resolution over Europe (Hagemann et al., 2020);

196   • the coupler: OASIS3-MCT, coupling the three models with a coupling frequency of 30 minutes, and
197   using the SCRIP library to interpolate fields between different model grids (Craig et al., 2017) ;

198   We report in figure 2↑ a graphical summary of different grids. Further details of its implementation, tuning,
199   and performances are described in (Storto et al., 2023).

200

201   **3   Methods**

202   Many schemes to reconstruct the skin SST diurnal variations rely on the existence of a cool skin and a warm
203   layer, respectively in the upper micrometers and few meters of the ocean, whose dynamics strongly depends on
204   wind conditions and solar radiation extinction within the upper ocean. To explain the rationale behind the
205   developments in our new method, we need to recap here some elements of this theory, which is mostly based
206   on Zeng and Beljaars, 2005 (named ZB05 hereafter) work.
207   We start from the one-dimensional heat transfer equation in the ocean:

208
$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(K_w + k_w\right)\frac{\partial T}{\partial z} + \frac{1}{\rho_w c_w}\frac{\partial R}{\partial z} \tag{1}$$

209 in which the subscript w refers to water properties, T is seawater temperature (K), K_w $(m^2 s^{-1})$ is the turbulent

210 diffusion coefficient, k_w $(m^2 s^{-1})$ is the molecular thermal conductivity, ρ_w $(Kg\,m^{-3})$, c_w $(J\,Kg^{-1}K^{-1})$ are respectively

211 seawater density and heat capacity per unit volume, R $(W\,m^{-2})$ is the net solar radiation flux, defined as positive

212 downward.

3.1 Cool Skin

214 We assume that there exists an oceanic molecular sublayer of depth δ, where K_w is negligible, and

215 temperature can be assumed constant in time, since it is always cooler than temperature of the underlying

216 seawater (Donlon et al., 2007, Zeng and Beljaars, 2005). Then integration of eqn. (1) gives, $\forall z \in [0, -\delta]$

$$k_w \frac{\partial T}{\partial z} + \frac{1}{\rho_w c_w}[R(z) - R_s] - k_w \overbrace{\cancel{\frac{\partial^2 T}{\partial z^2}}}^{\mathcal{O}(z^2)} = const,$$

217 (2)

218 where R_s is the net solar radiation at the surface (constant, open-ocean albedo, since the Mediterranean Sea is

219 an ice-free basin), assuming this constant to be the top boundary condition at $z = 0$:

$$\rho_w c_w k_w \frac{\partial T}{\partial z}\bigg|_{z=0} = Q = LH + SH + LW,$$

220 (3)

221 in which LH, SH, LW are respectively the surface fluxes of latent, sensible heat and net

222 long wave radiation.

223 Thus, eqn. (2) can be rewritten as

$$\rho_w c_w k_w \frac{\partial T}{\partial z} = Q + R_s - R(z).$$

224 (4)

225 Making a further integration we get the cool skin temperature difference:

$$T_s - T_{-\delta} = \frac{\delta}{\rho_w c_w k_w} \left(Q + f_s R_s\right),$$

226 (5)

227 where T_s and $T_{-\delta}$ are respectively the temperature at the upper (air-sea interface) and lower limits of the cool

228 skin layer, while f_s is the fraction of solar radiation absorbed in this layer:

229

$$f_s = \frac{1}{\delta} \int_{-\delta}^{0} \left(1 - \frac{R(z)}{R_s}\right) dz,$$

230

231 which depends on the way radiation gets absorbed within the cool skin. Being time-independent, the cool

232 skin temperature difference is a diagnostic variable in the scheme.

233 Eq. (5) is analogous to Saunders' model. Indeed, Saunders, 1967 was one of the first to construct a theory for

234 the ocean "cool skin" effect (already known from decades at those times), i.e. the observed temperature at the

235 air-sea interface is generally cooler than the temperature of the water at about 10 cm depth, especially during

236 nighttime. This effect takes place mainly because of the transfer of energy between the ocean and the

atmosphere, realized via heat loss and momentum transfers (wind stress). In a nutshell, at the end of its derivation (Saunders, 1967), he obtains the following expression for the temperature difference across the cool skin, ΔT_c (K):

$$\Delta T_c = \lambda \frac{Q v_w}{k_w (\tau/\rho_w)^{1/2}},\qquad(6)$$

where λ is the Saunders' proportionality constant, Q $(W\ m^{-2})$ has already been defined above, τ/ρ_w $(m^2\ s^{-2})$ is the kinematic stress (ratio between wind stress module and seawater density), and $v_w,$ $(m^2 s^{-1})$ k_w $(m^2 s^{-1})$ are respectively the kinematic viscosity and thermal conductivity of seawater. Saunders' formulation was originally conceived for low, nonzero wind conditions and neglecting the effect of solar radiation (which however recognized its role and added a discussion on how to account for it in the model only at the end of his paper). As noticed by Fairall et al. 1996, Artale et al., 2002 (named A02 hereafter), with a constant λ, eqn. (6) becomes problematic in limiting cases of low and very high wind speeds (greater than 7 m/s), because the wind stress in the denominator limits its validity. Thus, they A02 proposed to include a wind dependence in Saunders' constant, in order to still have a finite, nonzero cool skin to bulk temperature difference even when the wind speed goes to zero or becomes very high.

This scheme has proven to have good performances compared to other schemes also on a mooring site in the Pacific Ocean (Tu and Tsuang, 2005).

3.2 Warm Layer

Below the skin layer, turbulent transfer is much more effective, and k_w can be neglected in favor of K_w. Integrating eqn. (1) within the $[-d, -\delta]$ layer, we get:

$$\frac{\partial}{\partial t} \int_{-d}^{-\delta} T\ dz = \frac{Q + R_s - R(-d)}{\rho_w c_w} - K_w \frac{\partial T}{\partial z}\bigg|_{z=-d},\qquad(7)$$

where d is a reference depth which can be assumed as the depth at which the diurnal cycle can be omitted.

The turbulent diffusion coefficient can be expressed as (Large et al., 1994):

$$K_w = k u_{*w} (-z) / \phi_t \left(\frac{-z}{L}\right),\qquad(8)$$

in which $k = 0.4$ is the Von Karman constant, z is negative in the ocean, u_{*w} is the friction velocity in the water (this being the air friction velocity multiplied by the square root of air to sea density ratio), and the stability function ϕ_t discriminates between a stable and an unstable regime, depending on the sign of its argument, which is the ratio of the vertical coordinate to the Monin Obukhov length L: positive for the stable and negative for the unstable one. Assuming z to be negative in the ocean, the change of sign entirely depends on the Monin Obukhov length, which is a length characterizing the prevalence of buoyancy variations induced turbulence

268 over the one generated by wind shear effects. This in turn is strongly dependent on the sign of the net heat flux
269 Q. If $Q>0$, i.e. the ocean gains heat from the atmosphere, and we have the stable regime: the diffusion
270 coefficients decrease with increasing depth, favoring the downward heat transfer within the water column. The
271 opposite case, which favors transfer of heat from the ocean to the atmosphere, can be modeled in different ways
272 (see While et al., 2017 and references therein).

