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Climatological variations of total alkalinity and total inorganic carbon in the Mediterranean Sea surface waters

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Abstract

A compilation of several cruises data from 1998 to 2013 was used to derive polynomial fits that estimate total alkalinity (A_T) and total inorganic carbon (C_T) from measurements of salinity and temperature in the Mediterranean Sea surface waters. The optimal equations were chosen based on the 10-fold cross validation results and revealed that a second and third order polynomials fit the A_T and C_T data respectively. The A_T surface fit showed an improved root mean square error (RMSE) of ±10.6 µmol kg^{-1}. Furthermore we present the first annual mean C_T parameterization for the Mediterranean Sea surface waters with a RMSE of ±14.3 µmol kg^{-1}. Excluding the marginal seas of the Adriatic and the Aegean, these equations can be used to estimate A_T and C_T in case of the lack of measurements. The seven years averages (2005–2012) mapped using the quarter degree climatologies of the World Ocean Atlas 2013 showed that in surface waters A_T and C_T have similar patterns with an increasing eastward gradient. The surface variability is influenced by the inflow of cold Atlantic waters through the Strait of Gibraltar and by the oligotrophic and thermohaline gradient that characterize the Mediterranean Sea. The summer-winter seasonality was also mapped and showed different patterns for A_T and C_T. During the winter, the A_T and C_T concentrations were higher in the western than in the eastern basin, primarily due to the deepening of the mixed layer and upwelling of dense waters. The opposite was observed in the summer where the eastern basin was marked by higher A_T and C_T concentrations than in winter. The strong evaporation that takes place in this season along with the ultra-oligotrophy of the eastern basin determines the increase of both A_T and C_T concentrations.

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1 Introduction

The role of the ocean in mitigating climate change is well known as it absorbs about $2\,Pg\,C\,yr^{-1}$ of anthropogenic CO_2 (Wanninkhof et al., 2013). Worldwide measurements of surface seawater CO_2 properties are being conducted as they are important for advancing our understanding of the carbon cycle and the underlying processes controlling it. For instance, the buffer capacity of the CO_2 system varies with temperature, the distribution of total inorganic carbon and total alkalinity (Omta et al., 2011).

Our understanding of the open-ocean CO_2 dynamics has drastically improved over the years (Rödenbeck et al., 2013; Sabine et al., 2004; Takahashi et al., 2009; Watson and Orr, 2003). However our understanding of marginal seas such as the Mediterranean remains poor due to the limited measurements combined with the enhanced complexity of the land-ocean interactions. In the Mediterranean Sea, available measurements of the carbonate system are still scarce and only available in specific regions such as the Alboran Sea (Copin-Montégut, 1993), the Gibraltar Strait (Santana-Casiano et al., 2002), the Dyfamed time-series in the Ligurian Sea (Bégovic and Copin-Montégut, 2002; Copin-Montégut and Bégovic, 2002; Touratier and Goyet, 2009) and the Otranto Strait (Krasakopoulou et al., 2011). Large geographical repartition of CO_2 data are often confined to cruises with a short sampling period (Álvarez et al., 2014; Goyet et al., 2015; Rivaro et al., 2010; Schneider et al., 2007; Touratier et al., 2012). Numerical models have provided some insights of the carbon dynamics in the Mediterranean Sea (Cossarini et al., 2015; D'Ortenzio et al., 2008; Louanchi et al., 2009), but it remains important to constrain the system from in situ measurements to validate their output.

The scarcity of the CO_2 system measurements in the Mediterranean Sea make it difficult to constrain the CO_2 uptake in this landlocked area and also limits our understanding of the magnitude and mechanisms driving the natural variability on the ocean carbon system (Touratier and Goyet, 2009). Empirical modeling has been successfully used to study the marine carbon biogeochemical processes such as the estimation of

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biologically produced O_2 in the mixed layer (Keeling et al., 1993), estimation of global inventories of anthropogenic CO_2 (Sabine et al., 2004) and estimation of the $CaCO_3$ cycle (Koeve et al., 2014). Empirical algorithms were also used to relate limited A_T and C_T measurements to more widely available physical parameters such as salinity and temperature (Bakker et al., 1999; Ishii et al., 2004; Lee et al., 2006). The A_T and C_T fields can then be used to calculate $p CO_2$ fields and thus predict the CO_2 fluxes across the air–sea interface (McNeil et al., 2007).

Previous empirical approaches to constrain A_T in the Mediterranean Sea have only covered selected cruises (Schneider et al., 2007; Touratier and Goyet, 2009) or local areas such as the Dyfamed time-series station or the Strait of Gibraltar (Copin-Montégut, 1993; Santana-Casiano et al., 2002). As for C_T, empirical models have only been applied to data below the mixed layer depth (MLD) following the equation of Goyet and Davis (1997) at the Dyfamed time series station (Touratier and Goyet, 2009) or using the composite dataset from Meteor 51/2 and Dyfamed (Touratier and Goyet, 2011). Also Lovato and Vichi (2015) proposed an optimal multiple linear model for C_T using the Meteor 84/3 full water colum data. To the best of our knowledge the reconstruction of C_T in surface waters has not been yet performed in the Mediterranean Sea. This is probably due to the lack of measurements available for previous studies to capture the more complex interplay of biological, physical and solubility processes that drive the C_T variability in surface waters.

