Seasonal cycle of the mixed layer, the seasonal thermocline and the upper-ocean heat storage rate in the Mediterranean Sea derived from observations

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A B S T R A C T

We present a Mediterranean climatology (1° × 1° × 12 months) of the mixed layer and of the seasonal thermocline, based on a comprehensive collection of temperature profiles spanning 44 years (1969–2012). The database includes more than 190,000 profiles, merging CTD, MBT/XBT, profiling floats, and gliders observations. This data set is first used to describe the seasonal cycle of the mixed layer depth and temperature, together with the seasonal thermocline depth and averaged temperature, on the whole Mediterranean on a monthly climatological basis. Our analysis discriminates several regions with coherent behaviors, in particular the deep water formation sites, characterized by significant differences in the winter mixing intensity. Heat Storage Rate (HSR) is calculated as the time rate of change of the heat content due to variations in the temperature integrated from the surface down to the base of the seasonal thermocline. For the first time the quantification of heat storage rate in the upper-ocean, based only on in situ oceanographic data, is made for the whole Mediterranean. The spatial and temporal variability of the HSR in the Mediterranean Sea and its link with dynamic structures like oceanic gyres are also discussed.

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Introduction

The Mediterranean Sea is a semi-enclosed basin connected with the Atlantic (the Gibraltar Strait, ~300 m depth) and with the Black Sea (the Dardanelles Strait, ~100 m depth). It is composed of two main basins, the Western and the Eastern Mediterranean (WMED and EMED) separated by the strait of Sicily (~400 m depth), and eight sub-basins. The Alboran Sea (ALB), the Algerian Basin (ALG), the Northwestern Mediterranean (NWM, delimited to the South by the Balearic Islands and the Sardina) and the Tyrrenhenian Sea (TY) compose the WMED, while the Ionian Sea (IO), the Adriatic Sea (AD), the Aegean Sea (AG) and the Levantine Basin (LE) compose the EMED (Fig. 1).

In particular the Mediterranean Sea has different deep convection zones (in the West and in the East) and a well-defined overturning circulation (Wüst, 1961; Robinson et al., 2001) with distinct intermediate and deep water masses. The total Mediterranean heat and freshwater surface budgets over a long multi-year period are negative. These deficits of freshwater and heat are compensated by exchanges through the Strait of Gibraltar (positive net water and heat transports), where the inflow is composed by a relatively warm and fresh (15.4°C, 36.2 psu) upper water, and the outflow to the Atlantic is relatively cooler and saltier (13°C, 38.4 psu) (Bryden et al., 1994; Tsimplis and Bryden, 2000; Soto-Navarro et al., 2010; Criado-Aldeanueva et al., 2012).

Several past studies analyzed the climatological structure of the salinity and temperature fields of the Mediterranean Sea from observations, based on a variational inverse model (Brankart and Brasseur, 1998), or on inverse methods (Tziperman and Malanotte-Rizzoli, 1991), and even fewer studies estimated the heat content changes in the Mediterranean Sea (Krahmann et al., 2000; Matsoukas et al., 2005). In this work we estimate the heat content changes only in the upper layer of the Mediterranean, because the number of observational data is very large compared to deeper layers. We define the upper-ocean layer as a combination of an upper mixed layer, where temperature is almost vertically uniform, and a seasonal thermocline. The depth of the seasonal thermocline was determined as the depth of the temperature minimum on temperature profiles, except in cases of no distinguishable seasonal

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thermocline (in winter) where we used the mixed layer depth. The temperature minimum associated to the seasonal thermocline can be viewed as the mixed layer temperature (the vertically averaged temperature of the mixed layer) during deep winter ventilation of the water column. We have chosen to calculate heat storage rates down to the seasonal thermocline in order to not omit the heat stored below the mixed layer depth. We choose to work with the thermocline, rather than the pycnocline, because of the larger number of temperature profiles with respect to salinity profiles.

For the first time, climatologies of the seasonal thermocline depth and averaged temperature, together with climatology of the upper-ocean heat storage rate, will be provided for the Mediterranean Sea. In addition, 8 supplementary years of data (corresponding to more than 60,000 profiles, thanks to massive Argo deployments) are available since the computation of the last MLD climatology for the Mediterranean by D’Ortenzio et al. (2005), thus this work will provide also an updated version of the seasonal cycle of the mixed layer depth and temperature for the 1969–2012 period. As in D’Ortenzio et al. (2005), we have chosen to use simple averaging method to work with the thermocline, rather than the pycnocline, because of the larger number of temperature profiles with respect to salinity profiles.

The paper is organized as follows: First we describe the data sets and the method in the second section. Then we present our results in the third section, such as climatologies of the mixed layer depth and temperature, together with climatologies of the seasonal thermocline depth and averaged temperature, on the whole Mediterranean on a monthly climatological basis. Finally we discuss the spatial and temporal variability of the heat storage rate (HSR) in the Mediterranean Sea. Conclusions and perspectives are given in the fourth section.

Data sets and methods

Profile database

The primary source of data for this study is the Medar-MEDATLAS project (MEDAR Group, 2002). We also use data from the World Ocean Database (Conkright et al., 2002), from additional Italian (D’Ortenzio et al., 2005, http://www.mediterranean-marinedata.eu/moong/home.htm) and Spanish cruises (Puig et al., 2012), from the CORIOLIS data center (see Coriolis, http://www.coriolis.eu.org) and from deployments of gliders which are relatively new oceanographic platforms (Testor et al., 2010) carried out in the framework of several European and national projects (see EGO, http://www.ego-network.org). Gliders profiles are considered as vertical and are checked with the same quality control than Argo data.