273 Assuming a temperature ~~of~~ dependence, for $d \gg \delta$ of the form

$$T = T_{-\delta} - \left[\frac{z + \delta}{-d + \delta} \right]^{\nu} (T_{-\delta} - T_{-d}), \quad \nu \text{ empirical parameter} \tag{9}$$

275 eqn. (7) simplifies to

$$\frac{\partial}{\partial t} (T_{-\delta} - T_{-d}) = \frac{Q + R_s - R(-d)}{d \rho_w c_w} \frac{\nu + 1}{\nu} - \frac{(\nu + 1) k u_{*w}}{d \phi_t (d/L)} (T_{-\delta} - T_{-d}) \tag{10}$$

277 In ZB05 scheme (Zeng and Beljaars, 2005), eqs. (5, 10) are the coupled equations for the cool skin (diagnostic
278 part) and warm layer (prognostic part) respectively. Being time dependent, the determination of the warm layer
279 temperature difference at time t requires the knowledge of the one at the previous time step, and thus is the
280 prognostic variable in the scheme. Assumptions on the fraction of solar radiation within ~~this layer~~ the warm
281 layer and the cool skin depth usually follow Fairall et al., 1996 parameterization, whose detail are given in the
282 Supplementary Material section.

283

3.3 Solar transmission expression

285 The expression of the solar transmission in Zeng and Beljaars, 2005 is

$$\frac{R(-d)}{R_s} = \sum_{i=1}^{3} a_i e^{-d b_i}, \quad \begin{array}{l} (a_1, a_2, a_3) = (0.28, 0.27, 0.45), \\ (b_1, b_2, b_3) = (71.5, 2.8, 0.07) m^{-1}, \end{array}$$

287 following Soloviev formulation (Soloviev, 1982) (S82 in the following), which is very widely used in
288 atmosphere models (such as WRF, Skamarock et al., 2019).

289

290 So far this is not the only possibility: a formulation with 61 coefficients has been developed by Jerlov, 1968,
291 which is based on different water types classified based on chlorophyll concentration and particulates, for light
292 in the visible spectrum.

293

294 ~~A formulation~~Formulations with 9 coefficients (reported in Table 2) ~~has~~ have been proposed to include such
295 effects: for example ~~(~~Soloviev and Schlussel, 1996 use a different coefficient for the first term depending on
296 Jerlov's optical water type, while Gentemann et al., 2009 include solar angle in the parameterization, keeping
297 the value of the of the first coefficient as in the case of pure-water~~, Gentemann et al., 2009)~~. Without knowing
298 what the Jerlov water type is, what is currently implemented in NEMO is to take b_1 as ~~the first of them~~

299 ~~accounting for mean properties~~ the average between coefficients for~~of~~ I, IA, IB, II and III Jerlov's optical water

300 types. This formulation is widely employed in ocean models (such as in the optional skin SST routine of NEMO,

301 see While et al., 2017), with the reference depth d fixed to 3 m. So, the solar transmission ~~coefficient~~ follows

302 as:

$$\frac{R(-d)}{R_s} = \sum_{i=1}^{9} a_i e^{-db_i}$$
(13)

304 Ideally, one would like to have a reference depth representative of the one at which the transmission of solar

305 radiation is negligible, and if we take it as the depth at which transmission drops ~~by~~ to 1/e from its surface value,

306 we get a value which can be different from $d = 3$ m, as we can see from figure 3~~2~~a. Allowing for a realistic time

307 and space varying value of d represents the main novelty of our work.

308

309 From this viewpoint, choosing a value of $d = 3$ m while using the solar extinction formulation as in Soloviev,

310 1982 or Soloviev and Schlussel, 1996 would lead to underestimat~~ing~~e the penetration of solar radiation into the

311 warm layer. Another possibility, which constitutes our modification to the scheme already implemented in

312 NEMO ~~as in the case of the NEMO module for radiation calculations (Jerlov, 1968, Morel et al., 1989,~~

313 ~~Lengaigne et al., 2007)~~, is to reconstruct a chlorophyll profile from its surface values following what is already

314 implemented in the ~~as in the case of the~~ NEMO module for radiation calculations (Jerlov, 1968, Morel et al.,

315 1989, Lengaigne et al., 2007), and employ an R-G-B+Chl-a scheme to calculate radiation as a function of depth.

316 Then, ~~From~~from eqn. (13) with only 4 terms (one for chlorophyll, and three for R-G-B, expressed in lookup

317 tables), one can numerically derive the warm layer reference depth as the e-folding depth~~ of the light extinction~~

318 profile (see Fortran source files in the Zenodo repository, de Toma (2024)).~~using chlorophyll variations and the~~

319 ~~R-G-B light extinction coefficients taken from lookup tables in the source code~~.

320

321 This would give a constant transmission throughout the basin, but with a spatially and temporally varying e-

322 folding depth and defines our new prognostic scheme for skin SST warm layer calculation, thus embedding in

323 it the ocean color information coming from Chl-a. Everything else is left unchanged, both the refinements of

324 Takaya et al., 2010, which include the effect of Langmuir circulation and a modification of the Monin-Obukhov

325 similarity function under stable conditions (T10 hereafter), and the A02 model for cool skin, which has been

326 demonstrated to improve the scheme respectively under wavy and windy conditions.

3.3.1 E-folding depth estimates

328 Mediterranean Chlorophyll climatology data (see section 2.4) were re-gridded onto a 0.25 ° regular

329 longitude/latitude grid, and tabulated coefficients within NEMO were used to retrieve the transmission,

330 accounting for chlorophyll variations. E-folding depths then can be estimated as the depth at which transmission

331 drops ~~by~~ to 1/e from its surface value. It can be noticed from figure 3~~2~~b that also the e-folding depth varies with

332 seasonality, with typical values ranging from about 3 to 4.5 meters. This is the central point of our modification

333 to the prognostic scheme. In out setup we extracted pixelwise and at each time step of the NEMO model the e-
334 folding depth used within the prognostic scheme.

3.4 Overview of the simulations performed.

336 With the coupled ocean-atmosphere regional system we performed a set of four simulations, forced by
337 ERA5 in the atmosphere and ORAS5 (Zuo et al., 2018) in the ocean and covering three years (from 2019 to
338 2021), with hourly outputs (a synthesis is provided in Table 1). In cases where a skin SST scheme is active, we
339 substitute the SST, i.e. temperature on the first NEMO level, with the skin SST coming out from the scheme:

1. a control run, in which no skin SST prognostic scheme is activated, therefore the diurnal SST variations
    in the uppermost ocean layer (0.5 m thick) only come from the variability represented by the ocean model
    at about 0.5 m of depth, considering also the 0.5 hours frequency of the coupling. We will refer to this
    experiment in the following as *ctrlnoskin*;

2. a run in which the ZB05 scheme in WRF (Zeng and Beljaars, 2005) is active – we shall refer to this case
    in the following as *wrfskin*;

3. a run in which the existing scheme within NEMO, which employ the 9-coefficient parameterization for
    light extinction coefficients (Gentemann et al., 2009 – G09 hereafter), the scheme for the cool skin as
    modified in A02, and refinements of the stability function, in the warm layer formulation as in T10 – we
    shall refer to this as the *nemoskwrite* case;

4. a fourth simulation in which we modified the reference depth for the basis of the warm layer from $z = 3$
    m, to an e-folding depth (i.e. the depth at which radiation gets diminished by $1/e$ from its surface value),
    which is allowed to vary temporally and spatially because it is estimated from R-G-B light extinction
    coefficients and chlorophyll concentration (see section 3) below. We will refer to it as *modradnemo*,
    being the experiment where our modification to the skin SST scheme is implemented and tested.

355 The reason behind the choice of the above mentioned period of three years 2019-2021 is twofold: firstly, it
356 allows a validation against all the measurements from different data sources (satellite, drifters and objectively
357 analyzed profiles), and secondly, it is a good trade-off between the needs of keeping a reasonable computational
358 load, data volume for the analysis, and guarantees a minimal robustness of our finding, compared to a simulation
359 which covers just one year. However, we do not discard the possibility to extend the time coverage in our plans
360 for future works.

