IJEGEO

International Journal of Environment and Geoinformatics 11(3):147-155 (2024)

Research Article

Evaluation of Carbon Dioxide Partial Pressure and Air-Sea Fluxes in the Sea of Marmara

Seben Yucel*, **Nagihan Ersoy Korkmaz**, **Abdullah Aksu**, **Nuray Caglar Balkis**

Istanbul University, Marine Sciences and Management Institute, Chemical Oceanography Department, Istanbul

* Corresponding author: S. Yucel
E-mail: sebenyucel@istanbul.edu.tr

Received: 27.07.2024
Accepted 21.09.2024

How to cite: Yucel et al., (2024). Evaluation of Carbon Dioxide Partial Pressure and Air-Sea Fluxes in the Sea of Marmara, *International Journal of Environment and Geoinformatics (IJEGEO)*, 11(3): 147-155. doi. 10.30897/ijegeo.1523105

Abstract

The increasing impact of global warming has made it crucial to monitor greenhouse gases (GHG). Carbon dioxide (CO_2) is especially important due to its significant increase. Oceans and seas play a vital role in absorbing excess CO_2 from the atmosphere, but they are facing challenges such as ocean acidification and damage to coral reefs. Therefore, it's essential to thoroughly understand climate change and its human-caused effects and to take necessary precautions or improve existing measures. In this direction, in seawater, determining alkalinity is crucial for monitoring CO_2, along with other physical and chemical parameters, in order to help set carbon budgets and reduce environmental pressures. For this purpose, in this study in the Sea of Marmara, alkalinity and CO_2 measurements were conducted for the first time at stations CG2, CG3, MD26, and MD24. The CO_2 partial pressure (pCO_2) values ranged from 95 to 165 μatm at the surface of all stations, and total alkalinity values ranged from 1.075 to 1.46 mmol kg^{-1} at specified depths. Based on these measurements, flux values were calculated between −1.10 and −5.39 mmol m^2 day^{-1}, indicating that the Sea of Marmara acted as a net CO_2 sink.

Keywords: Air–sea flux, carbon dioxide, partial pressure, seawater, Sea of Marmara

Introduction

Global warming is a term used to describe the impact of human activities on the climate, specifically the burning of fossil fuels (coal, oil, and gas). These activities cause the release of significant amounts of GHG: Geenhouse gases (the most important one is CO2) into the atmosphere (Houghton, 2005). Carbon dioxide demands attention as a GHG due to its indisputable rise in concentration. To understand what controls its hence on the greenhouse effect, all processes that affect and are affected by its concentration in the atmosphere need to be taken into account. These processes involve interactions with the chemically buffered carbonate system in seawater and with vegetation, due to their vital role in photosynthesis (Keeling, 1997).

In recent years, interest in the carbonate system in the oceans has increased due to the effect of CO_2 as a GHG. Carbon can be exchanged rapidly with the atmosphere through physical processes (such as CO_2 gas dissolving into the water) and biological processes (such as the growth, death, and decay of plankton) and is transported to the deep ocean by sinking so it can be stored for a longer time (Dioxide, 1985; Lal, 2009).The regulation of seawater pH and the cycling of CO_2 between the biosphere, lithosphere, atmosphere, and ocean are both vital functions of the carbonate system and the major part of the carbon in the oceans occurs as a part of this system (Millero, 1996).

The presence of oceans is not just a factor, but a crucial determinant in shaping the behavior of climate over time. Temperature dynamics as well as CO_2 behavior are both heavily influenced by the oceans, and the chemical aspect of CO_2 behavior is particularly significant (Lindzen, 1994). Oceans play a critical role in regulating climate change and atmospheric CO_2; since the Industrial Revolution began, the ocean has absorbed an amount of carbon equivalent to approximately 40% of the anthropogenic carbon dioxide emitted by fossil fuel burning, and it is currently estimated that approximately 25% of anthropogenic CO_2 is absorbed by ocean annually (Gazioğlu et al., 2015; Takeshita et al., 2018). However, it also leads to fundamental changes in the chemistry of oceans. This is because CO_2 is also an acidic gas not only a GHG (Orr, 2011).

The pH of the ocean decreases as the amount of CO_2 dissolved in it increases, and this increase in CO_2 also leads to a decrease in the availability of carbonate (CO_2^{-3}) ions and reduces the saturation state of the major carbonate minerals that form shells (Kleypas et al., 2005). This decrease in pH and changes in fundamental chemical balances are often referred to as ocean acidification, which is also known as another CO_2 problem (Doney et al., 2009; Gazioğlu and Okutan, 2016).

In the last two decades, accurately measuring seawater CO_2 has become a priority for scientists who study how much CO_2 produced by human activities enters the ocean, its impact on ocean chemistry, and its distribution (Dickson et al., 2007). So, it is important to understand the dynamics of this system, and is also important to understand the relationship of climate change with the ocean. So measuring at least two of these four parameters, which are total alkalinity, pH, total CO_2, and partial pressure of CO_2 (Millero, 1996), the carbonate system in the oceans can be characterized (because the concentrations of the individual species cannot be directly measured in the carbon dioxide system in a solution) (Millero, 1996; Millero et al., 2002; Dickson et al., 2007). Dissolved inorganic carbon (DIC) is present in seawater in three main forms: bicarbonate ion (HCO_3^-), carbonate ion (CO_3^{2-}), and aqueous carbon dioxide ($CO_{2(aq)}$), which also includes carbonic acid (H_2CO_3) (Fabry et al., 2008). The alkalinity of seawater plays a pivotal role in determining the proportion of the DIC in seawater that exists as CO_2. Alkalinity is the primary control on how much DIC seawater can hold (Henson et al., 2024). The $CO_{2\ (aq)}$ concentration depends upon the chemical composition and the temperature of seawater. Also $CO_{2\ (aq)}$ amount is proportional to the partial pressure of CO_2 exerted by seawater (Feely et al., 2001).

