How high frequency atmospheric forcing impacts mesoscale eddy surface signature and vertical structure

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July 23, 2023

Abstract

Seasonal evolution of both surface signature and subsurface structure of a Mediterranean mesoscale anticyclones is assessed using the CROCO high-resolution numerical model with realistic background stratification and fluxes. In good agreement with remote-sensing and in-situ observations, our numerical simulations capture the seasonal cycle of the anomalies, induced by the anticyclone, both in the sea surface temperature (SST) and the mixed layer depth (MLD). The eddy signature on the SST shifts from warm-core in winter to cold-core in summer, while the MLD deepens significantly in the core of the anticyclone in late winter. Our sensitivity analysis shows that these dynamical properties can be accurately reproduced only if the resolution is high enough (˜1km for the horizontal with 100 vertical levels in a Mediterranean stratification) and if the atmospheric forcing contains high-frequency. In this configuration the deformation radius is explicitly resolved and the vertical mixing parametrized by the k-ε closure scheme is three times higher inside the eddy than outside the eddy. This differential mixing is explained by near-inertial waves, triggered by the high-frequency atmospheric forcing.Near-inertial waves propagate more energy inside the eddy because of the lower effective Coriolis parameter in the anticyclonic core. In addition to these high spatial and temporal resolution, SST retroaction on air-sea fluxes appears to be necessary to obtain marked eddy mixed layer depth anomaly.

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Key Points:

- Seasonal variations in SST and mixed layer anomalies of a Mediterranean anticyclone are retrieved in high resolution simulation
- The summer surface cold-core signatures are driven by differential vertical mixing due to near-inertial waves
- SST retroaction on air-sea fluxes is necessary to retrieve eddy-induced mixed layer anomalies

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Plain Language Summary

Mesoscale eddies are turbulent structures present in every regions of the world ocean, and accounting for a significant part of its kinetic energy budget. These structures can be tracked in time and recently revealed a seasonal cycle from in situ data. An anticyclone (clockwise rotating eddy in the northern hemisphere) is observed in the Mediterranean to be predominantly warm at the surface and to deepen the mixed layer in winter, but shifts to a cold-core summer signature. This seasonal signal is not yet understood and studied in ocean models. In this study we assess the realism of an anticyclone seasonal evolution in high resolution numerical simulations. Eddy surface temperature seasonal shift is retrieved and is linked to an increased mixing at the eddy core spontaneously appearing at high resolution (grid size $\sim 1km$) in the presence of high frequency atmospheric forcing. This increased mixed is due to the preferred propagation of near-inertial waves in the anticyclone due to its negative relative vorticity. Eddy-induced mixed layer depth anomalies also appear to be triggered by sea surface temperature retroaction on air-sea fluxes. These results suggest that present-day operational ocean forecasting models are too coarse to accurately retrieve mesoscale evolution.

1 Introduction

Mesoscale eddies are ubiquitous turbulent structures in the oceans, in thermal wind balance with a signature in density. Eddies have been observed for a long time, in particular in energetic regions such as the Gulf Stream rings (Richardson, 1980), the Kuroshio extension (Itoh & Yasuda, 2010), the Agulhas current retroflexion (Olson & Evans, 1986) , but also in the Mediterranean Sea (Mkhinini et al., 2014). Mesoscale anticyclones (respectively cyclones) are structures of negative (positive) density anomaly, identified with a sea surface height (SSH) elevation (depression) (Chelton et al., 2007). Eddies statistical descriptions really began with the availability of eddy automated detections based on gridded altimetry products (Doglioli et al., 2007; Chaigneau et al., 2009; Nencioli et al., 2010; Chelton, Schlax, & Samelson, 2011; Mason et al., 2014; Le Vu et al., 2018; Laxenaire et al., 2018). The first quantitative studies were done in a composite approach : many daily snapshots detections are colocated with eddy contours and gathered into a single annual mean eddy signature (Hausmann & Czaja, 2012; Everett et al., 2012). This approach combined with remote-sensing measurements provide an extensive view of eddies in various regions of the global ocean, with SST, sea surface salinity (Trott et al., 2019), chlorophyll (Chelton, Gaube, et al., 2011) and also meteorological variables (Frenger et al., 2013). Composite approach also allowed to reveal a modulation of air-sea fluxes at the eddy scale : in the Agulhas retroflexion region, (Villas Bôas et al., 2015) showed the total heat flux to the atmosphere to be enhanced over very strong and warm anticyclones. Similarly for the eddy vertical structure, gathering Argo profiles as a function of normalized distance to the eddy center, eddies were found to influence the mixed layer depth (MLD) (Sun et al., 2017; Gaube et al., 2019). Anticyclones have deeper MLD in their core, cyclones shallower MLD, with larger mixed layer anomalies in winter. Eddies were also observed to incorporate a significant seasonal cycle in their radius variations (Zhai et al., 2008) and their SST signature (Sun et al., 2019; Y. Liu et al., 2021). Anticyclones (respectively cyclones) usually identified as warm in surface, actually shift to cold (warm) signatures in summer in several regions of the world ocean (Sun et al., 2019; Moschos et al., 2022). This phenomenon is then referred to as 'inverse' SST signatures. (Moschos et al., 2022) showed that these 'inverse' signatures actually become predominant in summer in the Mediterranean Sea, a seasonal shift yet not properly understood.

The composite approach is nonetheless ill-suited to study eddy temporal variability due to the stacking of numerous observations in time. Recently Lagrangian approaches were developed to study eddies enabling to better track their temporal variability (Pessini et al., 2018; Laxenaire et al., 2020; Barboni et al., 2021). Using a Lagrangian approach, (Moschos et al., 2022) showed that the same individual anticyclones shift from a warm winter SST anomaly to a cold one in summer (and conversely for cyclone). With the additional Argo floats trapped in anticyclones, they further noticed that anticyclonic density anomaly remain warmer at depth while becoming colder in surface, leading to a smoother density gradient. Hence the hypothesis that this seasonal shift could be explained by a modulation of the vertical mixing by mesoscale eddies, anticyclones (cyclones) likely enhancing (decreasing) mixing in surface. Recent observations in the Mediterranean Sea of inside-anticyclone properties temporal evolution further revealed eddy mixed layer anomalies to be much larger than the composite approach mean value, reaching sometimes 300m (Barboni, Coadou-Chaventon, et al., 2023). MLD anomalies evolution was also shown to have evolution much faster than the month, with delayed restratification inside anticyclones. Mechanisms driving these MLD anomalies are also unexplained, but (Barboni, Coadou-Chaventon, et al., 2023) found it to be impacted by interactions with the anticyclone vertical structure.

An eddy modulation of vertical mixing was not proven so far but could be due to a modulation of near-inertial waves (NIW) propagation. NIW can not propagate at frequencies lower than the inertial frequency f due to Earth rotation (Garrett & Munk, 1972). However in the presence of a balanced flow, anticyclones (cyclones) having negative (positive) relative vorticity ζ locally shift this cut-off by an effective inertial frequency $f_e = f + \zeta/2$ (Kunze, 1985). Sub-inertial waves ($\omega \lesssim f$) can then remained trapped in anticyclones and supra-inertial waves ($\omega \gtrsim f$) can be expelled from cyclones. Consequently, NIW propagate more inside anticyclones, what was experimentally (D'Asaro, 1995) and numerically (Danioux et al., 2008, 2015; Asselin & Young, 2020) proven. This NIW trapping potential partly explains the interest in anticyclones rather than in cyclones, the other reason likely being that anticyclones are more stable in time (Arai & Yamagata, 1994; Graves et al., 2006), in particular for large structures (Perret et al., 2006), then more easily detected and trapping more often profilers (thus easing field campaigns). Several recent observations (Martínez-Marrero et al., 2019; Fernández-Castro et al., 2020) showed that mixing at depth is enhanced below anticyclones due to this more energetic NIW propagation. On the other hand numerical studies assumed extremely simplified set-up with constant wind (Danioux et al., 2008) or an idealized wind burst (Asselin & Young, 2020). They also looked at NIW propagation in an eddying field at short time scales, then without significant evolution of the eddies and stratification. Eddy-NIW interaction on longer time scales - eddy evolving time scales like months - in a varying stratification due to seasonal cycle has never been assessed so far. In particular the effect of this differential NIW propagation on eddies remains unknown and a gap remains between wave propagation and enhanced surface mixing.

Some recent studies started to assess eddy temporal evolution in high resolution regional models. In the Mediterranean Sea, Escudier et al. (2016) compared eddy size, drift and lifetime compared to eddies in altimetric observations. Mason et al. (2019) investigated these variables in assimilated operational models and additionally looked at MLD anomalies, but both were in a composite approach and did not look at eddy SST variations. More recently Stegner et al. (2021) performed an observation system simulation experiment on a 1/60°simulation of the Mediterranean sea and found great bias on size and strength for small eddy detections, but did not look at SST variations. Using the same simulation, an interesting method was developed by Ioannou et al. (2021), investigating differences in both trajectories, size and stratification of the Ierapetra anticyclonic eddy, but restricted to this particular case.

Both eddy SST anomalies seasonal shift and mixed layer depth anomalies remain poorly investigated so far in ocean models. If NIW propagation and eddy vertical structure are considered, grid resolution - both horizontal and vertical - and atmospheric forcing are likely key aspects to take into account. Indeed air-sea fluxes and near-inertia-gravity waves involve much shorter temporal and spatial scales, not reproduced even in eddy-permitting models at present stage. We then aim to assess the realism of an anticyclone seasonal signal in both surface and mixed layer using an idealized but high-resolution simulation and investigating driving physical processes. The goal is to assess the realism of the eddy temporal evolution compared to similar observations, in particular the retrieval of the surface signature seasonal cycle. In a first part we conduct a sensitivity analysis on horizontal grid cell. In a second part we study the sensitivity to atmospheric forcing frequency. Last, the effect of SST retroaction on air-sea fluxes is discussed.

2 Methods

2.1 Model set-up

Idealized numerical experiments are performed using the Coastal and Regional Ocean Community (CROCO) model. CROCO is based on the Regional Ocean Modeling System (ROMS) kernel (Shchepetkin & McWilliams, 2005). It uses a time splitting method between the fast barotropic mode and the slow baroclinic ones. Advection schemes are UP3 for horizontal and Akima-Splines for the vertical. Trying to conciliate realistic and idealized approach, we use double periodic conditions in a realistic stratification and on long timescale. The atmospheric forcing has realistic temporal variations but is spatially homogeneous. The only active tracer used is temperature. As a consequence, a linear state equation links density ρ and temperature T, with thermal expansion $T_c = 0.28kg.m^{-3}.K^{-1}$ and linear approximation close to $T_0 = 25$°C and $\rho_0 = 1026kg.m^{-3}$:

$$\rho = \rho_0 + T_c(T - T_0) \tag{1}$$

Discarding salinity effects is justified by the very weak salinity seasonal cycle in the Mediterranean Sea. The heat flux seasonal cycle is roughly $\pm 150\,W.m^{-2}$ (Pettenuzzo et al., 2010), whereas salinity fluxes are mostly driven by the evaporation minus precipitation balance, with a mean of roughly $10^3 mm/y$, a seasonal cycle maximal amplitude of $\Delta F = 4 \times 10^2 mm/y$ and river input being negligible (Mariotti, 2010). Considering a haline contraction coefficient of $S_c = 0.78kg.m^{-3}.PSU^{-1}$, a ΔF freshwater input would have a seasonal equivalent effect on buoyancy $Q_{eq} = \rho_0 c_p \frac{S_c}{T_c} S_0 \Delta F \approx 5\,W.m^{-2}$, indeed almost two orders of magnitude lower than Q_{tot}.

Grid

Simulation domain is double periodic, on the f-plane, with a flat bottom $H_{bot} = 3000m$. Horizontal extent is 200km in both directions, with horizontal resolution ranging between 4km and 500m, with respectively 25 to 150 vertical levels. Horizontal to vertical grid size ratio is kept roughly constant about 1000/3 in the upper layers, close to the Brunt-Vaisala to inertia frequency ratio. Coriolis parameter is $f = 9.0 \times 10^{-5} s^{-1}$. CROCO uses a σ terrain-following coordinate, the N vertical levels being modulated in time between bottom and sea surface height η. Constant depth level z_0 are stretched over thickness h_c with surface coefficient θ_s :

$$z = \eta + (\eta + H_{bot}) z_0 \tag{2}$$

$$z_0 = \frac{h_c \sigma + H_{bot} C_s(\sigma)}{h_c + H_{bot}} \quad \text{with} \quad C_s(\sigma) = \frac{1 - cosh\left(\theta_s \frac{\sigma - N}{N}\right)}{cosh(\theta_s) - 1} \tag{3}$$

With $N = 100$ levels, $h_c = 400m$ and $\theta_s = 8$, vertical grid steps are then close to $5m$ in the upper $200m$. $200m$ being the vertical scale of the thermocline, it ensures a maximal resolution in the upper ocean where seasonal variations occur (Houpert et al., 2015). This configuration has then a higher vertical resolution than previous similar studies ($N = 32$ $h_c = 250m$ and $\theta_s = 6.5$ for Escudier et al. (2016)) or operational models (Juza et al., 2016).

Turbulent closure

Mixing is parametrized through $k{-}\epsilon$ closure scheme (Rodi, 1987) using the generic length scale approach (Umlauf & Burchard, 2003). Turbulent kinetic energy k dissipates with rate ϵ and stability function c_v into an effective viscosity ν (respectively c_T and κ for diffusivity). No additional explicit mixing is added.

$$\nu = \frac{c_v k^2}{\epsilon} \quad \text{and} \quad \kappa = \frac{c_T k^2}{\epsilon} \tag{4}$$

As implemented in the CROCO model, a minimal k input is parameterized. Given that the minimal dissipation rate ϵ is set to $10^{-12} W.kg^{-1}$, the minimal k has to be set to $10^{-9} m^2.s^{-2}$ in order to retrieve a minimal diffusivity of $10^{-6} m^2.s^{-1}$ with a stability function of order unity. This diffusivity value is close to kinematic viscosity and thermal diffusivity for water (respectively $1{\times}10^{-6}$ and $1{\times}10^{-7}$ $m^2.s^{-1}$). This issue is also discussed by Perfect et al. (2020).

2.2 Background stratification and initial mesoscale anticyclone

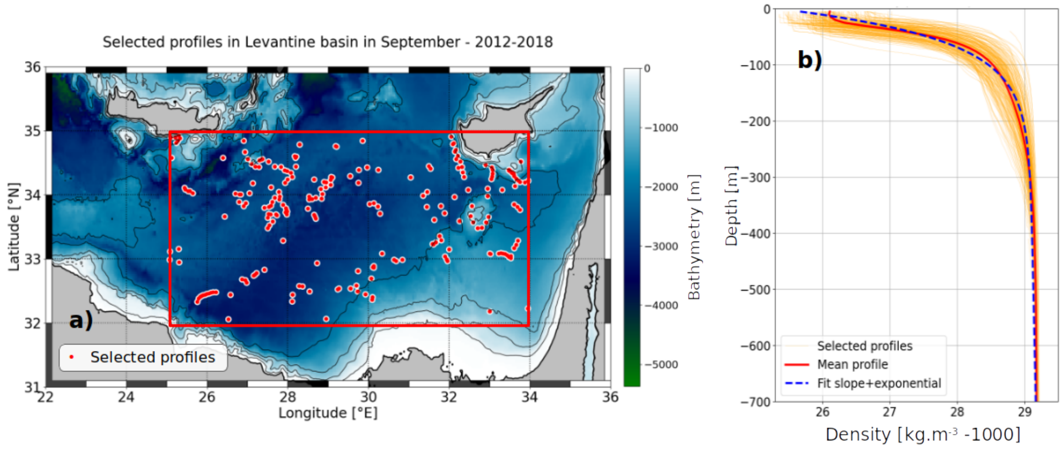

Figure 1. (a) Map showing the region of high long-lived anticyclones occurrence in the Levantine basin. The atmospheric fields used as input are averaged overt area delimited by the green frame. Black dots are the cast position of 242 selected in situ profiles identified as outside-eddy. Bathymetry is ETOPO1 data (Smith & Sandwell, 1997) with 0, 500, 1000 and 1500m isobaths. (b) Density vertical structure of selected profiles (orange thin lines), mean profile (red thick line) and fitted profile using Eq.5 (blue dashed).

A realistic background stratification is set from a climatological database gathering in situ delayed-time and near-real-time data from Copernicus Marine Environment Monitoring Service (Barboni, Stegner, et al., 2023). A region of interest is considered at the center of the Levantine Basin (25 to 34 °E and 32 to 35 °N, shown in Fig.1a). For background stratification we used only profiles in the region of interest, detected as outside-eddy using the DYNED eddy atlas dataset (see Barboni, Coadou-Chaventon, et al. (2023) for details), from 2012 to 2018 and for each year in September. Considering these criteria, 242 profiles are averaged into a mean stratification $\rho_b(z)$ fitted over the first 1000m with a linear slope S added to an exponential with vertical Z_T in the upper water column (Eq.5, see Fig.1b). September is chosen as the end of summer when the thermocline is marked and stratification gradient the strongest, allowing a better fit with exponential slope.

$$\rho_b(z) = \rho_1 + (\rho_s - \rho_1)exp\left(-\frac{z}{Z_T}\right) + Sz \tag{5}$$

Regression fit gave $\rho_1 = 1029.03kg.m^{-3}$, $\rho_s = 1025.3kg.m^{-3}$, $Z_T = 55m$, $S = 1.8{\times}10^{-4}kg.m^{-4}$. Corresponding baroclinic deformation radius is approximately $11km$. An initial density anomaly σ in geostrophic equilibrium is added to the background stratification. $\sigma(r,z)$ is azimuthally symmetric and has a Gaussian shape in the vertical direction and pseudo-Gaussian in the radial one, with radius R_m and vertical extent H :

$$\sigma(r,z) = \sigma_0 \frac{z}{H} exp\left(-\frac{1}{\alpha}\left(\frac{r}{R_m}\right)^{\alpha}\right) exp\left(-\frac{1}{2}\left(\frac{z}{H}\right)^2\right) \quad \text{with} \quad \sigma_0 = \frac{\rho_0 f V_m R_m e^{1/\alpha}}{gH} \tag{6}$$

The initial maximal speed radius R_m is 25 km, slightly more than twice the deformation radius but still smaller than the large long-lived Eastern Mediterranean anticyclones (Barboni, Coadou-Chaventon, et al., 2023), giving a Burger number ($Bu = R_d^2/R_m^2$) close to 0.2. Maximal speed is initially set to $V_m = 0.4\,m.s^{-1}$ giving a Rossby number ($Ro = V_m/R_m f$) of 0.16, but later decays around 0.1 due to eddy evolution. $Ro = 0.1$ is a standard value in the Mediterranean Sea (Ioannou et al., 2019). H is set to $100m$ on the same order as thermocline extent Z_T, and shape parameter $\alpha = 1.6$ ensures baroclinic stability (Carton et al., 1989; Stegner & Dritschel, 2000). Cyclogeostrophic correction is added following Penven et al. (2014).

2.3 Atmospheric heat forcing

Air-sea fluxes are computed with the Coupled Ocean–Atmosphere Response Experiment (COARE) 3.0 parametrization (Fairall et al., 2003), with improved accuracy for large wind speeds ($> 10m.s^{-1}$) encountered in high frequency forcing. Short wave heat flux Q_{SW} is distributed on the vertical following Paulson and Simpson (1977) transparency model with Jerlov water type I , consistent with very clear Mediterranean waters ($R = 0.58$, $\zeta_1 = 0.35m$, $\zeta_2 = 23m$):

$$I(z) = Q_{SW}\left(Rexp\left(-\frac{z}{\zeta_1}\right) + (1-R)exp\left(-\frac{z}{\zeta_2}\right)\right) \tag{7}$$

Upward long-wave heat flux $Q_{LW}^{\uparrow}$ computes the ocean SST (T_s) thermal loss using Stefan-Boltzmann black body law, with emissivity $\epsilon_{sb} = 98.5\%$ and $\sigma_{sb} = 5.6697 \times 10^{-8}\,W.m^{-2}.K^{-4}$, convention positive downwards :

$$Q_{LW}^{\uparrow} = -\epsilon_{sb}\sigma_{sb}T_s^4 \tag{8}$$

Latent heat flux Q_{Lat} and sensible heat flux Q_{Sen} also involves a direct SST retroaction:

$$Q_{Lat} = -\rho_a L_E C_E |V|(q_s - q_a) \quad ; \quad Q_{Sen} = -\rho_a c_p C_S |V|(T_s - T_{2m}) \tag{9}$$

With ρ_a air density, c_p air thermal capacity, L_E evaporation enthalpy and $|V|$ 10m wind speed. q_s and q_a are specific humidity for ocean and atmosphere at 2m respectively, saturated at T_s for q_s, related to saturated water pressure P_{sat} fro q_a : $q_a = 0.622 h_{2m} P_{sat}(T_{2m})/P_{SL}$ and $q_s = 0.98 \times 0.622 \times P_{sat}(T_s)/P_{SL}$. Factor 0.98 accounts for water vapor reduction caused by salinity (Sverdrup et al., 1942). Last, wind stress is computed from zonal and meridional winds (u and v) :

$$\tau_x = \frac{\rho_a}{\rho_0} C_D |u| u \quad \text{and} \quad \tau_y = \frac{\rho_a}{\rho_0} C_D |v| v \tag{10}$$

In equations 9-10, C_E, C_S and C_D are corresponding transfer coefficients considering the stability of the atmospheric boundary layer based on the Monin-Obukhov similarity theory. They are all on the order of 1×10^{-3} (Fairall et al., 2003).

