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FRANCIS DA SILVA LOPES

METEOROLOGICAL AND OCEANIC PARAMETERS AND THEIR IMPACTS ON THE BIOTA AND CLIMATE IN THE EQUATORIAL ATLANTIC

Recife
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Monografia apresentada ao Departamento de Oceanografia da Universidade Federal de Pernambuco como pré-requisito para a conclusão do curso de Bacharelado em Oceanografia.

Orientador: Profa. Dra. Dóris Regina Aires Veleda

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RESUMO

O objetivo deste trabalho é avaliar o impacto de parâmetros meteorológicos e oceânicos sobre a biota e o clima do Atlântico Equatorial (AE), através da relação entre a profundidade da camada de mistura oceânica (CMO) do Atlântico tropical e a precipitação no leste do Nordeste do Brasil (ENEB) e também avaliar a concentração de clorofila (Chl-a) em resposta à temperatura da superfície do mar (TSM), CMO e padrões de ventos no AE. Para atingir este objetivo, foram realizados: 1) um controle de qualidade no banco de dados oceanográfico dos flutuadores ARGO; 2) determinação da CMO com base nos métodos propostos na literatura; 3) uma identificação do método de CMO mais apropriado para o Atlântico Sudoeste Tropical (SWTA); 4) estimativa e análise da variabilidade espaço-temporal da CMO; 5) identificação da relação entre a CMO e as variáveis meteorológicas sobre a ENEB; 6) uma avaliação do índice climático do Atlântico Niño (ATL3); 7) avaliar a variabilidade espaço-temporal de Chl-a, SST, MLD e ventos sobre EA e 8) avaliar a relação dos índices climáticos com os padrões espaço-temporais dessas variáveis. Para determinar a variabilidade da CMO e como esta afeta os padrões de precipitação na ENEB, foram utilizados dados in situ de 10 anos dos flutuadores ARGO e dados de precipitação da APAC. Para a correlação entre o índice ATL3 e o Chl-a no EA, usamos 20 anos de dados mensais de SST e MLD do Sistema Global de Assimilação de Dados do Oceano (GODAS), as médias mensais das componentes zonal e meridional do vento da reanálise do ERA-Interim e da concentração de clorofila na superfície do mar a partir do sensor de visão ampla do Sea-view (SeaWiFS). Os resultados apresentam um método mais preciso para estimar a CMO na região, a partir na densidade. Detectou-se uma relação entre a variabilidade da CMO e ventos sobre o Atlântico Tropical Oeste com a precipitação em Pernambuco. 20 anos de dados vento, temperatura e correntes mostraram que os eventos do Atlântico Niño, contribuem para reduzir o Chl-a na EA. Por outro lado, durante os eventos do Atlântico Niña a concentração de Chl-a aumenta na parte oeste da bacia, e o transporte para o oeste de Chl-a é mais forte. Este transporte fornece Chl-a para a região oligotrófica ocidental do Atlântico Equatorial.

Palavras-chave: Atlântico equatorial. Variabilidade. Chl-a. CMO. Precipitação. Atlântico niño. ENEB.