One of the main methods commonly used to estimate the air-sea carbon flux is to measure the difference in CO_2 partial pressure between the atmosphere and ocean and multiply it by an estimate of the gas exchange coefficient (Wanninkhof, 2009; Takeshita et al., 2018). The air-sea gas exchange has long been of scientific interest due to its importance in the biogeochemical cycling of climate, health-related gaseous compounds, and weather (Wanninkhof et al., 2009). The oceans are considered a significant sink for CO_2. Therefore, gaining a better understanding of the net transport flux across the air-sea interface is crucial for comprehending the fate of this greenhouse gas emitted into the Earth's atmosphere (Takeshita et al., 2018).

In particular, the Marmara Sea, located within the borders of Turkiye, is exposed to various pressures both due to the influence of the densely populated provinces in which it is located and because it is the only waterway connecting the Mediterranean to the Black Sea through the Turkish Straits System (TSS). In addition, the carbon adsorption capacity of the Marmara Sea is not yet known. In this study, four different stations were selected in order to understand how the Marmara Sea, which has a geographically important location, reacts under the effects of GHGs, especially CO_2, and at what point it is in terms of carbon sequestration capacity, and to take the necessary measures to fight against global warming and climate change. As a preliminary study, alkalinity measurements were made for the first time, and based on these, partial pressure and flux calculations of carbon dioxide were made.

Materials and Methods

Study Area

The Sea of Marmara is located between the European and Asian continents, connected to the Mediterranean by the Çanakkale Strait and to the Black Sea by the Istanbul Strait (Figure 1) (Unlulata et al., 1990; Besiktepe et al., 1993; Besiktepe et al., 1994). The Sea of Marmara, which forms the TSS together with the Istanbul and Çanakkale Straits, and is located between the coordinates of 40°00' and 41°1'N and 26°15' and 29°55'E (Ozturk, 2002). The coastline of the Sea of Marmara is 972 km, with a surface area of 11,500 km^2 and a volume of 3,378 km^3 (Besiktepe et al., 1994; Polat et al., 1997; Ozturk, 2002).

There is a two-layer water system in the Marmara Sea; while the low salinity waters of the Black Sea (approximately 18‰) come from the Istanbul Strait to the Marmara Sea and form the upper layer, the salty water of the Mediterranean (approximately 38‰) come from the Çanakkale Strait and form the lower layer water (Unlulata et al., 1990; Besiktepe et al., 1994). A sharp density interface formed between these two layers at a depth of approximately 25 m separates them (Besiktepe et al., 1994). In this system, the residence time of the upper layer waters is 4-5 months, while the residence time of the lower layer waters is 6-7 years (Besiktepe et al., 1994; Algan et al., 2004). Four stations were selected as sampling stations in the study area (Figure 1, Table 1): MD24Y, MD26, CG2, and CG3. CG2 and CG3 stations are located around Balıkcı Island, which is 2 km away from Buyukada (the largest of the Prince Islands), MD26 station is located near Heybeliada, and MD24Y station is located on the coast of Yalova.

Table 1. Latitude and the longitude informations of the stations.

Statiton ID	**Latitudes** (°N)	**Longitudes** (°E)
MD26	40.84083	29.05433
CG2	40.81633	29.10983
CG3	40.8225	29.1125
MD24Y	40.70266	29.33383

Sampling

Sampling was carried out via R/V Alemdar II research vessel that belongs to Istanbul University Marine Sciences and Management Institute in September and November 2023. Seawater samples were taken at the specified stations from 3 different depths: surface, intermediate layer, and bottom, with Niskin bottles (General Oceanics Model 1010 Niskin Water Sampler, 5L) connected to the CTD + rosette system. As specially, alkalinity measurements were taken directly from the Niskin bottles with a glass apparatus that would prevent the seawater samples from coming into contact with air in 100 ml glass volumetric flasks that were previously washed with nitrogen gas.

Physicochemical Analysis

The physicochemical parameters such as pH, temperature, dissolved oxygen (DO), and salinity were determined along the water column. Temperature, salinity, and oxygen were measured with sensors connected to the CTD+Rosette system. pH measurements were performed by a BENTA210 Benchtop pH/Mv meter. DO was also measured by the Winkler method on board (modified with CTD data), and water samples were taken directly from the Niskin bottles to prevent any biological activity and contamination from the atmosphere (APHA-AWWA-WPCF, 1985).