After removal of duplicates and application of quality control procedures (elimination of profiles without data above 10 m below the surface, with constant temperature values, or with excessive temperature gradients; see details in De Boyer Montégut et al. (2004, Appendix A)), 140,083 profiles from 1969 to 2012 are kept for the analysis out of the initial 190,000. This database is composed by 45.8% of mechanical bathythermograph (MBT) and expandable bathythermograph (XBT/XCTD), 25.8% of conductivity–temperature–depth data (CTD from Research vessels cruises), 4.8% of Argo profiling floats data and 23.6% of EGO gliders data. This database is also composed of 74,934 salinity profiles (0.5% of XCTD, 47.2% of CTD, 8.9% of Argo profiling floats data and 43.4% of EGO gliders data). This represents more than 50,000 additional salinity profiles compared to the mixed layer climatology made by D’Ortenzio et al. (2005). This is mainly due to the increasing number of glider deployments (43 since 2006). However, the spatial distribution of these salinity profiles (often distributed along repeat-sections) is still not yet sufficient to have a horizontal description of a pycnocline climatology. The 110,000 supplementary temperature profiles, compared to the salinity profiles, is one of the main reason why we chose to work on the thermocline base, rather than on the pycnocline base.

Because XBT and MBT data compose almost 50% of our database and are known to be biased in temperature, we have paid a special attention in the correction of these data. The manufacturer’s product catalogues specify a depth accuracy of >1% of sample depth and a temperature accuracy of 0.1°C for MBT, and a depth accuracy of 5 m (0–250 m) or 2% below 250 m and a temperature accuracy of 0.2°C for XBT. The MBT are characterized by smaller and less time-variable biases compared to the XBT. Recently Gouretski and Koltermann (2007) discovered the existence of a globally time-dependent and systemic warm bias in XBT profiles, caused...
by depth error calculation and thermistor bias. Since that time, several authors have deduced a time-variable bias, modeling the bias as a depth error only (Wijffels et al., 2008; Ishii and Kimoto, 2009), a temperature bias (Levitus et al., 2009), or a combination (Gouretski and Reseghetti, 2010; Cheng et al., 2011; Hamon et al., 2012; Cowley et al., 2013).

In this work we correct the depth calculation and temperature biases in Bathythermograph data. The mechanical bathythermographs (MBT) data are corrected using the updated correction from Gouretski and Koltermann (2007) (http://www.nodc.noaa.gov/OC5/mbt-bias/gouretski_new.html), while the expendable bathythermographs (XBT) data are corrected using Cowley et al. (2013). The large database of over 4100 side-by-side deployments of XBTs and CTD data used in Cowley et al. (2013) allow them to separate out the pure temperature bias from depth error in a way that was not previously possible. The correction steps applied on our XBT data can be summarized by: (1) an identification of the appropriate correction depending of the probe type. If there is no information about XBT types (74,500 of the 86,000 profiles), the terminal depth is used to determine the probe type for data carried out before 1996. XBT data from 1996 to the present with no depth equation information (2000 profiles) are not included in the climatology calculation, since we do not know which depth equation was used (manufacturer or Hanawa fall rates equation, Hanawa et al., 1995); (2) convert to Hanawa fall rates if required; and (3) apply Cowley thermal gradient corrections. A more detailed description of the correction steps can be read on http://www.nodc.noaa.gov/OC5/XBT_BIAS/cowley.html.

Heat storage rate calculation

Moisan and Niiler (1998) derived the heat storage rate equation based on the conservation of mass equation and the conservation of heat equation without thermal conductivity term. Following this formalism, we can express the heat conservation equation integrated from the surface down to a chosen time- and space-dependent depth \( h = f(x,y,t) \) as follows:

\[
\frac{h \partial T_a}{\partial t} = -h \nabla_T \cdot \nabla - \nabla \cdot \left( \int_0^h \nabla T \, dz \right) - (T_a - T_{0, \text{h}}) \\
\times \left( \frac{\partial h}{\partial t} + \nabla_h \cdot \nabla_h + \mathbf{w} \cdot \mathbf{w}_{\text{h}} \right) + \frac{\text{NHF}}{\rho_c}.
\]

(E.1)

Here \( h \) is the depth level above which the depth-averaged temperature \( T_a \) and the depth-averaged horizontal velocity \( \mathbf{v} \) are calculated, \( \nabla \) is the horizontal gradient operator, \( \mathbf{v} \) is the horizontal divergence operator, \( \mathbf{v} \) is the deviation from the vertically averaged horizontal velocity (\( \mathbf{v} = \mathbf{v}_a + \mathbf{v} \)), \( T \) is the deviation from the vertically averaged temperature (\( T = T_a + T_{0, \text{h}} \)), and \( T_{0, \text{h}} \) and \( \mathbf{w}_{\text{h}} \) are the temperature and the vertical speed at the depth level \( h \). NHF is the net heat flux across the ocean surface, \( q_{\text{h}} \) is the net heat flux across the surface at depth \( -h \), and \( \rho \) and \( c_p \) are the mean density and specific heat of seawater.

The Heat Storage Rate (HSR, left term in (E.1)) can be expressed in terms of horizontal heat advection, vertical temperature/velocity covariance, entrainment processes at the depth \( h \) (deepening or shoaling of the \( h \) interface), horizontal advection through the sloping \( h \) interface, and vertical velocity at the base of the \( h \) interface), net surface heat flux adjusted for the amount of short wave radiation that penetrate the depth \( h \).

In their study, Moisan and Niiler (1998) defined the integration depth \( h \) as the depth of the isotherm whose temperature is one degree less than the annual coldest surface temperature in each region. This cannot be applied in the Mediterranean because most of the time, temperature can be colder in winter, at the surface than at any other depth. We choose an integration depth \( h \) corresponding to the bottom of the seasonal thermocline. Due to the weak vertical gradients at that depth, the net heat flux across the surface at depth \( -h \) is considered as negligible.