In this study we have compiled CO_2 system measurements from 14 cruises between 1995 and 2013, that allowed us to constrain an improved and new empirical algorithms for A_T and C_T in the Mediterranean Sea surface waters. We also evaluated the spatial and seasonal variability of the carbon system in the Mediterranean Sea surface waters, by mapping the 2005–2012 annual and seasonal averages of surface A_T and C_T using the quarter degree climatologies of salinity and temperature from the World Ocean Atlas 2013 (WOA13).

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2 Methods

2.1 Surface A_T and C_T data in the Mediterranean Sea

Between 1998 and 2013, there have been multiple research cruises sampling the seawater properties throughout the Mediterranean Sea. This includes parameters of the carbonate system more specifically A_T, pH and C_T and physico-chemical properties of in situ salinity, and temperature. However, the number of the nutrients concentrations was very limited. In this study we have compiled surface water samples between 0 and 10 m depth, totaling 490 and 426 measurements for A_T and C_T respectively (Table 1).

2.2 Polynomial model for fitting A_T and C_T data

Two polynomial equations for fitting A_T or C_T from salinity (S) alone or combined with sea surface temperature (T) in the surface waters (0–10 m) of the Mediterranean Sea were chosen from the results of the 10-fold cross validation method (Breiman, 1996; Stone, 1974). This type of analysis was previously performed by Lee et al. (2006) for global relationships of A_T with salinity and temperature. This model validation technique is performed by retaining a single subsample used for testing and training the algorithm on the 9 remaining subsamples. The cross validation process is then repeated 10 times. The best fit is chosen by computing the residuals from each regression model, and computing independently the performance of the selected optimal polynomial on the remaining subsets. The analysis was applied for polynomials of order 1 to 3, and the optimal equation was chosen based on the lowest Root Mean Square Error (RMSE) and the highest coefficient of determination (r^2).

To ensure the same spatial and temporal distribution of A_T and C_T polynomial fits we only selected stations where A_T and C_T were simultaneously measured (Table 1; Fig. 1). To validate the general use of the proposed parameterizations we tested the algorithms with measurements which are not included in the fits (Testing dataset). Hence for the A_T, 375 and 115 data points are used for the training and testing datasets re-

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spectively. For the C_T the training dataset is formed from 381 data points and the validation dataset is the same as the testing subset of the 10th fold (45 data points).

2.3 Climatological and seasonal mapping of A_T and C_T

The climatological and seasonal averages of salinity (Zweng et al., 2013) and temperature (Locarnini et al., 2013) in $1/4 \times 1/4°$ grid cells were downloaded from the World Ocean Atlas 2013 (WOA13). The seven years averages (2005–2012) and the summer-winter seasonality of A_T and C_T fields were mapped at 5 m depth by applying the respective derived algorithms in their appropriate ranges of S and T.

3 Results and discussion

3.1 Fitting A_T in the Mediterranean Sea surface waters

In the surface ocean the A_T variability is controlled by freshwater addition or the effect of evaporation, and salinity contributes to more than 80 % of the A_T variability (Millero et al., 1998). In the Mediterranean Sea, several studies have shown that the relationship between A_T and S is linear (Copin-Montégut, 1993; Copin-Montégut and Bégovic, 2002; Hassoun et al., 2015b; Rivaro et al., 2010; Schneider et al., 2007). In other studies, the sea surface temperature (T) has been included as an additional proxy for changes in surface water A_T related to convective mixing (Lee et al., 2006; Touratier and Goyet, 2011).

The results of the 10-fold cross validation analysis revealed that the optimal model for A_T is a second order polynomial in which A_T is fitted to both S and T (Table 2, Eq. 1). A linear relationship between A_T and S yields a higher RMSE (14.5 µmol kg^{-1}) and a lower r^2 (0.91) than Eq. (1). In a semi-enclosed basin such as the Mediterranean Sea, the insulation and high evaporation as well as the input of rivers and little precipitation leads to a negative freshwater balance (Rohling et al., 2009). The resulting

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anti-estuarine thermohaline circulation could explain the contribution of temperature to the A_T variability (Touratier and Goyet, 2011).

The residuals of training dataset used to generate the second order polynomial fit for A_T are presented in Fig. 2a. Most of the A_T residuals (340 over 375) were within a range of ±15 µmol kg^{-1} (1σ). However 35 residuals were high up to ±30 µmol kg^{-1} (1σ). Applying the A_T algorithm to the testing dataset (Fig. 2b), yields a mean residual of 0.91 ± 10.30 µmol kg^{-1} (1σ), and only 6 data points have residuals higher than ±15 µmol kg^{-1} (1σ).

The comparison of the RMSE as reported by other studies with that of Eq. (1) does not indicate if the parameterization developed here has advanced or not on previous attempts in the Mediterranean Sea. In that order, we independently applied each of the previous equations on the same training dataset used to develop Eq. (1) and then computed the RMSE and r^2 for every one (Table 3). The results show that Eq. (1) has a lower RMSE and a higher r^2 than all of the parameterizations presented in Table 3. For instance, the global relationship of Lee et al. (2006) applied to the dataset of this study yields a RMSE as high as ±40.50 µmol kg^{-1}. The RMSE of other studies developed strictly in the Mediterranean Sea varied from ±13.81 to ±26.11 µmol kg^{-1} using the equations of Touratier and Goyet (2011) and Schneider et al. (2007) respectively.

