The inorganic carbon cycle of the Red Sea

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Abstract

The aim of this thesis has been to improve the general scientific understanding of the marine inorganic carbon cycle of the Red Sea. The region houses an exceptional range of ecosystems and it is considered the most important repository of biodiversity in the world with vast calcification areas rich in coral reefs. Therefore, it is important to understand the interaction between biogeochemical processes and the Red Sea ecosystem. In spite of this fact, the area is poorly studied, and this is particularly true for the marine inorganic carbon cycle.

Prior to this work, no systematic sampling of biogeochemical variables had been done off the coast of Sudan. As part of this work, a new biogeochemical time series from the Sudanese coastal waters has been established, which represents the very first data on hydrography and inorganic carbon from this part of the Red Sea. The sampling site has been operated since 2007, and here is presented data over a period of 8 years, from 2007 to 2015. Consequently, there now exists a comprehensive dataset allowing the quantification of interannual to seasonal variations in biogeochemical variables, and identify important processes that control these changes. The establishment of such a baseline is important to compare future changes caused by climate and environmental changes when approaching the high CO₂ world.

The three papers in this thesis improve and renew our understanding of the Red Sea hydrography, the inorganic carbon cycle, how the variation goes, and what drives the observed changes.

A common assumption has been that the temperature and salinity variations in the Red Sea is solely a function of local heat and mass flux. However, results from the current work evidence that advection of temperature and salinity gradients also plays a role for establishing the temperature and salinity maxima in the area (Paper I). The finding of this paper also show that the annual temperature range is approximately 6°C, with
highest temperatures during summer and autumn. Changes in salinity lag that of temperature with about 3 months.

The inorganic carbonate measurements (Paper II) document for the very first time the seasonal as well as interannual variability of dissolved inorganic carbon (DIC) and total alkalinity (A_T) in this part of the ocean. This pristine dataset has been used to construct a baseline for the inorganic carbon cycle and constitutes an important reference for years to come. Higher A_T and DIC is measured during winter compared to the summer with an annual change of approximately 40 μmol kg^{-1} for A_T and 32 μmol kg^{-1} for DIC. A_T is mainly driven by physical processes such as advection and local evaporation (through salinity changes) as well as calcification, while changes in DIC are mainly a result of air-sea exchange and likely biological activity, and to a smaller degree along shore advection.

The Red Sea, which is situated in the sub-tropical and tropical area, has previously been regarded as a net annual source for atmospheric CO_2, but this view has to be revised. Paper III shows that pCO_2 is high during summer and autumn and low during spring and winter, with a seasonal amplitude of about 60 μatm. Consequently, the Sudanese coastal area acts as a source for atmospheric CO_2 during summer and autumn, while during winter and spring, the area is a sink for atmospheric CO_2. Over an annual cycle, the area is a net sink of atmospheric CO_2 of size 24.4 mmol CO_2 m^{-2} y^{-1}. The change from being a net annual source for atmospheric CO_2 to becoming a net sink likely occurred in the 2000s.
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1. Aim of the study

The main goal of this thesis has been to achieve a better understanding of the inorganic carbon cycle at the western Red Sea through unravelling the biogeochemical setting and examining the spatiotemporal variation of hydrography and carbon parameters. The environmental settings for the Red Sea is extreme in terms of biogeochemistry. The sea is considered an important calcification area with numerous coral reefs and thus its coastline is very vulnerable, but despite of this fact, there is a limited amount of scientific surveys in the area, which has resulted in a poorly understanding of the marine inorganic carbon cycle, in particular. The lack of data hinders adequate analyses of the interactions between the biogeochemical processes and the inorganic carbon cycle in the area. This is critical in light with the ongoing global warming, exemplified in a temperature increase in the Red Sea of 0.7 °C from 1985 to 2007 (Raitos et al., 2011), expanding oxygen–minimum zones in the tropical regions (Stramma et al., 2008), and the increasing atmospheric CO₂ (Le Quéré et al., 2016).