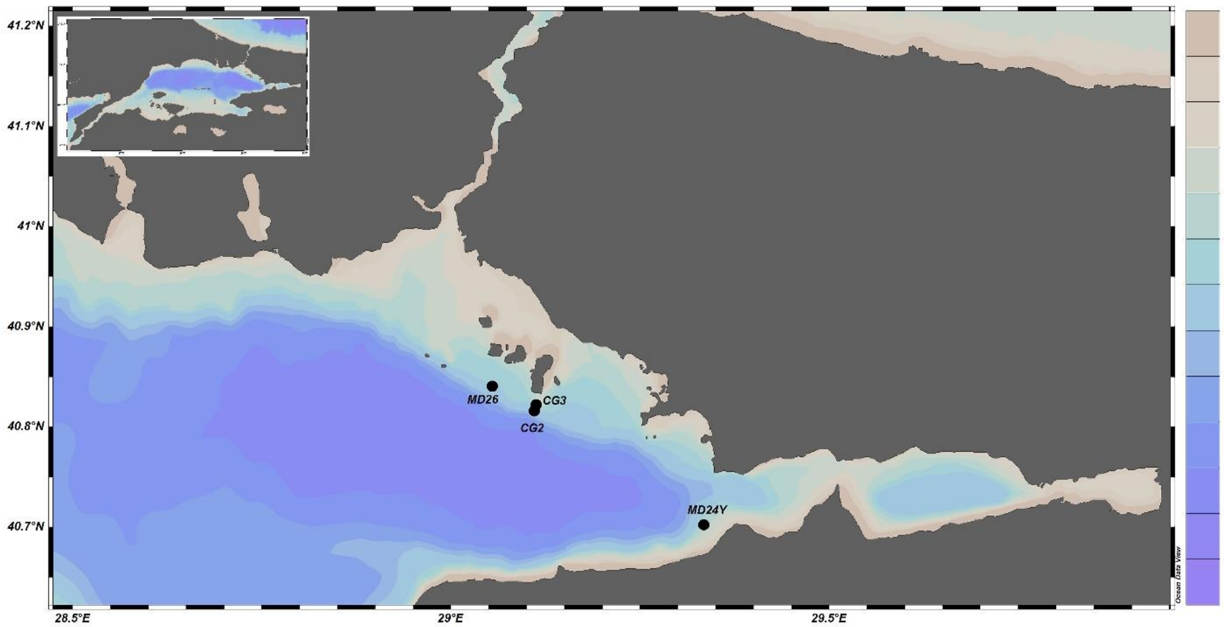

Fig. 1. The locations of the sampling stations in the Sea of Marmara.

Alkalinity measurement and calculation of partial pressure and flux of CO_2

The total alkalinity is determined by titrating the sample with a standard solution of a strong mineral acid. Methods using either an indicator or a pH-meter for the titration are described. For the titration, Metrohm Dosimat 876 Plus was used. Phenolphthalein and methyl orange were used as indicators. Also, a sulphuric acid solution was used as a strong acid (Suess, 1985; Hammer and Hammer, 2013; Dhoke, 2023).

For the calculation of phenolphthalein alkalinity as;

$$\text{Alkalinity } (mmolL^{-1}as\ CaCO_3) = \frac{1000*A*c}{V} \quad \text{(Eq.1)}$$

For the calculation of total alkalinity (A_t) as;

$$A_t(mmolL^{-1}as\ CaCO_3) = \frac{1000*B*c}{V} \quad \text{(Eq.2)}$$

In the given first and second equations respectively; where A is the amount of standard acid solution (mL) to reach the phenolphthalein endpoint or pH of the solution is 8.3. B is the amount of standard acid solution (mL) to reach the endpoint of methyl orange or pH of the solution is 4.5 or the second endpoint of titration. Also, c is the concentration of acid (mol L^{-1}), and V is the volume of the sample (mL).

To describe the CO_2 system in seawater, four parameters are measured pH, the total alkalinity (A_T), total carbon dioxide (C_t), and the (pCO_2) (Millero et al., 1993).

$$\begin{aligned} A_T &= A_C + [B(OH)_4^-] + [OH^-] - [H^+] \\ &= [HCO_3^-] + 2[CO_3^{2-}] + [B(OH)_4^-] + [OH^-] - [H^+] \end{aligned} \quad \text{(Eq.3)}$$

For the calculation of carbonate alkalinity (A_c) as;

$$A_c = A_t + \{H^+\} - \frac{K_W}{\{H^+\}} - \frac{B_t}{(1+\frac{\{H^+\}}{K_B})} \quad \text{(Eq.4)}$$

where H^+ is the concentration of hydrogen ion (mmol kg^{-1}); K_w is ionization constant of water (mmol kg^{-1}); B_t is borate alkanity (mmol kg^{-1}); K_B is boric acid alkalinity constant (mmol kg^{-1}) (Körtzinger, 1999; Grasshoff et al., 2009) .

$$C_t = [H_2CO_3] + [C0_2] + [HC0_3^-] + [CO_3^{2-}] \quad \text{(Eq.5)}$$

Total carbon dioxide (C_t) is calculated based on the following empirical formula;

$$C_t = \frac{A_C\,(\frac{\{H^+\}^2}{K_1+\{H^+\}+K_2})}{\{H^+\}+2K_2} \quad \text{(Eq.6)}$$

where K_1 is first dissociation constant of carbonic acid in seawater (mmol kg^{-1}); K_2 is second dissociation constant of carbonic acid in seawater (mmol kg^{-1}) (Körtzinger, 1999; Grasshoff et al., 2009) .

pCO_2 is calculated by the following equation;

$$p(CO_2) = \frac{A_C\,\{H^+\}^2}{K_0\,K_1(\{H^+\}+2K_2)} \quad \text{(Eq.7)}$$

The air-sea CO_2 fluxes (F, mmol m^{-2} d^{-1}) is calculated using the below equation;

$$F = k\,\alpha\,\Delta pCO_2 \quad \text{(Eq.8)}$$

where k is the velocity of gas transfer (cm h^{-1}); α is CO_2 gas solubility in seawater (mmol L^{-1} atm^{-1}); ΔpCO_2 is the variance between surface seawater and atmospheric pCO_2 (pCO_{2a}) (Sun et al., 2020).

The dry air atmospheric pCO_2 (pCO_{2a}, atm) is calculated by the following equation (Dickson et al., 2007; Takahashi et al., 2002; Takahashi et al., 2009; Padhy et al., 2016)

$$pCO_{2a} = xCO_{2a}\,(p_b - pH_2O) \qquad \text{(Eq.9)}$$

where xCO_{2a} is troposphere mole fraction (ppm); p_b is sea surface pressure (atm); pH_2O is saturated vapor pressure of water.

For the calculation of pH_2O as (Weiss and Price, 1980);

$$lnpH_2O = 24.4543 - 67.4509\left(\frac{100}{T}\right) - 4.8489\ln(\frac{T}{100}) - 0.000544\ S \qquad \text{(Eq.10)}$$

where T is temperature (Kelvin), S is the salinity (‰).

For the calculation of the velocity of gas transfer related to wind speed as (Wanninkhof, 2014);

$$k = 0.251\ U_{10}^2\ (\frac{Sc}{660})^{-1/2} \qquad \text{(Eq.11)}$$

where Sc is the Schmidt number of CO_2, U_{10} is the wind speed (m s^{-1}) at 10 m height. Wind speed data are collected from the earth nullschool data website (Earth Nullschool, 2024) Sc was calculated according to the empirical formula of Wanninkhof (2014);

$$Sc = 2116.8 - 136.25\ t + 4.7353\ t^2 - 0.092307\ t^3 + 0.0007555t^4 \qquad \text{(Eq.12)}$$

where t is the temperature (°C).