**Determination of the integration depth and the mixed layer depth**

**Definition of the integration depth**

In this work we estimate the heat storage rate only in the upper-ocean layer because we do not have enough deep data to

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**Fig. 2.** (a) The mean monthly temperature profiles from the 41–42°N, 4.5–5.5°E bin, the integration depth \( h \) is indicated for each month by a circle or a cross. The time series of (b) \( h \) and (c) \( T_a \) are shown for this bin.

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compute the heat storage rate integrated down to the bottom. In order to calculate HSR down to the base of the thermocline, we need to define first the thermocline in our profiles. We choose the integration depth $h$ as the depth where there is a local minimum temperature $T_{\text{min}}$ in the first 200 m. This temperature minimum can be viewed as the temperature of the mixed layer during deep winter ventilation of the water column. Even if the temperature minimum $T_{\text{min}}$ is nearly constant at $0.1^\circ C$ during the whole year (Fig. 2a), $h$ is not necessarily constant over the year. This seasonal variability of $h$ may be explained by the contribution of the different terms of (E.1).

In winter, the strong surface buoyancy losses increase the mixed layer depth (MLD) and the seasonal thermocline cannot be distinguishable on temperature profiles (Fig. 2a). Thus, a double criterion is chosen to calculate the integration depth $h$. For each individual profile, we calculate the seasonal thermocline base (corresponding to the depth where a temperature minimum occurred in the upper 200 m), and the MLD using a $\Delta T = 0.1^\circ C$ criterion and reference level at 10 m depth. In cases of no distinguishable seasonal thermocline (in winter), the integration depth is chosen to be the base of the MLD, otherwise the integration depth $h$ is chosen to be the depth of the seasonal thermocline (determined using a temperature minimum criterion).

**Assessment of the mixed layer depth criterion**

We assess our estimate of the mixed layer depth with other estimations based on temperature or density criteria used in previous studies. Unlike D’Ortenzio et al. (2005) who used a $\Delta T = 0.2^\circ C$ criterion, we choose a $\Delta T = 0.1^\circ C$ criterion to define the MLD on our profiles, because a finest temperature criterion reduces the difference between a MLD calculated with a temperature criterion and a MLD calculated on temperature–salinity profiles with a density criterion. Maps of the monthly mean differences between MLD calculated with a temperature criterion and MLD calculated with a density criterion are shown in Fig. 3. For this comparison, we use only the profiles with available temperature and salinity data (75,000). We calculate for two different $\Delta T$ criteria ($0.2^\circ C$ and $0.1^\circ C$) the MLD based on a temperature criterion and on a density criterion.

![Fig. 3. On the left (resp. right), monthly maps of the MLD difference between a $\Delta T = 0.2^\circ C$ (resp. $0.1^\circ C$) criterion and a variable $\Delta T$ criterion corresponding to a fixed $\Delta T$ decrease of $0.2^\circ C$ (resp. $0.1^\circ C$).](http://dx.doi.org/10.1016/j.pocean.2014.11.004)
Climatologies of \( h \) and \( T_{a} \)

The processing steps are summarized on the flowchart shown in Fig. 4. \( h \) and \( T_{a} \) are calculated from each single profile for the period 1969–2012 and are binned into a 1° latitude by 1° longitude grid. Median values are calculated for each box and for each month of each year. Hereafter we will refer to the median estimations as \( \overline{h_{b,y,m}} \) and \( \overline{T_{a,b,y,m}} \), where \( b \) correspond to the box index over the horizontal grid, \( y \) indicates the year between 1969 and 2012 and \( m \) loop over the month (from 1 to 12).

We smooth out short spatial fluctuations in our estimations of \( \overline{h_{b,y,m}} \) and \( \overline{T_{a,b,y,m}} \), by taking slightly displaced binned data and then taking the average value on the \( 0.5^{+} \times 0.5^{-} \) overlapping grid. So with 4 different \( 1^{+} \times 1^{-} \) grids we reconstructed estimations of \( \overline{h_{b,y,m}} \) and \( \overline{T_{a,b,y,m}} \), and the standard deviation \( \sigma(\overline{h_{b,y,m}}) \) and \( \sigma(\overline{T_{a,b,y,m}}) \) on a \( 0.5^{+} \times 0.5^{-} \) grid.

Finally, mean seasonal cycles are obtained in all b boxes by calculating the mean \( \overline{h_{a,b,y,m}} \), \( \overline{T_{a,b,y,m}} \), and the standard deviation \( \sigma(\overline{h_{b,y,m}}) \), \( \sigma(\overline{T_{a,b,y,m}}) \) of \( \overline{h_{b,y,m}} \) and \( \overline{T_{a,b,y,m}} \) over the years. Calculations are done only in boxes that contain at least 3 values for each climatological month \( m \). In our climatology definition, we choose to give the same weights for all years to avoid effects of oversampling during some specific years and to obtain a climatology less biased by the non-uniform sampling density in time. For example due to massive glider deployments like in 2007 and 2008 in the Northwestern Mediterranean, or due to the fact that glider data, largely deployed only since 2004 in the Mediterranean, may represent 80% of the available profiles in some boxes.

The number of available years used in the calculation of \( \overline{h_{a,b,y,m}} \), \( \overline{T_{a,b,y,m}} \) may be a source of error. A confidence interval for these mean parameters can be calculated by using the number of years of available data and the standard deviations shown in Section ‘Seasonal cycle of the mixed layer’, the integration depth \( h \) and the depth-averaged temperature \( T_{a} \). Applications of Student’s t test for 95% confidence limits give us uncertainty:

1. on \( \overline{h_{a,b,y,m}} \) that is about 15 m for most of the year, except in Winter in the deep water formation area (Gulf of Lions, Adriatic Sea, Aegean Sea) where uncertainty can be up to 200 m locally due to higher values of the standard deviation \( \sigma(\overline{h_{b,y,m}}) \) (discussed in detail in Section ‘Mixed layer depth and integration depth’);
2. on \( \overline{T_{a,b,y,m}} \) that is about 0.2 °C for most of the year. At the end of the summer, the uncertainty increase around 0.5 °C for most of the Mediterranean (up to 1.2 °C in October in the Ierapetra Gyre) due to a higher standard deviation \( \sigma(\overline{T_{a,b,y,m}}) \) (discussed in detail in Section ‘Mixed layer temperature and the upper-ocean depth-averaged temperature’).