With help of a pristine time series of hydrography and inorganic carbon data, the aim is to develop new and improved knowledge regarding the inorganic carbon cycle in this area.
2. The marine carbon system

2.1 Introduction

Due to the carbon emissions produced by the combustion of fossil fuels, production of cement, and land-use change, the atmospheric carbon dioxide (CO₂) concentration has increased exponentially from approximately 277 parts per million (ppm) at the beginning of industrial revolution in 1750 to 402 ppm in 2016. According to Le Quéré et al. (2016), combustion of fossil fuels globally emitted an amount of 9.3 ± 0.5 GtC y⁻¹ into the atmosphere while land-use change contributed 1.0 ± 0.5 GtC y⁻¹ during the period between 2006 and 2015. Ocean takes up about a quarter of the annual carbon emissions (2.6 ± 0.5 GtC y⁻¹) and the global residual terrestrial CO₂ sink is 3.1 ± 0.9 GtC y⁻¹. The growth rate of atmospheric CO₂ concentration for the same period is 4.5 ± 0.1 GtC y⁻¹ (Le Quéré et al., 2016) (see Fig.2.1).

Fig. 2.1 The global carbon dioxide budget with reservoirs (in GtC y⁻¹) for the period between 2006 and 2015. 1 GtC equals 10¹⁵ g C. Figure from Le Quéré et al. (2016).
The CO$_2$ continuously cycles between the atmosphere, land, and ocean, but the amount of CO$_2$ is not evenly distributed, and e.g. the ocean stores about 50 times more CO$_2$ than the atmosphere (Field and Raupach, 2004; Zeebe and Wolf-Gladrow, 2001). The CO$_2$ exchange between the surface ocean and atmosphere is faster than the exchange between the surface layer and deep layer of the ocean. The latter one takes from several hundreds to thousands of years because it is driven by slower processes such as deep-water formation and the marine biological production (Emerson and Hedges, 2008).

### 2.2 The carbon chemistry

When CO$_2$ is dissolved in seawater, it is converted to aqueous CO$_2$ and aqueous carbonic acid (H$_2$CO$_3$). This weak acid is dissociated in two steps producing one proton (H$^+$) and bicarbonate ion (HCO$_3^-$) in the first step and two protons and carbonate (CO$_3^{2-}$) ion in the second step (see Eq. 2.1):

$$CO_2 + H_2O \leftrightarrow H_2CO_3^* \leftrightarrow H^+ + HCO_3^- \leftrightarrow 2H^+ + CO_3^{2-} \quad (2.1)$$

where $K_0$ is Henry’s constant for CO$_2$, and $K_1$ and $K_2$ are the first and the second dissociation constants of carbonic acid. All constants depend on temperature, salinity, and pressure of seawater (Lueker et al., 2000; Mehrbach et al., 1973). The star (*) indicate aquatic solution. Additional CO$_2$ dissolved in seawater generates more H$^+$ ions which drive more CO$_3^{2-}$ to react with H$^+$ and produce HCO$_3^-$; known as the buffer reaction:

$$CO_2 + CO_3^{2-} + H_2O \leftrightarrow 2HCO_3^- \quad (2.2)$$

CO$_3^{2-}$ is also used to form calcium carbonate (CaCO$_3$), which is a building block for skeleton and shells of some marine organisms. The CaCO$_3$ saturation state describes if the water is supersaturated or under saturated with respect to CO$_3^{2-}$. 

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The marine carbon system is described by four carbon variables: total dissolved inorganic carbon (DIC), total alkalinity ($A_T$), pH, and fugacity of carbon ($fCO_2$). If two of the carbon variables are known in addition to the equilibrium constants, the other two variables can be calculated. The total dissolved inorganic carbon (DIC), sometimes called TCO$_2$, $C_T$, or $\sum CO_2$, is defined as the sum of inorganic carbon components (Dickson and Goyet, 1994):

$$DIC = [CO_2^*] + [HCO_3^-] + [CO_3^{2-}]$$ (2.3)

where $[CO_2^*]$ is the sum of aquatic CO$_2$ and H$_2$CO$_3$.