ABSTRACT

The objective of this work is the evaluate the impact of meteorogical and oceanic parameters on the biota and climate of the Equatorial Atlantic (EA), through the relation between the mixed layer depth (MLD) of tropical Atlantic and the rainfall on the East of Northeast of Brazil (ENEB) and the Chl-a concentration in response of the sea surface temperature (SST), MLD and winds patters in EA. To achieve this intent, specific goals were performed: 1) a quality control in the oceanographic database of the ARGO floats; 2) a determination of the MLD based on the most commonly methods that have been proposed in the previously studies; 3) an identification of the most appropriate MLD's method for Southwest tropical Atlantic (SWTA); 4) an estimation and analysis of the space-time variability of MLD; 5) identification of the relation between the MLD and the meteorological variables over the ENEB; 6) evaluation of the ATL-3 index; 7) evaluate the space-time variability of the Chl-a, SST, MLD and winds over EA and, 8) assess the relationship of the climatic indices to the space-time patterns of these variables. To determine the variability of the MLD and how its affect the rainfall patterns in the ENEB it was used 10 years of in situ data from ARGO and precipitation data from APAC. For the correlation between ATL-3 index and Chl-a in the EA, it was used 20 years of monthly SST, MLD data from Global Ocean Data Assimilation System (GODAS), and zonal and meridional components of the wind from ERA-Interim reanalysis and sea surface chlorophyll concentration from Sea-viewing Wide Field-of-view Sensor (SeaWiFS). The results show a most accurate method for estimating MLD, based on density. It was detected a relationship between the variability of MLD, winds and the rainfall in Pernambuco. 20 years wind, temperature and ocean currents have showed that the Atlantic Niño events contributed to reduce the Chl-a in the EA. On the contrary, during Atlantic Niña events the Chl-a concentration increases in the basin, and the westward transport of Chl-a is stronger. This transport supplies Chl-a to the western oligotrophic region in the Equatorial Atlantic.

Keywords: Equatorial Atlantic. Variability. Chl-a. MLD. Rainfall. Atlantic niño/niña. ENEB.

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LIST OF COMMON ACRONYMS AND ABBREVIATIONS

EA -	Equatorial Atlantic
TA -	Tropical Atlantic
SST -	Sea surface temperature
Chl-a -	Chlorophyll-a
ACT -	Atlantic cold tongue
ITCZ -	Intertropical Convergence Zone
MLD -	Mixed layer depth
ATL3 -	The Atlantic Niño/Niña index
NEB -	Northeastern Brazil
ENEB -	East of Northeastern Brazil
SWTA -	Southwest tropical Atlantic
GODAS -	Global Ocean Data Assimilation System
NOAA -	National Oceanic and Atmospheric Administration
OAR -	Oceanic and Atmospheric Research
ESRL -	Earth System Research Laboratory
PSD -	Physical Sciences Division
ECMWF -	European Centre for Medium-Range Weather Forecasts
SeaWiFS -	Sea-viewing Wide Field-of-view Sensor
OBPG -	Ocean Biology Processing Group
NASA -	National Aeronautics and Space Administration
MODIS -	Moderate Resolution Imaging Spectroradiometer
CTD -	Conductivity,temperature and detph (instrument)
APAC -	Pernambuco State Agency for Water and Climate
SAA -	South Atlantic anticyclone
ORAS4 -	Ocean Reanalysis System 4
DHN -	Department of Hydrography and Navigation
M1 -	Method 1
M2 -	Method 2
M3 -	Method 3
M4 -	Method 4
M5 -	Method 5
M6 -	Method 5
INMET -	National Institute of Meteorology
PE -	Pernambuco