Errors in the estimation of the seasonal cycle of the mixed layer depth and temperature may also be due to the spatial averaging used in the gridding procedure, for each month of each year. This can introduce a significant amount of error, particularly when observations are sparse. To reduce these errors, for each month of each year, individual values of temperature are compared to
the mean and standard deviation calculated from all temperature profiles made in the same box for the same month (defined by a monthly mean value of temperature $T_{a}^{m}$ and a standard deviation $\sigma(T_{a}^{m})$). Individual values are rejected in the calculation of $T_{a}^{m}$ if their deviation from $T_{a}^{m}$ is 3 times greater than $\sigma(T_{a}^{m})$.

Finally, another source of error is the effect of potential instrumental bias, but assuming this error is random for each year due to changes in the instruments and in the platform tracks (which ensure on average a relatively homogeneous spatio-temporal coverage), it should be restricted to individual monthly data set and should not affect the seasonal climatology. In addition, since the 1970s (date chosen for the beginning of our climatologies), the temperature sensors do not show strong bias (maximum value of 0.1–0.2 °C), compared to conductivity sensors that are more prone to drift over time.

MBT and XBT represent 45% of the database and their corrections did not affect significantly the basin-mean seasonal cycle of $h$, $T_{a}$ and HSR, since they are two order lower than the amplitude of their seasonal variations. To estimate the impact of the MBT/XBT corrections, we compute the climatologies of $h$, $T_{a}$ and HSR, following the processing steps summarized on the flowchart shown in Fig. 4. This gives us another climatology without MBT and XBT data corrections. Then, we perform the difference between the $1' \times 1' \times 12$ months climatologies of $h$, $T_{a}$ and HSR, calculated without MBT/XBT corrections, with the $1' \times 1' \times 12$ months climatologies calculated with MBT/XBT corrections. The monthly values for the basin-mean of the differences between the climatologies with the XBT/MBT corrections and the ones without corrections are comprised between: (1) 0.1 m and 2.3 m for $h$, and (2) $-0.08$ °C and $-0.05$ °C for $T_{a}$ and (3) $-2.1$ W m$^{-2}$ and $2.0$ W m$^{-2}$ for HSR.

**Climatologies of HSR**

The calculation of HSR is done only in boxes that contain a full 12-month time series of the $h^{m}$ and $T_{a}^{m}$ with at least 3 values for each month. The heat storage rate $HSR^{m}$ is calculated using centered derivative for each month $m$ and each grid point $b$:

$$HSR^{m} = \frac{\rho c_{p} \frac{T_{a}^{m+1} - T_{a}^{m-1}}{\Delta t} h^{m}}{H^{m}}$$

where $T_{a}^{m}$ is the mean temperature at the base of the integrated layer for each month $m$ and each box $b$, $\Delta t$ is the number of seconds between the center of the month $m+1$ and the center of the month $m-1$. For $m$ equal to the first month (resp. the last month) of the year, $m-1$ (resp. $m+1$) corresponds to the last month (resp. the first month) of the year.

Error propagation methods are overestimating uncertainties, so we use a Monte Carlo approach to estimate the uncertainties on HSR. We computed a Monte Carlo test (Krahmann et al., 2000) at each grid point $b$ and for each month $m$. Ten thousand estimations $HSR_{c}^{m}$ are calculated, wherein each estimation $X_i$ (respectively $Y_i$ and $Z_i$) come from random number chosen from normal distribution with mean $T_{a}^{m}$ (respectively $T_{a}^{m-1}$ and $h^{m}$) and standard deviation $\sigma(T_{a}^{m})$ (respectively $\sigma(T_{a}^{m-1})$ and $\sigma(h^{m})$). The mean heat storage rate ($HSR_{c}^{m}$) for these 10,000 realizations is calculated by the formula:

![Fig. 5. Number of years in which measurements were available for the computation of the climatology.](please cite this article in press as: Houpert, L., et al. Seasonal cycle of the mixed layer, the seasonal thermocline and the upper-ocean heat storage rate in the Mediterranean Sea derived from observations. Prog. Oceanogr. (2014), http://dx.doi.org/10.1016/j.pocean.2014.11.004)
\[
(HSR_{b,m}^{i}) = \frac{1}{N} \sum_{i=1}^{N} HSR_{b,m}^{i} = \frac{1}{N} \sum_{i=1}^{N} \rho_{p} \frac{X_i - Y_i}{A t} Z_i
\]

where \(N\) is the number of realizations. We use the same method for calculation of the standard deviation.

To justify why we choose random numbers from normal distributions in the Monte Carlo tests, we applied a Kolmogorov–Smirnov test (Gille, 2004) to compare the inter-annual values of the monthly averaged \(T_{b,m}^{i}\) (resp. \(h_{b,m}^{i}\)) to a normal distribution with mean \(T_{b,m}^{i}\) (respectively \(h_{b,m}^{i}\)) and standard deviation \(\sigma(T_{b,m}^{i})\) (respectively \(\sigma(h_{b,m}^{i})\)). The null hypothesis is that \(T_{b,m}^{i}\) (resp. \(h_{b,m}^{i}\)) came from a normal distribution. We obtain as results that in the two cases, we cannot reject this hypothesis at a 5% significant level.