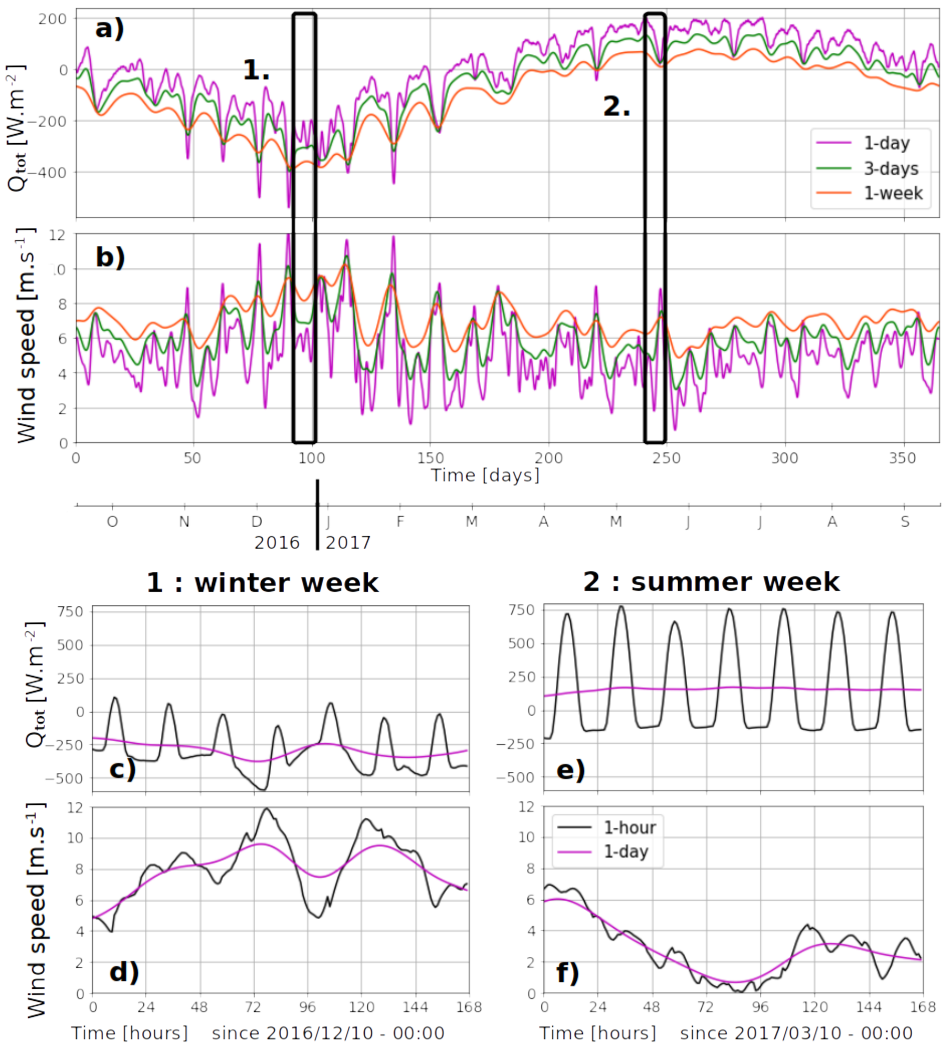

Figure 2. Net heat flux and wind speed for the 5 input timeseries, shown separately as diurnal cycle gives larger variations. (a) Net heat flux and (b) wind speed for the 1-day (magenta line), 3-days (green) and 1-week (orange) timeseries over one year. To enhance readability, 3-days and 1-week net heat flux are lowered by 20 and 40 $W.m^{-2}$ respectively. (c) 1-hour (black) and 1-day (magenta) net heat flux (respectively (d) for wind speed) in a winter week of 2016. (e) and (f) : same as (c) and (d) in a summer 2017 week.

ERA5 reanalysis input is used for atmospheric forcing. Fields are available with a 1 hour temporal resolution and 1/4°horizontal resolution (Hersbach et al., 2020). To focus on the temporal variability, forcing timeseries are spatially averaged over the Levantine basin (Fig.1a). Four forcing inputs with different temporal scales are tested : 1-hour, 1-day, 3-days and 1-week. The 1-hour forcing is the original ERA5 timeseries, the three later ones are Gaussian smoothing of the 1-hour timeseries with double-window size of 1, 3 and 7 days respectively, shown in Fig.2. One year of forcing from 15 September 2016 to 15 September 2017 runs cyclically for 2 years as forcing input, with root mean square wind speed $\overline{V} = 5.5m.s^{-1}$. 10m neutral wind from ERA5 is used for wind stress in Eq.10. To keep the same wind speed magnitude with varying wind frequency, smoothed timeseries for zonal and meridional winds ($\overline{u}$ and $\overline{v}$) have to be rescaled. The correction factor λ being $\gtrsim 1.1$ for 1-day timeseries, and $1.1 < \lambda < 2$ for 3-days and 1-week :

$$\tilde{u} = \lambda \overline{u}\,;\,\tilde{v} = \lambda \overline{v} \quad \text{with} \quad \lambda = \frac{\overline{\sqrt{u^2+v^2}}}{\sqrt{\overline{u}^2+\overline{v}^2}} \tag{11}$$

The same year is kept to avoid disturbance with interannual variations, which are strong for heat fluxes over the Mediterranean Sea (Mariotti, 2010; Pettenuzzo et al., 2010), but no significant variations were observed when selecting another year.

Forcing without surface temperature retroaction

A comparison experiment is run without SST retroaction on ocean-atmosphere fluxes. In this configuration, the net heat flux Q_{tot} from ERA5 directly forces the upper ocean layer, the short wave part Q_{SW} being still distributed on the vertical (Eq.7). Momentum fluxes are computed from Eq.10 with constant drag coefficient $C_D = 1.6 \times 10^{-3}$. The net heat flux Q_{tot} timeseries in ERA5 has daily amplitudes around $\pm 150 W.m^{-2}$ and an annual average of $-3.0\,W.m^{-2}$, consistent with the net evaporation of the Mediterranean sea (Mariotti, 2010). The net heat flux is then corrected by linearly decreasing the negative values to achieve a zero annual average, avoiding a drift of the mean stratification.

2.4 Eddy tracking indicators

Eddy shape, radius and intensity

Eddy detections are provided through the Angular Momentum Eddy Detection and Tracking Algorithm (AMEDA). AMEDA is a mixed velocity-altimetry approach, its relies on using primarily streamlines from a velocity field and identifying possible eddy centers computed as maxima of local normalized angular momentum (Le Vu et al., 2018). It was successfully used in several regions of the world ocean in altimetric data (Aroucha et al., 2020; Ayouche et al., 2021; Barboni et al., 2021), high frequency radar data (F. Liu et al., 2020) or numerical simulations (de Marez et al., 2021). In each eddy single observation (one eddy observed one day), AMEDA gives a center (which position is noted $\mathbf{X}_e$ hereafter) and two contours. The 'maximal speed' contour is the enclosed streamline with maximal speed (i.e. in the geostrophic approximation, with maximal SSH gradient) ; it is assumed to be the limit of the eddy core region where water parcels are trapped. The 'end' contour is the outermost closed SSH contour surrounding the eddy center and the maximal speed contour ; it is assumed to be the area of the eddy footprint, larger than just its core but still influenced by the eddy shear (Le Vu et al., 2018). The observed maximal speed radius R_m is then defined as the radius of the circle having an area equal to the maximal speed contour. Eddy detection through interpolated Level 4 SSH products leads to imperfections. It typically smooths gradients and then reduces observed geostrophic velocities (Amores et al., 2018; Stegner et al., 2021). To mimic those imperfections in the numerical simulations, AMEDA detections are performed on the 48h-averaged SSH field at model grid resolution.

Eddy SST signature δT, heat flux δQ, differential mixing ratio ξ and mixed layer anomaly

The anticyclone-induced SST signature δT is defined as the difference of SST between the eddy core SST_{in} and its periphery SST_{peri}. Adapting Moschos et al. (2022), SST_{in} is the average of the area centered on $\mathbf{X}_e(t)$ with radius $2/3R_m(t)$; SST_{peri} is the average on an annular area centered on $\mathbf{X}_e$ with radius between $2/3R_m(t)$ and $2R_m(t)$. Positive (negative) δT then indicates a warm-core (cold-core) signature. Similarly the induced signature on ocean-atmosphere fluxes is defined as δQ, with positive δQ for increased warming at the eddy core. Thermal heat flux feedback (THFF) is then defined as the linear regression of δQ as a function of δT over the second year of simulation (from 365 to 730 days, see Sect.3.3).

Differential mixing between the eddy core and outside-eddy are measured through the index ξ. Temperature vertical diffusivity κ computed by $k-\epsilon$ mixing closure from instantaneous history record is spatially averaged in the eddy core (κ^{AE}) and outside-eddy (κ^{Out}). The eddy core region corresponds here to the area around the eddy center with radius $2/3R_m(t)$. The outside-eddy region is defined as the area outside any 'end' contours detected by the tracking algorithm. Diffusivity spanning several orders of magnitude, differential mixing ξ is then evaluated as a vertical average of the ratio of these two quantities, typically using a depth $h = 20m$ to focus on the upper layers stratified in summer :

$$\xi = \frac{1}{h}\int_{-h}^{surf} \frac{\kappa^{AE}}{\kappa^{Out}} dz \tag{12}$$

Summer eddy SST signature magnitude $\overline{\delta T}$ is defined as the 30th δT percentile over the summer, and its spread as the difference between the 30th and the 10th percentiles (see results in Table 1). Similarly $\overline{\xi}$ is defined as the median of the ξ distribution over the summer, and its spread as the difference between the median and the 30th percentile. First and second summers are defined as 230 to 340 days and 590 to 700 days respectively, corresponding to the May to August period when a significant number of warm-core anticyclones are observed (Moschos et al., 2022).

Last, the MLD anomaly ΔMLD are defined for a given winter as the maximal difference reached between the MLD outside-eddy and the MLD inside-eddy, following (Barboni, Coadou-Chaventon, et al., 2023). In the following numerical experiments running for 2 years, the first winter is considered as a transient period not retained for analysis. ΔMLD is then computed only for the second winter, defined as 450 to 580 days, corresponding to the December to April period, when maximal MLD are reached in the Mediterranean Sea (Houpert et al., 2015).

3 Idealized simulations compared to observations

The temporal evolution of mesoscale eddies in the Levantine basin can be retrieved for several anticyclones where Argo floats remained trapped several months, as extensively studied in Barboni, Coadou-Chaventon, et al. (2023). A marked seasonal signal is detected in both SST and vertical structure. An example is shown in Fig.3 with a Ierapetra anticyclone. Ierapetra anticyclones are strong recurrent anticyclonic structures formed each year in the lee of Crete island (Ioannou et al., 2020). In the example shown below, δT index has a marked oscillation between a winter warm core and summer cold core. The weekly smoothed signature can be measured to about $\delta T \approx +0.7°C$ in both winters 2016-2017 and 2017-2018, and about $-0.3°C$ in summer 2017 (about $-0.2°C$ in summer 2018). The vertical structure could also be measured thanks to large Argo deployments (Fig.3h) ; due to errors in the salinity sensors, density in 2018 is estimated from temperature applying a linear regression using 2017 data. One can also notice the seasonal variations of the anticyclone maximal speed, with two maxima in late winter. This is consistent with kinetic energy inverse cascade maximal peak from submesoscale to mesoscale in kinetic energy distributions (Zhai et al., 2008; Steinberg et al., 2022), but it is still noticeable to have the same phenomenon tracking a single individual structure. Here the physical processes driving these observed seasonal variations are studied with numerical experiments, investigating sensitivity to horizontal resolutions, forcing frequency and SST retroaction on air-sea fluxes. Simulations are summarized in Table 1, the reference considered being 1km resolution with 1-hour forcing, 100 vertical levels with SST retroaction (run 1K100-1H in Table 1 below).

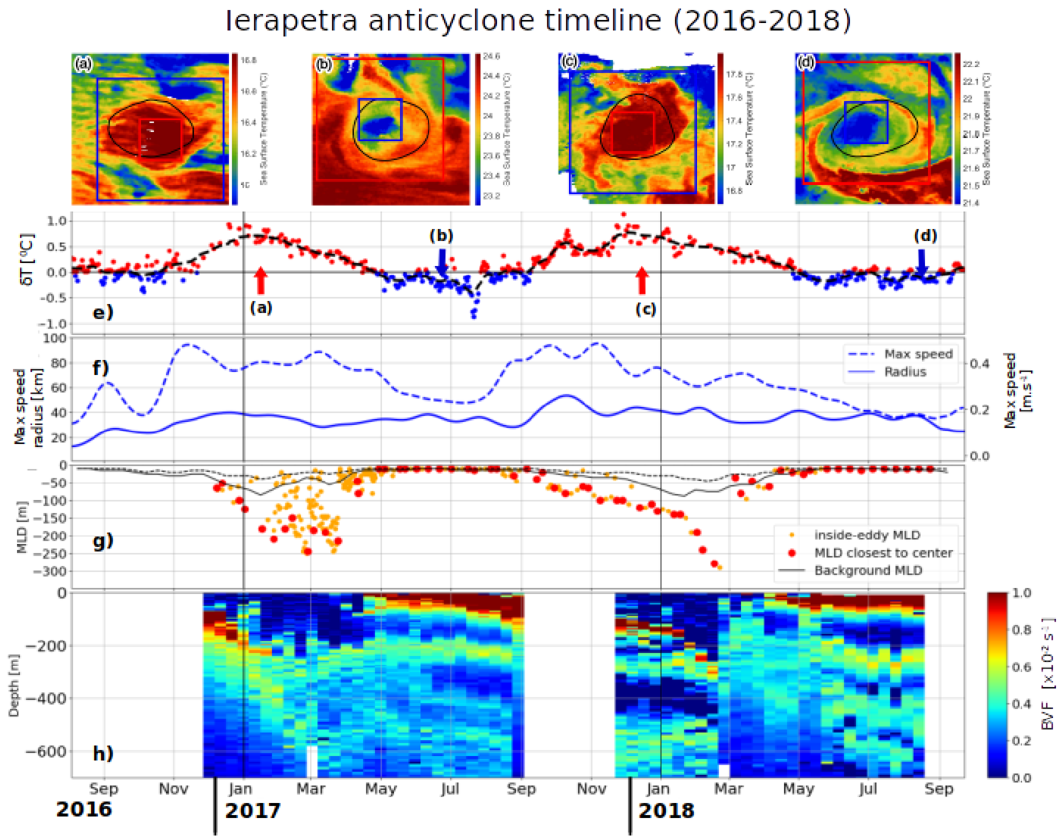

Figure 3. Temporal evolution of the Ierapetra anticyclone formed South-East of Crete in late summer 2016. Upper panels are high-resolution SST snapshots in (a) January 2017, (b) June 2017, (c) December 2017 and (d) July 2018, the maximal speed contour (see Sect.2.4 is in black line. (e) δT eddy SST anomaly, cold-core in blue and warm-core in red, with black dashed line showing the 5 days smoothed evolution. (f) Maximal speed V_m (dashed blue) and radius R_m (continuous blue) with 10 days smoothing. (g) MLD evolution inside the the anticyclone (dots, with red ones highlighting the closest to center), with outside-eddy background MLD in continuous black line (spread between dashed lines). (h) Brunt-Vaisala frequency.

3.1 Horizontal resolution sensitivity

The numerical simulation at 4km resolution and 25 vertical levels (run B in Table 1) reveals a few discrepancies with real observations. A horizontal resolution of 4km is close to operational oceanography models in the Mediterranean Sea (Juza et al., 2016). At the surface, despite seasonal variations of the eddy SST signatures (Fig.4a-c) and in the δT index (Fig.4f), summer 'inverse' signatures are not retrieved with no cold-core anticyclone. A steady erosion of the eddy strength is also noticeable, with a decrease in the maximal speed decreasing from $0.35 m.s^{-1}$ to $0.15 m.s^{-1}$ in 2 years, while its radius remains constant ($\approx 25 km$, Fig. 4e). Note that the initial maximal speed is set to $0.4 m.s^{-1}$ (see Sect.2.2) but the smoothing effect of time-averaging leads to a lower detected initial value of $0.35 m.s^{-1}$.

At depth, the mixed layer anomaly is significant, on the order of 50m (Fig.4g). Some bursts of differential mixing are observed in late winter from December to March when mixed layer instabilities and restratification processes can occur, with ξ reaching a few

[ht]

Table 1. Summary table of CROCO numerical experiments. Runs start in September of the atmospheric forcing timeseries. Thermal heat flux feedback (THFF), eddy SST anomaly index $\overline{\delta T}$ and differential mixing ratio $\overline{\xi}$ are defined in Sect.2.4, and $\overline{\xi}$ is computed over the upper 20m. Subscripts ($\overline{\xi}_1$,$\overline{\xi}_2$) refers to first and second summers defined as 230 to 340 days and 590 to 700 days respectively. ΔMLD refers only to the second winter defined as 450 to 590 days (see shades in Fig.4d-h).

Name	**Vertical levels**	**dx** (km)	**Freq**	**SST retroaction**	**THFF** ($W.m^{-2}.K^{-1}$)	$\overline{\delta \mathbf{T}_1}$ ($°C$)	$\overline{\delta \mathbf{T}_2}$ ($°C$)	$\overline{\xi}_1$	$\overline{\xi}_2$	**ΔMLD** (m)
1K100-1H	100	1	1-hour	Yes	-41.5 ± 1.3	-0.20 ± 0.10	-0.18 ± 0.04	3.05 ± 0.70	2.81 ± 0.74	51
2K50-1H	50	2	1-hour	Yes	-40.7 ± 1.0	-0.12 ± 0.14	-0.11 ± 0.06	1.54 ± 0.31	1.34 ± 0.22	63
4K25-1H	25	4	1-hour	Yes	-34.3 ± 1.8	0.01 ± 0.14	0.02 ± 0.10	1.10 ± 0.12	1.00 ± 0.12	48
05K150-1H	150	0.5	1-hour	Yes	-39.2 ± 1.4	-0.16 ± 0.10	-0.19 ± 0.06	2.58 ± 0.58	2.71 ± 0.45	91
1K100-1D	100	1	1-day	Yes	-42.1 ± 0.8	-0.21 ± 0.20	-0.31 ± 0.06	2.99 ± 0.44	3.34 ± 1.23	57
1K100-3D	100	1	3-days	Yes	-44.7 ± 1.0	-0.12 ± 0.14	-0.09 ± 0.03	1.41 ± 0.28	0.99 ± 0.09	70
1K100-1W	100	1	1-week	Yes	-41.0 ± 0.4	-0.05 ± 0.05	-0.03 ± 0.01	1.25 ± 0.14	1.02 ± 0.01	94
1K100-1H-NoSST	100	1	1-hour	No	-	-0.41 ± 0.16	-0.51 ± 0.00	2.60 ± 0.46	2.47 ± 0.25	10

times values higher than 2 (Fig. 4h). However no differential mixing is retrieved in summer. On the other hand, the anticyclone vertical structure is coarsely reproduced. The winter MLD cooling forms a homogeneous layer between 100 and 150m (Fig. 4i). These winter waters formed by convection do not reproduce the homogeneous subsurface anticyclone cores, separated by density jump or sharp temperature gradient such as the continuous temperature gradient in Fig.3h around 200m depth (see other examples in Fig.4-5 from Barboni, Coadou-Chaventon, et al. (2023)). The inability to reproduce this mesoscale subsurface lens is not surprising given the low vertical resolution, the vertical steps being on the order of $20m$ at $100m$ depth.

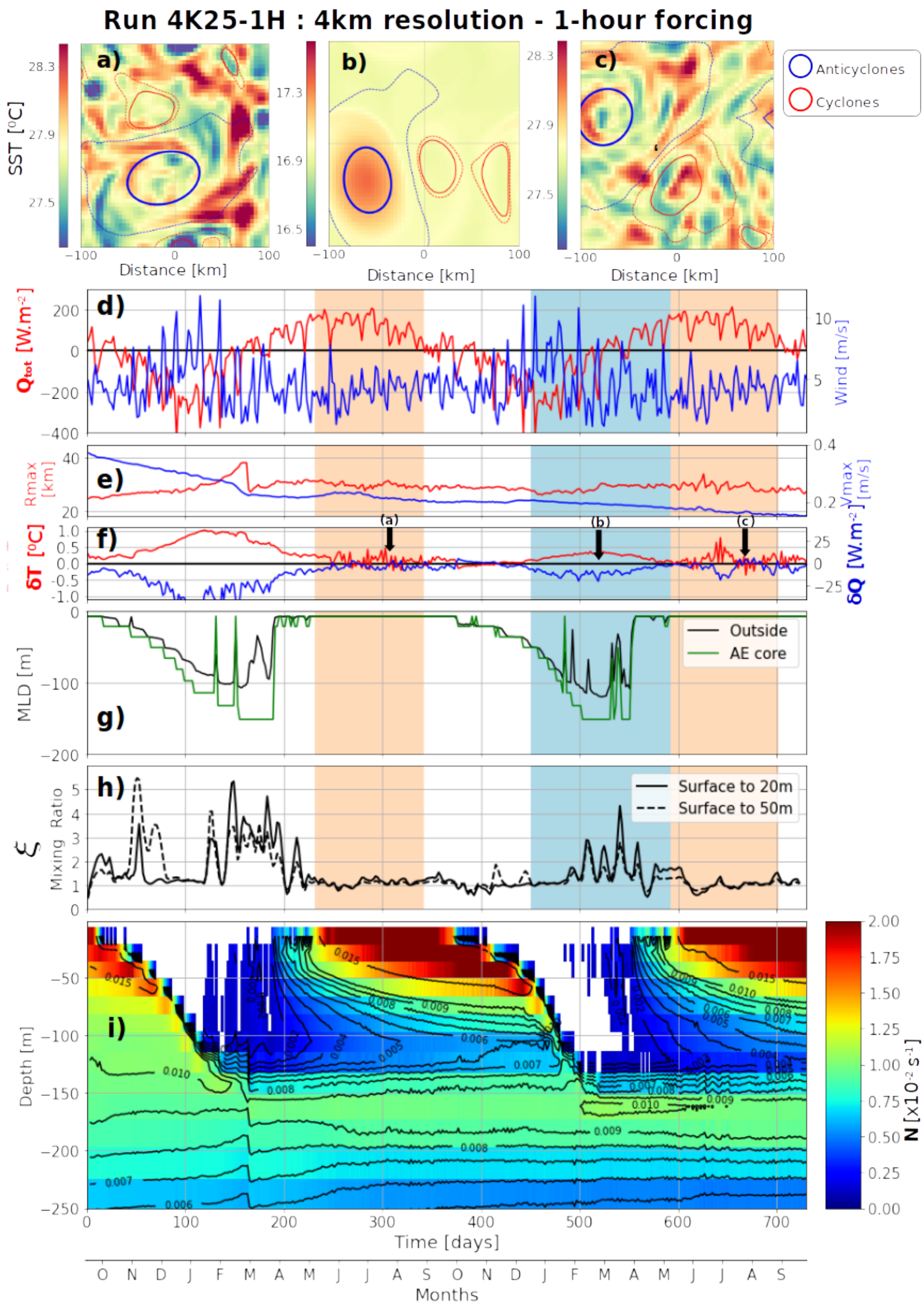

Figure 4. Simulation B from Table1. (a) SST snapshot in the first summer, (b) in the second winter, (c) in the second summer, with eddies detected by AMEDA in contours. The initial anticyclone is highlighted by a thicker line. (d) Net heat flux (red) and windspeed (blue). (e) $R_m(t)$ (red) and $V_m(t)$ (blue) from AMEDA. (f) SST anomaly index δT (red), respectively heat flux anomaly δQ, blue). (g) Mixed layer inside-eddy (green) and outside-eddy (black). (h) Differential mixing ratio ξ defined in Eq.12 with $h = 20m$ (solid) and $h = 50m$ (dashed line). (i) Inside-eddy stratification evolution shown with Brunt-Vaisala frequency ; contours are overlaid with interval $0.001s^{-1}$ and negative values are blanked. On panels c-h, summer periods are indicated by light red shades, winter by a light blue shade.