SUMMARY

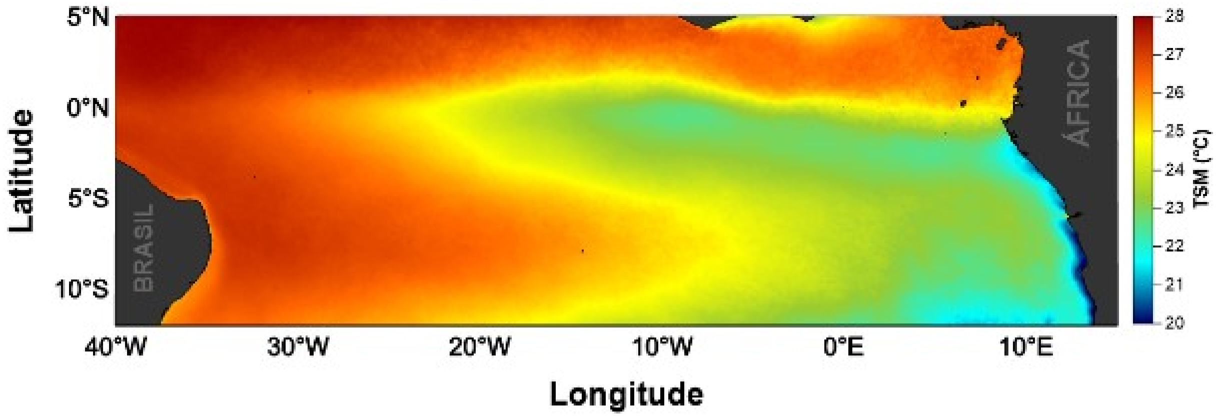
1 INTRODUCTION

In the Equatorial Atlantic (EA), the solar radiation is not a limiting factor for primary production (GRODSKY; CARTON; MCCLAIN, 2008) hence chlorophyll-a (Chl-a) concentration mainly depends on the amount of nutrients available in the ocean mixed layer (LONGHURST, 1993), which in turn is related to the wind patterns and the sea surface temperature (SST). Wind is one of the factors that drives upwelling and vertical mixing into the upper ocean, which consequently controls the ocean mixed layer depth (BABU *et al.*, 2004; KAHRU *et al.*, 2010; SHEN *et al.*, 2018). KAHRU *et al.* (2010), have found that the correlation between Chl-a and wind speed is generally negative in areas with deep mixed layers and positive in areas with shallow mixed layers. In general, the highest values of Chl-a in the eastern tropical Atlantic occur in coastal regions as a result of river discharge and upwelling (GRODSKY; CARTON; MCCLAIN, 2008) In situ data of the EA suggested a seasonal variability of primary productivity in the eastern EA (PÉREZ *et al.*, 2005).

The large pool of cooler SSTs that develops each spring just south of the equator, in the eastern EA, is called the Atlantic cold tongue (ACT). In May–June, the ACT develops rapidly when the Intertropical Convergence Zone (ITCZ) shifts northward over the eastern Atlantic, associated to the intensification of the southeast winds, and reaches its peak in July–August (CANIAUX *et al.*, 2011; OKUMURA; XIE, 2004). The ACT is reduced due to the weakening of the southerly cross–equatorial winds from September onward (DEPPENMEIER; HAARSMA; HAZELEGER, 2016). The southeast trade winds transport the coastal surface waters offshore, via the Ekman transport this induces the upwelling of cold and nutrient-rich subsurface waters (e.g. LUTJEHARMS; MEEUWIS, 1987) to a phytoplankton bloom that consumes nitrate down to the depth where light becomes limiting (PÉREZ *et al.*, 2005). When the trade winds are strong, from May to August, this Ekman transport of surface waters produces a deepening of the thermocline toward the west, resulting in a warm pool as well as a deeper mixed layer depth (MLD), in the western part of the basin, similar to the Bjerknes feedback (SIGNORINI *et al.*, 1999). Several authors have described the activity of Bjerknes feedback in the EA (e.g. DEPPENMEIER; HAARSMA; HAZELEGER, 2016; KEENLYSIDE; LATIF, 2007; LÜBBECKE; MCPHADEN, 2013, 2017; ZEBIAK, 1993).

In the EA (FIGURE 1), from the ocean's surface to the bottom it is possible to detect many different water masses, which are characterized by their specific temperature, salinity and organisms. An important distinction between the MLD and the other deeper layers of the ocean is that, the surface layer responds quickly to changes in meteorological forcings. However, the interior of the ocean, because of the velocity of circulation and high heat capacity of the water, responds more slowly to changes in boundary conditions. Therefore, the MLD has diurnal, seasonal and interannual variations, whereas in the deep ocean the changes occur in interannual, decadal and centennial time scales (MARSHALL; PLUMB, 2008).