This Monte Carlo approach, compared to a classical calculation of error propagation, specifies our estimation of the standard deviation of HSR, \(\sigma(HSR_{b,m}^{i})\). On average, the results from the Monte Carlo test reduce the standard deviation on \(HSR_{b,m}^{i}\) by 100 W m\(^{-2}\) compared to the error propagation method. In some places, like

![Mediterranean climatology of the upper-ocean layer h, determined by the depth of the seasonal thermocline, except in cases of no distinguishable seasonal thermocline (in winter) where the mixed layer depth was used.](image-url)
in the Ierapetra Gyre (South East of Crete), the standard deviation on $H_{SR}^b$ is reduced by 500 W m$^{-2}$ using the Monte Carlo method. This may be explained by the important standard deviation value for $T_a$ that increases the value of the standard deviation obtained by the error propagation method, particularly in the temperature derivative calculation.

Even if the estimations of $(H_{SR}^{b,m})$ are close to the estimations of $H_{SR}^b$ (the basin-average difference is less than 0.1 W m$^{-2}$), there are some discrepancies, up to 10 W m$^{-2}$ in January in the Gulf of Lions. They may be explained by the non-linearity of the $H_{SR}$ term. This area is a well-known place where deep ocean convection and dense shelf water cascading occur (CIESM, 2009; Durrieu de Madron et al., 2013). So, the high inter-annual variability associated with these Deep Water Formation (DWF) process is reflected in the high standard deviation of $h$, which is superior to 200 m from December to March. A maximum value of 670 m for the standard deviation is found in January in the Gulf of Lions with 13 years of data available in January (Fig. 5). Boxes with differences between $(H_{SR}^{b,m})$ and $H_{SR}^b$ also correspond to boxes where the standard deviation of $H_{SR}^{b,m}$, $\sigma(H_{SR}^{b,m})$, is significantly reduced.

Fig. 7. Mediterranean climatology of MLD, based on a temperature difference criterion of $\Delta T = 0.1 \, ^\circ\text{C}$ applied to individual profiles.
due to the Monte Carlo method. In the following, the term HSR will be used to refer to $\langle HSR^{b,m} \rangle$.

**Results and discussion**

*Seasonal cycle of the mixed layer, the integration depth $h$ and the depth-averaged temperature $T_a$*

**Mixed layer depth and integration depth**

After the processing stages described in the Section 'Determination of the integration depth and the mixed layer depth', we get the seasonal climatologies of $h$ and $T_a$. The depth of the seasonal thermocline (Fig. 6) shows mainly two seasons in the DWF area: the vertically mixed season from December to April with the absence of the seasonal thermocline, and the stratified season from May to November. $h$ shows maximal values between January and March due to the winter deepening of the mixed layer (Fig. 7), like in the previous median-MLD climatology of D’Ortenzio et al. (2005). In April the restratification of the water column is important: the MLD is not deeper than 30–40 m except in some coastal areas where it can reach 80 m. Maxima are found in wintertime in known places of DWF regions (Lascaratos et al., 1999) like the Gulf

![Maps of the standard deviation associated to the mesh box averages of the MLD shown in Fig. 7.](image)
of Lions (\( \sim 42^\circ N 5^\circ E \), Medoc Group, 1970) and the Southern Adriatic Sea (\( \sim 42^\circ N 18^\circ E \), Pollak, 1951), the Rhodes Gyre (\( \sim 36^\circ N 29^\circ E \), Ovchinnikov, 1984), in the Southern Aegean Sea (\( \sim 36^\circ N 25^\circ E \), Pollak, 1951) and in the Northern Aegean Sea (\( \sim 39^\circ N 25^\circ E \), Theocharis and Georgopoulos, 1993). In February, maximal monthly mean values of 425 m depth and 300 m depth are found respectively in the Gulf of Lions and in the Southern Adriatic Sea. There are others regions that present a local maximum of mixed layer depth (Fig. 7) and these regions are associated to anticyclonic circulation: the Southwest of Greece (\( \sim 36^\circ N 22^\circ E \) in February/March, Malanotte-Rizzoli et al., 1997) and the region South of Cyprus (\( \sim 34^\circ N 33^\circ E \) from January to March, Zodiatis et al., 2005). Although the region south-east of Crete (\( \sim 34^\circ N 27^\circ E \)) does not show a clear maximum in the monthly MLD climatology (Fig. 7), it is associated with higher standard deviation values (Fig. 8) certainly related to the presence of the anticyclonic Ierapetra Gyre (Menna et al., 2012). The presence of anticyclonic or cyclonic gyres is more evident on the climatology of the depth-averaged temperature, discussed later in this section. One can also distinguish a local MLD maximum in the North Ionian Sea, in a place where the upper-ocean circulation shows a decadal variability, known as the Bimodal Oscillating System (BiOS, Gačić et al., 2010).

Our monthly MLD climatology presents similarities with the one made by D’Ortenzio et al. (2005). We find the same kind of seasonal variations for the whole basin and we also find zones of deep winter ventilation in the same place than D’Ortenzio et al. (2005).
The difference of the MLT (Mixed Layer Temperature) between the two climatologies is on average less than 0.1 °C, but in some very specific locations (coastal zones), the difference reach 0.8 °C. Concerning the MLD, the main differences are found in winter (~20 m), except in the Gulf of Lions where the difference between the two MLD climatologies can be up to 250 m, particularly the maximum deepening of the mixed layer is not reached the same month (in February in our climatology). There are several sources of explanation for these differences:

1. the $\Delta T = 0.1$ °C criterion used in this work compared to the $\Delta T = 0.2$ °C of D’Ortenzio et al. (2005), that gives us shallower estimations of the MLD, particularly in March, which is more in agreement with an isopycnal definition of the MLD,
2. the grid and the estimator chosen for the climatology calculation (in this work we chose to maximizing the number of profiles in each box by making monthly-average over the year in 1° × 1° boxes, while D’Ortenzio et al. (2005) preferred to use median values on 0.5° × 0.5° boxes), and
3. the 8 additional years of measurements (from 2004 to 2012) used in our climatology, since 2004 the number of measurements increased in the whole Mediterranean, especially thanks to the numerous Argo and glider deployments made in addition to the “classical” XBT/CTD monitoring. Another difference from

![Fig. 10. Mediterranean climatology of the upper-ocean temperature $T_a$, determined by the depth-averaged temperature from the surface to the depth $h$.](image-url)
previous climatologies is the event of DWF in 2005–2006, which changed the stratification of the water column, and thus the mixed layer structure of the Gulf of Lions in the years to follow (Smith et al., 2008; Schroeder et al., 2008; Herrmann et al., 2010).