By applying directly the previous parameterizations to our training dataset, the calculated RMSE are significantly higher than the ones reported in their respective studies. For instance the reported RMSE in Lee et al. (2006) for sub-tropical oceanic regions is ±8 µmol kg^{-1} and that of Schneider et al. (2007) for the Meteor 51/2 cruise is ±4.2 µmol kg^{-1}. This shows that previous models were constrained by their spatial coverage, time span and used datasets. In fact the previous equations were calculated in local areas such as the Alboran Sea (Copin-Montégut, 1993), the Strait of Gibraltar (Santana-Casiano et al., 2002) or the Dyfamed Site (Copin-Montégut and Bégovic, 2002; Touratier and Goyet, 2009). On a large scale, equations were applied using limited datasets such as the Meteor 51/2 cruise in October–November 2001 (Schneider

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et al., 2007), the Transmed cruise in May–June 2007 (Rivaro et al., 2010) or the Meteor 51/2 and the Dyfamed time series station (Touratier and Goyet, 2011).

The proposed algorithm including surface data from multiple cruises, and on a large time span, presents a more global relationship to estimate A_T from S and T than the previously presented equations (Table 3). In Eq. (1), T and S contribute to 96 % of the A_T variability and the RMSE of $\pm 10.6\,\mu mol\,kg^{-1}$ presents a significant improvement of the spatial and temporal estimations of A_T in the Mediterranean Sea surface waters.

3.2 Fitting C_T in the Mediterranean Sea surface waters

The surface C_T concentrations are influenced by lateral and vertical mixing, photosynthesis, oxidation of organic matter and changes in temperature and salinity (Poisson et al., 1993; Takahashi et al., 1993). All these processes are directly or indirectly correlated with sea-surface temperature (Lee et al., 2000). Hence, the parameterization of C_T in surface waters includes both physical (S and T) and/or biological parameters (Bakker et al., 1999; Bates et al., 2006; Koffi et al., 2010; Lee et al., 2000; Sasse et al., 2013).

The results of the 10-fold cross validation analysis showed that a first order polynome fits C_T to S and T with an RMSE of $16.25\,\mu mol\,kg^{-1}$ and $r^2 = 0.87$. These values are comparable to the RMSE and r^2 found by previous empirical approaches applied in the Eastern Atlantic (Bakker et al., 1999; Koffi et al., 2010). However we found that a third order polynome improved the RMSE and r^2 of the equation compared to the first order fit (Table 4, Eq. 2). Hence we will retain the large dataset used to develop Eq. (2), where temperature and salinity contribute to 90 % of the C_T variability encountered in the Mediterranean Sea surface waters. The remaining 10 % could be attributed to the biological and air–sea exchange contributions to the C_T variability.

The C_T parameterization developed in this study (Table 4; Eq. 2) showed a higher uncertainty than that of A_T regarding both RMSE and r^2. In fact, the interpolation of C_T in the mixed layer adds a high uncertainty due to the seasonal variability. Also in surface

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waters the C_T are directly affected by air–sea exchange, and their concentrations will increase in response to the oceanic uptake of anthropogenic CO_2.

Previous models accounted for the anthropogenic biases in the C_T measurements by calculating the C_T rate of increase (Bates, 2007; Lee et al., 2000; Sasse et al., 2013; Takahashi et al., 2014). However in a study, Lee et al. (2000) also did not correct the C_T concentrations for regions above 30° latitude such as the Mediterranean Sea. In the following we will assess the importance of accounting or not for anthropogenic biases in the C_T measurements. In that order we dowloaded the monthly atmospheric pCO_2 concentrations measured from 1999 to 2013 at the Lampedusa Island Station (Italy) from the World Data Center for Green House Gases (http://ds.data.jma.go.jp/gmd/wdcgg/). Following the method described by Sasse et al. (2013), we corrected the C_T measurements to the nominal year of 2005 and applied the same 10-fold cross validation analysis using data with and without anthropogenic C_T corrections. We found that the RMSE of the C_T model trained using measurements with anthropogenic corrections is 13.9 $\mu mol\,kg^{-1}$, which is not significantly different from the model trained using measurements without anthropogenic corrections (Eq. (2); RMSE = 14.3 $\mu mol\,kg^{-1}$).

The yearly increase of C_T concentrations is difficult to assess due to the wide spatial distribution of the training dataset used to generate Eq. (2). Hence we will refer to the monthly C_T concentrations measured between 1998 and 2013 at the Dyfamed time-series station. We found that the rate of increase in C_T concentrations at the Dyfamed site was 0.99 $\mu mol\,kg^{-1}\,yr^{-1}$ (Fig. 3), which is consistent with the anthropogenic C_T correction rate used in the previous studies of Lee et al. (2000), Bates (2007) and Sasse et al. (2013).

The rate of increase in C_T concentrations of 0.99 $\mu mol\,kg^{-1}\,yr^{-1}$ as well as the RMSE difference of ±0.4 $\mu mol\,kg^{-1}$ between the two models (with or without anthropogenic corrections) are both smaller than the uncertainty of the C_T measurements of at least ±2 $\mu mol\,kg^{-1}$ (Millero, 2007). A recent study also showed that the uncertainty of the C_T measurements can be significantly higher than ±2 $\mu mol\,kg^{-1}$, as most laboratories

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reported values of C_T for the measures that were within a range of ±10 µmol kg^{-1} of the stated value (Bockmon and Dickson, 2015).