The $A_T$ is defined by Dickson (1981) as "the number of moles of hydrogen ion equivalent to the excess of proton acceptors over proton donors in one kilogram of sample":

$$A_T = [HCO_3^-] + 2[CO_3^{2-}] + [B(OH)_4^-] + [OH^-] + [HPO_4^{2-}] + 2[PO_4^{3-}] + [H_2SiO_4^-] + [NH_3] + [HS^-] - [H^+]_F - [HSO_4^-] - [HF] - [H_3PO_4]$$ (2.4)

where $[H^+]_F$ is the free concentration of H$^+$ ions. The major parts of total alkalinity in seawater is:

$$A_T = [HCO_3^-] + 2[CO_3^{2-}] + [B(OH)_4^-] + [OH^-] - [H^+]_F$$ (2.5)

The thermodynamic state of the acid-base system is described by the pH, which is defined as the negative logarithm of hydrogen ions concentration in gram atoms per liter:

$$pH = -\log[H^+]$$ (2.6)

The ocean can be considered as a natural buffer system and seawater at normal conditions is slightly basic ($\approx 8.1$) (Zeebe, 2012). At this pH, HCO$_3^-$ is the major component of DIC (88.6%) followed by CO$_3^{2-}$ (10.9%) while $[CO_2^*]$ only takes about 0.5% (Fig. 2.2).
Fig. 2.2 Concentrations of the different inorganic carbon components as a function of pH. The dissociation constants of Eq. 2.1 are presented in the figure. $[H_2CO_3]^*$ equals $[CO_3]^*$, which is mentioned in the text. Figure from Sarmiento and Gruber (2006).

The partial pressure of CO$_2$ ($pCO_2$) is describing the amount of CO$_2$ in gas phase that is in equilibrium with seawater. According to Sarmiento and Gruber (2006), Eq. 2.1 and its dissociation constants can be reformulated to express the $pCO_2$ as:

$$pCO_2 = \frac{k_2}{k_0 \cdot k_1} \frac{[HCO_3^-]^2}{[CO_3^{2-}]}$$

(2.7)

The fugacity of CO$_2$ ($fCO_2$) differs from $pCO_2$ by taking into account the non-ideal behavior of CO$_2$ and the difference between them is less than 0.5% (Dickson and Goyet, 1994). Zeebe and Wolf-Gladrow (2001) has described the relation between $pCO_2$ and $fCO_2$ as:

$$fCO_2 = pCO_2 \exp(P \frac{B + 2\delta}{RT})$$

(2.8)
where $P$ is the total atmospheric pressure, $B$ is the first virial coefficient of CO$_2$, $\delta$ is the cross virial coefficient, $R$ is the gas constant, and $T$ is absolute temperature. The unit for these parameters are: $f$CO$_2$ and $p$CO$_2$ in $\mu$atm, $P$ in Pa (1 atm = 101325 Pa), $B$ and $\delta$ are in m$^3$ mol$^{-1}$, $R = 8.314$ J K$^{-1}$mol$^{-1}$, and $T$ in Kelvin. According to Weiss (1974), $B$ and $\delta$ can be determined, respectively, as:

$$B = (-1636.75 + 12.0408T - 3.27957 \cdot 10^{-2}T^2 + 3.16528 \cdot 10^{-5}T^3) \times 10^{-6}$$ (2.9)

$$\delta = (57.7 - 0.118T) \times 10^{-6}$$ (2.10)

The buffer capacity reflects the capacity of seawater to buffer changes in pH occurring because CO$_2$ is absorbed in the sea, and it can be quantified through the Revelle factor ($\gamma$). $\gamma$ describes how $f$CO$_2$ changes for a given change in DIC when $A_T$ is constant:

$$\gamma = \frac{\Delta f_{CO_2}}{\Delta DIC} \cdot \frac{DIC}{f_{CO_2}}$$ (2.11)