Standard deviations values of the mixed layer depth are generally below 30 m for most of the basin from September to December (Fig. 8). In winter and in April, the standard deviation values increase strongly throughout the whole basin, reaching the maximum values in the Gulf of Lions area (550 m). The irregular intensity and occurrence of the DWF processes (Mertens and Schott, 1998) is mainly responsible for these important standard deviation values. As an example, in the Gulf of Lions, deep convection can reach the bottom some years like in 2009, 2010, 2011 and 2012, while other years the winter mixing do no reach depths larger than 200 m like in 2007 or 1000 m like in 2008. Moreover, recent measurements showed that from one year to another the deepening of the mixed layer did not start exactly at the same time and the time lag, for the mixed layer to deepen beyond the Levantine Intermedi-
The LIW located at 200–400 m depth, can be up to 1 month and a half (Houpert et al., 2014).

The standard deviation of $h$, $\sigma(h^m)$, can be seen in Fig. 9. $\sigma(h^m)$ is inferior to 10 m from May to December in the east of the EMED, is comprised between 10 and 50 m in the east Alboran Sea, Algerian Basin and Ionian Sea, and shows higher values comprised between 30 and 50 m in the Tyrrhenian Sea and in the northwestern Mediterranean. From January to April, the standard deviation is important in the regions of DWF and, as for the MLD, is due to the irregular intensity and occurrence of the DWF processes. For most of the Mediterranean, zones of high standard deviation values are the results of variable intra box estimations of $h$ due to the meso-scale activity associated with fronts, or DWF processes.

The northern Ionian Sea is a region that presents large standard deviations of either the mixed layer (Fig. 8) or the thermocline depth (Fig. 9) which are not associated to dense water formation. This feature is clearly seen from January to March and may be explained by the decadal variability in the upper-ocean circulation of the Ionian Sea. To describe the recurrent reversals of the ocean upper-layer circulation, Gacic et al. (2010) proposed a feedback mechanism (BiOS) between the redistribution of water masses, related to variations in the thermohaline properties of the Southern Adriatic, and inver-

![Fig. 12. Mediterranean climatology of the thermocline slope.](image-url)
sions of the Ionian circulation. These decadal changes between a cyclonic and an anticyclonic circulation in the Ionian Sea involve also decadal changes in the upper ocean stratification (MLD, seasonal thermocline). In anticyclonic gyres the seasonal thermocline tends to be pushed down while in cyclonic gyres the seasonal thermocline tends to be pushed up. This decadal variability may explain why large standard deviation of either MLD or the seasonal thermocline depth are found in the northern Ionian Sea.

Mixed layer temperature and the upper-ocean depth-averaged temperature

The depth-averaged temperature \( T_a \) is presented in Fig. 10. On average over the whole Mediterranean, the depth-averaged temperature in the upper-ocean is going from 14.6 °C to 16.8 °C in September. One can clearly notice the contrast between the Western and the Eastern Mediterranean, with a colder basin-mean depth-averaged temperature in the WMED (from 13.7 °C in March to 15.9 °C in September) than in the EMED (from 15.3 °C in March to 17.5 °C in September). Another striking fact is the strong temperature anomalies along the Cretan Arc (at the junction of the Aegean Sea with the Ionian and Levantine basins, Fig. 1), from July to January, corresponding to known anticyclonic (warm anomaly) and cyclonic (cold anomaly) gyres (from West to East: the anticyclonic Pelopys Gyre (PC), the cyclonic West Cretan Gyre (WCG), the anticyclonic Ierapetra Gyre (IG), the cyclonic Rhodes Gyre (RG)).

The seasonal cycle of \( T_a \) in the center of RG is comprised between 0.46 °C and 0.75 °C on average on the whole Mediterranean. Maximum values (up to 2 °C) can be found in the east of the IG in October (Fig. S1). These higher standard deviations may be explained by the interannual variability of the position and the dimension of the gyres, already pointed out by Marullo et al. (1999).

Unlike the depth-averaged temperature \( T_a \), the mixed layer temperature presents a more pronounced seasonal cycle (Fig. 11), with minimal values below 13 °C found in February in the Adriatic Sea and in the Gulf of Lions, and maximal values reaching 28.0 °C are found in August in the east of the Levantine Basin. The basin-mean of the mixed layer temperature is going from 14.8 °C in February to 24.9 °C in August for the Mediterranean Sea, from 14.0 °C in February to 24.2 °C in August for the WMED, and from 15.4 °C in March to 25.5 °C in August for the EMED. The seasonal variability of the surface layer, strongly impacted by the solar heat flux, masks the upper-ocean temperature anomaly associated with the presence of the anticyclonic and cyclonic gyres in the EMED. They are more easily distinguishable by their depth average temperature \( T_a \) (Fig. 10).

The standard deviation of the mixed layer temperature (Fig. S2) presents more variability than the standard deviation of the depth integrated temperature \( T_a \), particularly in summer when the mixed layer is shallow and thus is more sensitive to the temporal and spatial variability of air–sea exchanges.