4 Results

362 In this section, we present skill scores against satellite, drifters and temperature profiles data (see section 2)
363 from the set of the simulations performed, aimed at characterizing the impacts of our modified skin SST scheme.
364 we compare simulations outputs with data from different sources (see section 2), to assess methods

365 ~~performances and impacts of our modifications.~~ Since we are mainly acting to improve skin SST diurnal

366 variations reconstruction in the ocean component, the main focus is on the difference between the nemoskwrite

367 and modradnemo, while the ctrlnoskin and wrfskin ones are included as further reference elements (the latter

368 being not directly comparable because the atmospheric model sees the ocean foundation SST and employ the

369 scheme just to diagnose the skinSST).

370

4.1 Comparison with CMEMS MED DOISST

372 We calculated the mean diurnal warming amplitude in each season as the seasonally averaged diurnal

373 warming amplitudes (diurnal warming amplitude being defined for each day as the difference between daytime

374 maximum and nighttime minimum of SST), which can be cast into the following equation:

$$\langle \mathrm{DWA} \rangle_{\mathrm{seas}} = \frac{1}{N_{\mathrm{seas}}} \sum_{i=1}^{N_{\mathrm{seas}}} \left\{ \max_{h_i \in [10:00,18:00]} \mathrm{SST}\,(h_i) - \min_{h_i \in [00:00,06:00]} \mathrm{SST}\,(h_i) \right\},$$

$$\langle \mathrm{DWA} \rangle_{\mathrm{seas}} = \frac{1}{N_{\mathrm{seas}}} \sum_{i=0}^{N_{\mathrm{seas}}} \left\{ \max_{h_i \in [10:00,18:00]} - \min_{h_i \in [00:00,06:00]} \right\} \mathrm{SST}\,(h_i),$$

$$(14)$$

377 where seas = DJF, JJA, MAM, SON is the given season, N_{seas} ~~Nseas~~ is the number of days in that particular

378 season and h_i ~~h_i~~ is~~is~~ the local time in hours for any given day.

379 Seasonally averaged diurnal warming amplitudes are shown in figure 4~~3~~. On average, the maximum amplitude

380 is reached in summer, with the wrfskin simulation peaking at about 3 K, thus overestimating the mean diurnal

381 cycle compared to CMEMS MED DOISST (the monthly biases with respect to CMEMS foundation SST both

382 in the western and the eastern part of the Mediterranean Sea stay below 1 K year-round for ~~every~~ all of the

383 simulations performed – see figure S1 in Supplementary Materials). The nemoskwrite simulation yields a

384 pattern very similar to CMEMS MED DOISST in summer, but underestimates the signal in the remaining

385 seasons. Outside the Summer season, our modifications yield a slight improvement (see modradnemo, last row

386 of figure 4~~3~~). As expected, the control run in which no skin SST method is active, generally underestimates the

387 diurnal signal everywhere. Compared to nemoskwrite, the modradnemo simulation improves JJA mean diurnal

388 warming amplitude, especially over the Southern Mediterranean Sea, while in central and Northern part of the

389 basin tends to overestimate the signal by about 0.5 K with respect to CMEMS-DOI data. Furthermore, a general

390 underestimation is present also in DJF, with the modradnemo simulation showing the smallest differences with

391 respect to CMEMS-DOI data.

392 The spatial average over the whole Mediterranean domain is shown i~~i~~n figure 5~~4~~, confirming the general

393 underestimation of the control run and the overestimation of the wrfskin ~~(ZB05 scheme)~~ and modradnemo in

394 all seasons except winter.

Computing spatial averagesSpatially averaging highlights that modradnemo slightly improves the mean diurnal warming amplitude signal during wintertime, our modification brings improvement, especially during wintertime, while in all the other seasons the best agreement is gained by using the nemoskwrite setup (ZB05 with T10 and A02 modifications), at least according to the verification against CMEMS MED DOISST.

On a monthly timescale, figure 65 confirms that the control simulation tends generally to have a negative bias of the diurnal amplitude, for the whole simulated period. The wrfskin (ZB05 scheme) shows a warm bias during summertime months, shown just as a reference. The comparison between nemoskwrite (ZB05+A02+T10) and modradnemo (chl e-folding depth) shows improvement of our scheme (modradnemo) over the old one (nemoskwrite) especially in May, but not in April, June, July, August and September, despite in the rest of the period the amplitude of the bias is slightly reduced.

4.2   Comparison with iQuam Star HR-Drifters

The bias with respect to drifter measurements averaged over drifters positions as a function of the month and time of the day is shown in figure 76. All the schemes present a systematic cool bias in autumn (SON) for most of the hours of the day. During April and June, the modradnemo simulation significantly reduced the warm bias with respect to observations, compared to the nemoskwrite case, keeping it however generally positive. This is quite reasonable, since drifters measurement can be thought representative of a depth which can also be can be also below the subskin level (typically of the order of some centimeters). Consistently with figure 65, the wrfskin has a larger positive bias than modradnemo in June.

Further, as shown by figure 87, the bias between CMEMS MED DOISST and drifters is generally positive anytime except in late spring/summer and autumn during nighttime. This pattern arises because of the composite effect of having a temperature representative of the subskin level where and when there are data from radiometers, and a temperature of about 1 m depth from the MEDFS system as first guess of the optimal interpolation over cloudy regions (Pisano et al., 2022). However, the modified scheme significantly reduces the difference, yielding a bias closer to the one of CMEMS MED DOISST with respect to drifters, especially during April, which is the month in which the number of observations from drifters is definitely larger.

4.3   Comparison with EN4 objective analysis

Bias corrected vertical profiles gathered in an objective analysis were used to assess differences across schemes along the water column. To summarize we report here only a macro subdivision into the eastern and the western Mediterranean Sea, respectively in figures 98, 109. Model outputs were remapped on the same vertical and horizontal grid. Looking at the mean profile averaged over all grid points in the given area, the agreement is better for all simulations during summertime months, both for the eastern and the western region (see figs. 98c, 109c), showing in particular that the modradnemo simulation outperforms the nemoskwrite one. This is also true for the wintertime season in the eastern Mediterranean (see fig.98b). On the other hand, in the

western Mediterranean all simulations tend to overestimate the signal, with our modified scheme doing a better job with respect to the nemoskwrite case, with an average profile which is about 0.4°C closer to the EN4 profile. However, below about 80 m depth differences across schemes vanish.

Looking in more detail at the RMSE on the top 15 m depth between each simulation and EN4 as a function of the month and more detailed region subdomains shown in figure 110a, we can see how in general all simulations present the same pattern for the region outside of Gibraltar Strait, which can be thought an effect related to the presence of the relaxation to horizontal boundary conditions, while for all the remaining regions and months the control run, the wrfskin and the modradnemo present a similar pattern, with the modradnemo reducing the RMSE in most of the regions and for most of the months, especially with respect to nemoskwrite, and this is particularly true over the central Mediterranean Sea, in regions like Thyrrenian and Adriatic Seas.

4.4 Heat fluxes and vertical propagation

In this section we aim to characterize the differences of each scheme with respect to the control simulation. We do this by specifically looking at the seasonality of Mixed Layer Depth (MLD), vertical profiles of temperature in specific months and regions, and via the comparison of the net surface heat fluxes over the whole Mediterranean Sea.