According to Takahashi et al. (1993), the Revelle factor is high (approximately 14) in cold polar and subpolar surface water and low (ca. 8) in warm surface water in tropical and subtropical areas. A global Revelle factor of 10 indicates that 1% change in DIC will drive the surface $f$CO$_2$ to change by about 10%. The current $\gamma$ values are higher by one unit compared to the values prior to the industrial revolution (Sabine et al., 2004). Increasing surface $f$CO$_2$ drives the $\gamma$ values to increase, thus the surface ocean has become less able to absorb additional CO$_2$.

### 2.3 The main processes controlling the marine carbon cycle

In general, the processes described below have been assessed and make up the theoretical basis in the general carbon cycle for the three papers presented in this thesis.
2.3.1 Solubility pump

The solubility of any gas in seawater depends on temperature and salinity (Gordon and Jones, 1973). Therefore, cold water at high latitudes contain more CO\textsubscript{2} in equilibrium with the atmosphere than warm water at lower latitudes. By help of the global thermohaline circulation (Broecker, 1991), dissolved CO\textsubscript{2} sinks towards deep layers through deep-water formation at high latitudes, while at low latitudes, upwelling brings carbon rich deep water to the surface, which is warm and has low gas solubility, and thus, CO\textsubscript{2} is emitted into the atmosphere. This physical process (called solubility pump) takes long time (hundreds of years) and constantly exchange CO\textsubscript{2} between ocean and atmosphere.

2.3.2 Biological pump

The biological pump can be split into two parts: organic carbon pump and calcium carbonate counter pump.

The organic carbon pump is described through the primary production taking place in the surface water and remineralization of organic matter which occurs in sub-surface and deep waters. During primary production, the phytoplankton use aquatic CO\textsubscript{2} from the surface water and transform it into organic matter through photosynthesis:

\[
106CO_2 + 122H_2O + 16HNO_3 + H_3PO_4 \xrightarrow{\text{sunlight}} (CH_2O)_{106}(NH_3)_{16} H_4PO_4 + 138O_2 \quad (2.12)
\]

A minor part of the sinking organic particles is trapped in the sediment, while most of the organic matter is remineralized by bacteria and regenerated into CO\textsubscript{2} and nutrients. Upwelling of deep water brings water rich in carbon and nutrients into the surface (Fig. 2.3).
The calcium carbonate counter pump is described through the production of calcium carbonate in the surface layer and dissolution in the deep waters. The coral reefs and many planktonic organisms such as Coccolithophorids uses CaCO$_3$ to form their shells and skeletons according to

$$Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + CO_2 + H_2O$$ (2.13)

In the deep water, where the conditions are more acidic, the CaCO$_3$ shells are dissolved:

$$H^+ + CaCO_3 \rightarrow Ca^{2+}(aq) + CO_3^{2-}(aq) + H^+ \rightarrow HCO_3^-$$ (2.14)
unit, and only to a minor degree decreases pH and increases $f_{CO_2}$ (Fig. 2.4) (Zeebe and Wolf-Gladrow, 2001).

![Diagram](image)

Fig. 2.4 Effects of the different processes on DIC, $A_T$, pH, and $[CO_2]$ (from Zeebe and Wolf-Gladrow, 2001).

### 2.3.3 Air-sea gas exchange

Air-sea gas exchange affects the DIC and pH (Fig. 2.4), and the direction of CO$_2$ fluxes between air and sea depends on differences in CO$_2$ concentration ($f_{CO_2}$ or $pCO_2$) between atmosphere and surface ocean, as well as the wind at sea surface, the temperature and to a small degree the salinity. The CO$_2$ flux is determined as

$$ F = SK (f_{CO_2}^{seawater} - f_{CO_2}^{atmosphere}) $$

(2.15)

where $S$ is solubility of CO$_2$ in seawater (mol kg$^{-1}$atm$^{-1}$), which depends on temperature and salinity of the surface water (Weiss, 1974). $K$ is the gas transfer velocity, which depends on the molecular diffusivity, kinematic viscosity, and turbulence at the air-
water interface. $K$ is commonly parameterized as a function of wind speed because the wind speed is important for the turbulence.