Climatology of the seasonal thermocline

Once the climatology of the depth of the seasonal thermocline base is done, one can easily produce a climatology of the seasonal thermocline slope. We defined the seasonal thermocline slope as \( \frac{T_a - M LT}{MLT} \), where \( T_a \) is the temperature at the base of the seasonal thermocline, \( MLT \) is the temperature at the base of the mixed layer, \( h \) is the depth of the seasonal thermocline and \( MLD \) is the mixed layer depth.

The monthly variability of the seasonal thermocline slope (Fig. 12) is characterized by a global decrease of the slope from September to February and an increase from April to August. This seasonal cycle is in good agreement with the seasonal cycle of the mixed layer (Fig. 7). Maxima are found in August, essentially in the Tyrrenhian Sea (0.08–0.1 °C m⁻¹) and around the Balearic Islands (0.07–0.08 °C m⁻¹). One can also notice that some regions like the DWF zones in the Northwestern Mediterranean, in the Adriatic Sea, in the Aegean Sea or the Rhodes Gyre, present a less pronounced seasonal thermocline than the other parts of the basin. These results are expected as those zones are known as the main zones of intermediate and deep water formation in the Mediterranean Sea (Schoeder et al., 2012). As the water column is well mixed almost every year in these zones, the local stratification and consequently the thermocline is weaker locally as illustrated for example in Herrmann et al. (2010).

Seasonal cycle of HSR on average over the Mediterranean

For the whole Mediterranean, the annual value of the basin-mean of HSR is estimated at \(-1.9 ± 4.0 \text{ W m}^{-2}\) for the 1969–2012 period. This does not establish a clear warming trend in the upper-ocean in contrast to Rixen et al. (2005) who showed a warming of the surface layer since the 1980s, but is not in contradiction with their results due to our uncertainties on HSR. Fig. 13 displays the basin-mean climatological annual cycle for HSR from ocean observations over the Mediterranean. One can see that the climatological annual cycle for HSR has a clear seasonal signal with minimal and maximal values being about \(-165.8 ± 4.5 \text{ W m}^{-2}\) in December, and \(+127.8 ± 3.5 \text{ W m}^{-2}\) in June.

Variability of local heat storage rates

The spatial pattern of the seasonal cycle of the HSR is presented in Fig. 14. The basin variability of the upper-ocean heat content changes is dominated by a seasonal cycle oscillating between minimal value found in December, and maximal value in June (Fig. 13 and Section ‘Seasonal cycle of HSR on average over the Mediterranean’). Zero values are reached in February/March and in September/October, due to extrema in the seasonal cycle of the depth-averaged temperature. The standard deviation of HSR, shown in Fig. 15, is highly variable: from 25 to 50 W m⁻² in March in the Tyrrenhian Sea to more than 300 W m⁻² in January in the Gulf of Lions, in fall in Ierapetra Gyre. Location of high standard deviation values for HSR are related to high standard deviation of \( h \) (e.g. in

![Fig. 13. Seasonal cycle of the basin-mean HSR.](image-url)
winter in DWF zones, Fig. 9) and of $T_o$ (e.g. in summer/fall, especially in the Ierapetra Gyre, Fig. S1).

Fig. 16 represents the seasonal cycle of the monthly HSR anomalies from the basin mean ($HSR_{ano}$). $HSR_{ano}$ presents significant local variations, some as high as 200 W m$^{-2}$. A positive $HSR_{ano}$ associated with a positive basin-mean (respectively a negative $HSR_{ano}$ associated with a negative basin-mean) indicates regions gaining (resp. losing) heat more quickly than the whole basin. In contrast, a positive $HSR_{ano}$ associated with a negative basin-mean (or the opposite) indicate region losing heat less quickly than the whole basin. These local modulations of HSR can be due to different forcings: (1) surface net heat flux, (2) horizontal heat advection, and (3) entrainment mixing at the interface depth, which cannot be calculated with our dataset. Local variations of $HSR_{ano}$ seem to be related to specific dynamic structures, like oceanic gyres.

The Alboran Sea

Coherent patterns can be distinguished from the map of monthly HSR anomalies (Fig. 16). From October to January in the Alboran Sea, anomalies are between 75 and 100 W m$^{-2}$ higher than the basin mean which is negative ($-53.0$, $-112.3$, $-166.7$ and $-155.5$ W m$^{-2}$), while in summer anomalies are between 50 and 100 W m$^{-2}$ lower than the basin mean which is positive (from +126.6 W m$^{-2}$ in May to +15.3 W m$^{-2}$ in September). Two hypoth-

![Fig. 14. Mediterranean climatology of the Heat Storage Rate HSR.](image-url)
eses may explain this pattern. First, the proximity of this sub-basin to the Gibraltar Strait and thus to the inflow of warmer or colder water coming from the Atlantic Ocean, may explain why seasonal variations of HSR are reduced compared to the rest of the Mediterranean Sea. The second hypothesis may be due to more net lateral heat flux from this region to adjacent regions, preventing heat from being store locally.

The Balearic Sea

High HSR values can be seen along the 40° parallel in the WMED between the Spain mainland and Sardinia, and through the Balearic Islands, from June to August (Fig. 14), with larger HSR values (up to 260 W m⁻²) than in the rest of the WMED. These high values for HSR are also found by looking $\text{HSR}_{\text{ano}}$ (Fig. 16) which presents positive values (between −50 and −100 W m⁻²). These positive anomalies indicate that this region is gaining heat quickly than the rest of the basin. At that time of the year, this might be due to a more pronounced heat convergence in the upper-ocean.

This region has the particularity to be a place where a strong thermal front in winter separates warm Atlantic surface water from the South to cold Mediterranean surface water from the North (Lacombe and Tchernia, 1972). The regional circulation...
around the Balearic Islands is linked to the Northern Current which carries down surface water from the Gulf of Lions along the continental slope of the Iberian Peninsula into the Balearic subbasin (Font et al., 1988). In the South, the cyclonic recirculation of anticyclonic eddies in the Algerian Basin carries Atlantic Water from the Algerian boundary Current to the South of the Balearic Islands (Taupier-Letage and Millot, 1988; Puillat et al., 2002).