Between 1998 and 2013, the C_T concentrations measured at the Dyfamed time-series station showed a slightly increasing trend ($r^2 = 0.05$). The increase in C_T concentrations in response to elevated atmospheric CO_2, was masked by the high seasonal variations. For example, during the year 1999 the variation in C_T concentrations reached as high as 100 µmol kg^{-1} (Fig. 4a). Also there is a clear seasonal cycle of surface waters C_T in the Dyfamed station (Fig. 4b). In the summer, the C_T starts to increase gradually to reach a maximum of 2320 µmol kg^{-1} during the winter season, after which a gradual decrease is observed to reach a minimum of 2200 µmol kg^{-1} by the end of spring. The seasonal cycle can be explained by the counter effect of temperature and biology on the C_T variations. During the spring, the increasing effect of warming of pCO_2 is counteracted by the photosynthetic activity that lowers the C_T. During the winter, the decreasing effect of cooling on pCO_2 is counteracted by the upwelling of deep waters rich in C_T (Hood and Merlivat, 2001; Takahashi et al., 1993). This shows that the C_T concentrations in surface waters were more affected by the seasonal variations than by anthropogenic forcing.

Considering the small differences in RMSE obtained by the two models, the uncertainties in the C_T measurements and the clear signal of the seasonal variations; no corrections were made to account for the rising atmospheric CO_2 concentrations. In regions above 30° latitude such as the Mediterranean Sea, the corrections of C_T are small considering that the outcropping of deep isopycnal surfaces dilutes the anthropogenic CO_2 throughout the water column (Lee et al., 2000). Also the dynamic overturning circulation in the Mediterranean Sea plays an effective role in absorbing the anthropogenic CO_2 and transports it from the surface to the interior of the basins (Hassoun et al., 2015a; Lee et al., 2011).

The residuals of the dataset used to generate the third order polynomial fit for C_T are presented in Fig. 5a. Most of the C_T residuals (330 over 381) were within a range of ±18 µmol kg^{-1} (1σ). In contrast only few residuals (12 over 381) reached up to

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±50 µmol kg^{-1} (1σ). Applying the C_T algorithm to the testing dataset (Fig. 5b), yields a mean residual of 1.48 ± 19.80 µmol kg^{-1} (1σ) which is close to the uncertainties of our C_T relationship. The high residuals observed in this study are consistent with the results of the optimal multiple linear regression performed by Lovato and Vichi (2015), where the largest discrepancies between observations and reconstructed data were detected at the surface layer with RMSE higher than ±20 µmol kg^{-1}.

Considering the high uncertainties of the C_T measurements, the seasonal variations and the anthropogenic forcing; Eq. (2) presents the first parametrization for C_T in the Mediterranean Sea surface waters, with an RMSE of ±14.3 µmol kg^{-1} (1σ) and a r^2 = 0.90 (Table 4, Eq. 2).

3.3 Spatial and seasonal variability of A_T and C_T in surface waters

The ranges of the 2005–2012 average annual climatologies of the World Ocean Atlas 2013 (WOA13) are from 35.91 to 39.50 for S and from 16.50 to 23.57 °C for T (Locarnini et al., 2013; Zweng et al., 2013). However a wider range is observed for the seasonal climatologies, especially during the winter season where T ranges from 9.05 to 18.43 °C. The estimations of A_T and C_T in surface waters from Eqs. (1) and (2) respectively are only applicable in the appropriate ranges of $T > 13$ °C and $36.3 < S < 39.65$. Hence the surface waters A_T and C_T concentrations were mapped only where T and S were within the validity range of Eqs. (1) and (2) respectively (Tables 2 and 4). Excluding few near-shore areas and the influence of cold Atlantic Waters in winter, the ranges in which Eqs. (1) and (2) can be applied are within those of the climatological products of T and S of the WOA13.

The mapped climatologies for 2005–2012 at 5 m depth showed a strong increase in the eastward gradient for both A_T and C_T with the highest concentrations always found in the Eastern Mediterranean (Fig. 6). The minimum values of 2400 µmol kg^{-1} for A_T and 2100 µmol kg^{-1} for C_T are found near the Strait of Gibraltar and the maximum values of 2650 and 2300 $\mu mol\,kg^{-1}$ are found in the Levantine and Aegean sub-basin for A_T and C_T respectively.

The A_T parameterization of this study detects a clear signature of the alkaline waters entering through the Strait of Gibraltar that remains traceable to the Strait of Sicily as also shown by Cossarini et al. (2015). In the eastern basin the positive balance between evaporation and precipitation contributes to the increasing surface A_T. Local effects from some coastal areas such as the Gulf of Gabes and riverine inputs from the Rhone and Po River are also detected.