FIGURE 1: Study Area



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POLO *et al.* (2015) characterized the interannual variability of SST in the Tropical Atlantic by two main modes: the zonal mode or Atlantic Niño and the meridional mode (CARTON *et al.*, 1996; KUSHNIR *et al.*, 2006; RUIZ-BARRADAS; CARTON; NIGAM, 2000; ZEBIAK, 1993). These main modes are associated to the interannual variations of SST in EA, (CARTON *et al.*, 1996; KUSHNIR *et al.*, 2006; RUIZ-BARRADAS; CARTON; NIGAM, 2000; XIE, S. P.; CARTON, 2004; ZEBIAK, 1993). The meridional mode of the Atlantic is associated with the surface wind speed, SST and evaporation (CHANG; JI; LI, 1997; XIE, S.-P., 1999). The zonal mode is represented by the Atlantic Niño as a result of the air-sea interaction, in which the equatorial thermocline movement and the ocean wave propagation have extreme importance (ZHU; HUANG; WU, 2012). This mechanism is defined as Bjerknes feedback (CARTON; HUANG, 1993; HANDOH; BIGG, 2000; HUANG *et al.*, 1997; XIE, S. P.; CARTON, 2004; ZEBIAK, 1993). The events associated with strong positive and negative interannual anomalies of SST are considered as Atlantic Niño and Niña, respectively (BURLS *et al.*, 2011; MARIN *et al.*, 2009). These anomalies, which are similar to the Pacific El Niño, have been reported im previous studies (BINET; GOBERT; MALOUEKI, 2001; HISARD, 1980; SHANNON *et al.*, 1986).

The Atlantic Niño/Niña can be characterized through the ATL3 index, defined as the anomalies of averaged SST in the region between 20ºW – 0ºE and 3ºS – 3ºN (ZEBIAK, 1993). The different timescales of SST variability in the tropics have a strong effect on the marine ecosystem (LÜBBECKE *et al.*, 2014), and therefore a strong impact on the balance of life in the oceans. The understanding of the mechanisms that control the SST is of high importance improving their predictability (LÜBBECKE; MCPHADEN, 2017), and also for a better management of the ecosystems.

In the past years, several programs have conducted studies in the Tropical Atlantic, for their important role in climate change and seeking a better understanding of the mechanisms that cause climatic anomalies on the Brazilian Northeast (ANDREOLI; KAYANO, 2007). Several studies have already shown the relationship between SST in the Tropical Atlantic and the climate in Northeastern Brazil (NEB) (FOLLAND; PALMER; PARKER, 1986; MOURA, ANTONIO D.; SHUKLA, 1981).

The interannual rainfall variability in the East of Northeastern Brazil (ENEB) is associated to the anomalies of the position and intensity of the ITCZ, which responds to the positive anomalies of the SST of the South Atlantic, according to MOURA et al., (1981) and (NOBRE, 1993). The ITCZ also causes high values of precipitation from February to April (KOUSKY, 1979; RODRIGUES *et al.*, 2011) in the western, northern and central parts of the NEB. However, the main atmospheric phenomenon that causes intense rainfall in the ENEB is the easterly waves, which propagate from the African continent towards the coast of Brazil. Easterly waves are associated with instabilities in the atmospheric pressure fields, by approximately 3000 or 1500 meters (700 or 850 milibars), altering vorticity at these levels and causing convergence of ocean's moisture to high levels (SILVA *et al.*, 2018). These atmospheric instabilities arrive on the coast of the ENEB causing extreme rainfall events. This rainfall variability in the ENEB is mainly controlled by the SST and MLD of the tropical Atlantic.

The MLD thickness is associated with the heat content and the mechanical inertia of the oceanic surface layer, that interacts directly with the atmosphere, for these reasons, estimating the MLD with precision is crucial for understanding the dynamics of the oceans and climate change (CHU *et al.*, 2011).