Compared to the Mixed Layer Depth climatology from 1969 to 2013 (Houpert et al., 2015a, Houpert et al., 2015b, section 2.7), all of the tested schemes seemsseem to have a similar impact on Mixed Layer Depth's seasonality, with larger differences with respect climatological values being mostly located in the Eastern Mediterranean Sea and during wintertime/spring (Figure 121). It may seem from this picture that there's not such a huge change to prefer one method over the other considered in this paper, and this may also be because of the short period simulated (2019-2021). Figure 132 show how our modified scheme allows more (less) vertical propagation of the diurnal signal during summer (winter) with respect to schemes with constant e-folding depth in all central regions of the Mediterranean domain (regions 2, 3, 4 as defined in figure 110a), when all of them are referenced to the control simulation temperature daily minimum.
Indeed, from figure 132b, we can see that when all the temperature profiles for each simulation are referenced to the ctrlnoskin daily minimum, there is a much wider diurnal warming signal for most of all the considered depths level, with modradnemo representing an intermediate situation between the wrfskin and the nemoskwrite simulation. This is probably due to the inclusion of chlorophyll-interactive variations, which allow for a better representation of the variability of the mixed layer dynamics.
Estimates of the mean Mediterranean heat exchange between ocean and atmosphere based on previous studies range from -11 to +22 W/m^2, with an evident dominance of negative estimates, i.e., heat loss from the ocean to the atmosphere (Jordà et al., 2017, Pettenuzzo et al., 2010). Some other studies suggest that the Mediterranean heat budget is close to a neutral value, -1 W/m^2 (Ruiz et al., 2008) or +1 W/m^2 (Criado et. al., 2012). Many factors can contribute to such wide variability among different estimates, such as differences in the

parameterizations employed, initial and boundary conditions, and the way the physical processes, especially through the Strait of Gibraltar are modeled (Macdonalds et al., 1994, Gonzales, 2023).

As shown by table 3, all simulations on an annual basis give a negative, non-closed balance for the net surface heat flux, and modifications to include skinSST, performing very similarly one to another, bring the budget by 1.5 W/m^2 closer to zero, while ERA5 data show a positive net surface heat flux close to 5 W/m^2. However, all estimates fall into the (large uncertain) literature-based estimates. On seasonal timescales, the inclusion of skinSST diurnal variations has the following effects:

- less net heat loss to the atmosphere during wintertime with respect to the control run (wrfskin differing from the ctrlnoskin by about 6W/m^2, while nemoskwrite and modradnemo having a similar impact, with a difference of about 4W/m^2 with respect to the control run);

- in springtime, all simulations show a positive imbalance, with the highest difference with respect to the control run of about 1 W/m^2 in the modradnemo simulation;

- during summer, our modified scheme brings on average about 3 W/m^2 more than the control simulation into the basin, yielding an estimate which is closer to ERA5;

- in autumn, our scheme cools down more than the control (about 2 W/m^2), being the farthest simulation from ERA5 estimate, while traditional schemes tend to have a less negative net heat input.

All seasons except spring show larger difference with respect to ERA5 fluxes, with underestimation in summer, and overestimation during winter and autumn, resulting in a bias of about 10 W/m^2 with respect to the net heat flux annual budget in ERA5.

5 Summary and Conclusions

In this paper we studied the sensitivity of a regional coupled ocean-atmosphere-hydrological discharge regional model on the Mediterranean Sea to prognostic schemes for skin sea surface temperature. Specifically, we developed a new scheme which allows for spatial and temporal variations of the warm layer's extent according to seawater's transparency conditions. This is possible by using tabulated solar extinction coefficients already used in the ocean model, and inverting the functional form which determines how the solar radiation varies along the vertical direction to find the depth at which this latter drops by to 1/e from its surface value.

We simulated the period 2019-2021, analyzing hourly model outputs, and comparing aggregated results with satellite, objectively analyzed and drifters data. Overall, the comparison with data shows that the new scheme improves what is already implemented in NEMO, e.g. mean diurnal warming amplitudes are closer to satellite observations in winter, spring and autumn, not being much worse than other existing schemes in summer, at least looking at maps of mean diurnal warming amplitude grouped by seasons. Looking to the typical

499 temperature profile in both the eastern and the western Mediterranean Sea, non-negligible differences across
500 schemes stay confined in the topmost 20m (100m) of depth during summertime (wintertime). Regionally,
501 typical profiles are warmer than EN4 observation year-round for western regions (regions -1,1,2) especially in
502 winter, while regions in the east show a smaller RMSE in the topmost meters for basically all the regions and
503 months when comparing modradnemo to nemoskwrite. The Adriatic Sea has a systematically higher RMSE
504 with respect to EN4 in all the tested methods, for the whole period considered. In the central regions, the new
505 scheme penetrates temperature anomalies more (less) during summer (winter) months, having a less intense
506 mean diurnal warming amplitude signal in summer, especially over the upper few meters (the converse holds
507 for wintertime values). Therefore, with respect to the ctrlnoskin simulation, nemoskwrite shows the coldest
508 signal, the wrskin the hottest, and our modification modradnemo constitutes the middle situation, with milder
509 summer and winter than the control run. Our interpretation is that within modradnemo, the chl-interactive e-
510 folding depths allow, where and when necessary, the warm layer to become a little deeper than in the already
511 existing scheme (nemoskwrite), depending on chl-variations. For these space-time points, solar penetration is
512 increased and so it tends to make warmer the warm layer. Therefore, future research efforts should be devoted
513 to the better characterization of this aspect, especially to understand if the modified vertical penetration of heat
514 has some particular effect on the dynamics of the mixed layer (see Song and Yu, 2017 and references therein).
515 Furthermore, testing the implementation within NEMO of more sophisticated radiative transfer models (such
516 as the one of Ohlmann and Siegel 2000), or the development of deep learning based parameterizations are
517 underway as future research efforts. On a long-term perspective, the method needs to be tested also in other
518 areas and for longer periods, which can increase the results' certainty and allow for usage in investigating
519 impacts on relevant climate large-scale phenomena, where the role of an improved diurnal warming signal
520 could be more relevant (Bernie et al., 2007, Bernie et al., 2008). These includes phenomena and physical
521 processes such as propagation of Marine Heat Waves (MHW) or deep water formation and deep convective
522 events.
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524

Code and data availability

The NEMO ocean model code (v4.0.7) is available at https://forge.ipsl.jussieu.fr/nemo/wiki.

The WRF atmospheric model code (v4.3.3) is available at https://github.com/wrf-model/WRF.

The HD hydrological discharge model (v5.1) is available at https://zenodo.org/record/5707587#.Y-0VQ3bMKUk.

The frozen version of the MESMARv1 code used in this manuscript is available at: https://doi.org/10.5281/zenodo.7898938.

CMEMS MED DOISST Data downloaded from CMEMS portal.

Chlorophyll data are freely available from CMEMS portal.

The iQuam data version of this study used is V2.1, downloaded from the National Environmental Satellite, Data, and Information Service Satellite Applications and Research NOAA NESDIS STAR portal.

Gridded analyses of EN4 profiles are distributed from the MetOffice Hadley Centre Observations (we used version 4.2.1).

ERA5 data are freely available after registration on the Climate Data Store (CDS) by Copernicus Climate Change Service (C3S).

MLD data are distributed on a 0.25 degree regular grid, and freely available from the Sea Open Scientific Data Publication SEANOE portal.

Minimal data and scripts used within the manuscript to reproduce the figures in the manuscript are available at this link:
https://zenodo.org/records/10451206

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561   and WRF interfaces to the OASIS coupler. This Research is supported by ICSC – National Research Center for
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563   dell'Università e della Ricerca through the NextGenerationEU programme.