There are numerous relationships of $K$ in the literature (Liss and Merlivat, 1986; Wanninkhof, 1992; Wanninkhof and McGillis, 1999; Nightingale et al., 2000; Sweeney et al., 2007), and in this thesis, the one from Nightingale et al. (2000) is used:

$$K = \left(0.222U_{10}^2 + 0.333U_{10}\frac{Sc}{660}\right)^{-1/2} \quad (2.16)$$

where $U_{10}$ is wind speed at 10 m above the sea surface, and $Sc$ is the Schmidt number, which show the ratio between kinematic viscosity and molecular diffusivity.
3. Description of the study area

The Red Sea separates the northeastern Africa from the Arabian Peninsula, and linking between the tropical and sub-tropical parts of the ocean. It is about 1930 km long, on average 200 km wide, and represent an area of approximately $0.46 \times 10^6 \text{ km}^2$. The Red Sea is connected with the Gulf of Aden and the Indian Ocean through the narrow strait of Bab Al Mandab (BAM) (Fig. 3.1). In the northern part, the Red Sea is divided into two main branches; Gulf of Aqaba and Gulf of Suez, the latter is linked with the Mediterranean Sea via the Suez Canal. The bottom topography of the Red Sea is naturally wedge shaped with relatively large maximum depth (2920 m) in the central of the basin. The average depth of Bab Al Mandab strait is about 300 m while the Gulf of Suez has a relatively flat bottom with depth about 60 m. The Gulf of Aqaba is a deep basin with narrow shelves and a mean depression depth of about 1200 m (Morcos, 1970; Patzert, 1974; Edwards, 1987; Maillard and Soliman, 1986).

The climate of the Red Sea is influenced by two wind regimes (Fig. 3.1). North of 19° N the north-northwesterly wind (NNW), which is controlled by eastern Mediterranean weather systems, is active throughout the year. During the summer, the NNW wind extends as far south as the BAM Strait. The area south of 14°N is influenced by Indian Monsoon system, which switches between south-southeasterly wind (SSE) during winter and NNW during summer. The area between 14°N and 19°N is characterized as a convergent zone for the wind field for most of the year, see Fig. 3.1 (Morcos, 1970; Pedgley, 1974; Patzert, 1974).

The average sea surface temperature of the Red Sea is about 26°C in the north and 30°C in the south during summer, while during winter, the temperature is 2-4°C lower. The highest surface temperatures (28°C - 34°C) are found in the south-central parts of the Red Sea where the wind field is convergent and thus weak for most of the year (Sofianos and Johns, 2003). Because of high evaporation, low precipitation, and supply of relative fresh water from the south, the surface salinity of the Red Sea is increasing from approximately 36.5 psu in the south to more than 41.0 psu in the north. Deeper
than about 300 m, the temperature is about 21.5°C and salinity about 40.5 psu all the way to the bottom (Edwards, 1987).

Fig. 3.1 Map showing the location and bathymetry of the Red Sea. Bab Al Mandab Strait (BAM), Gulf of Suez (GS), and Gulf of Aqaba (GA) are indicated in the figure. Arrows refer to wind directions: NNW wind = yellow arrows; SSE wind = red arrows. Arrows to the right in the figure indicate summer situation, while those to the left are winter situation. Locations of cyclonic and anticyclonic gyres are also indicated.