Results

Temperature, salinity, and pH are important factors that affect the partial pressure of CO_2 in seawater (Kao et al., 2023; Mathis et al., 2012). pH can affect the solubility of CO_2 because it affects the carbonate balance in seawater. When the pH value increases, CO_2 partial pressure and flux decrease (Cao et al., 2023). The increase in water temperature affects the solubility of CO_2 gas. Increasing water temperature can increase primary production and thus reduce pCO_2 (Wang et al., 2017). In this study, alkalinity was measured in seawater, and calculations were made using temperature, salinity, and pH values. The physicochemical properties of the study area are shown in Figure 2. The method application determined in accordance with the purpose of the study was optimized for the Sea of Marmara.

When the September and November salinity and temperature (T-S) profiles of four stations are examined, the double-layer structure of the Marmara Sea is seen (Unluata, 1990; Besiktepe, 1994). In the T-S profiles of the MD24Y station located at the southern exit of the Gulf of Izmit, the upper layer of water originating from the Black Sea is seen in the first 16-18 meters in both September and November measurements. The water mass with a salinity value of 30 psu, representing the Marmara Sea interface, is at ~19 and 21 meters in September and November, respectively. It is seen that the intermediate layer thickens in September compared to November due to the effect of vertical mixing (Muftuoglu, 2008).

The minimum dissolved oxygen (DO) located just above the Mediterranean water can be associated with organic material that has precipitated and accumulated in this layer. The presence of a cold intermediate layer remaining from the winter month in September is noteworthy. The presence of a cold intermediate layer is also seen in the other three stations, although it loses its prominence. The intermediate layer observed in November and the approximately 1°C excess temperature in the water column up to a depth of 60 meters indicate that the Mediterranean-origin water entering from the Çanakkale Strait in this month settled in this layer in the entire Marmara Sea due to its relatively low density.

At station MD26, which reflects similar characteristics to the MD24Y profile, a uniform structure in terms of both salinity and temperature is noticeable starting from 70 meters. The fluctuating structure in the DO value again becomes regular after 70 meters and decreases. The relatively warm surface water at stations CG2 and CG3, located in the south of the islands, contains less DO than the water immediately below, depending on the solubility (Muftuoglu, 2008).

In September, the highest A_c and C_t were measured at station MD26 in the bottom, respectively 1.42 mmol kg^{-1} and 1.336 mmol kg^{-1}. The lowest values were observed at the same station but on the surface, respectively 1.07 mmol $kg^{-1}$and 0.924 mmol kg^{-1}. The highest pCO_2 was observed at the bottom water of CG3 station, 386 µatm, and the lowest pCO_2 was observed at the surface water of MD26 station, 95 µatm (Table 2). In November, the highest A_c and C_t were measured at station MD26 in the surface, respectively 1.46 mmol kg^{-1} and 1.266 mmol kg^{-1}. The lowest values were 1.13 mmol kg^{-1} and 1.051 mmol kg^{-1} at the CG3 station in the intermediate layer, respectively.

The highest pCO2 was observed at the bottom water of MD26, 279 µatm, and the lowest pCO2 was observed at the surface water of MD26 station, 109 µatm (Table 3). In the results obtained, although alkalinity values varied between 1.07–1.46 mmol kg^{-1}, no major differences were found. The lowest pCO_2 value was observed at MD26 station in September (95 µatm), and the highest value was observed at CG2 station in September (165 µatm). However, the opposite situation exists at the MD26 station. It is thought that this may be due to the change in primary production between two months. It can be said that CO_2 uptake increased due to the fact that the weather conditions were harsher in November and the flux amount became more negative (Wanninkhof and Triñanes, 2017).