Two hypotheses may explain the persistence of the positive HSR anomaly during the summer around the Balearic Islands. First this region is under the influence of the Western Mediterranean Intermediate Water (WIW, previously named as Winter Intermediate Water), this mode water is formed in winter in the whole northwestern Mediterranean Sea and spreads southwards following the general circulation (Millot, 1999; Fuda et al., 2000). The accumulation of these cold and fresh lenses of WIW in the Balearic Sea may explain why HSR anomalies are lower in that region at the end of the winter and during spring. Then the persistence of the high HSR anomaly in summer may be explain by the horizontal heat fluxes from relative warm surrounding water to these cold water.

The other hypothesis may have its source in the dramatic and frequent changes of the circulation around the Balearic Islands in summer (Monserrat et al., 2008). The Northern Current may be blocked when reaching the Ibiza Channel by anticyclonic chan-
nel-size eddies and then recirculates with a cyclonic way in the Balearic Sea without significant transport of water through the Ibiza Channel (Castellón et al., 1990). These channel-size eddies are composed of cold and relative fresh water (Pinot et al., 2002), corresponding to WIW characteristics. In the North of the Balearic Islands, the recirculation of the Northern Current joins the eastward Balearic Current, while in the South, the flow due to recirculation of eddies into the Algerian Basin is westward (Testor et al., 2005). So, this summer anticyclonic circulation around the Balearic Islands might keep isolated relative cold subsurface water (WIW) from the surrounding warm water and induced a horizontal heat transfer. These two hypothesis may explain why we have positive HSR anomalies in summer around the Balearic Islands.

Cyclonic and anticyclonic gyres of the Cretan Arc

In the EMED, strong negative HSR anomalies from November to February and positive anomalies from June to September are located in places of known anticyclonic gyres (the PG and the IG, discussed Section ‘Seasonal cycle of the mixed layer, the integration depth \( h \) and the depth-averaged temperature \( T_a \)). Unlike anticyclonic gyres, cyclonic gyres (the WCG and the RG, Section ‘Seasonal cycle of the mixed layer, the integration depth \( h \) and the depth-averaged temperature \( T_a \)) present positive HSR anomalies from December to February and negative anomalies in July and August.

The two anticyclonic gyres have a more permanent signature on the seasonal climatology of HSR (particularly the IG) than cyclonic gyres in winter and summer. This may be explained by the structure of gyres: in anticyclonic gyres the thermocline is pushed down, unlike cyclonic ones where the thermocline is pushed up. So, anticyclonic gyres tend to expand the layer that exchanges heat with the atmosphere. The “warm core” structure of anticyclonic gyres is responsible for temperature difference between the gyre core and the relative colder surrounding waters, which may induce a heat transport, particularly in winter when this horizontal temperature gradient is more pronounced.

Conclusions

We present here a new climatology of the mixed layer depth and temperature, and of the thermocline depth on a 1° × 1° grid, based on recent data collected between 1969 and 2012 containing more than 140,000 profiles. The depth and the temperature of the mixed layer and the thermocline, together with the thermocline slope, revealed well known Mediterranean circulation features. This new mixed layer climatology is calculated using a \( \Delta T = 0.1 \, ^\circ\text{C} \) criterion that is more in agreement with an isopycnal definition of the MLD, compared to the \( \Delta T = 0.2 \, ^\circ\text{C} \) criterion used in previous studies. The 8 additional years of measurements relative to D’Orentzio et al. (2005) make this climatology, the most comprehensive and up to date climatology of the MLD for the Mediterranean. In addition our climatology of the seasonal thermocline slope gives for the first time a description of the seasonal variations of the upper-ocean temperature stratification for the Mediterranean Sea.

This work is the first quantification of heat storage rate in the upper-ocean for the whole Mediterranean based only on in situ oceanographic data, thereby providing a new benchmark in particular for the development of ocean models. The annual mean value of the basin-averaged HSR calculated from ocean observations is 2.0 ± 4 W m⁻². The basin-averaged HSR presents a clear seasonal signal with minimal and maximal values being about -169.3 ± 4.7 W m⁻² in December and +125.2 ± 3.9 W m⁻² in June respectively. Future comparisons between this climatology of HSR and other heat fluxes climatologies may be useful to test heat fluxes parametrization and thus improve the accuracy of the models, particularly in the context of long term climate simulations.

This work also highlights the implication of known Mediterranean circulation patterns (like the anticyclonic and cyclonic gyres in the Eastern Mediterranean, or the circulation around the Balearic Islands) in the spatial and temporal variability of the heat storage rate. Local modulations of HSR can be due to different forcings (surface net heat flux, horizontal heat advection, and entrainment mixing at the interface depth) that cannot be calculated with our dataset. So future works would also be to use complementary datasets, such as other climatologies deduced from surface geostrophic currents estimations (Rio et al., 2007; Poulain et al., 2012), or from surface flux datasets (OAFlux, HOAPS, NOC) to estimations other term in the heat budget equation, like the horizontal heat advection term, or the NHF term.

Finally, the limitations of our studies (analyses made only for the upper-ocean, period covered, uncertainties) stress the need of sustained observing systems (like repeated cruises, Argo profiling floats, mooring lines and gliders), if we want to better estimate the heat storage rate in the deep ocean, and the heat fluxes in key regions like the Gibraltar Strait, the Alboran Sea, the Deep Water Formation zones, the anticyclonic and cyclonic gyres in the Ionian and Levantine Basin. This sustained effort of observation will also allow us to distinguish and/or confirm any long-term trend in temperature and salinity in the ocean.

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Appendix A. Supplementary material

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.pocean.2014.11.004.

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