564

565   *Author Contributions.* VdT and AS conceived the study and designed the experiments to conduct, VdT
566   Performed the simulation and data analysis, data downloading and writing of the first draft, VdT, DC, YH, CY,
567   VA, AP, DC, RS and AS equally contributed to discuss and interpret the results, finalizing the draft.

568

569   *Competing Interests.* All authors declare they have no competing interests.

Figures
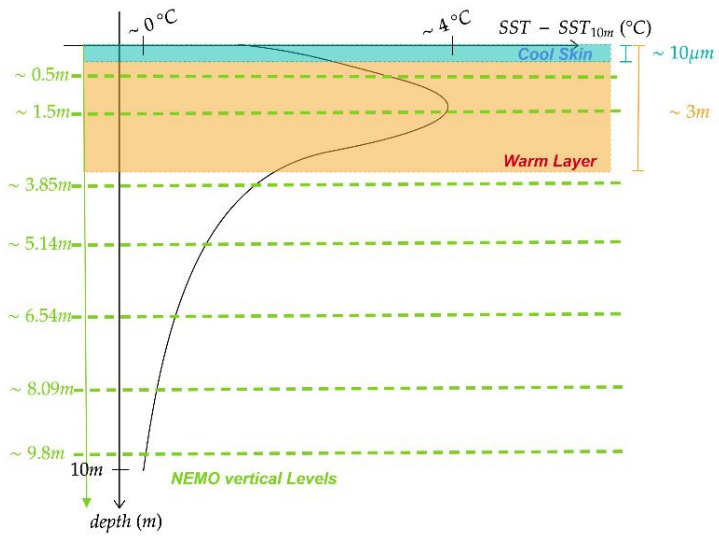
Figure 1. Sketch of the cool skin and warm layer adapted from Donlon et al., 2007. Vertical discretization of NEMO levels is shown in green (not perfectly in scale with the underlying y-axis).

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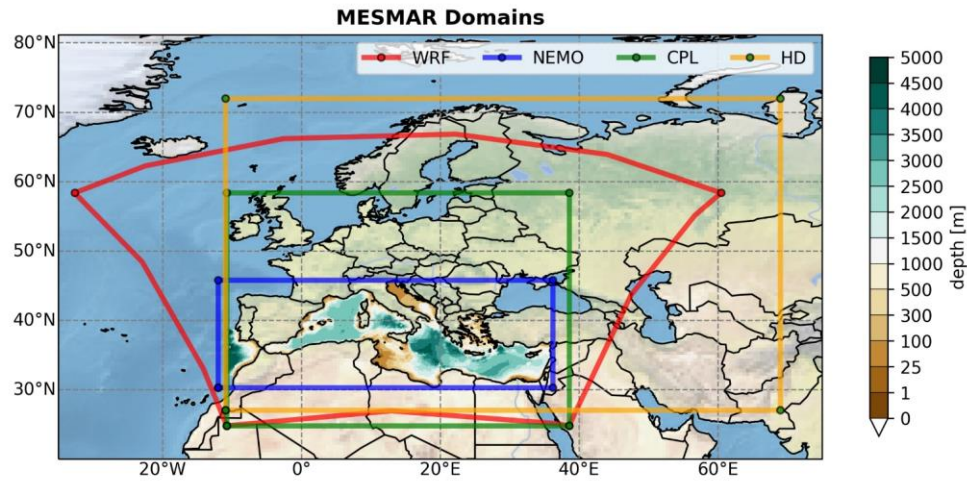
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Figure ~~1~~2: The modeling system domain: WRF, NEMO, HD and boundaries for the coupling mask are respectively in red, blue, orange, and green. Contour filled plot shows the ocean model bathymetry.

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584 **Figure** 32: Panel 2a shows two different formulations frequently used for the transmission coefficient
585 expression: the red curve shows the formulation of Soloviev, 1982, while the green curve the one defined in
586 Soloviev and Schlussel, 1996. Panel 2b shows e-folding depth estimates from Mediterranean Chlorophyll
587 climatology of Volpe et al., 2019: lowest values touch the 2.5 meters.

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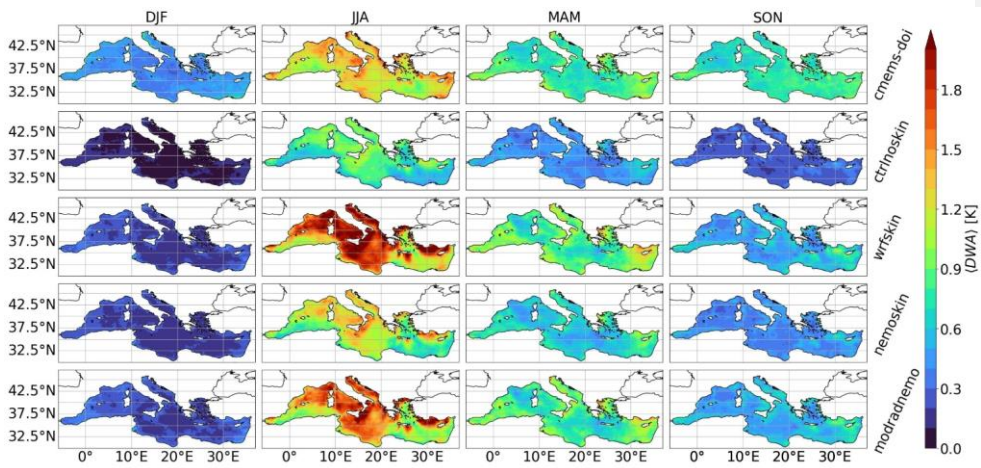
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591 **Figure** ~~3~~4: Mean diurnal warming amplitude averaged over seasons (on columns), for each case (row): the first
592 row is the CMEMS MED DOISST data, followed in order by the control simulation, wrfskin, nemoskwrite and
593 modradnemo.