There are four dominant water masses in the Red Sea, all affected by two major masses in the Gulf of Aden. During wintertime, the Gulf of Aden Surface Water (GASW) enters the Red Sea through the BAM Strait as a result of southeasterly winds (Morcos, 1970; Pedgley, 1974; Patzert, 1974; Quadfasel and Baudner, 1993). At deeper layers, the Red Sea Overflow Water (RSOW), which is a mixture of Red Sea Intermediate Water (RSIW) and Red Sea Deep Water (RSDW), is observed to flow out of the Red Sea throughout the year (Sofianos and Johns, 2003; Yao et al., 2014 a; b). During
summer, when northwesterly winds extend as far south as the BAM Strait, an outflow of Red Sea Surface Water (RSSW) is induced while south of the BAM Strait, southwesterly winds provoke an upwelling of Gulf of Aden Intermediate Water (GAIW), as it moves towards the Red Sea (Morcos, 1970; Patzert, 1974; Smeed, 1997).

The overall circulation of the Red Sea is influenced by the monsoon winds. Surface currents during winter flow from the Indian Ocean northwards throughout the Red Sea while the direction of currents during summer, is reversed i.e. flowing southwards to the Indian Ocean as deep currents. The surface circulation in the south (around 15°N) is featured by anticyclonic rotation during winter and cyclonic rotation during summer. Serval cyclonic and anticyclonic gyres are distributed along the north-south axis of the Red Sea, and the strength, size and location of these gyres vary with time (Fig.3.1) (Morcos and Soliman, 1974; Quadfasel and Bauner, 1993; Yao et al., 2014 a; b).
4. Objectives

The main objective of this thesis is to explore the inorganic carbon cycle of the Red Sea, which is poorly known with respect to seasonal and interannual variability. The main objectives has been to:

- establish and maintain a time series to document seasonal and interannual changes.
- understand the coastal physical oceanographic setting.
- determine the drivers of biogeochemical variability at multiple time scales.

The thesis is divided into three topics, which are covered in separate papers. The three papers are based upon the new time series of discrete and continuous data collected in the Sudanese coastal waters off Port Sudan during the period 2007-2015.

Generally, there are few studies focusing on temporal variations of temperature or salinity within the Red Sea, and more specific, there is no study, which fully has dealt with the dynamics responsible for the observed temperature and salinity variations. **Paper I** aims towards getting a better understanding on the temporal variability of ocean physics based on temperature and salinity as well as pointing at the mechanisms responsible for these variations. **Paper II** aims to unravel the seasonality of the marine inorganic carbon cycle, represented by DIC and $A_T$, the relationship with the hydrography, and the drivers causing the observed variability of DIC and $A_T$.

**Paper III** focus on how to determine the air-sea flux of CO$_2$ over an annual cycle using continuous atmospheric and oceanic pCO$_2$ measurements of moored autonomous sensors deployed in our study area. Further, the seasonal variations and drivers of the oceanic pCO$_2$ variability have been identified.
5. Summary

Combined, the three papers in this thesis improve and renew our understanding of the Red Sea inorganic carbon cycle, the hydrography, how it varies, and what drives the observed changes.

A new time-series of discrete and continuous data from the Sudanese coast shows that, in contradiction to previous knowledge, advection in addition to local heat and mass flux drive the temperature and salinity variability (Paper I). The surface temperature (SST) is at the highest value during summer and autumn and at lowest during winter, with a seasonal amplitude of approximately 6°C. The salinity lagged temperature with about 3 months, and the seasonal amplitude was about 1.1 psu. Validated satellite-derived SST data confirmed the above temperature findings, which was also confirmed by computed T arrived from the local heat flux when taking into account the adjustment for advection and mixing. The impact of alongshore advection on seasonal distribution of temperature and salinity has been estimated using a simple model based on gradient features of temperature, salinity and geostrophic surface velocity. The geostrophic surface velocity is computed from sea level anomaly field. SSS increases from south to north and the maximum SST zone is located south of the area of study, in the central Red Sea. The geostrophic current directs form south to north during spring and summer introducing relatively warm and fresh water from the south. During autumn and winter, the geostrophic current reverses and brings waters, which is cold and more saline towards the study area. The close match between estimated and observed seasonal temperature and salinity supports the conclusion that the observed seasonal T and S signals off Port Sudan are largely the product of local heat and mass flux and alongshore advection.