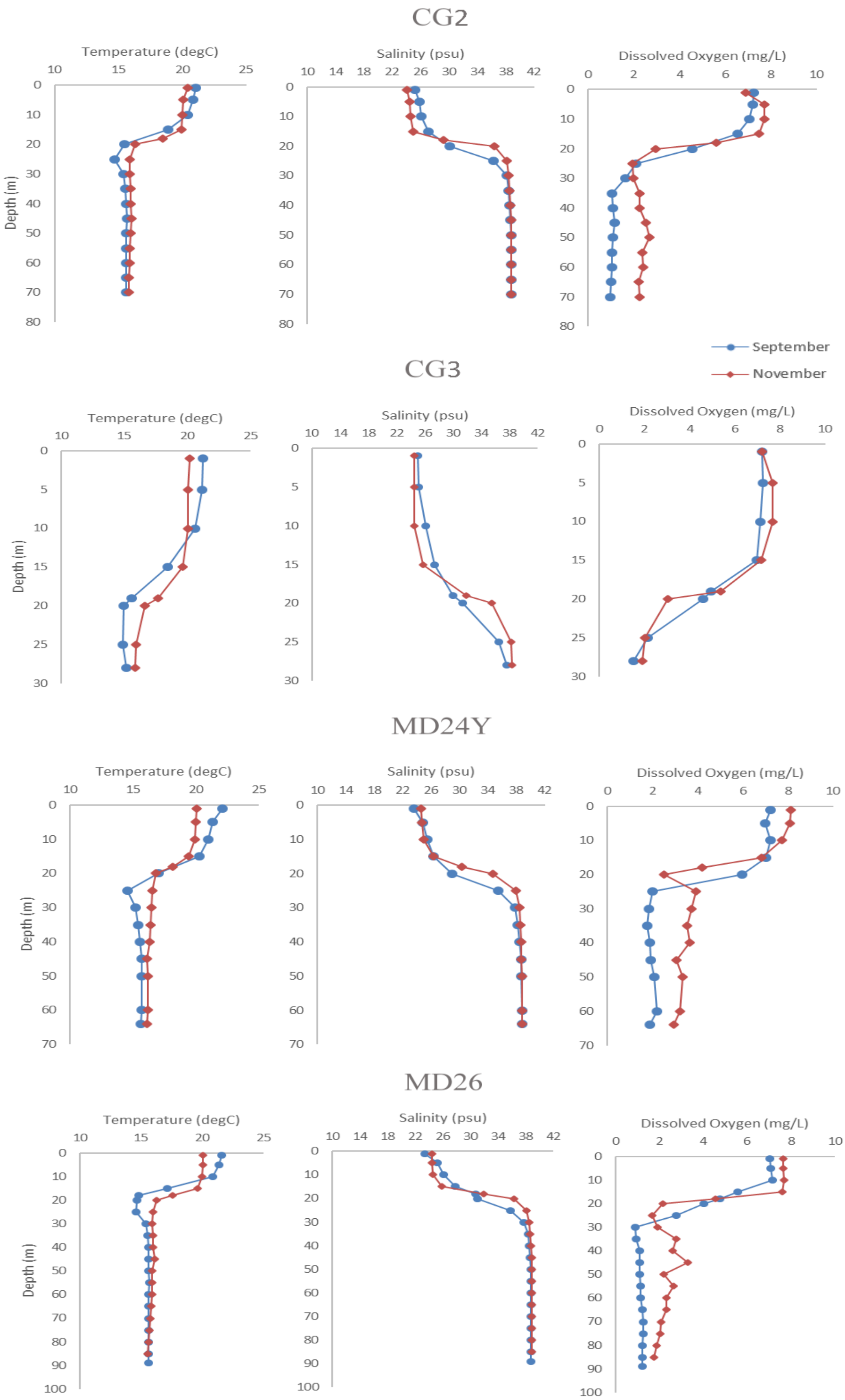

Fig. 2. The graphs show the temperature (°C), salinity (PSU), and Dissolved Oxygen (mg L^{-1}) values versus depth in the stations CG2, CG3, MD24Y, and MD26. The blue line represents the September data, and the red line represents November.

Table 2. The Ac (mmol kg^{-1}), C_t (mmol kg^{-1}), pCO_2 (µatm), pH, and Flux (mmol m^{-2} day^{-1}) values in September in the Sea of Marmara.

Station ID	Sampling Depth (m)	Sampling Time	Ac (mmol kg^{-1})	C_t (mmol kg^{-1})	pCO_2 (µatm)	pH	Flux (mmol m^{-2} day^{-1})
MD26	0.5	September	1.07	0.924	95	8.23	-3.55
MD26	19	September	1.28	1.212	280	7.86	
MD26	88	September	1.42	1.336	316	7.86	
MD24Y	0.5	September	1.19	1.065	143	8.10	-2.34
MD24Y	19	September	1.19	1.100	190	7.98	
MD24Y	66	September	1.17	1.099	264	7.86	
CG2	0.5	September	1.35	1.213	165	8.10	-1.10
CG2	19	September	1.36	1.276	276	7.91	
CG2	28	September	1.35	1.266	313	7.85	
CG3	0.5	September	1.35	1.199	138	8.15	-1.21
CG3	19	September	1.36	1.294	339	7.82	
CG3	25	September	1.37	1.309	386	7.76	

Table 3. The Ac (mmol kg^{-1}), C_t (mmol kg^{-1}), pCO_2 (µatm), pH, and Flux (mmol m^{-2} day^{-1}) values in November in the Sea of Marmara.

Station ID	Sampling Depth (m)	Sampling Time	Ac (mmol kg^{-1})	C_t (mmol kg^{-1})	pCO_2 (µatm)	pH	Flux (mmol m^{-2} day^{-1})
MD26	0.5	November	1.46	1.266	109	8.26	-5.39
MD26	16	November	1.33	1.180	135	8.13	
MD26	85	November	1.23	1.164	279	7.83	
MD24Y	0.5	November	1.41	1.253	137	8.18	-1.77
MD24Y	18	November	1.16	1.090	240	7.89	
MD24Y	60	November	1.21	1.086	140	8.08	
CG2	0.5	November	1.41	1.224	116	8.24	-4.26
CG2	18	November	1.34	1.221	183	8.03	
CG2	28	November	1.21	1.136	255	7.86	
CG3	0.5	November	1.44	1.260	114	8.24	-4.32
CG3	19	November	1.13	1.051	204	7.93	
CG3	28	November	1.25	1.173	263	7.86	

High C_t concentration was associated with saltier water, and low C_t was associated with less salty water (Sun et al., 2020). In this study, an increase in Ct towards the lower layer was observed at all stations in September (Figure 3). This is because saltier Mediterranean waters dominate the lower layer of the Sea of Marmara. However, in November, Ct decreased slightly towards the bottom layer at all stations. (Figure 3). Some of the organic matter entering the aquatic environment is biologically mineralized (Capelle et al., 2020; Mann et al., 2012). Another portion of the organic matter rapidly sinks to the bottom (seabed), which may mineralize into CO_2 if sufficient O_2 is available. Organic matter that remains unmineralized is buried in sediment (Capelle et al., 2020). The relatively slight decrease in Ct values from the surface to the bottom water at the study stations in November can be explained by the entry of high organic matter into the receiving environment this month and the fact that some of this organic matter was mineralized with O_2 while some remained buried in the sediment without being mineralized.