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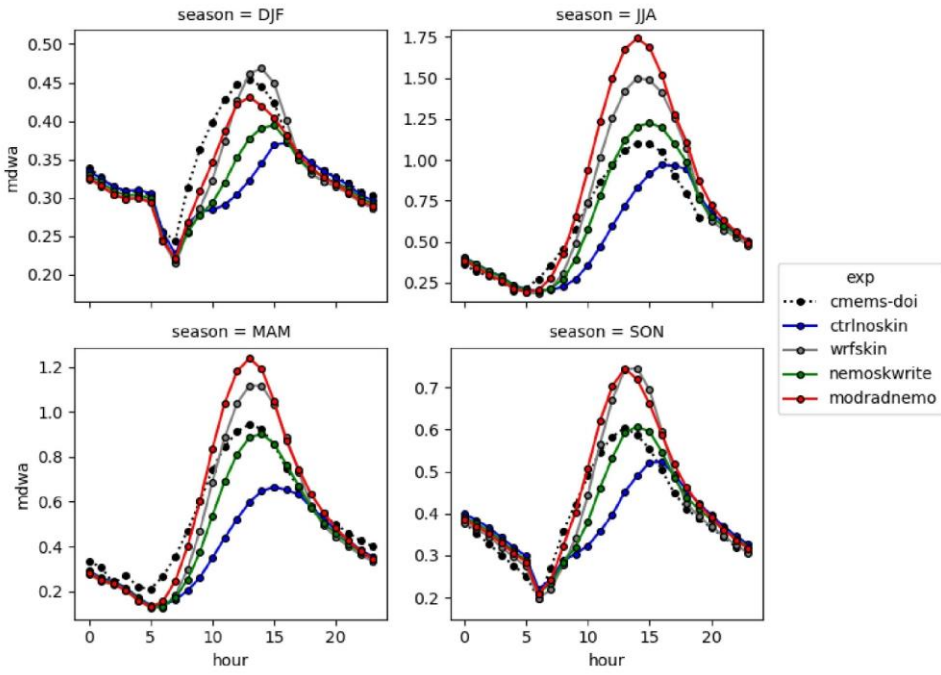
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598 **Figure 54**: Seasonality of the diurnal cycle averaged over the whole Mediterranean Sea, masking out regions
599 in time and space where the percentage of model data in CMEMSDOI is greater than 50%.
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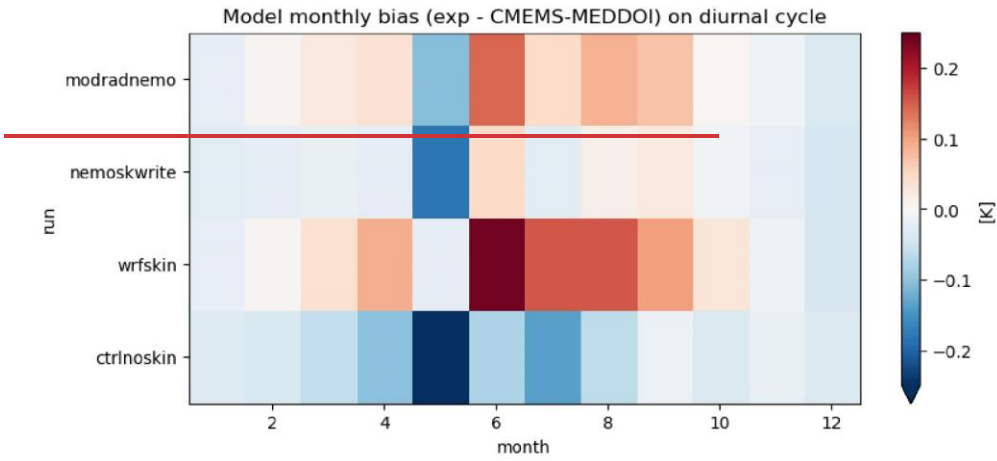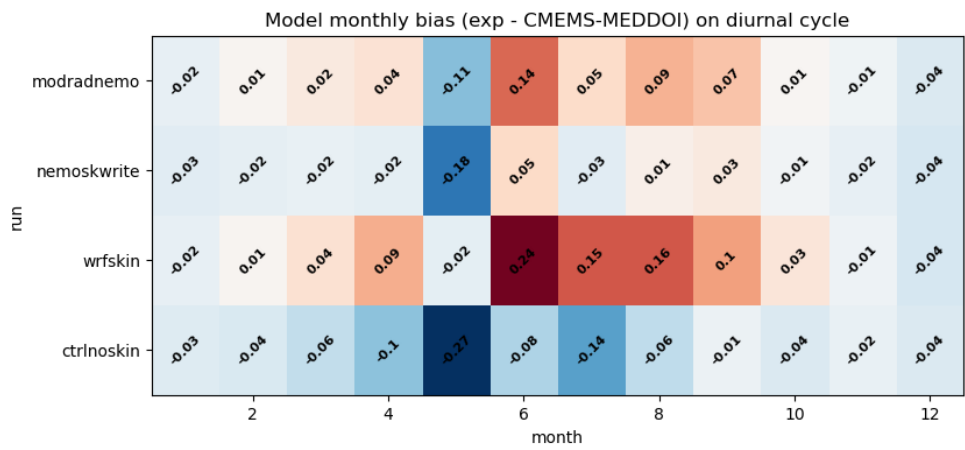
Figure 65: Monthly averaged values for the time series of spatial mean diurnal cycle over the Mediterranean Sea (bias with respect to CMEMS MED DOISST)

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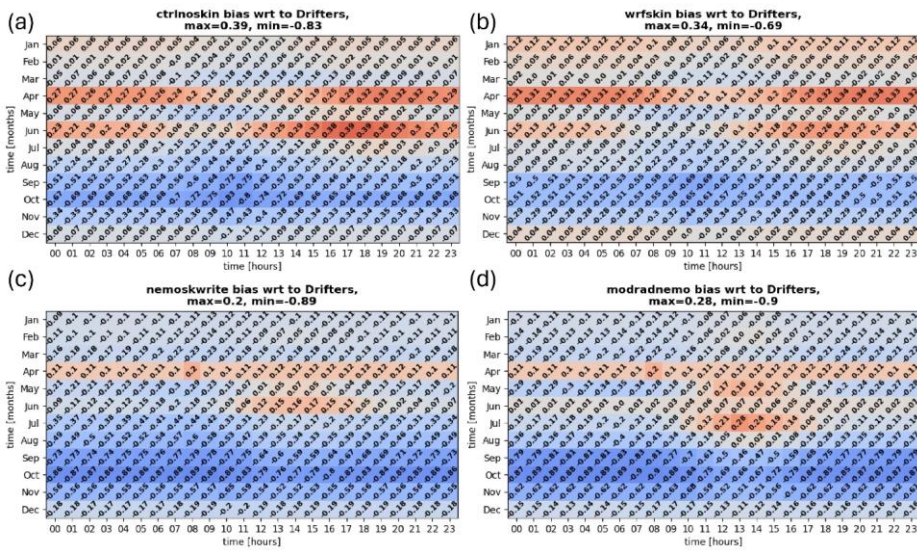
607
608 **Figure 76**: Bias with respect to measurements averaged over drifters' locations as a function of the month and
609 the time of the day. Panels 6a, 6b, 6c, 6d show respectively the results for all the simulations carried out in the
610 present study. Confidence on these numbers can be supported by the numbers of measurements reported in table
611 S1.
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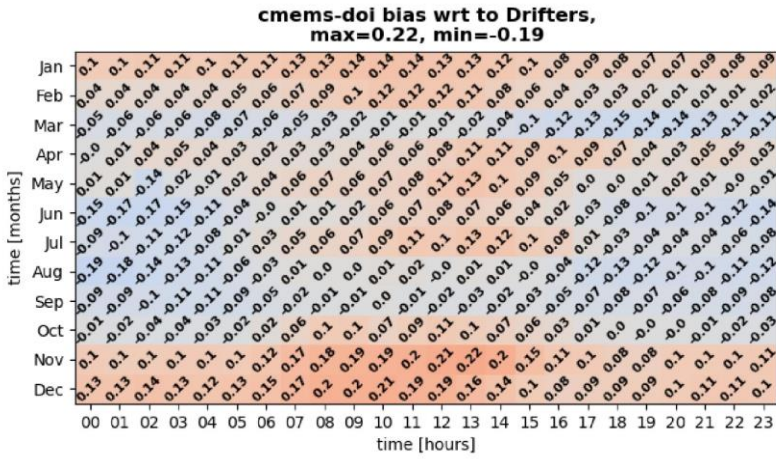
Figure 8~~7~~: Bias with respect to measurements averaged over drifters' locations as a function of the month and time of the day, for CMEMS MED DOISST data.