Our results for surface A_T have a similar spatial pattern and range as the annual climatology of Cossarini et al. (2015) which simulates surface A_T values from 2400 to 2700 $\mu mol\,kg^{-1}$. The main difference is marked in the upper ends of the Adriatic and Aegean sub-basins where our algorithm predicts A_T values around 2400–2500 $\mu mol\,kg^{-1}$, whereas the analysis of Cossarini et al. (2015) yields a maximum of 2700 $\mu mol\,kg^{-1}$ in these regions. Regressions in regions of river input indicate a negative correlation between alkalinity and salinity (Luchetta et al., 2010); hence Eastern marginal seas are expected to have high A_T due to the freshwater inputs (Cantoni et al., 2012; Souvermezoglou et al., 2010). This shows the sensitivity of our algorithms to temperature and salinity especially in areas that are more influenced by continental inputs such as the Po River and inputs of the Dardanelle in the northern Adriatic and northern Aegean respectively (Fig. 6a).

At the surface, the basin wide distributions of C_T are affected by physical processes and their gradient is similar to that of A_T (Fig. 6b). The lowest C_T concentrations are found in the zone of the inflowing Atlantic water and increases toward the East in part due to evaporation as also shown by Schneider et al. (2010). Our results for surface C_T have a similar range as the optimal linear regression performed by Lovato and Vichi (2015) which estimates surface C_T values from 2180 to 2260 $\mu mol\,kg^{-1}$. Moreover, the results show that the Mediterranean Sea is characterized by C_T values that are much higher (100–200 $\mu mol\,kg^{-1}$ higher) than those observed in the Atlantic Ocean at the same latitude (Key et al., 2004).

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As a consequence of uptake of atmospheric CO_2, the eastward pCO_2 increase is parallel to that of C_T (D'Ortenzio et al., 2008). For example the Ionian and Levantine sub-basin are characterized by a pCO_2 as high as 470 µatm (Bégovic, 2001), whereas the Algerian sub-basin is characterized by a much lower pCO_2 of 310 µatm (Calleja et al., 2013). The high pCO_2 and C_T encountered in the eastern basin make it a permanent source of atmospheric CO_2 (D'Ortenzio et al., 2008; Taillandier et al., 2012). Overall the western basin has a lower surface C_T content than the eastern basin which could be explained by the eastward decrease of the Mediterranean Sea trophic gradient (Lazzari et al., 2012). The higher rate of inorganic carbon consumption by photosynthesis in the western basin can lead to the depletion of C_T in the surface waters, whereas the ultra-oligotrophic state in the eastern basin can lead to a high remineralization rate that consumes oxygen and enriches surface waters with C_T (Moutin and Raimbault, 2002).

The magnitude of the seasonal variability between summer and winter for A_T and C_T is shown in Fig. 7. Unlike the seven years averages, the seasonal climatological variations (2005–2012) of A_T have different spatial patterns than those of C_T. Overall the summer-winter time differences for A_T have an increasing eastward gradient (Fig. 7a). The largest magnitudes are marked in the Alboran sub-basin with differences reaching up to $-80\,\mu mol\,kg^{-1}$; the negative difference implies that during the winter inflowing surface Atlantic water has higher A_T concentrations than in summer. Higher winter than summer time A_T concentrations are also observed in the Balearic, Ligurian and the South-western Ionian sub-basins but with a less pronounced seasonality ($\sim -30\,\mu mol\,kg^{-1}$). For these three sub-basins, the C_T has a higher summer-winter magnitude than A_T ($\sim -70\,\mu mol\,kg^{-1}$). The winter cooling of surface waters increases their density and promotes a mixing with deeper water. Thus, the enrichment in winter time likely reflects the upwelling of deep waters that have accumulated A_T and C_T from the remineralization of organic matter, respiration and the dissolution of $CaCO_3$. The seasonality is more pronounced for C_T, which likely reflects the stronger response of C_T to biological processes than A_T (Takahashi et al., 1993).

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In the Algerian sub-basin and along the coasts of Tunisia and Libya, the seasonality is inversed with higher A_T and C_T concentrations prevailing in the summer. The African coast is an area of coastal downwelling during the winter season. However, during summer the coastal upwelling appears in response to turning of the wind near the coast toward the West (Bakun and Agostini, 2001). In general, the magnitude of the A_T seasonal variability is higher in summer than in winter for the eastern basin and more particularly in the Ionian and Levantine sub-basins. During this season strong evaporation takes place and induce an increase of A_T concentrations (Schneider et al., 2007). In the eastern basin, the high evaporation during the summer has a smaller effect on the C_T, and magnitudes reach their maxima in the Levantine sub-basin ($\sim +20\,\mu\text{mol}\,\text{kg}^{-1}$). During winter time the western basin and South East of Sicily appear to be dominated by higher C_T concentrations than the rest of the eastern basin, where the summer C_T concentrations are prevailing (Fig. 7b). During winter the high C_T concentrations that coincide with low SST in the western basin, could result from the deepening of the mixed layer and could be enhanced by the upwelling associated with the Tramontane-Mistral winds that blow from the southern of France and reach the Balearic Islands and the Spanish coast.

4 Summary

The A_T and C_T algorithms are derived from a compilation of 490 and 426 quality controlled surface measurements respectively, collected between 1999 and 2013 in the Mediterranean Sea. A second order polynomial relating A_T to both S and T yielded a lower RMSE ($\pm 10.4\,\mu\text{mol}\,\text{kg}^{-1}$) and a higher r^2 (0.96) than a linear fit deriving A_T from S alone. This confirmed the important contribution of temperature to the A_T variability. Hence, temperature should be included in future algorithms to help better constrain the surface A_T variations. The proposed second order polynomial had a lower RMSE than other studies when we applied their respective algorithms to the same training dataset.