The same numerical set-up with a finer 1km horizontal resolution (run 1K100-1H in Table 1) shows a net contrast with the previous coarser simulation. This simulation has a 1km horizontal grid size and 100 levels with same stretching parameters giving vertical grid steps close to 3m in the upper 200m. A summer 'inverse' eddy surface temperature is clearly retrieved with 1-hour frequency heat and momentum forcing. As shown in Fig.5 in this configuration, a clear anticyclonic cold-core SST signature is observed in summer (Fig.5a), switching back to a winter warm-core SST the next winter (Fig.5b) and appearing again in the second summer (Fig.5c). This anticyclone surface seasonal oscillation can clearly be tracked by δT (Fig.5f). $\overline{\delta T}$ reached about $-0.2°C$ in the both summers (see Table 1) with spikes of $\delta T \approx -0.5°C$ and maximal value around $+0.4°C$ in winter. Considering anticyclonic cold-core signatures statistics in the Mediterranean Sea ((Moschos et al., 2022) in particular their Fig.5b) $\delta T \approx -0.2°C$ is a low but standard value, anticyclone SST anomalies typically not being colder than $-0.5°C$. This cold-core summer signature goes along with a mixing increase in the upper layers at the eddy core, measured by a diffusivity in summer more than twice as high inside the eddy core as outside. Sensibility of the ξ indicator is shown on Fig.5h, with ξ averaged over the upper 20m or 50m, the first case leading to ξ values higher than 4 in summer despite some variability. This enhanced mixing seems to be confined in the upper layers, as ξ decreases to approximately 1 as soon as the mixed layer deepens, but it increases again to similar values during the second summer.

At depth the maximal mixed layer anomaly reaches about 50m (Fig.5h), very close to the value of the simulation at 4km resolution. However the vertical structure is better reproduced at 1km, and in particular between 100 and 150m deep the $5\times10^{-3}s^{-1}$ stratification isoline closes in December, 4 months later than in the 4km simulation (in August, see Fig.4i). This means that homogeneous waters formed at depth in the first winter restratify more slowly. Eddy decay in time is also slower on maximal speed : after 2 years the anticyclone velocity is about $0.3m.s^{-1}$ with 1km resolution compared to $0.15m.s^{-1}$ with 4km (Fig.4e). Sharp density gradients are smoothed in a coarser simulation, leading to unrealistic temporal evolution of the anticyclones vertical structure. Surface (SST) or depth-integrated (maximal geostrophic speed) measurements are then not accurately reproduced at a spatial resolution of 4km.

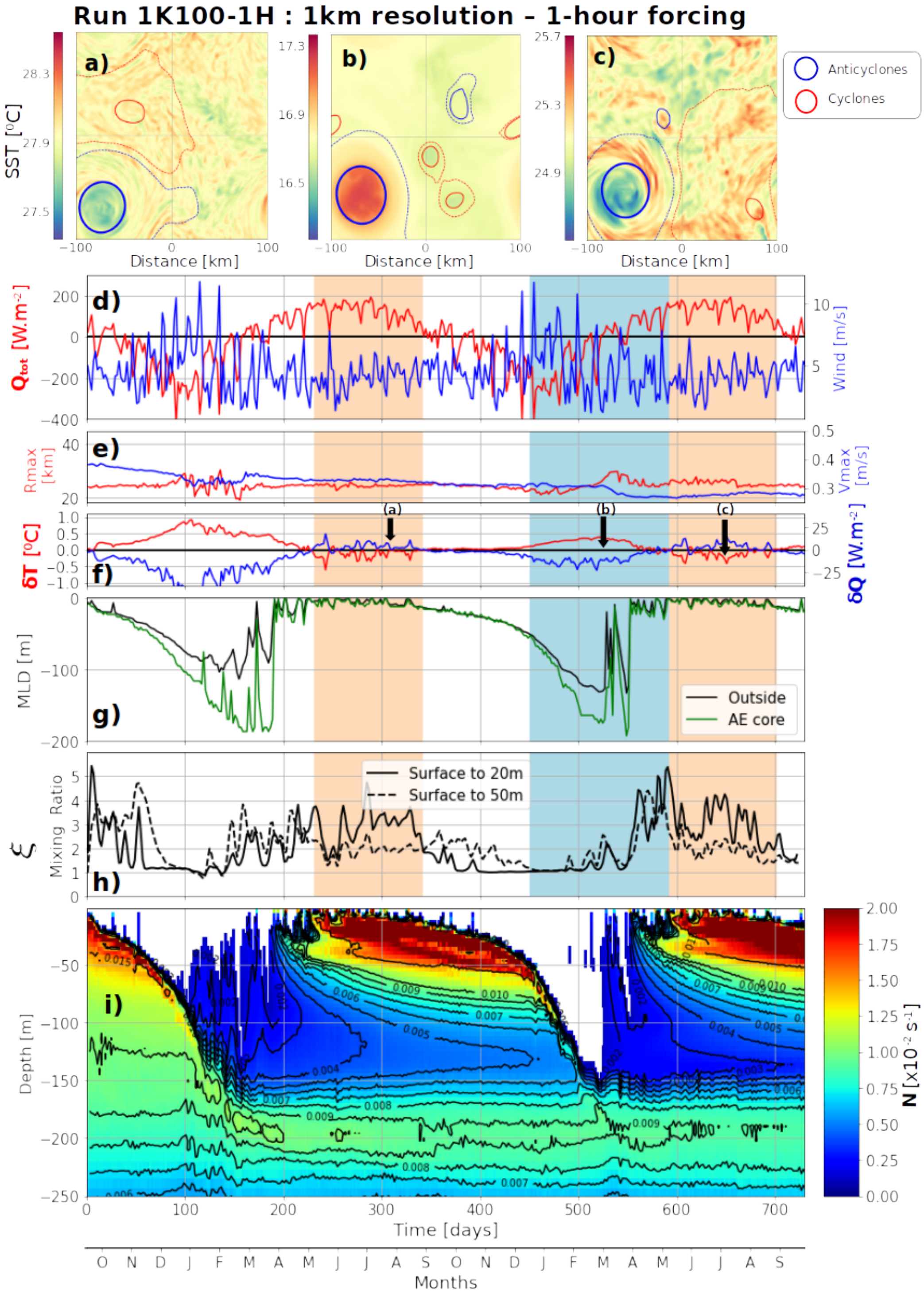

Figure 5. Simulation 1K100-1H from Table1. Same as in Fig.4 but with a 1km horizontal resolution.

An experimental series with the same numerical set-up is performed, increasing horizontal resolution from 4km to 500m (runs 1K100-1H, 2K50-1H, 4K25-1H, 05K150-1H) and vertical resolution accordingly. The horizontal to vertical resolution ratio is kept close to the Brunt-Vaisala to inertial frequencies ratio. It reveal that summer anticyclonic cold-core signature $\overline{\delta T}$ and differential mixing $\bar{\xi}$ both continuously increase when decreasing the grid cell (see Fig.6c). Summer eddy SST inversions are then consistently correlated with an increased mixing. In addition a convergence behavior is observed for mixing at 1km with 100 levels to $\bar{\xi} \approx 3$, as no further mixing is obtained increasing the resolution to 500m and 150 levels. Differential mixing appearing at 1km resolution implies that small scale processes, smaller than the eddy size are at stake. 1km horizontal resolution with a baroclinic first deformation radius around 11km entails that deformation radius to be explicitly resolved, which is not entirely effective for resolution of 2km or larger, similarly to other numerical studies (Marchesiello et al., 2011; Soufflet et al., 2016). On the other hand in winter very similar δT are retrieved at all resolution, with a maximum around $+0.4^{\circ}C$ (Fig.6a) and similar THFF suggesting that winter thermal loss is less affected by horizontal resolution. THFF slightly decreases for lower resolution, likely due to smoothing effect of strong SST patterns.

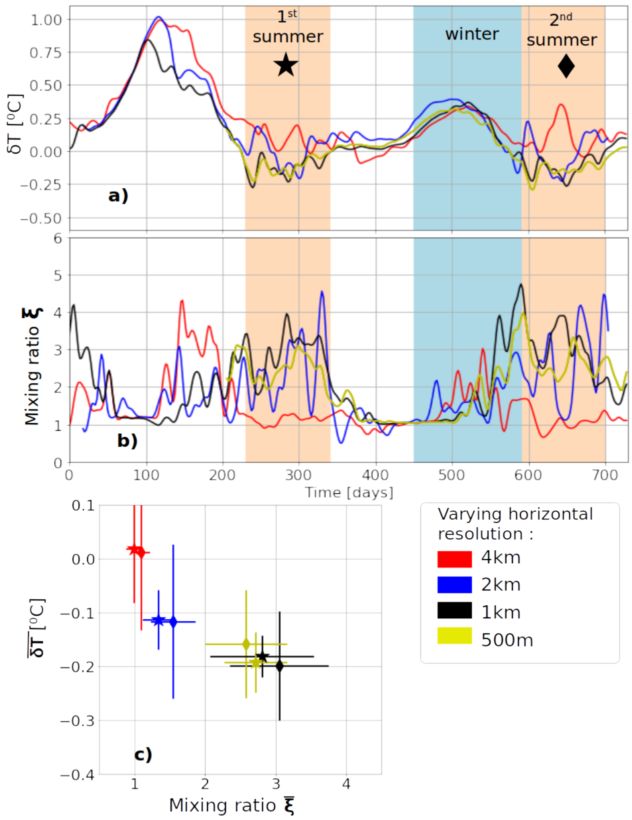

Figure 6. (a) δT and (b) ξ timeseries for experiments 1K100-1H, 2K50-1H, 4K25-1H, 05K150-1H listed in Tab.1 with SST retroaction on air-sea fluxes and varying horizontal resolution frequency. 2-days Gaussian smoothing is applied and summer periods are shaded in light red, winter in light blue. Due to computer memory issues, the first transient winter at 500m resolution was not recorded. (c) Summer-averaged eddy-induced SST anomalies ($\overline{\delta T}$) and mixing ratio ($\bar{\xi}$), with stars for the first summer and diamonds for the second one. Errorbars are ξ spread (30^{th} percentile) over the same period.

For the eddy-induced mixed layer anomaly, similar values are obtained from 4km to 1km resolution ($\Delta MLD \approx 50m$), but a larger $\Delta MLD = 91m$ is retrieved at 500m resolution. This effect could be due to the partial resolution of sub-mesoscale processes such as mixed layer instabilities (Boccaletti et al., 2007; Capet et al., 2008). Maximal background mixed layer deepens when resolution gets finer down to 1km resolution (see Fig.4g and 5g), in consistence with previous experiments (Couvelard et al., 2015). At 500m resolution, a closer look at the MLD evolution inside- and outside-eddy shows that the outside-eddy MLD restratified earlier in run 05K150-1H (in March) than in run 1K100-1H (in April) due to restratification beginning at submesoscale with mixed layer instabilities (Fig.7b). But in both cases inside-eddy MLD reached the same depth (193m, see Fig.7e-f). This suggests that maximal mixed layer inside-eddy indeed reached a maximum driven by air-sea cooling, while restratification outside-eddy occurred too late in run 1K100-1H because vertical buoyancy fluxes are too weak (Capet et al., 2008). Compared to Mediterranean MLD climatology, a restratification in April is indeed quite late (Houpert et al., 2015).

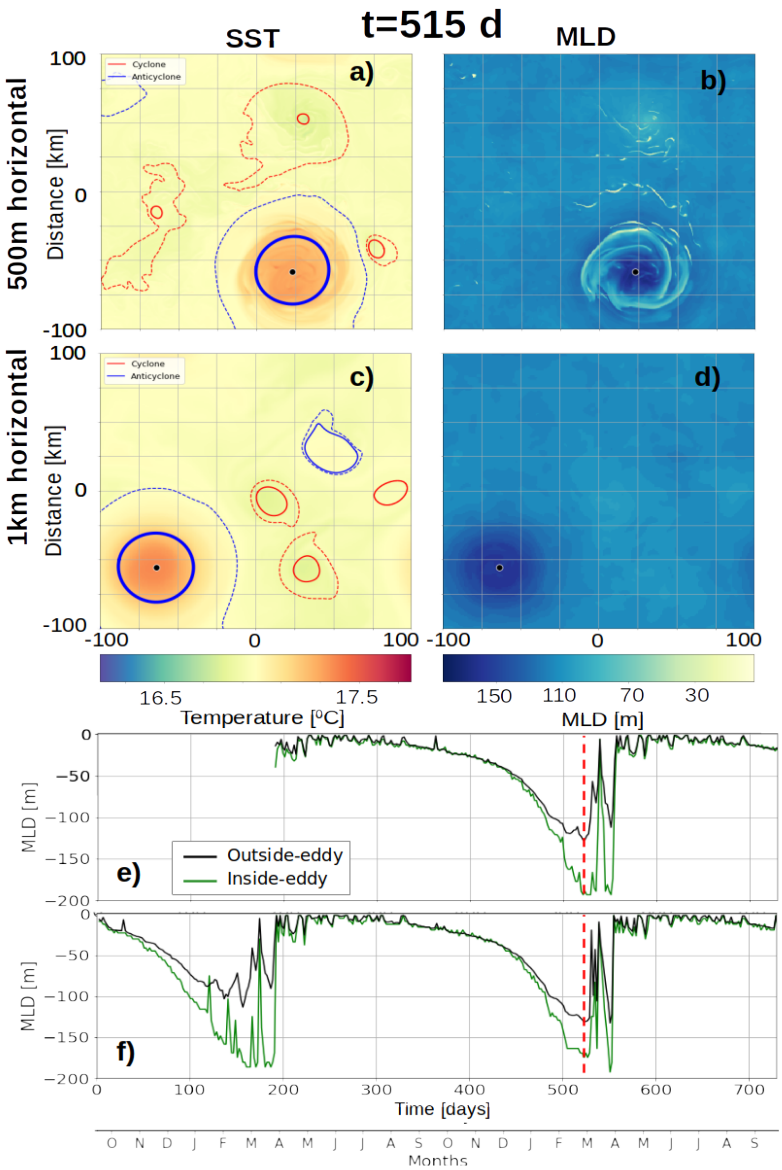

Figure 7. (a) SST with anticyclones and cyclones as in Fig.4 (the initial anticyclone has thicker contour) in 05K150-1H simulation. (b) MLD in 05K150-1H. (c) and (d) : same as (a) and (b) but in 1K100-1H simulation. (e) MLD time series inside-anticyclone (green) and outside-eddy (black) for the 05K150-1H simulation, a red dashed line indicates the time step shown in panels (a)-(d). Due to memory issues, the first transient winter was not recorded. (f) Same as (e) in 1K100-1H simulation.

Mixing patterns over the vertical in the 1km resolution simulation are also consistent with observations. Anticyclones were recently observed to enhance mixing at depth through the propagation of trapped near-inertial internal waves in their core. In studies from Martínez-Marrero et al. (2019) and Fernández-Castro et al. (2020), in situ measurements revealed lower dissipation rate ϵ in anticyclonic homogeneous core than in the neighboring background, and enhanced ϵ below at depth. In our numerical experiments, both diffusivity κ (Fig.8c) and dissipation rate ϵ (Fig.8e) match this feature, with enhanced mixing in summer below the anticyclone, up to one order of magnitude larger from 200 to 300m depth. The anticyclone subsurface core revealed by thick isopycnal displacement on Fig.8e, also shows locally reduced ϵ between 100 and 200m. Fig.8e is then a striking reproduction of dissipation rate section obtained by Fernández-Castro et al. (2020) (see in particular their Fig.5f). However those in situ measurements could not compare outside- and inside-eddy mixing close to the surface, because the value range for ϵ would be too large with surface processes a lot more powerful than deep ocean ones. Numerical simulation enables to reveal that anticyclones also enhance mixing in near surface, with higher ϵ and κ just above the homogeneous core, in the upper 50 meters. The differential mixing ratio ξ previously shown in anticyclone time series then accurately measures a surface-enhanced mixing.

The seasonal cycle of eddy SST signature is then effectively reproduced at 1km horizontal resolution, close to observed value for the example shown above (Fig.3e). eddy SST seasonal shift correlates with increased mixing at the anticyclone core, in consistence with Moschos et al. (2022) hypothesis. This differential mixing is absent at 4km, but appears through $k-\epsilon$ mixing parametrization and converges at 1km resolution.

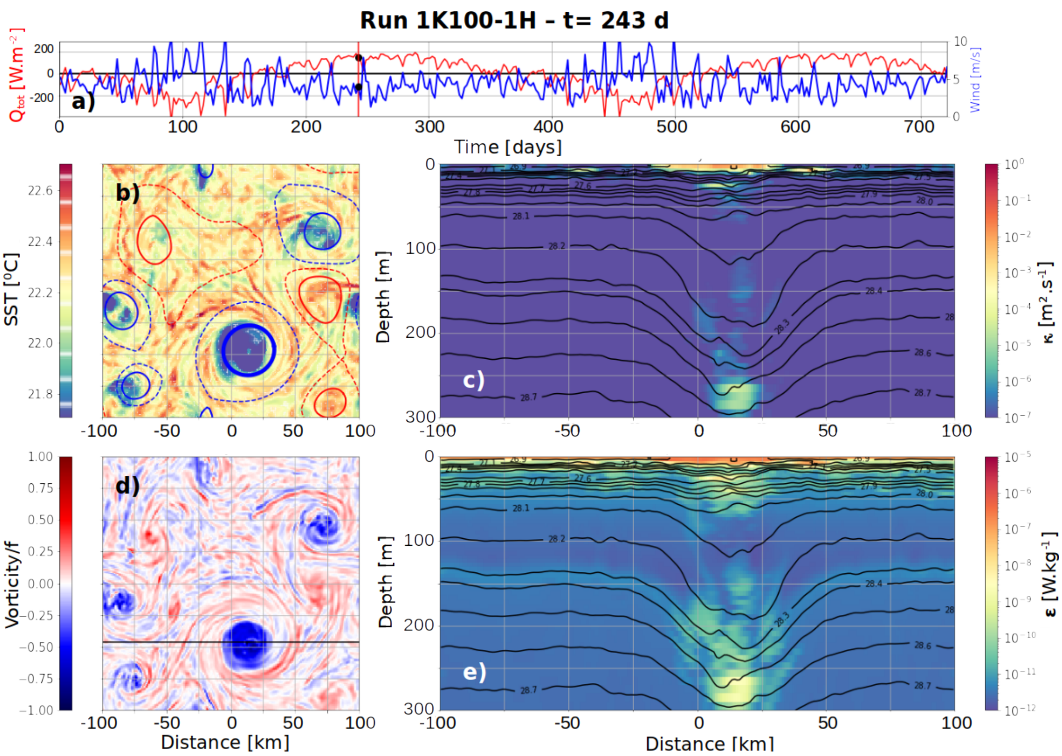

Figure 8. Snapshot at $t = 243\,d$ for the 1K100-1H simulation (see Fig.5). (a) Wind speed (blue line) and Q_{tot} (red lines) timeseries. (b) SST and (d) surface vorticity normalized by f with eddy detections as in Fig.4 (initial anticyclone has a thicker contour). (c) κ and (e) ϵ vertical sections in the upper 300m with logarithmic color scales, in both case the colorbar lower bound is the minimal possible value (see Sect.2.1). Isopycnals are added in black lines.

3.2 Forcing frequency sensitivity

Sensitivity of the eddy SST signature δT and differential mixing ξ to forcing temporal resolution is investigated by progressively removing high frequencies from the atmospheric inputs. These experiences are summarized as 1K100-1D to 1K100-1W in Table 1, using 1-day, 3-days and 1-week atmospheric timeseries respectively. δT and differential mixing ξ timeseries for these experiments are shown in Fig.9a-b. Significantly cold SST signatures ($\overline{\delta T} \lesssim -0.2^{\circ}C$) are obtained together with strong mixing ($\overline{\xi} \approx 3$) for 1-hour and 1-day frequency, but no significant differential mixing is retrieved ($1 < \overline{\xi} < 1.5$) for all lower forcing frequencies (Fig.9c). This threshold behavior is a strong result and shows that spontaneous appearance of differential mixing is driven by small scale and high frequency features. With a Coriolis parameter $f = 9.0 \times 10^{-5}s^{-1} = 1.24cpd$, the inertial period is about 19h, the 1-day forcing can then partly trigger near-inertial waves.

The relationship between $\overline{\delta T}$ and $\overline{\xi}$ is however less clear than for the resolution sensitivity analysis (Fig.6). No differential mixing is observed for forcing frequencies lower than 1 day, but summer cold-core signatures are still found ($\overline{\delta T} \gtrsim 0.1^{\circ}C$, see Table1), even for the 1-week forcing. δT timeseries clearly show for all frequencies a marked seasonal signal (Fig.6a). In particular a significant warm winter signature is always observed, with stable maximal value at $\delta T \approx +0.4^{\circ}C$. In the same context a surprising result is the summer averaged $\overline{\delta T}$ being colder on average at 1-day than 1-hour forcing, despite similar differential mixing. Temporal evolution of eddy SST anomalies reveals this effect to be caused by a larger oscillation of the eddy surface signature (Fig.9a) about $\pm 0.2^{\circ}C$, hence larger errorbars at 1-day on Fig.9c. This suggests that other mechanisms not triggered by high frequency winds also contribute to the eddy SST seasonal cycle. If no differential vertical mixing is observed but if seasonal variations of the anticyclone SST (and hence surface density) is found, one can only hypothesize the role of lateral exchanges. Despite some tries, we were unsuccessful in quantifying eddy lateral exchanges following a varying $R_m(t)$ contour. No particular asymmetric wave modes was observed on SST snapshots, discarding the hypothesis of vortex Rossby waves (Guinn & Schubert, 1993; Montgomery & Kallenbach, 1997).

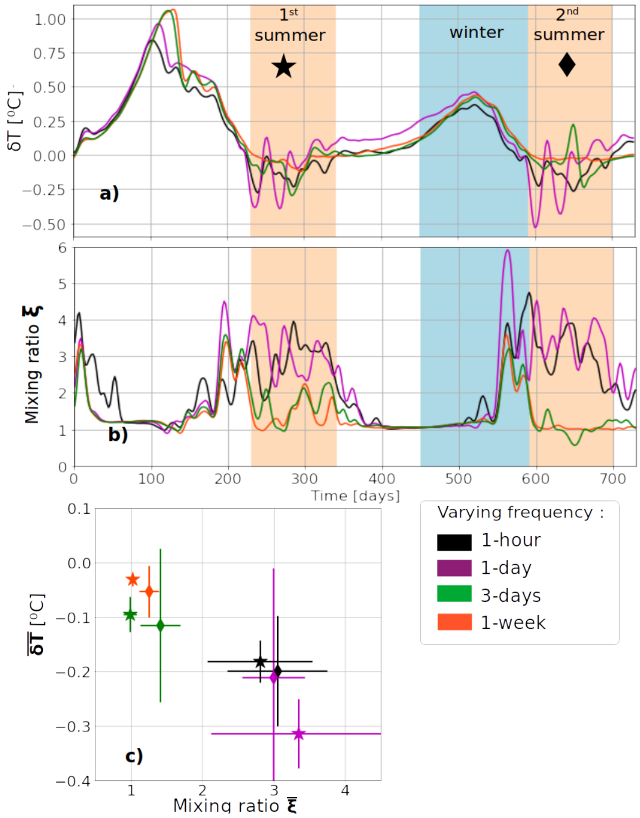

Figure 9. (a) δT and (b) ξ timeseries for experiments 1K100-1H, 1K100-1D, 1K100-3D and 1K100-1W listed in Tab.1 with SST retroaction on air-sea fluxes and varying forcing frequency. 2-days Gaussian smoothing is applied, summer periods are shaded in light red, winter in light blue. (c) Summer-averaged eddy-induced SST anomalies ($\overline{\delta T}$) and mixing ratio ($\overline{\xi}$), with stars for the first summer and diamonds for the second one.