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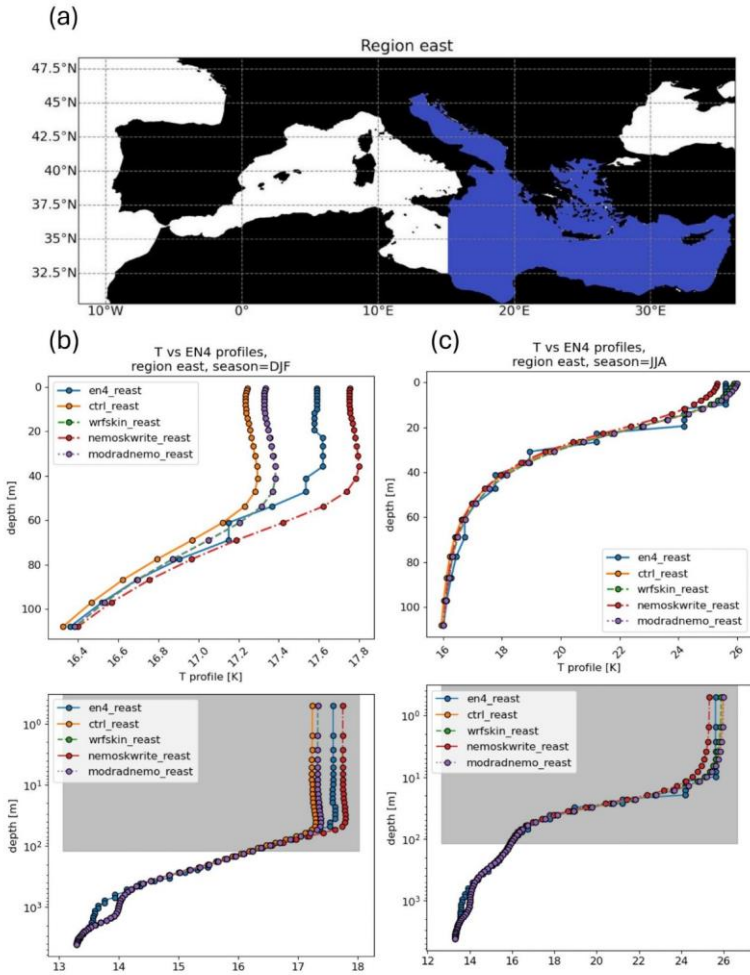
620
621 **Figure 98**: Spatial average of profiles within the eastern Mediterranean Sea, during winter and summer. Panel
622 98a shows the eastern region, while 98b, 98c show respectively wintertime and summertime spatially averaged
623 profiles within the top 100 *m* in the upper part, on the bottom the whole depth range on a logarithmic scale.
624

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Figure 10~~9~~: Spatial average of profiles within the eastern Mediterranean Sea, during winter and summer. Panel 10~~9~~a shows the eastern region, while 10~~9~~b, 10~~9~~c show respectively wintertime and summertime spatially-averaged profiles within the top 100 *m* in the upper part, on the bottom the whole depth range on a logarithmic scale.