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In this study we propose an improved and global relationship to estimate the A_T spatial and temporal variations in the Mediterranean Sea surface waters.

The C_T parameterization is a first attempt to estimate the surface variations in the Mediterranean Sea. A third order polynomial is suggested to fit the C_T to T and S with a RMSE of $\pm 14.3\,\mu mol\,kg^{-1}$. The biological contributions to the C_T variations were less pronounced than the physical processes. The contributions of to the physical processes and biology to the C_T variability were 90 and 10 % respectively. In terms of anthropogenic forcing, the C_T rate of increase of $0.99\,\mu mol\,kg^{-1}\,yr^{-1}$ was significantly lower than the uncertainty of the measurements than can reach $\pm 10\,\mu mol\,kg^{-1}$ between different laboratories. Moreover the C_T concentrations were more affected by the seasonal variations than the increase of atmospheric CO_2.

We propose to use Eqs. (1) and (2) for the estimation of surface A_T and C_T in the Mediterranean Sea when salinity and temperature of the area are available and are in the appropriate ranges of the equations. However in the Eastern marginal seas especially the northern Adriatic and northern Aegean there is a need to develop a more specific equation that minimizes the errors in these areas. Hence, it is important to enrich the existing dataset by an extensive sampling program such as the Med-SHIP initiative (CIESM, 2012) in order to improve the modeling of the carbonate system over the whole Mediterranean Sea.

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Table 1. List of available carbonate system datasets for the Mediterranean Sea.

Dataset	Period	Carbonate system parameters	Reference
Prosope	Sep–Oct 1999	A_T and pH	Bégovic and Copin (2013)
Meteor 51/2	Oct–Nov 2001	A_T and C_T	Schneider and Roether (2007)
Meteor 84/3	Apr 2004	A_T, C_T and pH	Tanhua et al. (2012)
Carbogib 2–6	2005–2006	A_T and pH	Huertas (2007a–e)
Gift 1–3	2005–2006	A_T and pH	Huertas (2007f–h)
Transmed	Jun 2007	A_T and pH	Rivaro et al. (2010)
Sesame IT-4	Mar–Apr 2008	A_T and C_T	SeaDataNet
Boum	Jun–Jul 2008	A_T and C_T	Touratier et al. (2012)
Pacific–Celebes	2007–2009	A_T and C_T	Hydes et al. (2012)
Moose-GE	May 2010	A_T and C_T	SeaDataNet
Hesperides	May 2013	A_T	Perez et al. (2013)
MedSEA	May 2013	A_T and C_T	Goyet et al. (2015)
Dyfamed time-series	1998–2013	A_T and C_T	Oceanological Observatory of Villefranche-sur-Mer

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Table 2. Second order polynomial fit to derive A_T from salinity and temperature in the Mediterranean Sea surface waters.

Polynomial fit	N	r^2	RMSE (μmol kg^{-1})
Eq. (1): $A_T = 2558.4 + 49.83(S - 38.2) - 3.89(T - 18) - 3.12(S - 38.2)^2 - 1.06(T - 18)^2$ $T > 13\,^{\circ}$C and $36.30 < S < 39.65$	375	0.96	10.6

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Table 3. Performance of the different parameterizations for the estimation of A_T applied independently to the training dataset of this study.

Region	Parameterization	RMSE (μmol kg^{-1})	r^2	Reference
Alboran Sea	$A_T = 94.85(S) - 1072.6$	±16.61	0.92	Copin-Montégut (1993)
Dyfamed site	$A_T = 93.99(S) - 1038.1$	±16.31	0.92	Copin-Montégut and Bégovic (2002)
Strait of Gibraltar	$A_T = 92.28(S) - 968.7$	±16.48	0.92	Santana-Casiano et al. (2002)
Mediterranean Sea	$A_T = 73.7(S) - 285.7$	±26.11	0.68	Schneider et al. (2007)
Dyfamed site	$A_T = 99.26(S) - 1238.4$	±18.53	0.91	Touratier and Goyet (2009)
Western Mediterranean	$A_T = 95.25(S) - 1089.3$	±16.97	0.92	Rivaro et al. (2010)
Eastern Mediterranean	$A_T = 80.04(S) - 499.8$	±14.58	0.91	
Mediterranean Sea	$A_T = 1/(6.57 \times 10^{-5} + 1.77 - 10^{-2})/S - (5.93 - -10^{-4}(\ln\theta)/\theta^2)$	±13.81	0.92	Touratier and Goyet (2011)
Global relationship (Sub-tropics)	$A_T = 2305 + 58.66(S-35) + 2.32(S-35)^2 + 1.41(T-20) + 0.04(T-20)^2$	±40.50	0.26	Lee et al. (2006)

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Table 4. Third order polynomial fit to derive C_T from salinity and temperature in the Mediterranean Sea surface waters.