As pH value decreases, pCO_2 increases. In September and November, pH values decrease towards the lower layers and accordingly pCO2 values increase at all stations. However, a decrease in pCO2 value was observed at station MD24Y in November. The reason is that the lower layer pH value (8.08) was higher than the intermediate layer pH value (7.89) (Table 3). In the study conducted by Sun et al. in 2010, pCO_2 values (137 - 591 µatm) in Bering Sea surface water were higher than the values determined in this study (Sun et al., 2020). Kao et al. in 2018, pCO_2 values in surface water in the Southeast China Sea (305 - 325 µatm) were higher than the results of this study (Kao et al., 2023). CO_2 fluxes were calculated based on pCO_2 and daily wind speeds for different stations. Since the pCO_2 in the atmosphere is higher than the pCO2 in water, the calculated flux values gave negative results (−1.10 to −5.39 mmol m^{-2} day^{-1}) (Tables 2 and 3). The negative flux meaning is the net CO_2 exchange from the atmosphere to the sea (Zhai etal., 2005). So It was concluded that there was a transfer of CO_2 from the atmosphere to the sea at the time of sampling. The flux values of determined in the studies of Sun et al. (2010) and Kao et al. (2018), which is similar to the value determined in this study (Kao et al., 2023; Sun et al., 2020).

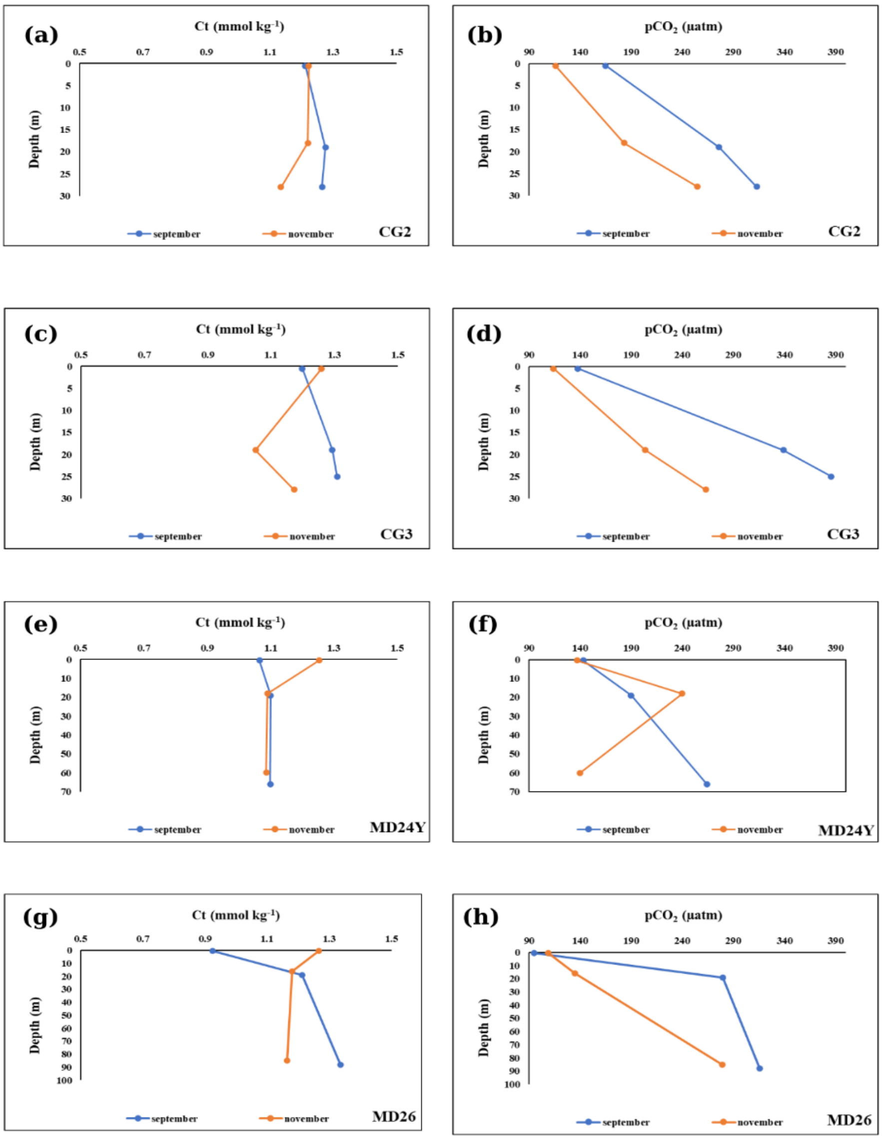

Figure 3. Ct (mmol kg-1) and pCO2 (µatm) distributions at the surface, intermediate, and bottom depths of different stations of the Marmara Sea in September (blue line) and November (orange line).

Conclusion

pCO2 is an important parameter needed to calculate air-sea flux and gain information about ocean acidification. In this study, the distribution of pCO2 and air-sea flux in the surface waters of the Sea of Marmara was investigated in September and November 2023. To determine pCO2 and air-sea flow, alkalinity was measured in seawater and calculations were made using temperature, salinity and pH values. The method application determined in accordance with the purpose of the study was optimized for the Sea of Marmara. At the selected station in the Sea of Marmara CG2, CG3, MD26, and MD24Y for the first time alkalinity measurments and calculations of pCO2 and fluxes were done. According to the pCO2 and air-sea flux results obtained, it was concluded that the CO2 sink dominated in the Sea of Marmara at the sampling time. This study is important as it is the first study conducted in this context and contains the first data. It is one of the first steps taken to find out how the Sea of Marmara behaves in terms of CO2 uptake by showing its condition at the specified stations and at the time the sampling was carried out. In the future, more frequent and long-term studies are required to evaluate the effects of the variability of physicochemical and biological parameters and global warming on the marine environment.

Acknowledgements

The scientific research expedition of this work was supported by the Scientific Research Projects Coordination Unit of Istanbul University (Project Number: 39478, 39143, FHZ-2024-40850, FBA-2023 -39478). Also, we appreciate project coordinators Prof. Dr. Cem Gazioğlu and Prof. Dr. Nazlı Demirel Erol for their valuable contributions. Additionally, we thank to R/ V ALEMDAR II captains and technicians for their help.