Also, the MLD is an important control for Chl-a, once the depth of maximum of Chl-a is trapped just below the MLD (Lavigne et al., 2015). Several authors have highlighted the phytoplankton biomass response to climate changes (DUNSTAN *et al.*, 2018; TROLLE *et al.*, 2014; WINDER; SOMMER, 2012). Understanding the chlorophyll variability will improve predictions of how climate change will affect the carbon uptake of the ocean (IRWIN; FINKEL, 2008). Moreover, under this climate change scenario, it also seems of great importance to understand the connectivity between climate and ecological events, which could allow a more efficient forecast of foraging areas and distribution of highly migratory species (MONLLOR-HURTADO; PENNINO; SANCHEZ-LIZASO, 2017).

2 OBJECTIVE

The main objective of this work is the evaluate the impact of meteorogical and oceanic parameters on the biota and climate of the Equatorial Atlantic, thought the relation between the MLD of tropical Atlantic and the rainfall on the ENEB; and the Chl-a concentration in response of the SST, MLD and winds patters in EA. To achieve this intent, specific goals were performed: 1) a quality control in the oceanographic database of the ARGO floats; 2) a determination of the MLD based on the most common methods that have been proposed in the previous studies; 3) an identification of the most appropriate MLD's method for Southwest tropical Atlantic (SWTA); 4) an estimation and analysis of the spatial-temporal variability of MLD; 5) an identification of the relation between the MLD and the meteorological variables over the ENEB; 6) an evaluation of the ATL-3 index; 7) evaluate the Chl-a, SST, MLD and winds space-time variability over EA and 8) assess the relationship of the climatic indices to the space-time patterns of these variables.

3 METHODS

For the correlation between ATL-3 index and Chl-a in the EA, we used 20 years (January 1998 to December 2017) of monthly sea surface temperature (SST) and mixed layer depth (MLD) data from Global Ocean Data Assimilation System (GODAS) a reanalysis database, provided by the NOAA/OAR/ESRL PSD (BEHRINGER *et al.*, 2016), Boulder, Colorado, USA, available from http://www.esrl.noaa.gov/psd/, and with a spatial resolution of 1º by 1º, enhanced to 1/3º in latitude. GODAS has 40 levels with a 10-meter resolution in the upper 200 meters. For the same period, we analyzed monthly zonal and meridional components of the wind, 0.25º x 0.25º, from ERA-Interim reanalysis, produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), at the http://www.ecmwf.int/. We also examined monthly sea surface chlorophyll concentration from Sea-viewing Wide Field-of-view Sensor (SeaWiFS) Ocean Color Data, 9km resolution, from January 1998 to December 2002, and the dataset from Moderate Resolution Imaging Spectroradiometer (MODIS), also with 9km resolution, from January 2003 to December 2017, and provided by Ocean Biology Processing Group (OBPG)/NASA, available at https://oceancolor.gsfc.nasa.gov/.

The monthly Chl-a climatologies were calculated by the monthly average of SeaWiFS and MODIS data, thus obtaining the seasonal patterns of the Chl-a in the EA. The Chl-a monthly anomalies were calculating by removing the monthly mean climatology. Monthly wind climatologies were obtained from the monthly ERA-Interim data, including monthly anomalies for the zonal and meridional components, and surface wind speed. Monthly climatologies of SST and MLD were constructed to characterize their seasonal patterns in the EA. Interannual patterns were observed by calculations of monthly anomalies. The GODAS MLD is based on potential density and can be defined as the depth where the buoyancy difference with the surface level is equal to 0.03 $Kg.m^{-3}$.

In the SWTA, for the detected the MLD, from the *in situ* data, we used the salinity and temperature profiles of the ARGO float and data collected from the CTD in the cruise, performed in the region by the DHN, with these database was possible to evaluate and to determinate which MLD’s calculation method was more effective and cohesive at the region. Data from the Ocean Reanalysis System, version 4 (ORAS4)

(BALMASEDA, 2013), were used to understand the seasonal variability of the MLD and SST; we analyzed monthly zonal and meridional components of the wind, 0.5º x 0.5º from ERA-Interim reanalysis, produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), and http://www.ecmwf.int/ (BERRISFORD et al., 2009), and also monthly precipitation data were obtained from 132 meteorological stations, along the coast of the state of Pernambuco, produced by the Pernambuco State Agency for Water and Climate (APAC), all data were collected for the period of 11 years (January 2000 to December 2010).