Near-inertial internal waves are investigated using Fourier transforms on vertical speed anomalies in run 1K100-1H. We focus on a single vertical level at 20m in near-surface where the enhanced mixing occurs (see Fig.8c). Transforms are computed only in the second summer (590 to 700 simulated days) with a 1-hour sampling frequency. Following Babiano et al. (1987), inside-eddy spectrum is performed keeping only the inside-eddy area (around the eddy center with radius $2/3R_m(t)$) and the remaining area is set to 0 before performing the Fourier transform. Similarly outside-eddy spectrum is performed blanking all value inside any eddy contours. The results clearly show a differential effect inside-eddy vertical kinetic energy density revealing a second powerful peak at the effective inertial frequency $f_e = f + \zeta/2 \approx 1.0cpd$, lower than the inertia frequency (Fig.10a). Outside-eddy spectrum (Fig.10b) shows only one peak at the inertial frequency, and internal waves cannot propagate at lower frequencies due to the f-cut-off (Garrett & Munk, 1972). Normalizing by the investigated area, total vertical kinetic energy per unit surface is indeed higher inside the anticyclone ($4.19{\times}10^{-14}m^2.s^{-2}/m^2$) than outside-eddy ($1.64{\times}10^{-14}m^2.s^{-2}/m^2$) due to these powerful subinertial internal waves. An assumption of this method is however to assume that both inside- and outside-eddy areas roughly keep the same area, which is verified. This result is consistent with (Kunze, 1985) theory and recent numerical works (Danioux et al., 2015; Asselin & Young, 2020) subinertial waves ($\omega \lesssim f$) can be trapped in the anticyclone due to the locally lower absolute vorticity, and enhance mixing while breaking as proposed by Fernández-Castro et al. (2020).

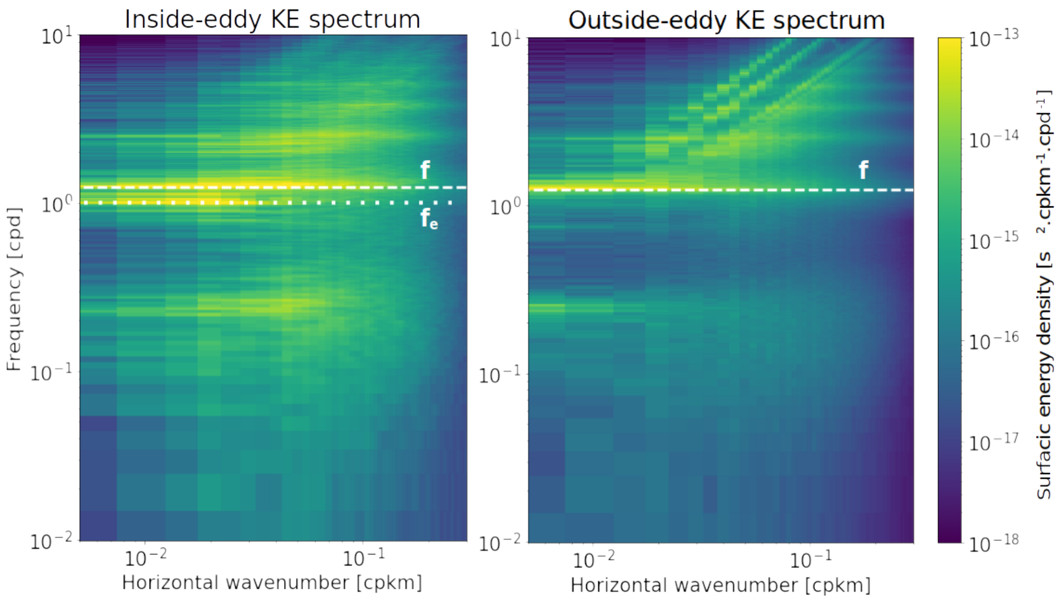

Figure 10. (a) Inside-eddy and (b) outside-eddy vertical kinetic energy density spectrum at 20m depth. For comparison, spectrum are normalized by the area of interest. Analysis performed on simulation 1K100-1H with 1-hour sampling. Normal (respectively effective) inertial frequencies $f = 1.24cpd$ ($f_e \approx 1.0cpd$) are highlighted by a white dashed (dotted) line.

3.3 Air-sea fluxes sensitivity

Sensitivity of the anticyclone temporal evolution to air-sea fluxes components is further investigated. A $1km$ resolution simulation experiment is run similarly as the 1K100-1H simulation without applying SST retroaction on air-sea fluxes (see Sect.2.3, run 1K100-1H-NoSST in Table 1). Although quite unrealistic, this experiment enables to check if the eddy SST anomaly seasonal shift and differential mixing observed in previous simulations are triggered by air-sea fluxes retroaction. Time series for SST reveals that eddy SST anomalies seasonal oscillation is retrieved without SST retroaction (Fig.11a-c), and summer cold-core signatures are even stronger : $\delta T \approx -0.8^{\circ}C$ the first summer and $\approx -0.8^{\circ}C$ the second one (Fig.11f). Simultaneously, differential mixing reaches $\xi \approx 3$, approximately the same value as run 1K100-1H (Fig.11h). This confirms that differential eddy mixing triggering the eddy SST variations is not linked to air-sea fluxes retroaction. However this feedback can modulate and dampen the δT seasonal cycle leading to reduced anomalies.

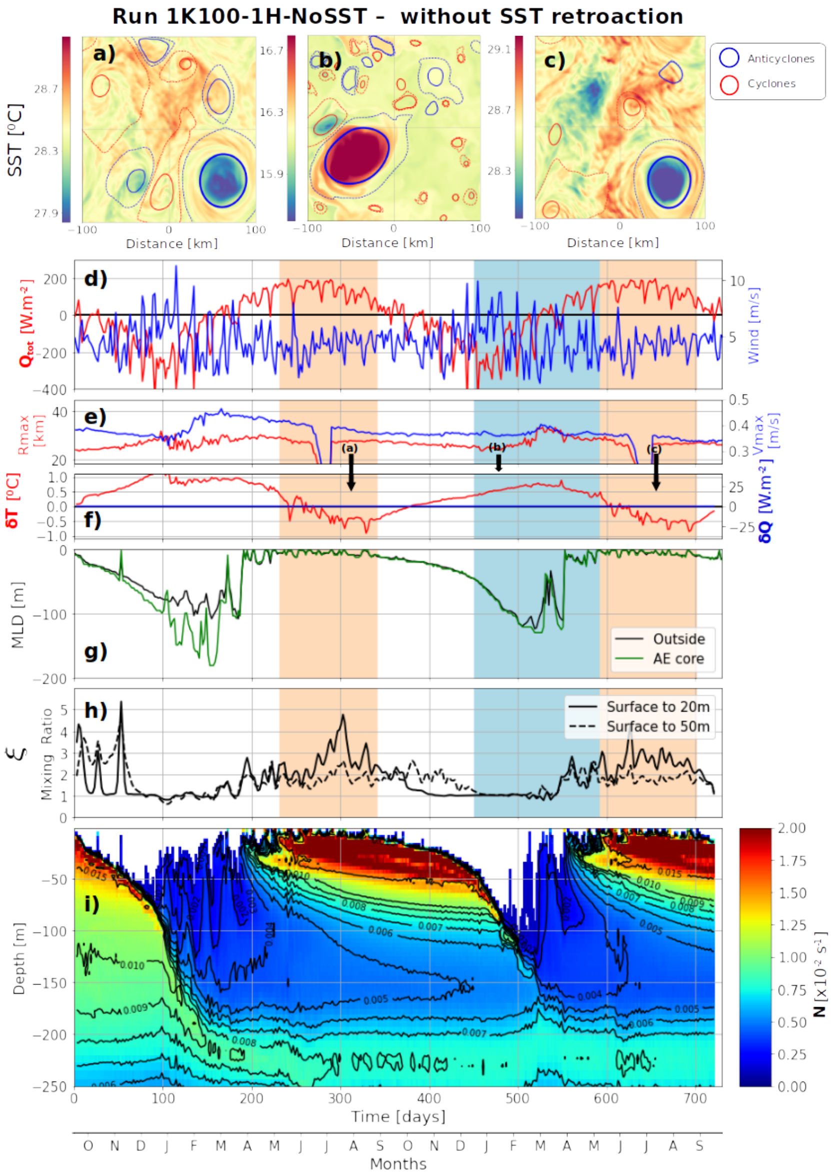

Figure 11. Simulation 1K100-1H-NoSST from Table 1. Same as in Fig.4 but without SST retroaction on air-sea fluxes. Discontinuities in R_{max} and V_{max} in panel (e) are due to the anticyclone crossing twice the grid borders.

SST retroaction acting as a negative feedback on SST anomalies can be analytically expected as linear. The derivative of each heat component with respect to T_s is indeed approximately constant (T_s being in Kelvin in Eq.13). Transfer coefficients C_E and C_S are indeed much more dependent on wind speed than on temperature, varying roughly about 0.2 with a T_s change of $1K$. The most sensitive case is a low air-sea temperature difference with weak wind, in which the boundary layer can switch from stable to unstable conditions (see for instance Fig.A1b from Pettenuzzo et al. (2010)). Assuming C_E and C_S are roughly constant with respect to temperature one gets :

$$\frac{\partial Q_{LW}^{\uparrow}}{\partial T_s} = -4\epsilon_{sb}\sigma_{sb}T_s^3 \approx -6\,W.m^{-2}.K^{-1} \tag{13}$$

$$\frac{\partial Q_{Lat}}{\partial T_s} \approx -\frac{\rho_a L_E C_E |V| 0.610}{P_{SL}} \frac{dP_{sat}}{dT_s} \approx -3 \times 10^1\, W.m^{-2}.K^{-1} \tag{14}$$

$$\frac{\partial Q_{Sen}}{\partial T_s} = -\rho_a c_p C_S |V| \approx -1 \times 10^1\, W.m^{-2}.K^{-1} \tag{15}$$

Altogether a thermal feedback on the order of $\frac{dQ_{tot}}{dT_s} \approx -4 \times 10^1 W.m^{-2}.K^{-1}$ is then expected, mostly driven by latent heat flux. THFF in Table 1 is computed only on the whole simulated year (from 365 to 730 days) and a value of $\approx -40\, W.m^{-2}.K^{-1}$ is retrieved with a simple SST retroaction, in consistence with Eq.13 to 15. This value is relatively constant in our simulations, slightly decreasing for coarser resolution and lower forcing frequencies (see Table 1). $\partial C_E/\partial T_s$ and $\partial C_S/\partial T_s$ being also positive, taking this into account in Eq.14 leads to a even higher THFF estimate. THFF for the 1K100-1H simulation, defined here as δQ as a function of δT is shown in Fig.12. The obtained thermal feedback is consistent with previous estimates in coupled climate model : Ma et al. (2016) found a higher THFF ranging between 40 and $56 W.m^{-2}.K^{-1}$ but in the specific area of very warm eddies of the Kuroshio extension region. Moreton et al. (2021) found THFF ranging between 35 and $45\, W.m^{-2}.K^{-1}$ over mesoscale eddies. They however used a composite approach in a model coupled with atmosphere and maximal oceanic resolution of 1/12°, for effective radius about $40km$. A coupled atmosphere layer is expected to further dampen the total THFF, taking into account other feedbacks than SST, in particular evaporation. Humidity is expected to increase over warm eddy, consequently decreasing the latent heat flux driving evaporation, whereas we applied a uniform h_{2m} field. Similar THFF in our simulations compared to coupled ocean-atmosphere models suggests that our results would not change significantly with more complex heat flux retroaction.

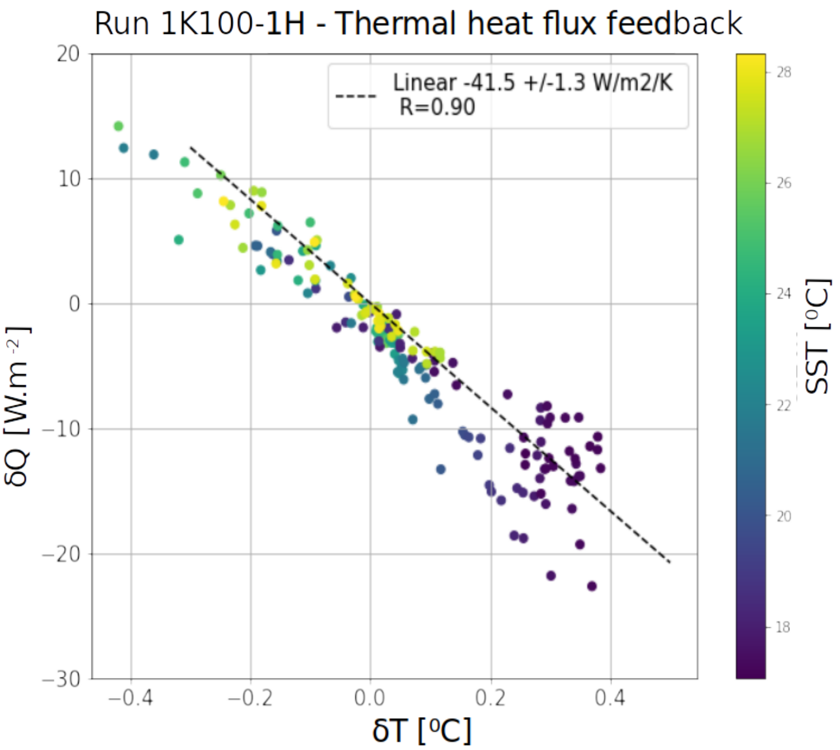

Figure 12. Thermal heat flux feedback in run 1K100-1H on the 2^{nd} simulated year, with linear regression as dashed black line, δQ and δT are from Fig.5f. Regression coefficient and parameters are indicated in the legend.

Without SST retroaction on air-sea fluxes, the most important difference from run 1K100-1H is the MLD anomaly variations. Outside-eddy, mixed layer evolution is very similar in runs 1K100-1H and 1K100-1H-NoSST reaching about 100m at its winter maximum, but the eddy MLD anomaly is an order of magnitude smaller ($\Delta MLD = 10m$, see Fig.11h). With no THFF, the MLD deepens at the same rate outside- and inside-eddy. Winter MLD deepening can be computed estimating the thermal loss ΔT, assuming a linear thermal linear stratification $\partial_z T$:

$$MLD = \frac{\Delta T}{\partial_z T} \tag{16}$$

The thermal loss is the integration of the heat flux over winter duration D. Assuming stratification is at first order the same outside- and inside-eddy, MLD anomaly is then driven by heat flux lateral gradients :

$$\Delta MLD = \frac{D}{\rho_0 c_p \partial_z T} \delta Q \tag{17}$$

In the 1K100-1H with SST retroaction on air-sea fluxes, δQ is positive in winter reaching about $+15 W.m^{-2}$ over 4 months. This leads to an estimate $\Delta MLD \approx 2 \times 10^1 m$. This estimate should then be the eddy MLD anomaly contribution from THFF alone, but a lot higher difference is obtained between run 1K100-1H and 1K100-1H-NoSST. The main assumption in Eq.17 is that $\partial_z T$ is roughly the same inside- and outside-eddy. This is true in the upper layers where stratification is mostly the seasonal thermocline (see isopycnals in Fig.8c-d). At depth lower than 100m however, the anticyclone constitutes a more homogenized layer and this assumption should not hold as MLD should deepen faster inside-eddy, even with no SST retroaction. The very low ΔMLD found with no THFF then suggests that thermal feedback may also impact inside-eddy stratification. An example of inside-eddy MLD faster deepening is shown in Fig.3g : the MLD connects in February 2018 with the layer homogenized the previous winter and reaching quickly about 300m. Such mixed layer deepening acceleration is partly retrieved in run 1K100-1H around 500 days, with a MLD jump of about 30m (Fig.5g) inside-eddy but only about 10m outside-eddy. This coincides with the mixing of the subsurface homogenized layer formed in the first winter, despite diffusion (stratification isolines progressively closing, Fig. 5i) as discussed earlier.

ΔMLD is however still relatively weak compared to the 200 to 300m MLD anomalies observed in Mediterranean anticyclones (Barboni, Coadou-Chaventon, et al., 2023). Two main hypotheses can be proposed, the first being that some interannual variability is needed. The second hypothesis is that layers homogenized by winter MLD progressively restratify at depth in summer due to numerical diffusion. MLD in the following winter will then have to break this numerical stratification. This second hypothesis entails that the vertical grid is not enough refined yet to correctly preserve homogenized layers from one winter to another. From the comparison between runs 1K100-1H and 1K100-1H-NoSST shows that SST retroaction on air-sea fluxes is necessary to obtained eddy MLD anomalies, but quantitative description deserves further research and ΔMLD is not only driven by fluxes gradients at the eddy scale.

Conclusions

Idealized numerical experiment at high horizontal resolution and high frequency atmospheric forcing are able to qualitatively and quantitatively retrieve SST signature seasonal cycle for a mesoscale anticyclone. Starting from a surface intensified mesoscale anticyclone at $Ro \approx 0.16$, seasonal oscillations of the eddy SST anomalies are recovered with an 1km resolution, hourly atmospheric forcing and SST retroaction on air-sea fluxes. Retrieved eddy anomalies are a warm winter SST feature at $\delta T \approx +0.5^{\circ}C$ and a cold summer SST at $\delta T \approx -0.2^{\circ}C$, in consistence with in situ observations. The shift from warm winter SST signature to summer cold one is partly explained by an increased vertical mixing in the anticyclone upper layers. This differential mixing is due to higher NIW energy propagation well captured through the $\kappa - \epsilon$ mixing parametrization.

A sensitivity analysis reveals that this differential mixing depends on the grid resolution. Model diffusivity near the surface is then consistently 3 times higher in summer inside-eddy than outside for horizontal resolution of 1km or smaller. This resolution corresponds to an explicitly resolved first baroclinic deformation radius. Sensitivity to the forcing frequency is investigated by progressively removing high frequencies from the atmospheric input fields. A threshold behavior is observed when forcing frequency is lower than a day, then differential mixing dramatically vanishes with no significant summer cold-core anticyclonic SST. Vertical kinetic energy signing internal wave propagation indeed reveals a second powerful peak at $\omega = 1.0cpd$ inside the anticyclone in near-surface, corresponding to the effective inertial frequency and responding to high frequency forcing. This peak is absent outside-eddy because the cut-off inertia frequency $f = 1.24cpd$ is higher. Such an analysis suggests a significant impact of the eddy vorticity as cut-off frequency in allowing or not the selective NIW propagation. Weaker eddy SST seasonal oscillations are also retrieved in the absence of high frequently forcing and consequently without differential mixing (3-days and 1-week experiments). This highlights that other contributions might participate to these eddy SST signatures, in particular lateral exchanges. A new question for future research opened by this eddy-modulated mixing is how it depends on the eddy vorticity and size.

SST retroaction on air-sea fluxes is not found to be responsible of eddy SST signatures seasonal shift, as the seasonal oscillation is retrieved with and without air-sea fluxes parametrization. However this retroaction is logically found to dampen the SST anomalies, and then reduces eddy anomalies magnitude in both summer and winter. The average thermal heat flux feedback of our mesoscale anticyclone is approximately $40W.m^{-2}.K^{-1}$, in consistence with analytical derivation and previous studies.

Significant eddy-induced mixed layer anomaly $\Delta MLD \approx 50m$ are found at 1km horizontal resolution, only in the presence of SST retroaction on fluxes. Linear MLD anomaly analysis suggests that the thermal feedback is only responsible for about half of the MLD anomaly. Further analysis should then investigate how SST retroaction impacts inside-eddy stratification. MLD anomalies do not completely converge at 1km as larger anomalies are obtained with a 500m resolution due to restratification beginning outside-eddy driven by submesoscale instabilities, despite similar maximal mixed-layer at the anticyclone core. No restratification delay is clearly observed, but it could occur at even higher resolution inside the anticyclone because the balanced density gradients inhibits mixed layer instabilities there. This hypothesis is consistent with observations (Barboni, Coadou-Chaventon, et al., 2023) but would deserve more investigation in the future. This result is also important as the mixed layer is a significant driver of atmospheric and bio-geochemical exchanges, and the explicit resolution of submesoscale processes might be needed to accurately reproduce their interaction with eddies (Capet et al., 2008; Lévy et al., 2018). An important result is still that significant ΔMLD is retrieved only when SST exerts a retroaction on air-sea fluxes, but the quantitative description of its evolution would deserve more analysis.

This is the first time that subinertial waves concentration in anticyclones is linked to an increased mixing in near surface, spontaneously retrieved through the $k{-}\epsilon$ mixing closure. Mixing modulation by eddies suggests a strong scale interactions between subinertial internal waves ($\omega \lesssim f$) and the mesoscale ($\omega \ll f$). Differential mixing triggered by high frequency winds is an important result highlighting the need of both fine resolution and atmospheric forcing at sufficiently high frequency to correctly reproduce mesoscale eddies evolution. At present stage, global operational models do not have the spatial resolution to capture these phenomena. According to this study, 1/120 °resolution with 100 vertical levels would then be necessary to reproduce accurately mesoscale temporal evolution.

Open Research Section

In-situ profiles colocalized with mesoscale eddies database is available at `https://doi.org/10.17882/93077`. AMEDA eddy tracking algorithm is open source and available at `https://github.com/briaclevu/AMEDA`. ERA5 atmospheric reanalysis are publicly available at `https://doi.org/10.24381/cds.adbb2d47`. The CROCO code is publicly available at `https://www.croco-ocean.org/`.

Acknowledgments

Authors gratefully acknowledge *Ifremer* and *Service Hydrographique et Océanographique de la Marine* for their use of the *Datarmor* computing facility. Authors acknowledge Evangelos Moschos (*Amphitrite*) for the reuse of his figures as snapshots in Fig.3a-d. Authors also acknowledge fruitful discussions with Clément Vic (*Ifremer*), in particular the comparison with observations in Fig.8 and 10.