Polynomial fit	N	r^2	RMSE ($\mu mol\,kg^{-1}$)
Eq. (2): $C_T = 2234 + 38.15(S - 38.2) - 14.38(T - 17.7) - 4.48(S - 38.2)^2 - 1.43(S - 38.2)(T - 17.7) + 9.62(T - 17.7)^2 - 1.10(S - 38.2)^3 + 3.53(T - 17.7)(S - 38.2)^2 + 1.47(S - 38.2)(T - 17.7)^2 - 4.61(T - 17.7)^3$ $T > 13\,^\circ C$ and $36.30 < S < 39.65$	381	0.90	14.3

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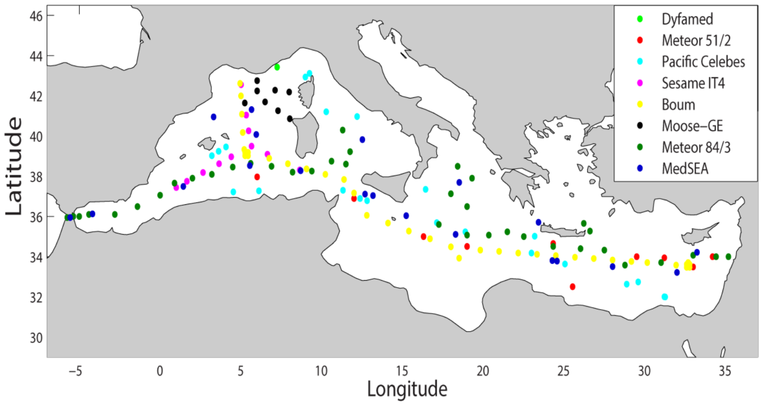

Figure 1. Spatial distribution of data points used to initiate the fits of A_T and C_T.

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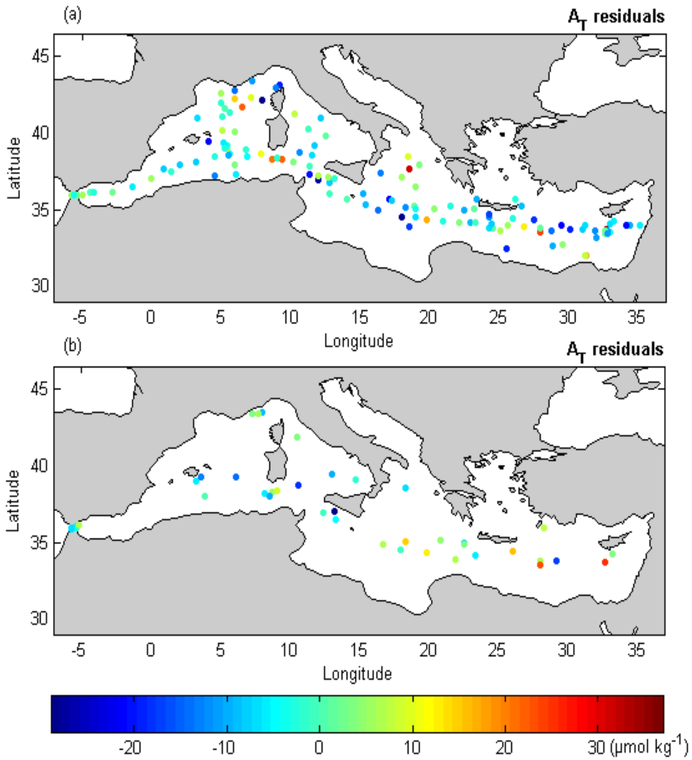

Figure 2. Map of the residuals of the A_T algorithm (Table 1; Eq. 1) applied the **(a)** training and **(b)** testing datasets.

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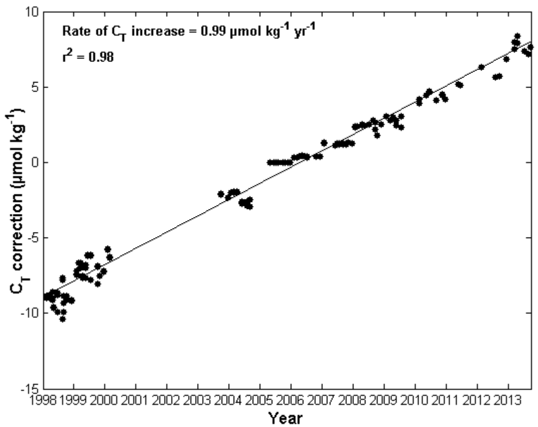

Figure 3. Rate of increase applied to correct the C_T measurements in reference to the year 2005.

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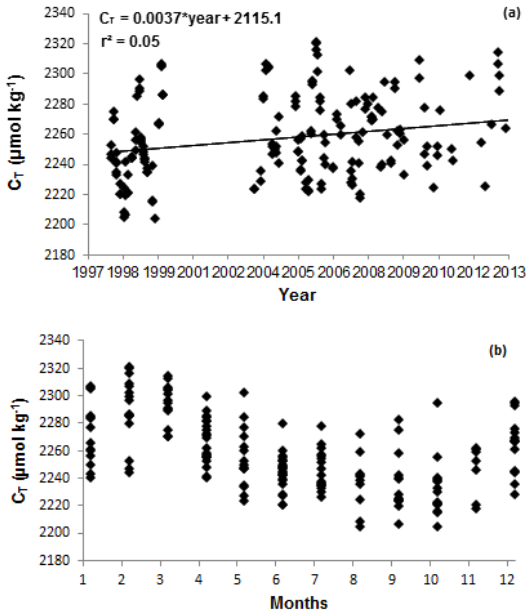

Figure 4. (a) Temporal and **(b)** seasonal variations of C_T measured at the Dyfamed time-series station between 1998 and 2013.