References

Algan, O., Balkis, N., Çağatay, M. N., Sari, E. (2004). The sources of metal contents in the shelf sediments from the Marmara Sea, Turkey. *Environmental Geology, 46,* 932-950. doi:10.1007/s00254-004-1104-2

APHA-AWWA-WPCF. (1985). Standard methods for the analysis of water and waste water. In: American Public Health Association Washington.

Besiktepe, S., Sur, H., Özsoy, E., Latif, M., Oguz, T., Ünlüata, Ü. (1994). The circulation and hydrography of the Marmara Sea. *Progress In Oceanography, 34*, 285-334. doi:10.1016/0079-6611(94)90018-3

Beşiktepe, Ş., Özsoy, E., Ünlüata, Ü. (1993). Filling of the Marmara Sea by the Dardanelles lower layer inflow. *Deep Sea Research Part I: Oceanographic Research Papers, 40*(9), 1815-1838.

Cao, X., Wu, Q., Wang, W., Wu, P. (2023). Carbon dioxide partial pressure and its diffusion flux in karst surface aquatic ecosystems: a review. *Acta Geochimica, 42*(5), 943-960. doi:10.1007/s11631-023-00625-7

Capelle, D. W., Kuzyk, Z. Z. A., Papakyriakou, T., Guéguen, C., Miller, L. A., Macdonald, R. W. (2020). Effect of terrestrial organic matter on ocean acidification and CO2 flux in an Arctic shelf sea. *Progress In Oceanography, 185*, 102319. doi.10.1016/j.pocean.2020.102319

Dhoke, S. (2023). Determination of alkalinity in the water sample: a theoretical approach. *Chemistry Teacher International, 5,* 283-290. doi:10.1515/cti-2022-0052

Dickson, A., Chris, S., Christian, J. R. (2007). Guide to Best Practices for Ocean CO2 Measurements. *Guide to Best Practices for Ocean CO2 Measurements, 3*.

Dioxide, A. C. (1985). the Global Carbon Cycle. *Department of Energy ER-0239, Washing*.

Doney, S. C., Fabry, V. J., Feely, R. A., Kleypas, J. A. (2009). Ocean Acidification: The Other CO2 Problem. *Annual Review of Marine Science, 1*(Volume 1, 2009), 169-192. doi.10.1146/annurev.marine.010908.163834

Earth Nullschool. (2024). A Global Map Of Wind, Weather, And Ocean Conditions.

Fabry, V. J., Seibel, B. A., Feely, R. A., Orr, J. C. (2008). Impacts of ocean acidification on marine fauna and ecosystem processes. *ICES Journal of Marine Science, 65*(3), 414-432. doi:10.1093/icesjms/fsn048

Feely, R. A., Sabine, C. L., Takahashi, T., & Wanninkhof, R. (2001). Uptake and Storage of Carbon Dioxide in the Ocean: The Global CO2 Survey. *Oceanography, 14*, 18-32.

Gazioğlu, C., Müftüoğlu, A. E., Demir, V., Aksu, A., et al. (2015). Connection between Ocean Acidification and Sound Propagation. *International Journal of Environment and Geoinformatics,* 2(2), 16-26. https://doi.org/10.30897/ijegeo.303538

Gazioğlu, C., Okutan, V. (2016). Underwater Noise Pollution at the Strait of Istanbul (Bosphorus). *International Journal of Environment and Geoinformatics,* 3(3), 26-39. https://doi.org/10.30897/ijegeo.306478

Grasshoff, K., Kremling, K., Ehrhardt, M. (2009). *Methods of seawater analysis*: John Wiley & Sons.

Hammer Sr, M. J., Hammer Jr, M. J. (2013). *Water and Wastewater Technology: Pearson New International Edition*: Pearson Education.

Henson, S., Baker, C. A., Halloran, P., McQuatters-Gollop, A., Painter, S., Planchat, A., Tagliabue, A. (2024). Knowledge Gaps in Quantifying the Climate Change Response of Biological Storage of Carbon in the Ocean. *Earth's Future, 12*(6), e2023EF004375. doi:https://doi.org/10.1029/2023EF004375

Houghton, J. (2005). Global warming. *Reports on Progress in Physics, 68*(6), 1343-1403. doi:DOI:101088/0034-4885/68/6/R02

Kao, K.-J., Huang, W.-J., Chou, W.-C., Gong, G.-C., Weerathunga, V. (2023). Factors Controlling the Sea Surface Partial Pressure of Carbon Dioxide in Upwelling Regions: A Case Study of the Southern East China Sea Before and After Typhoon Maria. *Journal of Geophysical Research: Oceans, 128*(2), e2022JC019195. doi:https://doi.org/10.1029/2022JC019195

Keeling, C. D. (1997). Climate change and carbon dioxide: An introduction. *Proceedings of the National Academy of Sciences, 94*(16), 8273-8274. doi:doi:10.1073/pnas.94.16.8273

Kleypas, J. A., Feely, R. A., Fabry, V. J., Langdon, C., Sabine, C. L., Robbins, L. L. (2005). Impacts of ocean acidification on coral reefs and other marine calcifiers: a guide for future research. In *Report of a workshop held* (Vol. 18, No. 2005, p. 20).

Körtzinger, A. (1999). Determination of carbon dioxide partial pressure (p(CO2)). In *Methods of Seawater Analysis* (pp. 149-158).

Lal, R. (2009). Sequestering Atmospheric Carbon Dioxide. *Critical Reviews in Plant Sciences, 28*(3), 90-96. doi:10.1080/07352680902782711

Lindzen, R. S. (1994). Climate Dynamics and Global Change. *Annual Review of Fluid Mechanics, 26*(Volume 26, 1994), 353-378. doi:https://doi.org/10.1146/annurev.fl.26.010194.002033

Mann, P. J., Davydova, A., Zimov, N., Spencer, R. G. M., Davydov, S., Bulygina, E., Holmes, R. M. (2012). Controls on the composition and lability of dissolved organic matter in Siberia's Kolyma River basin. *Journal of Geophysical Research: Biogeosciences, 117*(G1). doi:https://doi.org/10.1029/2011JG001798

Mathis, J. T., Pickart, R. S., Byrne, R. H., McNeil, C. L., Moore, G. W. K., Juranek, L. W., Feely, R. A. (2012). Storm-induced upwelling of high pCO2 waters onto the continental shelf of the western Arctic Ocean and implications for carbonate mineral saturation states. *Geophysical Research Letters, 39*(7). doi. 10.1029/2012GL051574

Millero, F. J. (1996). *Chemical oceanography*: CRC press.