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How high frequency atmospheric forcing impacts mesoscale eddy surface signature and vertical structure

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Key Points:

- Seasonal variations in SST and mixed layer anomalies of a Mediterranean anticyclone are retrieved in high resolution simulation
- The summer surface cold-core signatures are driven by differential vertical mixing due to near-inertial waves
- SST retroaction on air-sea fluxes is necessary to retrieve eddy-induced mixed layer anomalies

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Abstract
Seasonal evolution of both surface signature and subsurface structure of a Mediterranean mesoscale anticyclones is assessed using the CROCO high-resolution numerical model with realistic background stratification and fluxes. In good agreement with remote-sensing and in-situ observations, our numerical simulations capture the seasonal cycle of the anomalies, induced by the anticyclone, both in the sea surface temperature (SST) and the mixed layer depth (MLD). The eddy signature on the SST shifts from warm-core in winter to cold-core in summer, while the MLD deepens significantly in the core of the anticyclone in late winter. Our sensitivity analysis shows that these dynamical properties can be accurately reproduced only if the resolution is high enough ($\sim 1km$ for the horizontal with 100 vertical levels in a Mediterranean stratification) and if the atmospheric forcing contains high-frequency. In this configuration the deformation radius is explicitly resolved and the vertical mixing parametrized by the $k-\epsilon$ closure scheme is three times higher inside the eddy than outside the eddy. This differential mixing is explained by near-inertial waves, triggered by the high-frequency atmospheric forcing.Near-inertial waves propagate more energy inside the eddy because of the lower effective Coriolis parameter in the anticyclonic core. In addition to these high spatial and temporal resolution, SST retroaction on air-sea fluxes appears to be necessary to obtain marked eddy mixed layer depth anomaly.

Plain Language Summary

Mesoscale eddies are turbulent structures present in every regions of the world ocean, and accounting for a significant part of its kinetic energy budget. These structures can be tracked in time and recently revealed a seasonal cycle from in situ data. An anticyclone (clockwise rotating eddy in the northern hemisphere) is observed in the Mediterranean to be predominantly warm at the surface and to deepen the mixed layer in winter, but shifts to a cold-core summer signature. This seasonal signal is not yet understood and studied in ocean models. In this study we assess the realism of an anticyclone seasonal evolution in high resolution numerical simulations. Eddy surface temperature seasonal shift is retrieved and is linked to an increased mixing at the eddy core spontaneously appearing at high resolution (grid size $\sim 1km$) in the presence of high frequency atmospheric forcing. This increased mixed is due to the preferred propagation of near-inertial waves in the anticyclone due to its negative relative vorticity. Eddy-induced mixed layer depth anomalies also appear to be triggered by sea surface temperature retroaction on air-sea fluxes. These results suggest that present-day operational ocean forecasting models are too coarse to accurately retrieve mesoscale evolution.

1 Introduction

Mesoscale eddies are ubiquitous turbulent structures in the oceans, in thermal wind balance with a signature in density. Eddies have been observed for a long time, in particular in energetic regions such as the Gulf Stream rings (Richardson, 1980), the Kuroshio extension (Itoh & Yasuda, 2010), the Agulhas current retroflexion (Olson & Evans, 1986) , but also in the Mediterranean Sea (Mkhinini et al., 2014). Mesoscale anticyclones (respectively cyclones) are structures of negative (positive) density anomaly, identified with a sea surface height (SSH) elevation (depression) (Chelton et al., 2007). Eddies statistical descriptions really began with the availability of eddy automated detections based on gridded altimetry products (Doglioli et al., 2007; Chaigneau et al., 2009; Nencioli et al., 2010; Chelton, Schlax, & Samelson, 2011; Mason et al., 2014; Le Vu et al., 2018; Laxenaire et al., 2018). The first quantitative studies were done in a composite approach : many daily snapshots detections are colocated with eddy contours and gathered into a single annual mean eddy signature (Hausmann & Czaja, 2012; Everett et al., 2012). This approach combined with remote-sensing measurements provide an extensive view of eddies in various regions of the global ocean, with SST, sea surface salinity (Trott et al., 2019), chlorophyll (Chelton, Gaube, et al., 2011) and also meteorological variables (Frenger et al., 2013). Composite approach also allowed to reveal a modulation of air-sea fluxes at the eddy scale : in the Agulhas retroflexion region, (Villas Bôas et al., 2015) showed the total heat flux to the atmosphere to be enhanced over very strong and warm anticyclones. Similarly for the eddy vertical structure, gathering Argo profiles as a function of normalized distance to the eddy center, eddies were found to influence the mixed layer depth (MLD) (Sun et al., 2017; Gaube et al., 2019). Anticyclones have deeper MLD in their core, cyclones shallower MLD, with larger mixed layer anomalies in winter. Eddies were also observed to incorporate a significant seasonal cycle in their radius variations (Zhai et al., 2008) and their SST signature (Sun et al., 2019; Y. Liu et al., 2021). Anticyclones (respectively cyclones) usually identified as warm in surface, actually shift to cold (warm) signatures in summer in several regions of the world ocean (Sun et al., 2019; Moschos et al., 2022). This phenomenon is then referred to as 'inverse' SST signatures. (Moschos et al., 2022) showed that these 'inverse' signatures actually become predominant in summer in the Mediterranean Sea, a seasonal shift yet not properly understood.

The composite approach is nonetheless ill-suited to study eddy temporal variability due to the stacking of numerous observations in time. Recently Lagrangian approaches were developed to study eddies enabling to better track their temporal variability (Pessini et al., 2018; Laxenaire et al., 2020; Barboni et al., 2021). Using a Lagrangian approach, (Moschos et al., 2022) showed that the same individual anticyclones shift from a warm winter SST anomaly to a cold one in summer (and conversely for cyclone). With the additional Argo floats trapped in anticyclones, they further noticed that anticyclonic density anomaly remain warmer at depth while becoming colder in surface, leading to a smoother density gradient. Hence the hypothesis that this seasonal shift could be explained by a modulation of the vertical mixing by mesoscale eddies, anticyclones (cyclones) likely enhancing (decreasing) mixing in surface. Recent observations in the Mediterranean Sea of inside-anticyclone properties temporal evolution further revealed eddy mixed layer anomalies to be much larger than the composite approach mean value, reaching sometimes 300m (Barboni, Coadou-Chaventon, et al., 2023). MLD anomalies evolution was also shown to have evolution much faster than the month, with delayed restratification inside anticyclones. Mechanisms driving these MLD anomalies are also unexplained, but (Barboni, Coadou-Chaventon, et al., 2023) found it to be impacted by interactions with the anticyclone vertical structure.

An eddy modulation of vertical mixing was not proven so far but could be due to a modulation of near-inertial waves (NIW) propagation. NIW can not propagate at frequencies lower than the inertial frequency f due to Earth rotation (Garrett & Munk, 1972). However in the presence of a balanced flow, anticyclones (cyclones) having negative (positive) relative vorticity ζ locally shift this cut-off by an effective inertial frequency $f_e = f + \zeta/2$ (Kunze, 1985). Sub-inertial waves ($\omega \lesssim f$) can then remained trapped in anticyclones and supra-inertial waves ($\omega \gtrsim f$) can be expelled from cyclones. Consequently, NIW propagate more inside anticyclones, what was experimentally (D'Asaro, 1995) and numerically (Danioux et al., 2008, 2015; Asselin & Young, 2020) proven. This NIW trapping potential partly explains the interest in anticyclones rather than in cyclones, the other reason likely being that anticyclones are more stable in time (Arai & Yamagata, 1994; Graves et al., 2006), in particular for large structures (Perret et al., 2006), then more easily detected and trapping more often profilers (thus easing field campaigns). Several recent observations (Martínez-Marrero et al., 2019; Fernández-Castro et al., 2020) showed that mixing at depth is enhanced below anticyclones due to this more energetic NIW propagation. On the other hand numerical studies assumed extremely simplified set-up with constant wind (Danioux et al., 2008) or an idealized wind burst (Asselin & Young, 2020). They also looked at NIW propagation in an eddying field at short time scales, then without significant evolution of the eddies and stratification. Eddy-NIW interaction on longer time scales - eddy evolving time scales like months - in a varying stratification due to seasonal cycle has never been assessed so far. In particular the effect of this differential NIW propagation on eddies remains unknown and a gap remains between wave propagation and enhanced surface mixing.

Some recent studies started to assess eddy temporal evolution in high resolution regional models. In the Mediterranean Sea, Escudier et al. (2016) compared eddy size, drift and lifetime compared to eddies in altimetric observations. Mason et al. (2019) investigated these variables in assimilated operational models and additionally looked at MLD anomalies, but both were in a composite approach and did not look at eddy SST variations. More recently Stegner et al. (2021) performed an observation system simulation experiment on a 1/60°simulation of the Mediterranean sea and found great bias on size and strength for small eddy detections, but did not look at SST variations. Using the same simulation, an interesting method was developed by Ioannou et al. (2021), investigating differences in both trajectories, size and stratification of the Ierapetra anticyclonic eddy, but restricted to this particular case.

Both eddy SST anomalies seasonal shift and mixed layer depth anomalies remain poorly investigated so far in ocean models. If NIW propagation and eddy vertical structure are considered, grid resolution - both horizontal and vertical - and atmospheric forcing are likely key aspects to take into account. Indeed air-sea fluxes and near-inertia-gravity waves involve much shorter temporal and spatial scales, not reproduced even in eddy-permitting models at present stage. We then aim to assess the realism of an anticyclone seasonal signal in both surface and mixed layer using an idealized but high-resolution simulation and investigating driving physical processes. The goal is to assess the realism of the eddy temporal evolution compared to similar observations, in particular the retrieval of the surface signature seasonal cycle. In a first part we conduct a sensitivity analysis on horizontal grid cell. In a second part we study the sensitivity to atmospheric forcing frequency. Last, the effect of SST retroaction on air-sea fluxes is discussed.

2 Methods

2.1 Model set-up

Idealized numerical experiments are performed using the Coastal and Regional Ocean Community (CROCO) model. CROCO is based on the Regional Ocean Modeling System (ROMS) kernel (Shchepetkin & McWilliams, 2005). It uses a time splitting method between the fast barotropic mode and the slow baroclinic ones. Advection schemes are UP3 for horizontal and Akima-Splines for the vertical. Trying to conciliate realistic and idealized approach, we use double periodic conditions in a realistic stratification and on long timescale. The atmospheric forcing has realistic temporal variations but is spatially homogeneous. The only active tracer used is temperature. As a consequence, a linear state equation links density ρ and temperature T, with thermal expansion $T_c = 0.28 kg.m^{-3}.K^{-1}$ and linear approximation close to $T_0 = 25$°C and $\rho_0 = 1026 kg.m^{-3}$:

$$\rho = \rho_0 + T_c(T - T_0) \tag{1}$$

Discarding salinity effects is justified by the very weak salinity seasonal cycle in the Mediterranean Sea. The heat flux seasonal cycle is roughly $\pm 150\, W.m^{-2}$ (Pettenuzzo et al., 2010), whereas salinity fluxes are mostly driven by the evaporation minus precipitation balance, with a mean of roughly $10^3 mm/y$, a seasonal cycle maximal amplitude of $\Delta F = 4 \times 10^2 mm/y$ and river input being negligible (Mariotti, 2010). Considering a haline contraction coefficient of $S_c = 0.78 kg.m^{-3}.PSU^{-1}$, a ΔF freshwater input would have a seasonal equivalent effect on buoyancy $Q_{eq} = \rho_0 c_p \frac{S_c}{T_c} S_0 \Delta F \approx 5\, W.m^{-2}$, indeed almost two orders of magnitude lower than Q_{tot}.

Grid

Simulation domain is double periodic, on the f-plane, with a flat bottom $H_{bot} = 3000m$. Horizontal extent is 200km in both directions, with horizontal resolution ranging between 4km and 500m, with respectively 25 to 150 vertical levels. Horizontal to vertical grid size ratio is kept roughly constant about 1000/3 in the upper layers, close to the Brunt-Vaisala to inertia frequency ratio. Coriolis parameter is $f = 9.0 \times 10^{-5} s^{-1}$. CROCO uses a σ terrain-following coordinate, the N vertical levels being modulated in time between bottom and sea surface height η. Constant depth level z_0 are stretched over thickness h_c with surface coefficient θ_s :

$$z = \eta + (\eta + H_{bot}) z_0 \tag{2}$$

$$z_0 = \frac{h_c \sigma + H_{bot} C_s(\sigma)}{h_c + H_{bot}} \quad \text{with} \quad C_s(\sigma) = \frac{1 - cosh\left(\theta_s \frac{\sigma - N}{N}\right)}{cosh(\theta_s) - 1} \tag{3}$$

With $N = 100$ levels, $h_c = 400m$ and $\theta_s = 8$, vertical grid steps are then close to $5m$ in the upper $200m$. $200m$ being the vertical scale of the thermocline, it ensures a maximal resolution in the upper ocean where seasonal variations occur (Houpert et al., 2015). This configuration has then a higher vertical resolution than previous similar studies ($N = 32$ $h_c = 250m$ and $\theta_s = 6.5$ for Escudier et al. (2016)) or operational models (Juza et al., 2016).

Turbulent closure

Mixing is parametrized through $k{-}\epsilon$ closure scheme (Rodi, 1987) using the generic length scale approach (Umlauf & Burchard, 2003). Turbulent kinetic energy k dissipates with rate ϵ and stability function c_v into an effective viscosity ν (respectively c_T and κ for diffusivity). No additional explicit mixing is added.

$$\nu = \frac{c_v k^2}{\epsilon} \quad \text{and} \quad \kappa = \frac{c_T k^2}{\epsilon} \tag{4}$$

As implemented in the CROCO model, a minimal k input is parameterized. Given that the minimal dissipation rate ϵ is set to $10^{-12} W.kg^{-1}$, the minimal k has to be set to $10^{-9} m^2.s^{-2}$ in order to retrieve a minimal diffusivity of $10^{-6} m^2.s^{-1}$ with a stability function of order unity. This diffusivity value is close to kinematic viscosity and thermal diffusivity for water (respectively $1{\times}10^{-6}$ and $1{\times}10^{-7}$ $m^2.s^{-1}$). This issue is also discussed by Perfect et al. (2020).

2.2 Background stratification and initial mesoscale anticyclone

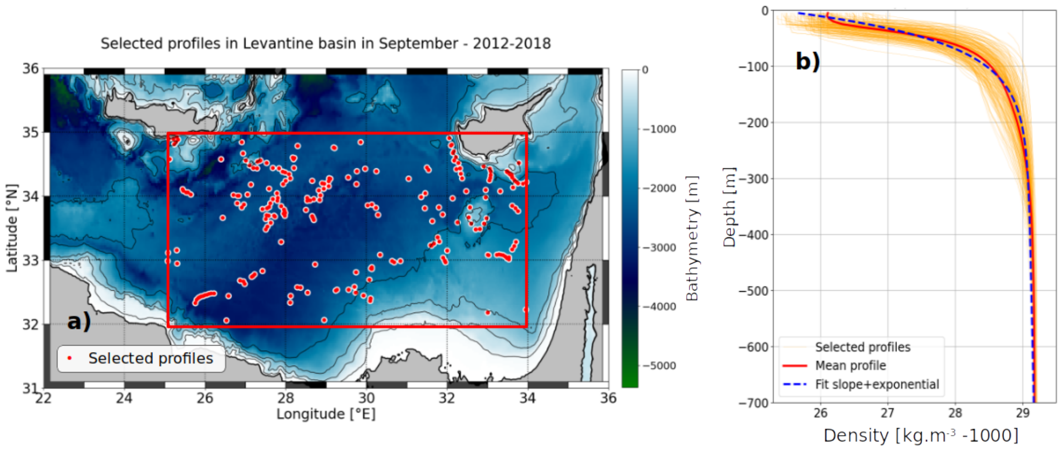

Figure 1. (a) Map showing the region of high long-lived anticyclones occurrence in the Levantine basin. The atmospheric fields used as input are averaged overt area delimited by the green frame. Black dots are the cast position of 242 selected in situ profiles identified as outside-eddy. Bathymetry is ETOPO1 data (Smith & Sandwell, 1997) with 0, 500, 1000 and 1500m isobaths. (b) Density vertical structure of selected profiles (orange thin lines), mean profile (red thick line) and fitted profile using Eq.5 (blue dashed).

A realistic background stratification is set from a climatological database gathering in situ delayed-time and near-real-time data from Copernicus Marine Environment Monitoring Service (Barboni, Stegner, et al., 2023). A region of interest is considered at the center of the Levantine Basin (25 to 34 °E and 32 to 35 °N, shown in Fig.1a). For background stratification we used only profiles in the region of interest, detected as outside-eddy using the DYNED eddy atlas dataset (see Barboni, Coadou-Chaventon, et al. (2023) for details), from 2012 to 2018 and for each year in September. Considering these criteria, 242 profiles are averaged into a mean stratification $\rho_b(z)$ fitted over the first 1000m with a linear slope S added to an exponential with vertical Z_T in the upper water column (Eq.5, see Fig.1b). September is chosen as the end of summer when the thermocline is marked and stratification gradient the strongest, allowing a better fit with exponential slope.

$$\rho_b(z) = \rho_1 + (\rho_s - \rho_1)exp\left(-\frac{z}{Z_T}\right) + Sz \tag{5}$$

Regression fit gave $\rho_1 = 1029.03kg.m^{-3}$, $\rho_s = 1025.3kg.m^{-3}$, $Z_T = 55m$, $S = 1.8{\times}10^{-4}kg.m^{-4}$. Corresponding baroclinic deformation radius is approximately $11km$. An initial density anomaly σ in geostrophic equilibrium is added to the background stratification. $\sigma(r,z)$ is azimuthally symmetric and has a Gaussian shape in the vertical direction and pseudo-Gaussian in the radial one, with radius R_m and vertical extent H :

$$\sigma(r,z) = \sigma_0 \frac{z}{H} exp\left(-\frac{1}{\alpha}\left(\frac{r}{R_m}\right)^{\alpha}\right) exp\left(-\frac{1}{2}\left(\frac{z}{H}\right)^2\right) \quad \text{with} \quad \sigma_0 = \frac{\rho_0 f V_m R_m e^{1/\alpha}}{gH} \tag{6}$$

The initial maximal speed radius R_m is 25 km, slightly more than twice the deformation radius but still smaller than the large long-lived Eastern Mediterranean anticyclones (Barboni, Coadou-Chaventon, et al., 2023), giving a Burger number ($Bu = R_d^2/R_m^2$) close to 0.2. Maximal speed is initially set to $V_m = 0.4\,m.s^{-1}$ giving a Rossby number ($Ro = V_m/R_m f$) of 0.16, but later decays around 0.1 due to eddy evolution. $Ro = 0.1$ is a standard value in the Mediterranean Sea (Ioannou et al., 2019). H is set to $100m$ on the same order as thermocline extent Z_T, and shape parameter $\alpha = 1.6$ ensures baroclinic stability (Carton et al., 1989; Stegner & Dritschel, 2000). Cyclogeostrophic correction is added following Penven et al. (2014).

2.3 Atmospheric heat forcing

Air-sea fluxes are computed with the Coupled Ocean–Atmosphere Response Experiment (COARE) 3.0 parametrization (Fairall et al., 2003), with improved accuracy for large wind speeds ($> 10m.s^{-1}$) encountered in high frequency forcing. Short wave heat flux Q_{SW} is distributed on the vertical following Paulson and Simpson (1977) transparency model with Jerlov water type I , consistent with very clear Mediterranean waters ($R = 0.58$, $\zeta_1 = 0.35m$, $\zeta_2 = 23m$):

$$I(z) = Q_{SW}\left(Rexp\left(-\frac{z}{\zeta_1}\right) + (1-R)exp\left(-\frac{z}{\zeta_2}\right)\right) \tag{7}$$

Upward long-wave heat flux $Q_{LW}^{\uparrow}$ computes the ocean SST (T_s) thermal loss using Stefan-Boltzmann black body law, with emissivity $\epsilon_{sb} = 98.5\%$ and $\sigma_{sb} = 5.6697\times 10^{-8}\,W.m^{-2}.K^{-4}$, convention positive downwards :

$$Q_{LW}^{\uparrow} = -\epsilon_{sb}\sigma_{sb}T_s^4 \tag{8}$$

Latent heat flux Q_{Lat} and sensible heat flux Q_{Sen} also involves a direct SST retroaction:

$$Q_{Lat} = -\rho_a L_E C_E|V|(q_s - q_a) \quad ; \quad Q_{Sen} = -\rho_a c_p C_S|V|(T_s - T_{2m}) \tag{9}$$

With ρ_a air density, c_p air thermal capacity, L_E evaporation enthalpy and $|V|$ 10m wind speed. q_s and q_a are specific humidity for ocean and atmosphere at 2m respectively, saturated at T_s for q_s, related to saturated water pressure P_{sat} fro q_a : $q_a = 0.622h_{2m}P_{sat}(T_{2m})/P_{SL}$ and $q_s = 0.98\times 0.622\times P_{sat}(T_s)/P_{SL}$. Factor 0.98 accounts for water vapor reduction caused by salinity (Sverdrup et al., 1942). Last, wind stress is computed from zonal and meridional winds (u and v) :

$$\tau_x = \frac{\rho_a}{\rho_0}C_D|u|u \quad \text{and} \quad \tau_y = \frac{\rho_a}{\rho_0}C_D|v|v \tag{10}$$

In equations 9-10, C_E, C_S and C_D are corresponding transfer coefficients considering the stability of the atmospheric boundary layer based on the Monin-Obukhov similarity theory. They are all on the order of 1×10^{-3} (Fairall et al., 2003).