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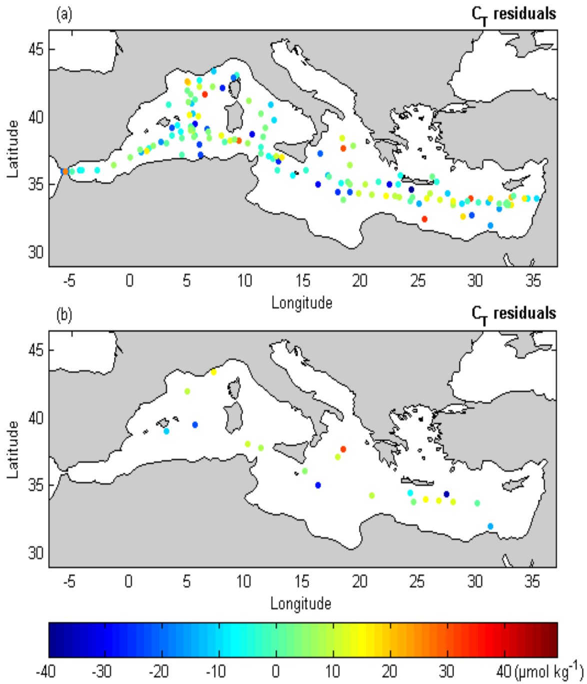

Figure 5. Comparison of the predicted C_T values from the C_T algorithm given in Table 1 – Eq. (2) with measurements which are **(a)** included or **(b)** excluded when deriving the fit.

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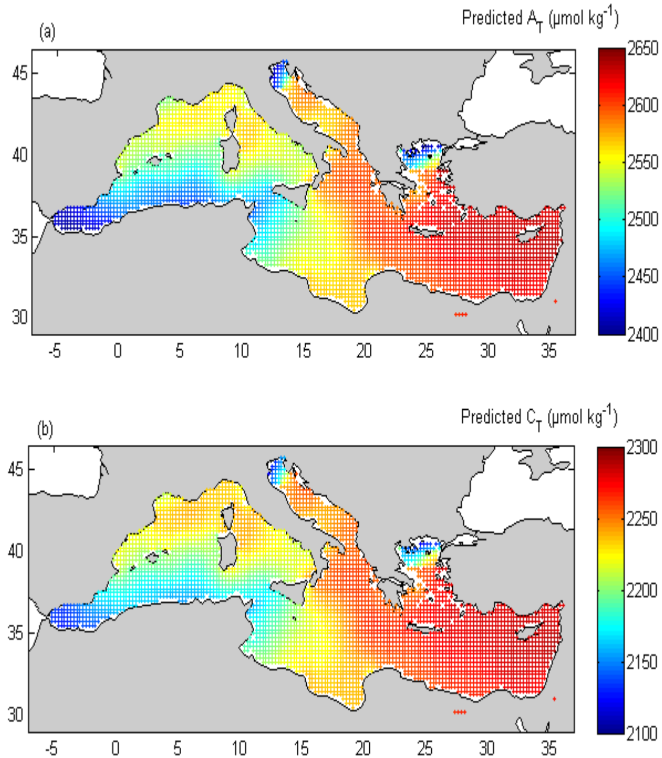

Figure 6. The seven years averages spatial variability of **(a)** surface A_T predicted from Eq. (1) and **(b)** surface C_T predicted from Eq. (2), applied to the 2005–2012 climatological fields of S and T from the WOA13.

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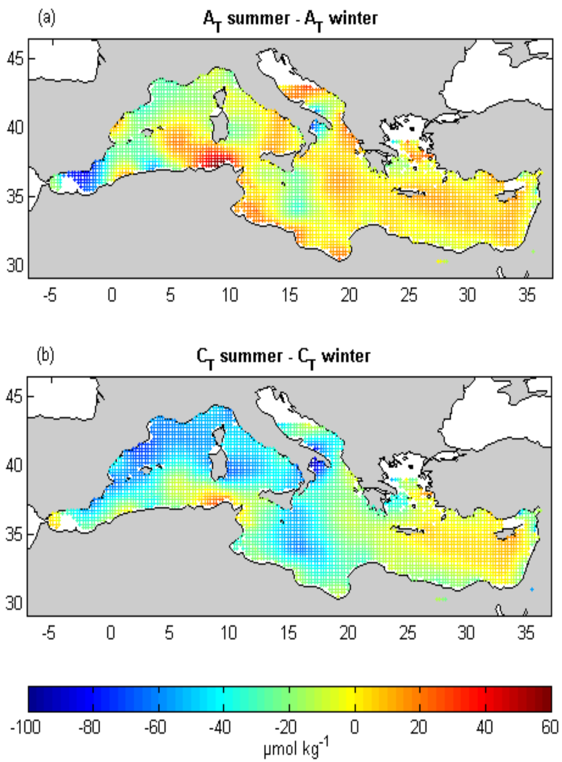

Figure 7. Distribution of the summer-winter differences of **(a)** surface A_T predicted from Eq. (1) and **(b)** surface C_T predicted from Eq. (2), applied to the 2005–2012 climatological fields of S and T from the WOA13.

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