A study of the seasonal variations of \(A_T\) and DIC off the Sudanese coast (Paper II) shows that \(A_T\) and DIC are high during winter and low during summer with an average annual variability of 40 \(\mu\text{mol kg}^{-1}\) for \(A_T\) and 32 \(\mu\text{mol kg}^{-1}\) for DIC. The seasonal signals are associated with the maxima and minima of salinity described in Paper I. Advection of water, and thus changing salinity is an important factor controlling \(A_T\) variations,
while the observed change in DIC is primarily controlled by air-sea gas exchange, through change in temperature, and, very likely, by biological production. The remaining changes of DIC are caused by alongshore advection during autumn and winter (Paper I). Seasonality for DIC and $A_T$ have been reconstructed using the observed DIC-SST and $A_T$-SSS relationships together with SST satellite data and SSS-advection from Paper I. The calculated results fit the observed variability of DIC and $A_T$. The interannual change of the surface $A_T$ were twice as high as the seasonal variation whereas for surface DIC, the interannual changes were found to be less than the seasonal changes.

An annual cycle of oceanic $pCO_2$ between October 2014 and October 2015 is presented in Paper III. The highest values occur during summer-autumn and lowest during winter-spring, with a seasonal amplitude of approximately 60 μatm overlaid a high frequency signal of about 10 μatm. More than half of the variability of oceanic $pCO_2$ is driven by SST changes, which is in line with findings in Paper II that half of the observed change in DIC is due to temperature driven air-sea CO$_2$ exchange. The $pCO_2$-SST relationship throughout a year has an elliptical shape, which confirms that beside the temperature influence, other processes also contribute in controlling $pCO_2$ changes, e.g. along-coast advection described in Paper I. Based on oceanic $pCO_2$ and atmospheric data, the area is a net annual sink for atmospheric CO$_2$ of size 24.4 mmol CO$_2$ m$^{-2}$ y$^{-1}$. During summer and autumn, the area act as a source for atmospheric CO$_2$, while during winter and spring the area is a sink for atmospheric CO$_2$. The air-sea gas exchange was examined for the period between 1977-2015. It shows that the area most likely switched from being a net annual source area for atmospheric CO$_2$ to becoming a net annual sink during the 2000s.

Most of the work done is pristine and for this reason important, and a useful baseline is made for a region with extreme condition in term of physical, chemical, and ecological properties. Through this baseline, future changes in climate and environment and their predicted trends and impacts can be documented and assessed. In this regard, sustaining the coastal time series initiated by this study will be of extreme importance.
6. Future plans

In light of the vulnerable environment of the Red Sea as well as its natural extreme environmental settings, there are several topics, which are important and interesting to explore in the future. Multiple stressors, e.g. increasing temperature, deoxygenation, and rising atmospheric CO$_2$ concentration, affect the ocean and its ecosystems. Warming of the ocean will e.g. decrease the gas solubility, and thus, reduce oceanic oxygen concentration, which has been observed at several locations (Shepherd et al., 2017, and references therein). However, the current understanding and implications of such a deoxygenation is not yet resolved.

Further, exploration of precipitation and dissolution of calcium carbonate, CaCO$_3$, has not been sufficiently discussed in this thesis due to lack of time and data scarceness. The Red Sea is a region with high CaCO$_3$ production, and recent studies by Takahashi et al. (2014), Steiner et al. (2014), and Elageed et al. (“Oxygen and alkalinity utilization rates in the Red Sea”, manuscript in preparation) show that such a production affects A$_T$ through calcification within the pelagic layer and by corals.

At present, ocean acidification is not an imminent threat for the Red Sea. The area is super saturated with respect to calcium carbonate, and thus has relatively high resistance to Ocean Acidification (Elageed, 2010; Omer, 2010). However, this might change in the future, and continuous monitoring of the marine carbon cycle is important.
7. Bibliography