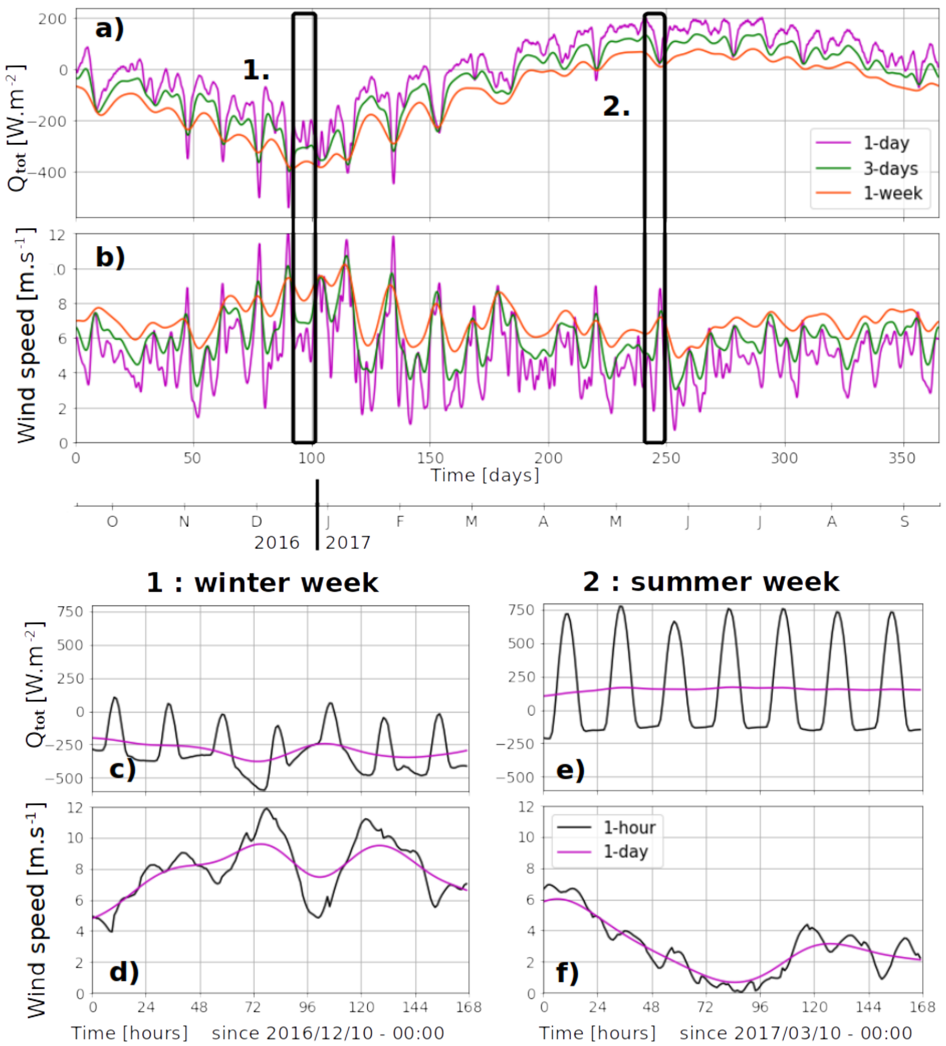

Figure 2. Net heat flux and wind speed for the 5 input timeseries, shown separately as diurnal cycle gives larger variations. (a) Net heat flux and (b) wind speed for the 1-day (magenta line), 3-days (green) and 1-week (orange) timeseries over one year. To enhance readability, 3-days and 1-week net heat flux are lowered by 20 and 40 $W.m^{-2}$ respectively. (c) 1-hour (black) and 1-day (magenta) net heat flux (respectively (d) for wind speed) in a winter week of 2016. (e) and (f) : same as (c) and (d) in a summer 2017 week.

ERA5 reanalysis input is used for atmospheric forcing. Fields are available with a 1 hour temporal resolution and 1/4°horizontal resolution (Hersbach et al., 2020). To focus on the temporal variability, forcing timeseries are spatially averaged over the Levantine basin (Fig.1a). Four forcing inputs with different temporal scales are tested : 1-hour, 1-day, 3-days and 1-week. The 1-hour forcing is the original ERA5 timeseries, the three later ones are Gaussian smoothing of the 1-hour timeseries with double-window size of 1, 3 and 7 days respectively, shown in Fig.2. One year of forcing from 15 September 2016 to 15 September 2017 runs cyclically for 2 years as forcing input, with root mean square wind speed $\overline{V} = 5.5 m.s^{-1}$. 10m neutral wind from ERA5 is used for wind stress in Eq.10. To keep the same wind speed magnitude with varying wind frequency, smoothed timeseries for zonal and meridional winds ($\overline{u}$ and $\overline{v}$) have to be rescaled. The correction factor λ being $\gtrsim 1.1$ for 1-day timeseries, and $1.1 < \lambda < 2$ for 3-days and 1-week :

$$\tilde{u} = \lambda \overline{u}\,;\, \tilde{v} = \lambda \overline{v} \quad \text{with} \quad \lambda = \frac{\overline{\sqrt{u^2 + v^2}}}{\sqrt{\overline{u}^2 + \overline{v}^2}} \tag{11}$$

The same year is kept to avoid disturbance with interannual variations, which are strong for heat fluxes over the Mediterranean Sea (Mariotti, 2010; Pettenuzzo et al., 2010), but no significant variations were observed when selecting another year.

Forcing without surface temperature retroaction

A comparison experiment is run without SST retroaction on ocean-atmosphere fluxes. In this configuration, the net heat flux Q_{tot} from ERA5 directly forces the upper ocean layer, the short wave part Q_{SW} being still distributed on the vertical (Eq.7). Momentum fluxes are computed from Eq.10 with constant drag coefficient $C_D = 1.6 \times 10^{-3}$. The net heat flux Q_{tot} timeseries in ERA5 has daily amplitudes around $\pm 150 W.m^{-2}$ and an annual average of $-3.0\, W.m^{-2}$, consistent with the net evaporation of the Mediterranean sea (Mariotti, 2010). The net heat flux is then corrected by linearly decreasing the negative values to achieve a zero annual average, avoiding a drift of the mean stratification.

2.4 Eddy tracking indicators

Eddy shape, radius and intensity

Eddy detections are provided through the Angular Momentum Eddy Detection and Tracking Algorithm (AMEDA). AMEDA is a mixed velocity-altimetry approach, its relies on using primarily streamlines from a velocity field and identifying possible eddy centers computed as maxima of local normalized angular momentum (Le Vu et al., 2018). It was successfully used in several regions of the world ocean in altimetric data (Aroucha et al., 2020; Ayouche et al., 2021; Barboni et al., 2021), high frequency radar data (F. Liu et al., 2020) or numerical simulations (de Marez et al., 2021). In each eddy single observation (one eddy observed one day), AMEDA gives a center (which position is noted $\mathbf{X}_e$ hereafter) and two contours. The 'maximal speed' contour is the enclosed streamline with maximal speed (i.e. in the geostrophic approximation, with maximal SSH gradient) ; it is assumed to be the limit of the eddy core region where water parcels are trapped. The 'end' contour is the outermost closed SSH contour surrounding the eddy center and the maximal speed contour ; it is assumed to be the area of the eddy footprint, larger than just its core but still influenced by the eddy shear (Le Vu et al., 2018). The observed maximal speed radius R_m is then defined as the radius of the circle having an area equal to the maximal speed contour. Eddy detection through interpolated Level 4 SSH products leads to imperfections. It typically smooths gradients and then reduces observed geostrophic velocities (Amores et al., 2018; Stegner et al., 2021). To mimic those imperfections in the numerical simulations, AMEDA detections are performed on the 48h-averaged SSH field at model grid resolution.

Eddy SST signature δT, heat flux δQ, differential mixing ratio ξ and mixed layer anomaly

The anticyclone-induced SST signature δT is defined as the difference of SST between the eddy core SST_{in} and its periphery SST_{peri}. Adapting Moschos et al. (2022), SST_{in} is the average of the area centered on $\mathbf{X}_e(t)$ with radius $2/3R_m(t)$; SST_{peri} is the average on an annular area centered on $\mathbf{X}_e$ with radius between $2/3R_m(t)$ and $2R_m(t)$. Positive (negative) δT then indicates a warm-core (cold-core) signature. Similarly the induced signature on ocean-atmosphere fluxes is defined as δQ, with positive δQ for increased warming at the eddy core. Thermal heat flux feedback (THFF) is then defined as the linear regression of δQ as a function of δT over the second year of simulation (from 365 to 730 days, see Sect.3.3).

Differential mixing between the eddy core and outside-eddy are measured through the index ξ. Temperature vertical diffusivity κ computed by $k{-}\epsilon$ mixing closure from instantaneous history record is spatially averaged in the eddy core (κ^{AE}) and outside-eddy (κ^{Out}). The eddy core region corresponds here to the area around the eddy center with radius $2/3R_m(t)$. The outside-eddy region is defined as the area outside any 'end' contours detected by the tracking algorithm. Diffusivity spanning several orders of magnitude, differential mixing ξ is then evaluated as a vertical average of the ratio of these two quantities, typically using a depth $h = 20m$ to focus on the upper layers stratified in summer :

$$\xi = \frac{1}{h}\int_{-h}^{surf} \frac{\kappa^{AE}}{\kappa^{Out}} dz \tag{12}$$

Summer eddy SST signature magnitude $\overline{\delta T}$ is defined as the 30th δT percentile over the summer, and its spread as the difference between the 30th and the 10th percentiles (see results in Table 1). Similarly $\overline{\xi}$ is defined as the median of the ξ distribution over the summer, and its spread as the difference between the median and the 30th percentile. First and second summers are defined as 230 to 340 days and 590 to 700 days respectively, corresponding to the May to August period when a significant number of warm-core anticyclones are observed (Moschos et al., 2022).

Last, the MLD anomaly ΔMLD are defined for a given winter as the maximal difference reached between the MLD outside-eddy and the MLD inside-eddy, following (Barboni, Coadou-Chaventon, et al., 2023). In the following numerical experiments running for 2 years, the first winter is considered as a transient period not retained for analysis. ΔMLD is then computed only for the second winter, defined as 450 to 580 days, corresponding to the December to April period, when maximal MLD are reached in the Mediterranean Sea (Houpert et al., 2015).

3 Idealized simulations compared to observations

The temporal evolution of mesoscale eddies in the Levantine basin can be retrieved for several anticyclones where Argo floats remained trapped several months, as extensively studied in Barboni, Coadou-Chaventon, et al. (2023). A marked seasonal signal is detected in both SST and vertical structure. An example is shown in Fig.3 with a Ierapetra anticyclone. Ierapetra anticyclones are strong recurrent anticyclonic structures formed each year in the lee of Crete island (Ioannou et al., 2020). In the example shown below, δT index has a marked oscillation between a winter warm core and summer cold core. The weekly smoothed signature can be measured to about $\delta T \approx +0.7°C$ in both winters 2016-2017 and 2017-2018, and about $-0.3°C$ in summer 2017 (about $-0.2°C$ in summer 2018). The vertical structure could also be measured thanks to large Argo deployments (Fig.3h) ; due to errors in the salinity sensors, density in 2018 is estimated from temperature applying a linear regression using 2017 data. One can also notice the seasonal variations of the anticyclone maximal speed, with two maxima in late winter. This is consistent with kinetic energy inverse cascade maximal peak from submesoscale to mesoscale in kinetic energy distributions (Zhai et al., 2008; Steinberg et al., 2022), but it is still noticeable to have the same phenomenon tracking a single individual structure. Here the physical processes driving these observed seasonal variations are studied with numerical experiments, investigating sensitivity to horizontal resolutions, forcing frequency and SST retroaction on air-sea fluxes. Simulations are summarized in Table 1, the reference considered being 1km resolution with 1-hour forcing, 100 vertical levels with SST retroaction (run 1K100-1H in Table 1 below).

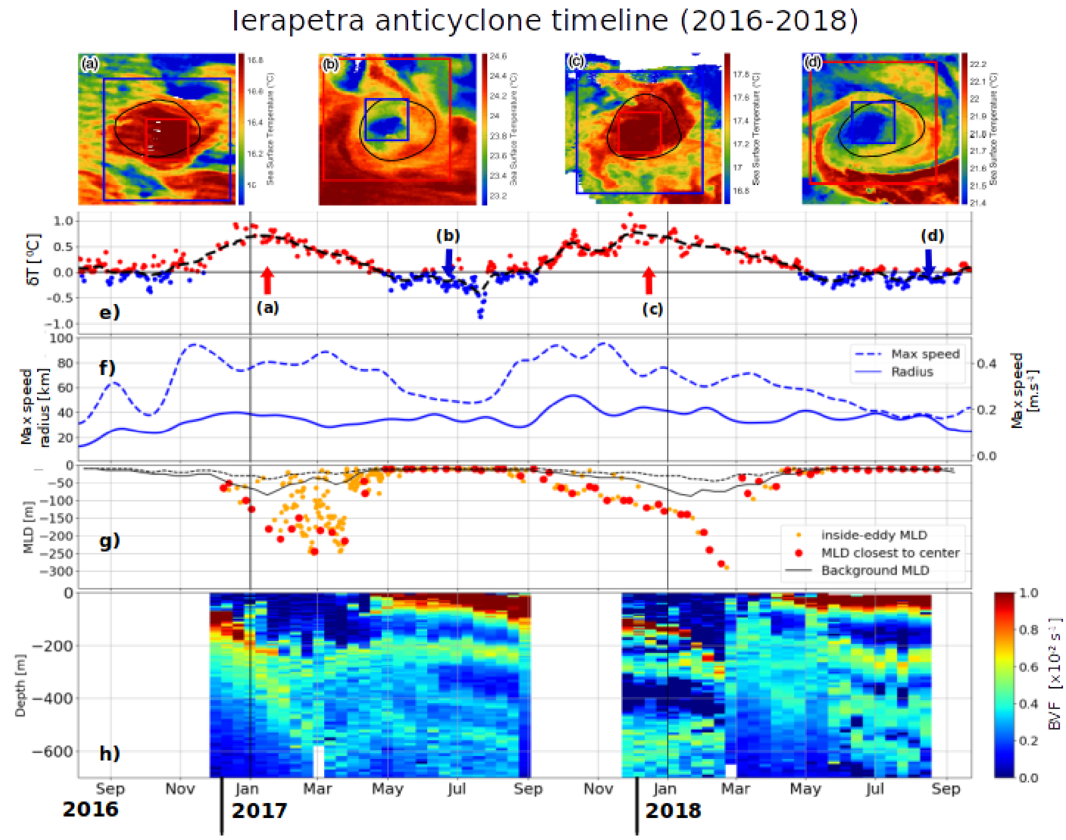

Figure 3. Temporal evolution of the Ierapetra anticyclone formed South-East of Crete in late summer 2016. Upper panels are high-resolution SST snapshots in (a) January 2017, (b) June 2017, (c) December 2017 and (d) July 2018, the maximal speed contour (see Sect.2.4 is in black line. (e) δT eddy SST anomaly, cold-core in blue and warm-core in red, with black dashed line showing the 5 days smoothed evolution. (f) Maximal speed V_m (dashed blue) and radius R_m (continuous blue) with 10 days smoothing. (g) MLD evolution inside the the anticyclone (dots, with red ones highlighting the closest to center), with outside-eddy background MLD in continuous black line (spread between dashed lines). (h) Brunt-Vaisala frequency.

3.1 Horizontal resolution sensitivity

The numerical simulation at 4km resolution and 25 vertical levels (run B in Table 1) reveals a few discrepancies with real observations. A horizontal resolution of 4km is close to operational oceanography models in the Mediterranean Sea (Juza et al., 2016). At the surface, despite seasonal variations of the eddy SST signatures (Fig.4a-c) and in the δT index (Fig.4f), summer 'inverse' signatures are not retrieved with no cold-core anticyclone. A steady erosion of the eddy strength is also noticeable, with a decrease in the maximal speed decreasing from $0.35 m.s^{-1}$ to $0.15 m.s^{-1}$ in 2 years, while its radius remains constant ($\approx 25 km$, Fig. 4e). Note that the initial maximal speed is set to $0.4 m.s^{-1}$ (see Sect.2.2) but the smoothing effect of time-averaging leads to a lower detected initial value of $0.35 m.s^{-1}$.

At depth, the mixed layer anomaly is significant, on the order of 50m (Fig.4g). Some bursts of differential mixing are observed in late winter from December to March when mixed layer instabilities and restratification processes can occur, with ξ reaching a few

[ht]

Table 1. Summary table of CROCO numerical experiments. Runs start in September of the atmospheric forcing timeseries. Thermal heat flux feedback (THFF), eddy SST anomaly index $\overline{\delta T}$ and differential mixing ratio $\overline{\xi}$ are defined in Sect.2.4, and $\overline{\xi}$ is computed over the upper 20m. Subscripts ($\overline{\xi}_1$,$\overline{\xi}_2$) refers to first and second summers defined as 230 to 340 days and 590 to 700 days respectively. ΔMLD refers only to the second winter defined as 450 to 590 days (see shades in Fig.4d-h).

Name	**Vertical levels**	**dx** (km)	**Freq**	**SST retroaction**	**THFF** ($W.m^{-2}.K^{-1}$)	$\overline{\delta \mathbf{T}}_\mathbf{1}$ ($°C$)	$\overline{\delta \mathbf{T}}_\mathbf{2}$ ($°C$)	$\overline{\xi}_\mathbf{1}$	$\overline{\xi}_\mathbf{2}$	**ΔMLD** (m)
1K100-1H	100	1	1-hour	Yes	-41.5 ± 1.3	-0.20 ± 0.10	-0.18 ± 0.04	3.05 ± 0.70	2.81 ± 0.74	51
2K50-1H	50	2	1-hour	Yes	-40.7 ± 1.0	-0.12 ± 0.14	-0.11 ± 0.06	1.54 ± 0.31	1.34 ± 0.22	63
4K25-1H	25	4	1-hour	Yes	-34.3 ± 1.8	0.01 ± 0.14	0.02 ± 0.10	1.10 ± 0.12	1.00 ± 0.12	48
05K150-1H	150	0.5	1-hour	Yes	-39.2 ± 1.4	-0.16 ± 0.10	-0.19 ± 0.06	2.58 ± 0.58	2.71 ± 0.45	91
1K100-1D	100	1	1-day	Yes	-42.1 ± 0.8	-0.21 ± 0.20	-0.31 ± 0.06	2.99 ± 0.44	3.34 ± 1.23	57
1K100-3D	100	1	3-days	Yes	-44.7 ± 1.0	-0.12 ± 0.14	-0.09 ± 0.03	1.41 ± 0.28	0.99 ± 0.09	70
1K100-1W	100	1	1-week	Yes	-41.0 ± 0.4	-0.05 ± 0.05	-0.03 ± 0.01	1.25 ± 0.14	1.02 ± 0.01	94
1K100-1H-NoSST	100	1	1-hour	No	-	-0.41 ± 0.16	-0.51 ± 0.00	2.60 ± 0.46	2.47 ± 0.25	10

times values higher than 2 (Fig. 4h). However no differential mixing is retrieved in summer. On the other hand, the anticyclone vertical structure is coarsely reproduced. The winter MLD cooling forms a homogeneous layer between 100 and 150m (Fig. 4i). These winter waters formed by convection do not reproduce the homogeneous subsurface anticyclone cores, separated by density jump or sharp temperature gradient such as the continuous temperature gradient in Fig.3h around 200m depth (see other examples in Fig.4-5 from Barboni, Coadou-Chaventon, et al. (2023)). The inability to reproduce this mesoscale subsurface lens is not surprising given the low vertical resolution, the vertical steps being on the order of $20m$ at $100m$ depth.

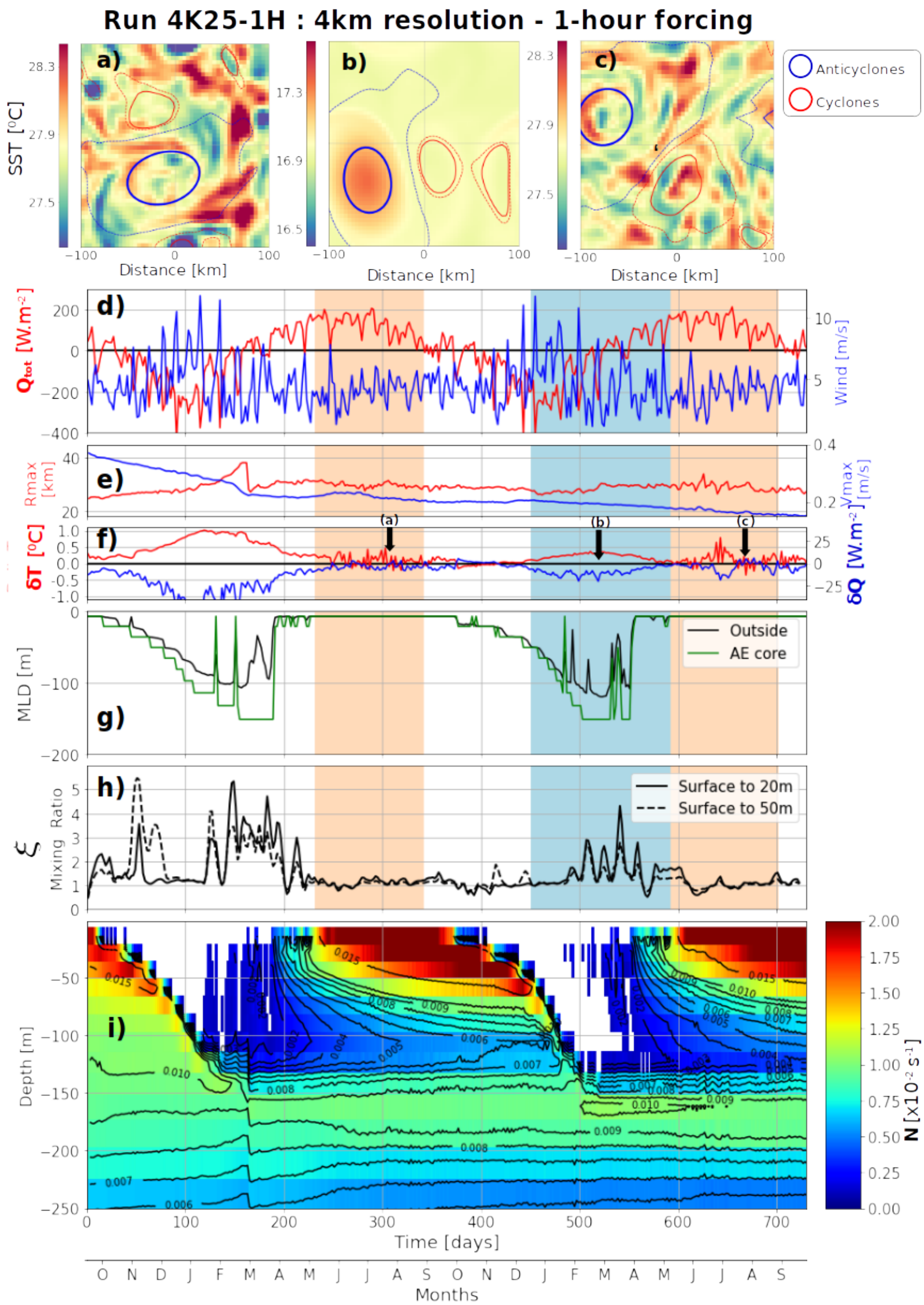

Figure 4. Simulation B from Table1. (a) SST snapshot in the first summer, (b) in the second winter, (c) in the second summer, with eddies detected by AMEDA in contours. The initial anticyclone is highlighted by a thicker line. (d) Net heat flux (red) and windspeed (blue). (e) $R_m(t)$ (red) and $V_m(t)$ (blue) from AMEDA. (f) SST anomaly index δT (red), respectively heat flux anomaly δQ, blue). (g) Mixed layer inside-eddy (green) and outside-eddy (black). (h) Differential mixing ratio ξ defined in Eq.12 with $h = 20m$ (solid) and $h = 50m$ (dashed line). (i) Inside-eddy stratification evolution shown with Brunt-Vaisala frequency ; contours are overlaid with interval $0.001s^{-1}$ and negative values are blanked. On panels c-h, summer periods are indicated by light red shades, winter by a light blue shade.