Millero, F. J., Pierrot, D., Lee, K., Wanninkhof, R., Feely, R., Sabine, C. L., Takahashi, T. (2002). Dissociation constants for carbonic acid determined from field measurements. *Deep Sea Research Part I: Oceanographic Research Papers, 49*(10), 1705-1723. doi:https://doi.org/10.1016/S0967-0637(02)00093-6

Millero, F. J., Zhang, J.-Z., Lee, K., Campbell, D. M. (1993). Titration alkalinity of seawater. *Marine Chemistry, 44*(2), 153-165. doi.10.1016/0304-4203(93)90200-8

Müftüoğlu, E. (2008). *Hydrodynamic structure of the bays and the estuary of the Sea of Marmara.* Ph. D. Dissertation. Institute of Marine Sciences and Management, Istanbul

Orr, J. (2011). Recent and future changes in ocean carbonate chemistry. *Ocean Acidification*, 41-66.

Ozturk, B. (2002). The Marmara Sea, a Link between the Mediterranean and the Black Sea. In E. Leppäkoski, S. Gollasch, S. Olenin (Eds.), *Invasive Aquatic Species of Europe. Distribution, Impacts and Management* (pp. 337-340). Dordrecht: Springer Netherlands.

Padhy, P. C., Nayak, R. K., Dadhwal, V. K., Salim, M., Mitra, D., Chaudhury, S. B., Dutt, C. B. S. (2016). Estimation of Partial Pressure of Carbon Dioxide and Air-Sea Fluxes in Hooghly Estuary Based on In Situ and Satellite Observations. *Journal of the Indian Society of Remote Sensing, 44*(1), 135-143. doi:10.1007/s12524-015-0459-z

Polat, S. C., Tuğrul, S., Çoban, Y., Basturk, O., Salihoglu, I. (1997). Elemental composition of seston and nutrient dynamics in the Sea of Marmara. *Hydrobiologia, 363*(1), 157-167. doi:10.1023/A:1003117504005

Suess, M. J. (1985). *Examination of water for pollution control: A reference handbook. Physical, chemical and radiological examination*. United States: Pergamon Books Inc.,Elmsford, NY.

Sun, H., Gao, Z., Qi, D., Chen, B. s., Chen, L., Cai, W.-J. (2020). Surface seawater partial pressure of CO2 variability and air-sea CO2 fluxes in the Bering Sea in July 2010. *Continental Shelf Research, 193*, 104031. doi:https://doi.org/10.1016/j.csr.2019.104031

Takahashi, T., Sutherland, S. C., Sweeney, C., Poisson, A., Metzl, N., Tilbrook, B., Nojiri, Y. (2002). Global sea–air CO2 flux based on climatological surface ocean pCO2, and seasonal biological and temperature effects. *Deep Sea Research Part II: Topical Studies in Oceanography, 49*(9), 1601-1622. doi:https://doi.org/10.1016/S0967-0645(02)00003-6

Takahashi, T., Sutherland, S. C., Wanninkhof, R., et al., 2009). Climatological mean and decadal change in surface ocean pCO2, and net sea–air CO2 flux over the global oceans. *Deep Sea Research Part II: Topical Studies in Oceanography, 56*(8), 554-577. doi. 0.1016/j.dsr2.2008.12.009

Takeshita, Y., Johnson, K. S., Martz, T. R., Plant, J. N., Sarmiento, J. L. (2018). Assessment of Autonomous pH Measurements for Determining Surface Seawater Partial Pressure of CO2. *Journal of Geophysical Research: Oceans, 123*(6), 4003-4013. doi:https://doi.org/10.1029/2017JC013387

Ünlülata, Ü., Oğuz, T., Latif, M. A., Ersoy, E. (1990). On the Physical Oceanography of the Turkish Straits. In L. J. Pratt (Ed.), *The Physical Oceanography of Sea Straits* (pp. 25-60). Dordrecht: Springer Netherlands.

Wang, X. F., Yuan, X. Z., Chen, H., He, Y. X., Luo, Z., Liu, L., He, Z. Y. (2017). [Review of CO(2) and CH(4) Emissions from Rivers]. *Huan Jing Ke Xue, 38*(12), 5352-5366. doi:10.13227/j.hjkx.201701049

Wanninkhof, R. (2014). Relationship between wind speed and gas exchange over the ocean revisited. *Limnology and Oceanography: Methods, 12*, 351-362. doi:10.4319/lom.2014.12.351

Wanninkhof, R., Triñanes, J. (2017). The impact of changing wind speeds on gas transfer and its effect on global air-sea CO2 fluxes. *Global Biogeochemical Cycles, 31*(6), 961-974. doi. 10.1002/2016GB005592

Wanninkhof, R., Asher, W. E., Ho, D. T., Sweeney, C., & McGillis, W. R. (2009). Advances in Quantifying Air-Sea Gas Exchange and Environmental Forcing*. *Annual Review of Marine Science, 1*(Volume 1, 2009), 213-244. doi. 10.1146/annurev.marine.010908.163742

Weiss, R. F., Price, B. A. (1980). Nitrous oxide solubility in water and seawater. *Marine Chemistry, 8*, 347-359. doi:10.1016/0304-4203(80)90024-9

Zhai, W., Dai, M., Cai, W.-J., Wang, Y., Hong, H. (2005). The partial pressure of carbon dioxide and air–sea fluxes in the northern South China Sea in spring, summer and autumn. *Marine Chemistry, 96*(1), 87-97.