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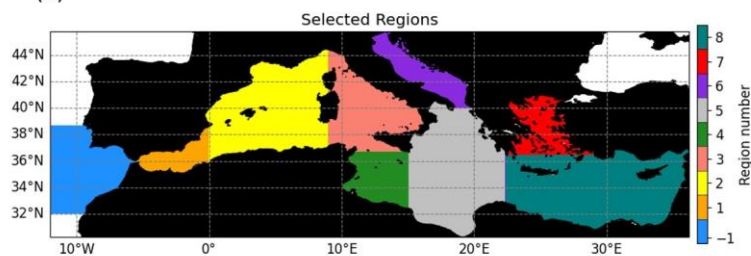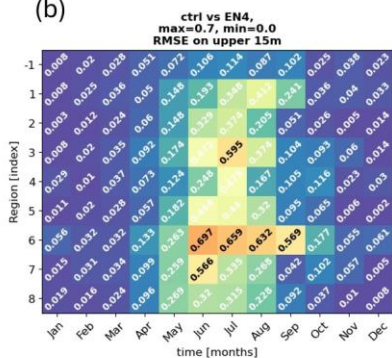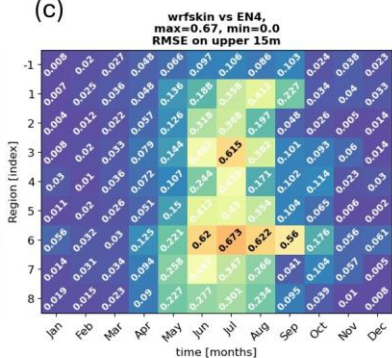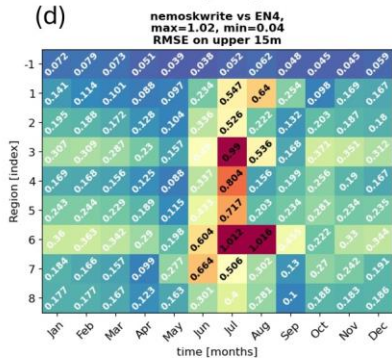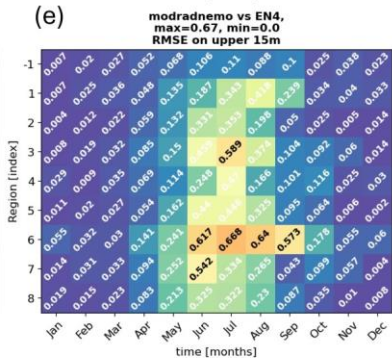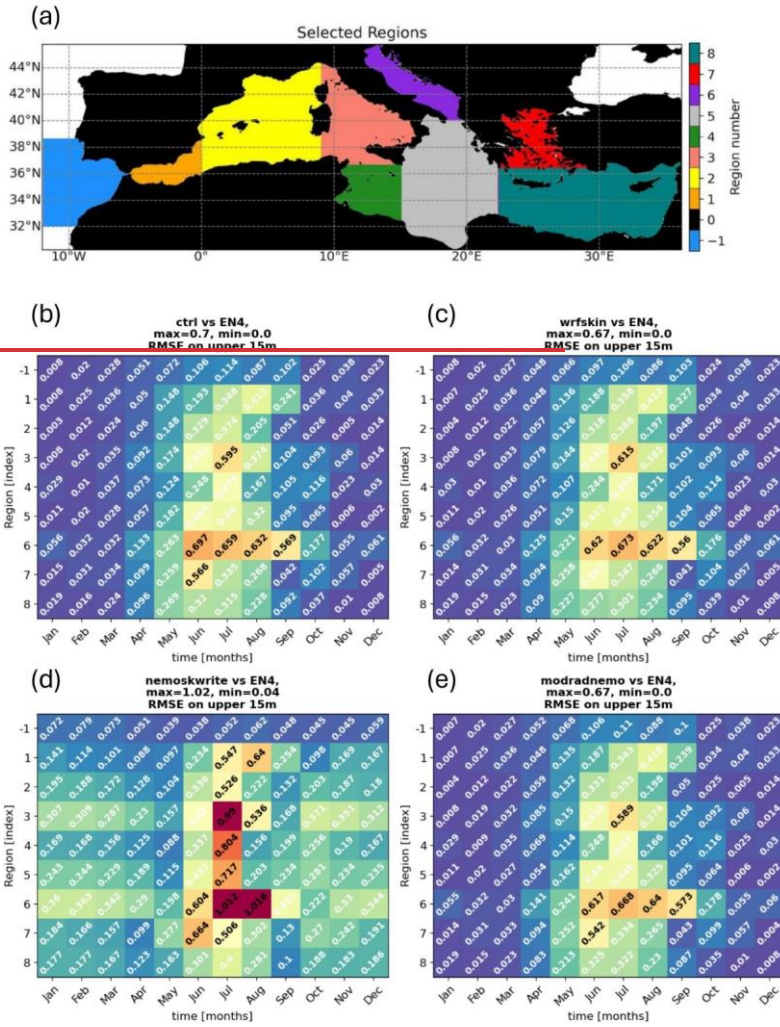
633
634 **Figure 110**: RMSE on the top 15m of the difference between regionally averaged profiles between each
635 simulation and EN4, displayed as a function of the region and the particular month. Division in regions is
636 reported in panel 110a, while 110b, 110c, 110d, 10e show respectively the results for all the simulations carried
637 out in the present study.
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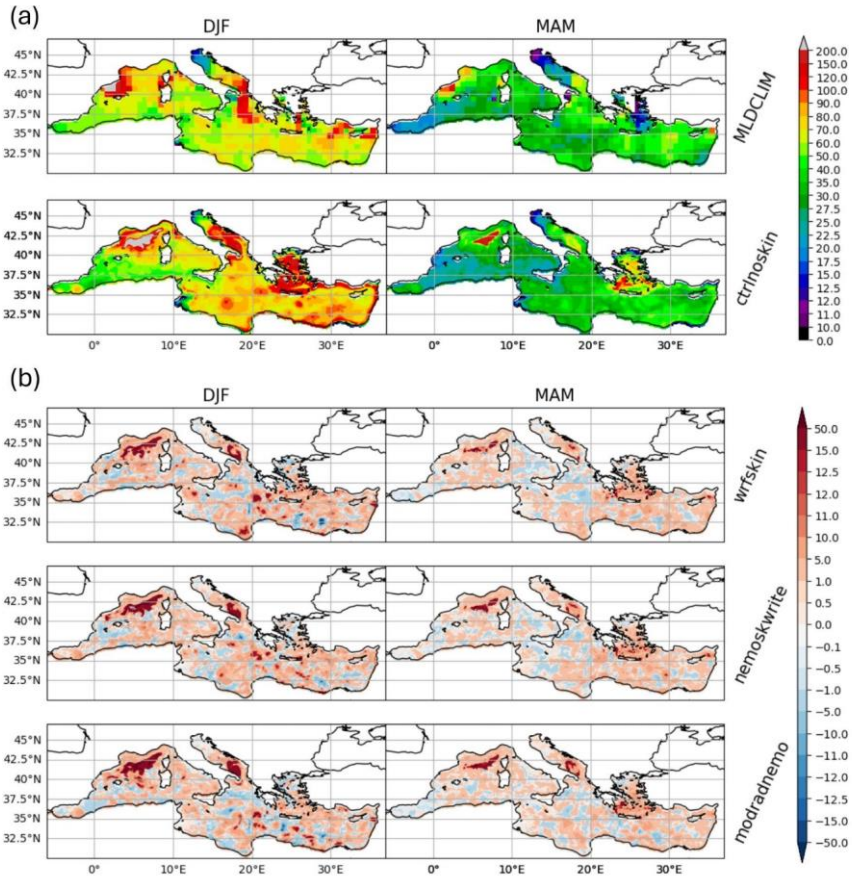
640 **Figure 12**: Maps of DJF, MAM of mixed layer depth for the climatology and for the control simulation in
641 panel (a). Panel (b) shows the difference of the control with respect to each simulation. Units are meters.
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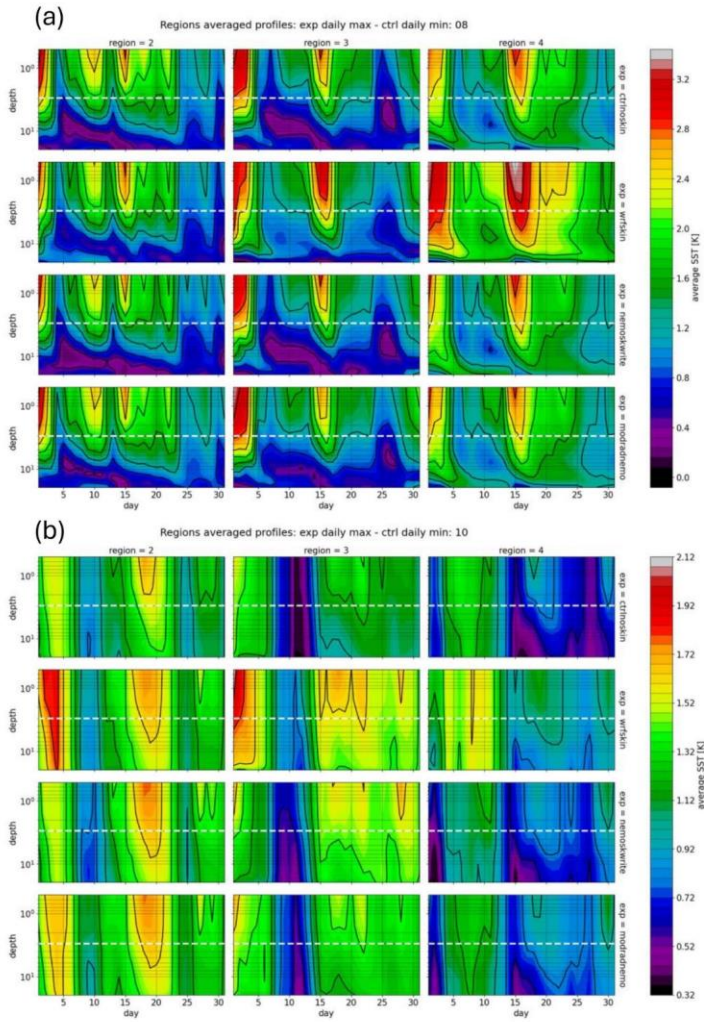
Figure 1<s>3</s>2: Hovmoller plots for spatial average of model outputs temperature profiles in the regions 2,3,4 as defined by figure 1<s>1</s>0a. Each row shows the difference between daily maxima for the given experiment minus the daily minima for the control simulation. The white dashed line traces the $z = 3m$ line of the depth used as reference for the base of the warm layer as in ZB05 scheme Zeng and Beljaars, 2005. Panel 1<s>3</s>2a shows August, panel 1<s>3</s>2b shows October.

666

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Tables

668

669

Simulation	Scheme active	Extinction coefficients in Warm Layer
ctrlnoskin	None	None
wrfskin	ZB05	SS82
nemoskwrite	ZB05+A02+T10	G09
modradnemo	ZB05+A02+T10	R-G-B + chl e-folding

670 **Table 1**: Overview of the simulations performed

671

Wavelength [μm]	i	a_i	$b_i \; [m^{-1}]$
0.3-0.6	1	0.2370	1.488×10^{-1}
0.6-0.9	2	0.3600	4.405×10^{-1}
0.9-1.2	3	0.1790	3.175×10^{1}
1.2-1.5	4	0.0870	1.825×10^{2}
1.5-1.8	5	0.0800	1.201×10^{3}
1.8-2.1	6	0.0246	7.937×10^{3}
2.1-2.4	7	0.0250	3.195×10^{3}
2.4-2.7	8	0.0070	1.279×10^{4}
2.7-3.0	9	0.0004	6.944×10^{4}

672 **Table 2**: Parameters for the Transmission coefficient following Soloviev and Schlüssel, 1996, in which the
673 first coefficients is the average between the one corresponding to I, IA, IB, II, and III Jerlov optical water types.
674 This is currently implemented in NEMO.

simulation	DJF	MAM	JJA	SON	Annual
ctrlnoskin	-173.31	133.92	75.56	-66.40	-7.55
wrfskin	-168.83	134.19	76.51	-65.87	-5.97
nemoskwrite	-169.28	133.79	76.77	-65.72	-6.10
modradnemo	-169.06	134.87	78.16	-68.13	-6.04
ERA5	-140.36	133.24	81.96	-53.46	5.35

675 **Table 3**: Averaged surface net heat flux over the Mediterranean Sea (W/m^2): seasonal and annual spatial
676 averaged mean values.

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