The same numerical set-up with a finer 1km horizontal resolution (run 1K100-1H in Table 1) shows a net contrast with the previous coarser simulation. This simulation has a 1km horizontal grid size and 100 levels with same stretching parameters giving vertical grid steps close to 3m in the upper 200m. A summer 'inverse' eddy surface temperature is clearly retrieved with 1-hour frequency heat and momentum forcing. As shown in Fig.5 in this configuration, a clear anticyclonic cold-core SST signature is observed in summer (Fig.5a), switching back to a winter warm-core SST the next winter (Fig.5b) and appearing again in the second summer (Fig.5c). This anticyclone surface seasonal oscillation can clearly be tracked by δT (Fig.5f). $\overline{\delta T}$ reached about $-0.2^{\circ}C$ in the both summers (see Table 1) with spikes of $\delta T \approx -0.5^{\circ}C$ and maximal value around $+0.4^{\circ}C$ in winter. Considering anticyclonic cold-core signatures statistics in the Mediterranean Sea ((Moschos et al., 2022) in particular their Fig.5b) $\delta T \approx -0.2^{\circ}C$ is a low but standard value, anticyclone SST anomalies typically not being colder than $-0.5^{\circ}C$. This cold-core summer signature goes along with a mixing increase in the upper layers at the eddy core, measured by a diffusivity in summer more than twice as high inside the eddy core as outside. Sensibility of the ξ indicator is shown on Fig.5h, with ξ averaged over the upper 20m or 50m, the first case leading to ξ values higher than 4 in summer despite some variability. This enhanced mixing seems to be confined in the upper layers, as ξ decreases to approximately 1 as soon as the mixed layer deepens, but it increases again to similar values during the second summer.

At depth the maximal mixed layer anomaly reaches about 50m (Fig.5h), very close to the value of the simulation at 4km resolution. However the vertical structure is better reproduced at 1km, and in particular between 100 and 150m deep the $5{\times}10^{-3}s^{-1}$ stratification isoline closes in December, 4 months later than in the 4km simulation (in August, see Fig.4i). This means that homogeneous waters formed at depth in the first winter restratify more slowly. Eddy decay in time is also slower on maximal speed : after 2 years the anticyclone velocity is about $0.3m.s^{-1}$ with 1km resolution compared to $0.15m.s^{-1}$ with 4km (Fig.4e). Sharp density gradients are smoothed in a coarser simulation, leading to unrealistic temporal evolution of the anticyclones vertical structure. Surface (SST) or depth-integrated (maximal geostrophic speed) measurements are then not accurately reproduced at a spatial resolution of 4km.

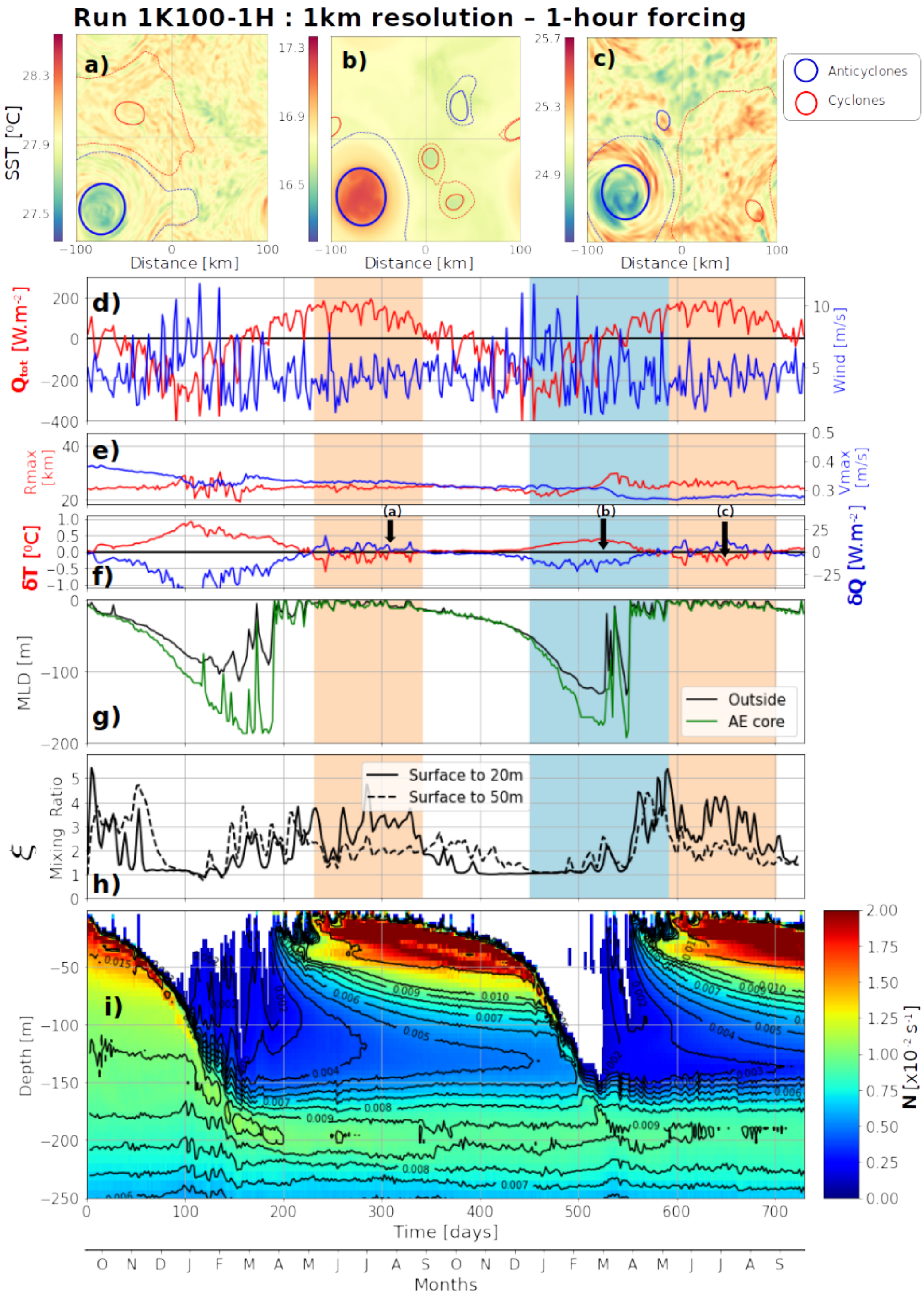

Figure 5. Simulation 1K100-1H from Table1. Same as in Fig.4 but with a 1km horizontal resolution.

An experimental series with the same numerical set-up is performed, increasing horizontal resolution from 4km to 500m (runs 1K100-1H, 2K50-1H, 4K25-1H, 05K150-1H) and vertical resolution accordingly. The horizontal to vertical resolution ratio is kept close to the Brunt-Vaisala to inertial frequencies ratio. It reveal that summer anticyclonic cold-core signature $\overline{\delta T}$ and differential mixing $\bar{\xi}$ both continuously increase when decreasing the grid cell (see Fig.6c). Summer eddy SST inversions are then consistently correlated with an increased mixing. In addition a convergence behavior is observed for mixing at 1km with 100 levels to $\bar{\xi} \approx 3$, as no further mixing is obtained increasing the resolution to 500m and 150 levels. Differential mixing appearing at 1km resolution implies that small scale processes, smaller than the eddy size are at stake. 1km horizontal resolution with a baroclinic first deformation radius around 11km entails that deformation radius to be explicitly resolved, which is not entirely effective for resolution of 2km or larger, similarly to other numerical studies (Marchesiello et al., 2011; Soufflet et al., 2016). On the other hand in winter very similar δT are retrieved at all resolution, with a maximum around $+0.4^{\circ}C$ (Fig.6a) and similar THFF suggesting that winter thermal loss is less affected by horizontal resolution. THFF slightly decreases for lower resolution, likely due to smoothing effect of strong SST patterns.

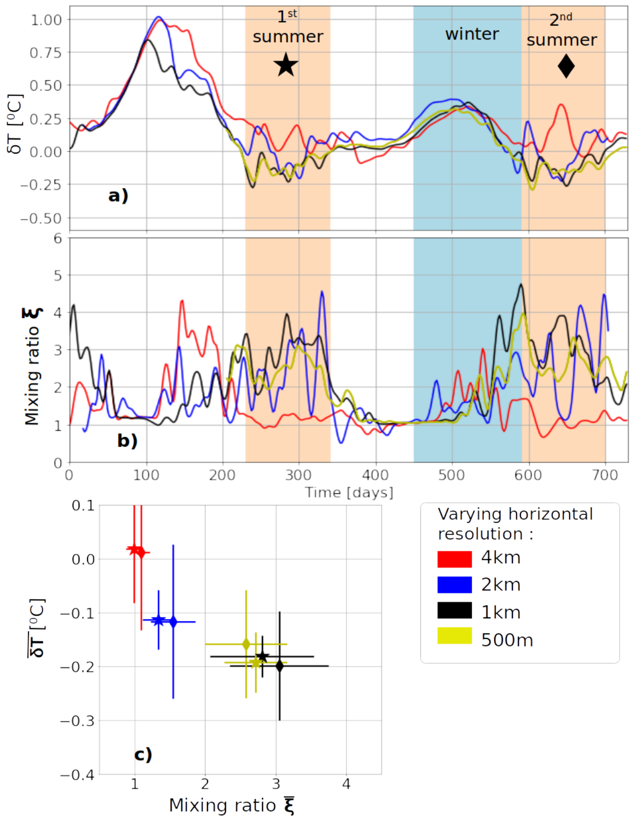

Figure 6. (a) δT and (b) ξ timeseries for experiments 1K100-1H, 2K50-1H, 4K25-1H, 05K150-1H listed in Tab.1 with SST retroaction on air-sea fluxes and varying horizontal resolution frequency. 2-days Gaussian smoothing is applied and summer periods are shaded in light red, winter in light blue. Due to computer memory issues, the first transient winter at 500m resolution was not recorded. (c) Summer-averaged eddy-induced SST anomalies ($\overline{\delta T}$) and mixing ratio ($\bar{\xi}$), with stars for the first summer and diamonds for the second one. Errorbars are ξ spread (30^{th} percentile) over the same period.

For the eddy-induced mixed layer anomaly, similar values are obtained from 4km to 1km resolution ($\Delta MLD \approx 50m$), but a larger $\Delta MLD = 91m$ is retrieved at 500m resolution. This effect could be due to the partial resolution of sub-mesoscale processes such as mixed layer instabilities (Boccaletti et al., 2007; Capet et al., 2008). Maximal background mixed layer deepens when resolution gets finer down to 1km resolution (see Fig.4g and 5g), in consistence with previous experiments (Couvelard et al., 2015). At 500m resolution, a closer look at the MLD evolution inside- and outside-eddy shows that the outside-eddy MLD restratified earlier in run 05K150-1H (in March) than in run 1K100-1H (in April) due to restratification beginning at submesoscale with mixed layer instabilities (Fig.7b). But in both cases inside-eddy MLD reached the same depth (193m, see Fig.7e-f). This suggests that maximal mixed layer inside-eddy indeed reached a maximum driven by air-sea cooling, while restratification outside-eddy occurred too late in run 1K100-1H because vertical buoyancy fluxes are too weak (Capet et al., 2008). Compared to Mediterranean MLD climatology, a restratification in April is indeed quite late (Houpert et al., 2015).

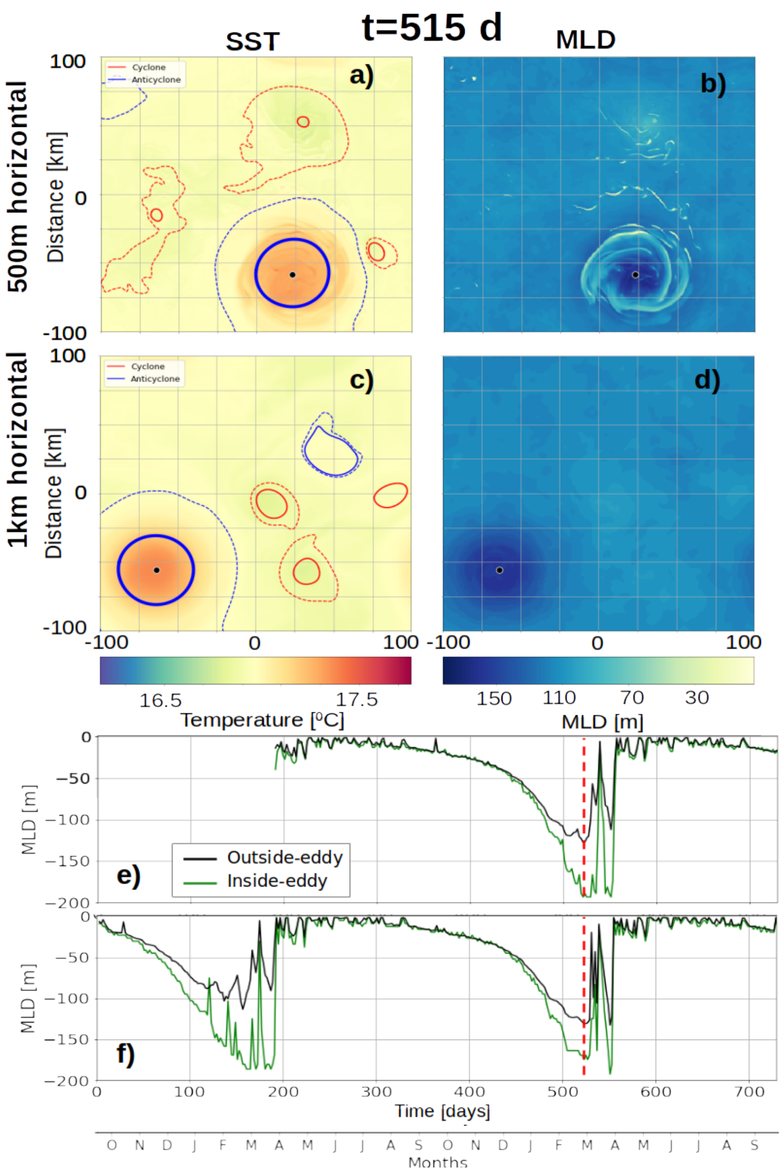

Figure 7. (a) SST with anticyclones and cyclones as in Fig.4 (the initial anticyclone has thicker contour) in 05K150-1H simulation. (b) MLD in 05K150-1H. (c) and (d) : same as (a) and (b) but in 1K100-1H simulation. (e) MLD time series inside-anticyclone (green) and outside-eddy (black) for the 05K150-1H simulation, a red dashed line indicates the time step shown in panels (a)-(d). Due to memory issues, the first transient winter was not recorded. (f) Same as (e) in 1K100-1H simulation.

Mixing patterns over the vertical in the 1km resolution simulation are also consistent with observations. Anticyclones were recently observed to enhance mixing at depth through the propagation of trapped near-inertial internal waves in their core. In studies from Martínez-Marrero et al. (2019) and Fernández-Castro et al. (2020), in situ measurements revealed lower dissipation rate ϵ in anticyclonic homogeneous core than in the neighboring background, and enhanced ϵ below at depth. In our numerical experiments, both diffusivity κ (Fig.8c) and dissipation rate ϵ (Fig.8e) match this feature, with enhanced mixing in summer below the anticyclone, up to one order of magnitude larger from 200 to 300m depth. The anticyclone subsurface core revealed by thick isopycnal displacement on Fig.8e, also shows locally reduced ϵ between 100 and 200m. Fig.8e is then a striking reproduction of dissipation rate section obtained by Fernández-Castro et al. (2020) (see in particular their Fig.5f). However those in situ measurements could not compare outside- and inside-eddy mixing close to the surface, because the value range for ϵ would be too large with surface processes a lot more powerful than deep ocean ones. Numerical simulation enables to reveal that anticyclones also enhance mixing in near surface, with higher ϵ and κ just above the homogeneous core, in the upper 50 meters. The differential mixing ratio ξ previously shown in anticyclone time series then accurately measures a surface-enhanced mixing.

The seasonal cycle of eddy SST signature is then effectively reproduced at 1km horizontal resolution, close to observed value for the example shown above (Fig.3e). eddy SST seasonal shift correlates with increased mixing at the anticyclone core, in consistence with Moschos et al. (2022) hypothesis. This differential mixing is absent at 4km, but appears through $k-\epsilon$ mixing parametrization and converges at 1km resolution.

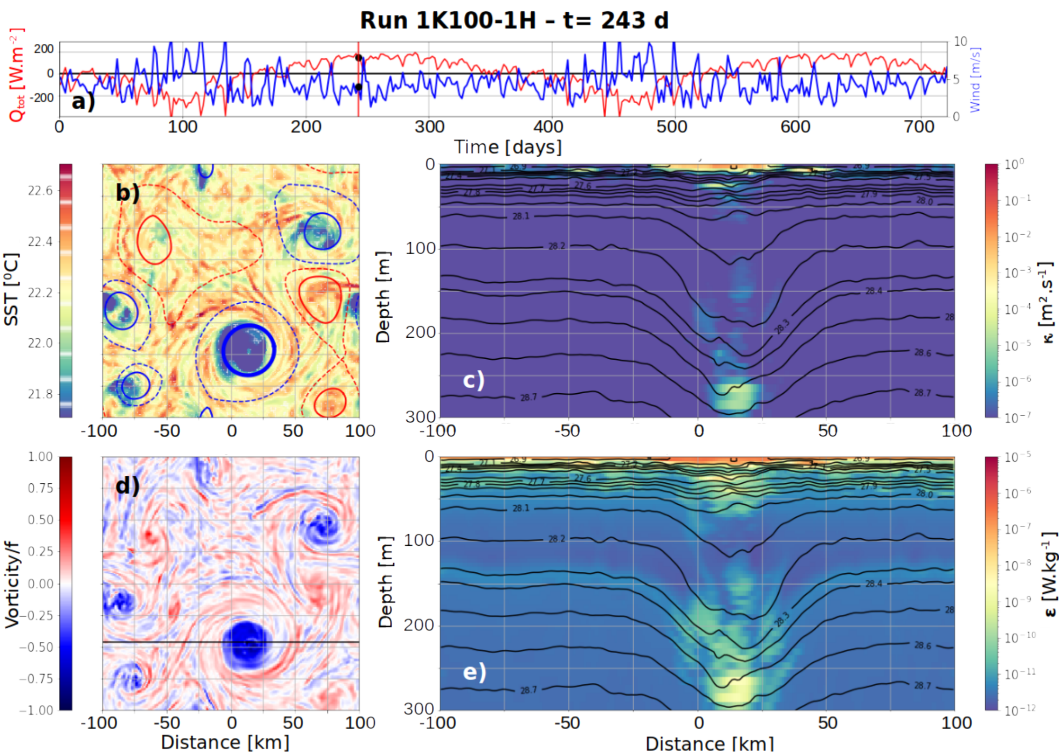

Figure 8. Snapshot at $t = 243\,d$ for the 1K100-1H simulation (see Fig.5). (a) Wind speed (blue line) and Q_{tot} (red lines) timeseries. (b) SST and (d) surface vorticity normalized by f with eddy detections as in Fig.4 (initial anticyclone has a thicker contour). (c) κ and (e) ϵ vertical sections in the upper 300m with logarithmic color scales, in both case the colorbar lower bound is the minimal possible value (see Sect.2.1). Isopycnals are added in black lines.

3.2 Forcing frequency sensitivity

Sensitivity of the eddy SST signature δT and differential mixing ξ to forcing temporal resolution is investigated by progressively removing high frequencies from the atmospheric inputs. These experiences are summarized as 1K100-1D to 1K100-1W in Table 1, using 1-day, 3-days and 1-week atmospheric timeseries respectively. δT and differential mixing ξ timeseries for these experiments are shown in Fig.9a-b. Significantly cold SST signatures ($\overline{\delta T} \lesssim -0.2^{\circ}C$) are obtained together with strong mixing ($\overline{\xi} \approx 3$) for 1-hour and 1-day frequency, but no significant differential mixing is retrieved ($1 < \overline{\xi} < 1.5$) for all lower forcing frequencies (Fig.9c). This threshold behavior is a strong result and shows that spontaneous appearance of differential mixing is driven by small scale and high frequency features. With a Coriolis parameter $f = 9.0 \times 10^{-5} s^{-1} = 1.24 cpd$, the inertial period is about 19h, the 1-day forcing can then partly trigger near-inertial waves.

The relationship between $\overline{\delta T}$ and $\overline{\xi}$ is however less clear than for the resolution sensitivity analysis (Fig.6). No differential mixing is observed for forcing frequencies lower than 1 day, but summer cold-core signatures are still found ($\overline{\delta T} \gtrsim 0.1^{\circ}C$, see Table1), even for the 1-week forcing. δT timeseries clearly show for all frequencies a marked seasonal signal (Fig.6a). In particular a significant warm winter signature is always observed, with stable maximal value at $\delta T \approx +0.4^{\circ}C$. In the same context a surprising result is the summer averaged $\overline{\delta T}$ being colder on average at 1-day than 1-hour forcing, despite similar differential mixing. Temporal evolution of eddy SST anomalies reveals this effect to be caused by a larger oscillation of the eddy surface signature (Fig.9a) about $\pm 0.2^{\circ}C$, hence larger errorbars at 1-day on Fig.9c. This suggests that other mechanisms not triggered by high frequency winds also contribute to the eddy SST seasonal cycle. If no differential vertical mixing is observed but if seasonal variations of the anticyclone SST (and hence surface density) is found, one can only hypothesize the role of lateral exchanges. Despite some tries, we were unsuccessful in quantifying eddy lateral exchanges following a varying $R_m(t)$ contour. No particular asymmetric wave modes was observed on SST snapshots, discarding the hypothesis of vortex Rossby waves (Guinn & Schubert, 1993; Montgomery & Kallenbach, 1997).

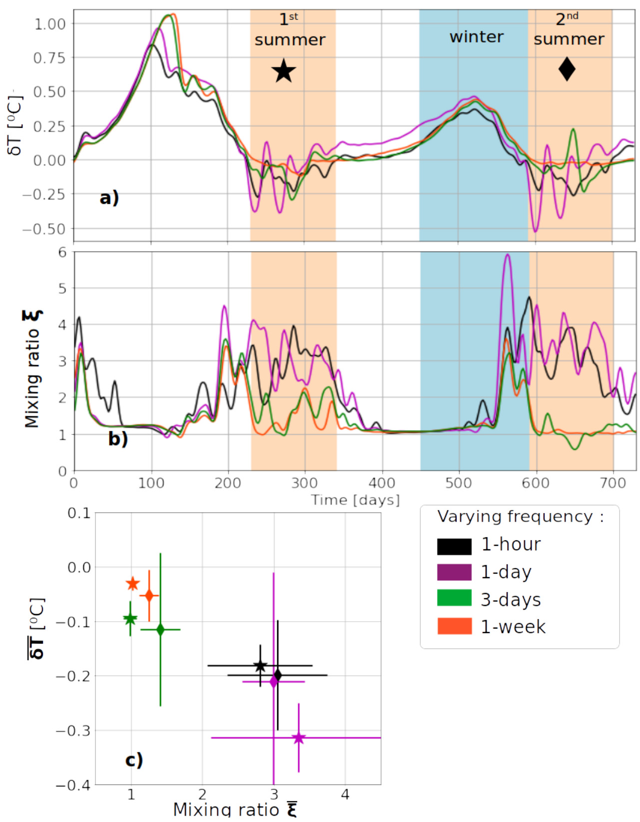

Figure 9. (a) δT and (b) ξ timeseries for experiments 1K100-1H, 1K100-1D, 1K100-3D and 1K100-1W listed in Tab.1 with SST retroaction on air-sea fluxes and varying forcing frequency. 2-days Gaussian smoothing is applied, summer periods are shaded in light red, winter in light blue. (c) Summer-averaged eddy-induced SST anomalies ($\overline{\delta T}$) and mixing ratio ($\overline{\xi}$), with stars for the first summer and diamonds for the second one.

Near-inertial internal waves are investigated using Fourier transforms on vertical speed anomalies in run 1K100-1H. We focus on a single vertical level at 20m in near-surface where the enhanced mixing occurs (see Fig.8c). Transforms are computed only in the second summer (590 to 700 simulated days) with a 1-hour sampling frequency. Following Babiano et al. (1987), inside-eddy spectrum is performed keeping only the inside-eddy area (around the eddy center with radius $2/3R_m(t)$) and the remaining area is set to 0 before performing the Fourier transform. Similarly outside-eddy spectrum is performed blanking all value inside any eddy contours. The results clearly show a differential effect inside-eddy vertical kinetic energy density revealing a second powerful peak at the effective inertial frequency $f_e = f + \zeta/2 \approx 1.0cpd$, lower than the inertia frequency (Fig.10a). Outside-eddy spectrum (Fig.10b) shows only one peak at the inertial frequency, and internal waves cannot propagate at lower frequencies due to the f-cut-off (Garrett & Munk, 1972). Normalizing by the investigated area, total vertical kinetic energy per unit surface is indeed higher inside the anticyclone ($4.19{\times}10^{-14}m^2.s^{-2}/m^2$) than outside-eddy ($1.64{\times}10^{-14}m^2.s^{-2}/m^2$) due to these powerful subinertial internal waves. An assumption of this method is however to assume that both inside- and outside-eddy areas roughly keep the same area, which is verified. This result is consistent with (Kunze, 1985) theory and recent numerical works (Danioux et al., 2015; Asselin & Young, 2020) subinertial waves ($\omega \lesssim f$) can be trapped in the anticyclone due to the locally lower absolute vorticity, and enhance mixing while breaking as proposed by Fernández-Castro et al. (2020).

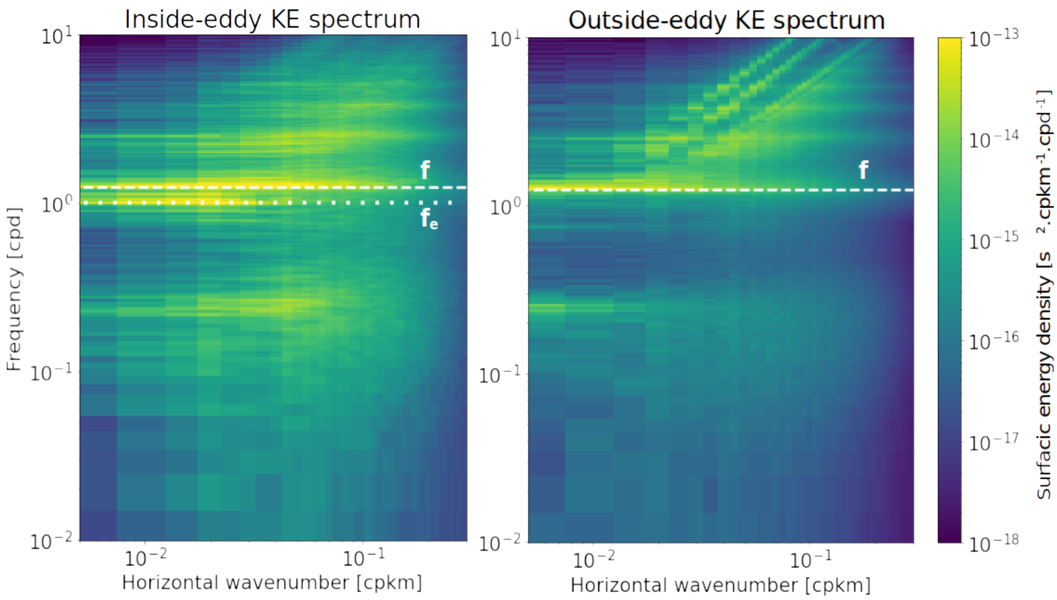

Figure 10. (a) Inside-eddy and (b) outside-eddy vertical kinetic energy density spectrum at 20m depth. For comparison, spectrum are normalized by the area of interest. Analysis performed on simulation 1K100-1H with 1-hour sampling. Normal (respectively effective) inertial frequencies $f = 1.24cpd$ ($f_e \approx 1.0cpd$) are highlighted by a white dashed (dotted) line.

3.3 Air-sea fluxes sensitivity

Sensitivity of the anticyclone temporal evolution to air-sea fluxes components is further investigated. A $1km$ resolution simulation experiment is run similarly as the 1K100-1H simulation without applying SST retroaction on air-sea fluxes (see Sect.2.3, run 1K100-1H-NoSST in Table 1). Although quite unrealistic, this experiment enables to check if the eddy SST anomaly seasonal shift and differential mixing observed in previous simulations are triggered by air-sea fluxes retroaction. Time series for SST reveals that eddy SST anomalies seasonal oscillation is retrieved without SST retroaction (Fig.11a-c), and summer cold-core signatures are even stronger : $\delta T \approx -0.8^{\circ}C$ the first summer and $\approx -0.8^{\circ}C$ the second one (Fig.11f). Simultaneously, differential mixing reaches $\xi \approx 3$, approximately the same value as run 1K100-1H (Fig.11h). This confirms that differential eddy mixing triggering the eddy SST variations is not linked to air-sea fluxes retroaction. However this feedback can modulate and dampen the δT seasonal cycle leading to reduced anomalies.

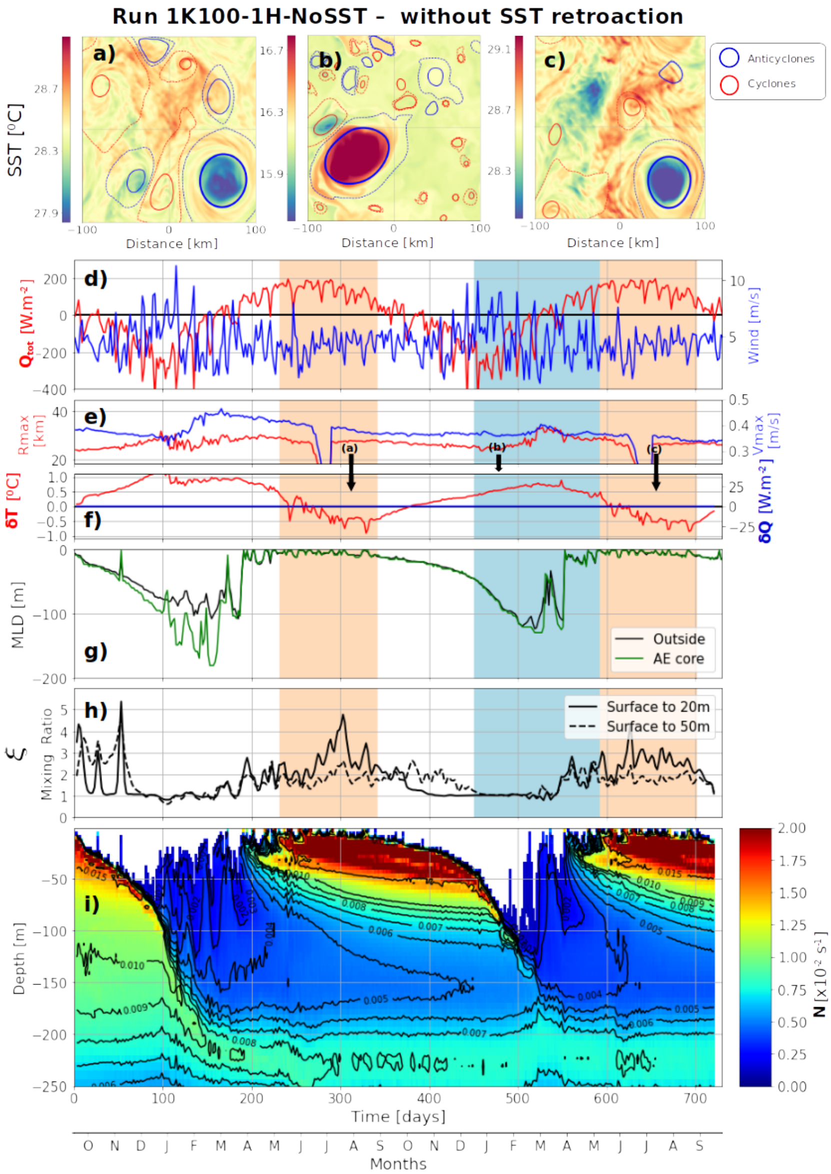

Figure 11. Simulation 1K100-1H-NoSST from Table 1. Same as in Fig.4 but without SST retroaction on air-sea fluxes. Discontinuities in R_{max} and V_{max} in panel (e) are due to the anticyclone crossing twice the grid borders.

SST retroaction acting as a negative feedback on SST anomalies can be analytically expected as linear. The derivative of each heat component with respect to T_s is indeed approximately constant (T_s being in Kelvin in Eq.13). Transfer coefficients C_E and C_S are indeed much more dependent on wind speed than on temperature, varying roughly about 0.2 with a T_s change of $1K$. The most sensitive case is a low air-sea temperature difference with weak wind, in which the boundary layer can switch from stable to unstable conditions (see for instance Fig.A1b from Pettenuzzo et al. (2010)). Assuming C_E and C_S are roughly constant with respect to temperature one gets :

$$\frac{\partial Q_{LW}^{\uparrow}}{\partial T_s} = -4\epsilon_{sb}\sigma_{sb}T_s^3 \approx -6\,W.m^{-2}.K^{-1} \tag{13}$$

$$\frac{\partial Q_{Lat}}{\partial T_s} \approx -\frac{\rho_a L_E C_E |V| 0.610}{P_{SL}} \frac{dP_{sat}}{dT_s} \approx -3 \times 10^1 \, W.m^{-2}.K^{-1} \tag{14}$$

$$\frac{\partial Q_{Sen}}{\partial T_s} = -\rho_a c_p C_S |V| \approx -1 \times 10^1 \, W.m^{-2}.K^{-1} \tag{15}$$

Altogether a thermal feedback on the order of $\frac{dQ_{tot}}{dT_s} \approx -4 \times 10^1 W.m^{-2}.K^{-1}$ is then expected, mostly driven by latent heat flux. THFF in Table 1 is computed only on the whole simulated year (from 365 to 730 days) and a value of $\approx -40 \, W.m^{-2}.K^{-1}$ is retrieved with a simple SST retroaction, in consistence with Eq.13 to 15. This value is relatively constant in our simulations, slightly decreasing for coarser resolution and lower forcing frequencies (see Table 1). $\partial C_E/\partial T_s$ and $\partial C_S/\partial T_s$ being also positive, taking this into account in Eq.14 leads to a even higher THFF estimate. THFF for the 1K100-1H simulation, defined here as δQ as a function of δT is shown in Fig.12. The obtained thermal feedback is consistent with previous estimates in coupled climate model : Ma et al. (2016) found a higher THFF ranging between 40 and $56 W.m^{-2}.K^{-1}$ but in the specific area of very warm eddies of the Kuroshio extension region. Moreton et al. (2021) found THFF ranging between 35 and $45 \, W.m^{-2}.K^{-1}$ over mesoscale eddies. They however used a composite approach in a model coupled with atmosphere and maximal oceanic resolution of 1/12°, for effective radius about $40km$. A coupled atmosphere layer is expected to further dampen the total THFF, taking into account other feedbacks than SST, in particular evaporation. Humidity is expected to increase over warm eddy, consequently decreasing the latent heat flux driving evaporation, whereas we applied a uniform h_{2m} field. Similar THFF in our simulations compared to coupled ocean-atmosphere models suggests that our results would not change significantly with more complex heat flux retroaction.

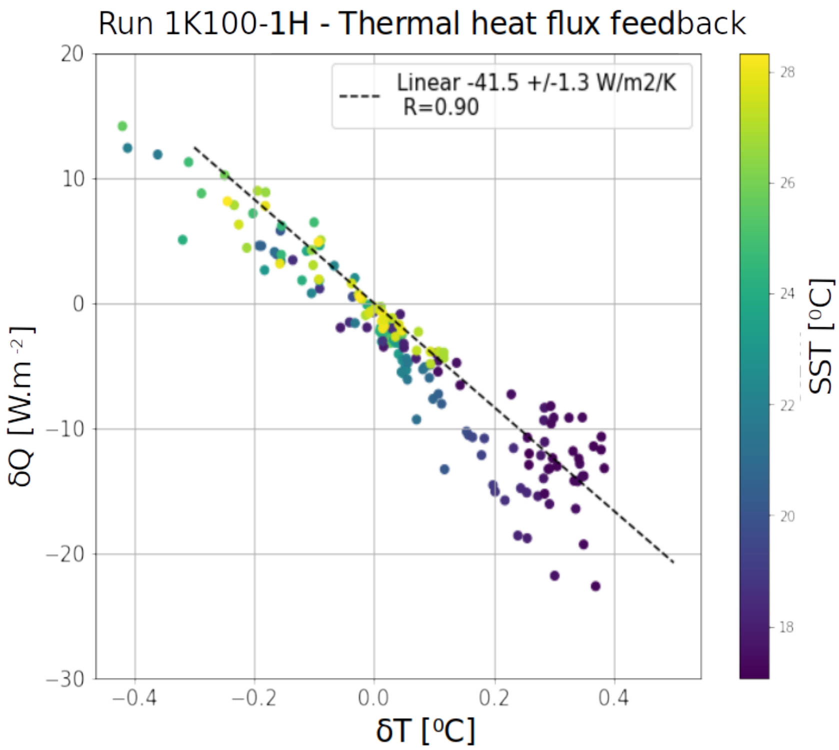

Figure 12. Thermal heat flux feedback in run 1K100-1H on the 2^{nd} simulated year, with linear regression as dashed black line, δQ and δT are from Fig.5f. Regression coefficient and parameters are indicated in the legend.

Without SST retroaction on air-sea fluxes, the most important difference from run 1K100-1H is the MLD anomaly variations. Outside-eddy, mixed layer evolution is very similar in runs 1K100-1H and 1K100-1H-NoSST reaching about 100m at its winter maximum, but the eddy MLD anomaly is an order of magnitude smaller ($\Delta MLD = 10m$, see Fig.11h). With no THFF, the MLD deepens at the same rate outside- and inside-eddy. Winter MLD deepening can be computed estimating the thermal loss ΔT, assuming a linear thermal linear stratification $\partial_z T$:

$$MLD = \frac{\Delta T}{\partial_z T} \tag{16}$$

The thermal loss is the integration of the heat flux over winter duration D. Assuming stratification is at first order the same outside- and inside-eddy, MLD anomaly is then driven by heat flux lateral gradients :

$$\Delta MLD = \frac{D}{\rho_0 c_p \partial_z T} \delta Q \tag{17}$$

In the 1K100-1H with SST retroaction on air-sea fluxes, δQ is positive in winter reaching about $+15W.m^{-2}$ over 4 months. This leads to an estimate $\Delta MLD \approx 2 \times 10^1 m$. This estimate should then be the eddy MLD anomaly contribution from THFF alone, but a lot higher difference is obtained between run 1K100-1H and 1K100-1H-NoSST. The main assumption in Eq.17 is that $\partial_z T$ is roughly the same inside- and outside-eddy. This is true in the upper layers where stratification is mostly the seasonal thermocline (see isopycnals in Fig.8c-d). At depth lower than 100m however, the anticyclone constitutes a more homogenized layer and this assumption should not hold as MLD should deepen faster inside-eddy, even with no SST retroaction. The very low ΔMLD found with no THFF then suggests that thermal feedback may also impact inside-eddy stratification. An example of inside-eddy MLD faster deepening is shown in Fig.3g : the MLD connects in February 2018 with the layer homogenized the previous winter and reaching quickly about 300m. Such mixed layer deepening acceleration is partly retrieved in run 1K100-1H around 500 days, with a MLD jump of about 30m (Fig.5g) inside-eddy but only about 10m outside-eddy. This coincides with the mixing of the subsurface homogenized layer formed in the first winter, despite diffusion (stratification isolines progressively closing, Fig. 5i) as discussed earlier.

ΔMLD is however still relatively weak compared to the 200 to 300m MLD anomalies observed in Mediterranean anticyclones (Barboni, Coadou-Chaventon, et al., 2023). Two main hypotheses can be proposed, the first being that some interannual variability is needed. The second hypothesis is that layers homogenized by winter MLD progressively restratify at depth in summer due to numerical diffusion. MLD in the following winter will then have to break this numerical stratification. This second hypothesis entails that the vertical grid is not enough refined yet to correctly preserve homogenized layers from one winter to another. From the comparison between runs 1K100-1H and 1K100-1H-NoSST shows that SST retroaction on air-sea fluxes is necessary to obtained eddy MLD anomalies, but quantitative description deserves further research and ΔMLD is not only driven by fluxes gradients at the eddy scale.

Conclusions

Idealized numerical experiment at high horizontal resolution and high frequency atmospheric forcing are able to qualitatively and quantitatively retrieve SST signature seasonal cycle for a mesoscale anticyclone. Starting from a surface intensified mesoscale anticyclone at $Ro \approx 0.16$, seasonal oscillations of the eddy SST anomalies are recovered with an 1km resolution, hourly atmospheric forcing and SST retroaction on air-sea fluxes. Retrieved eddy anomalies are a warm winter SST feature at $\delta T \approx +0.5^{\circ}C$ and a cold summer SST at $\delta T \approx -0.2^{\circ}C$, in consistence with in situ observations. The shift from warm winter SST signature to summer cold one is partly explained by an increased vertical mixing in the anticyclone upper layers. This differential mixing is due to higher NIW energy propagation well captured through the $\kappa - \epsilon$ mixing parametrization.

A sensitivity analysis reveals that this differential mixing depends on the grid resolution. Model diffusivity near the surface is then consistently 3 times higher in summer inside-eddy than outside for horizontal resolution of 1km or smaller. This resolution corresponds to an explicitly resolved first baroclinic deformation radius. Sensitivity to the forcing frequency is investigated by progressively removing high frequencies from the atmospheric input fields. A threshold behavior is observed when forcing frequency is lower than a day, then differential mixing dramatically vanishes with no significant summer cold-core anticyclonic SST. Vertical kinetic energy signing internal wave propagation indeed reveals a second powerful peak at $\omega = 1.0cpd$ inside the anticyclone in near-surface, corresponding to the effective inertial frequency and responding to high frequency forcing. This peak is absent outside-eddy because the cut-off inertia frequency $f = 1.24cpd$ is higher. Such an analysis suggests a significant impact of the eddy vorticity as cut-off frequency in allowing or not the selective NIW propagation. Weaker eddy SST seasonal oscillations are also retrieved in the absence of high frequently forcing and consequently without differential mixing (3-days and 1-week experiments). This highlights that other contributions might participate to these eddy SST signatures, in particular lateral exchanges. A new question for future research opened by this eddy-modulated mixing is how it depends on the eddy vorticity and size.

SST retroaction on air-sea fluxes is not found to be responsible of eddy SST signatures seasonal shift, as the seasonal oscillation is retrieved with and without air-sea fluxes parametrization. However this retroaction is logically found to dampen the SST anomalies, and then reduces eddy anomalies magnitude in both summer and winter. The average thermal heat flux feedback of our mesoscale anticyclone is approximately $40W.m^{-2}.K^{-1}$, in consistence with analytical derivation and previous studies.

Significant eddy-induced mixed layer anomaly $\Delta MLD \approx 50m$ are found at 1km horizontal resolution, only in the presence of SST retroaction on fluxes. Linear MLD anomaly analysis suggests that the thermal feedback is only responsible for about half of the MLD anomaly. Further analysis should then investigate how SST retroaction impacts inside-eddy stratification. MLD anomalies do not completely converge at 1km as larger anomalies are obtained with a 500m resolution due to restratification beginning outside-eddy driven by submesoscale instabilities, despite similar maximal mixed-layer at the anticyclone core. No restratification delay is clearly observed, but it could occur at even higher resolution inside the anticyclone because the balanced density gradients inhibits mixed layer instabilities there. This hypothesis is consistent with observations (Barboni, Coadou-Chaventon, et al., 2023) but would deserve more investigation in the future. This result is also important as the mixed layer is a significant driver of atmospheric and bio-geochemical exchanges, and the explicit resolution of submesoscale processes might be needed to accurately reproduce their interaction with eddies (Capet et al., 2008; Lévy et al., 2018). An important result is still that significant ΔMLD is retrieved only when SST exerts a retroaction on air-sea fluxes, but the quantitative description of its evolution would deserve more analysis.

This is the first time that subinertial waves concentration in anticyclones is linked to an increased mixing in near surface, spontaneously retrieved through the $k{-}\epsilon$ mixing closure. Mixing modulation by eddies suggests a strong scale interactions between subinertial internal waves ($\omega \lesssim f$) and the mesoscale ($\omega \ll f$). Differential mixing triggered by high frequency winds is an important result highlighting the need of both fine resolution and atmospheric forcing at sufficiently high frequency to correctly reproduce mesoscale eddies evolution. At present stage, global operational models do not have the spatial resolution to capture these phenomena. According to this study, 1/120 °resolution with 100 vertical levels would then be necessary to reproduce accurately mesoscale temporal evolution.

Open Research Section

In-situ profiles colocalized with mesoscale eddies database is available at `https://doi.org/10.17882/93077`. AMEDA eddy tracking algorithm is open source and available at `https://github.com/briaclevu/AMEDA`. ERA5 atmospheric reanalysis are publicly available at `https://doi.org/10.24381/cds.adbb2d47`. The CROCO code is publicly available at `https://www.croco-ocean.org/`.

Acknowledgments
Authors gratefully acknowledge *Ifremer* and *Service Hydrographique et Océanographique de la Marine* for their use of the *Datarmor* computing facility. Authors acknowledge Evangelos Moschos (*Amphitrite*) for the reuse of his figures as snapshots in Fig.3a-d. Authors also acknowledge fruitful discussions with Clément Vic (*Ifremer*), in particular the comparison with observations in Fig.8 and 10.

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