4 IMPACT OF ATLANTIC NIÑO AND NIÑA EVENTS ON THE CHLOROPHYLL-a CONCENTRATION IN THE EQUATORIAL ATLANTIC

The first manuscript was submitted to the *Frontiers in Marine Science* periodic, which analyses the connection between chl-a concentration and the ATL-3 index in the Equatorial Atlantic. This manuscript is described below:

Abstract

Chlorophyll-a variability is analyzed in the area 5ºN - 15ºS and 40ºW – 20ºE, using 20 years of monthly MODIS-Aqua satellite data. The zonal response of Chlorophyll-a to physical forcings was analyzed related to sea surface temperature (SST), mixed layer depth (MLD) and surface winds. Climatological monthly patterns show the SST feedback to the winds, evidencing the winds effect on the SST zonal gradient intensification during austral winter. The strongest Chl-a westward transport occurs from June to September, associated to a period of strong SST and MLD zonal gradient, covering part of the western equatorial Atlantic, known as an oligotrophic zone. At interannual timescales, the Atlantic Niño is responsible for a negative anomaly of Chl-a in most of the basin, while during Atlantic Niña years positive Chl-a anomalies reach the western equatorial Atlantic basin, west of 20ºW.

Introduction

Chlorophyll-a (Chl-a) biomass in the surface ocean is regulated by a complex interaction of physiological, oceanographic, and ecological factors, which in turn regulate the rates of primary production (HUNTER-CEVERA *et al.*, 2016; IRWIN; FINKEL, 2008; TRIMBORN *et al.*, 2015). This concept was initially introduced by RILEY (1946), that correlated the phytoplankton biomass as a function of temperature, nutrients, zooplankton and water depth. The concentration of chlorophyll has been widely used as a first-order indicator of the abundance and of biomass of oceanic phytoplankton (BOYCE *et al.*, 2014).

In the Equatorial Atlantic (EA), the solar radiation is not a limiting factor for primary production (GRODSKY; CARTON; MCCLAIN, 2008) hence Chl-a concentration mainly depends on the amount of nutrients available in the ocean mixed layer

(LONGHURST, 1993), which in turn is related to the wind patterns and the sea surface temperature (SST). Wind is one of the factors that drives upwelling and vertical mixing into the upper ocean, which consequently controls the ocean mixed layer depth (BABU *et al.*, 2004; KAHRU *et al.*, 2010; SHEN *et al.*, 2018). KAHRU et al. (2010) have found that the correlation between Chl-a and wind speed is generally negative in areas with deep mixed layers and positive in areas with shallow mixed layers. In general, the highest values of Chl-a in the eastern tropical Atlantic occur in coastal regions as a result of river discharge and upwelling (GRODSKY; CARTON; MCCLAIN, 2008). In situ data of the EA suggested a seasonal variability of primary productivity in the eastern EA (PÉREZ *et al.*, 2005).

The SST cold season in the eastern EA has values below 22°C, from June to October, due to the upwelling and is characterized by maximum Chl-a concentration at the surface. The warm period, from November to May, the SSTs vary between 27°C and 29°C, due to a weakening of the trade winds (CANIAUX *et al.*, 2011; XIE, S. P.; CARTON, 2004b). Several authors cite the existence of large variations in sea surface Chl-a concentration in the EA at seasonal scales (LONGHURST, 1993; LONGHURST *et al.*, 1995; MCCLAIN; FIRESTONE, 1993; RUDJAKOV, 1997; SIGNORINI *et al.*, 1999; YODER *et al.*, 1993). (PÉREZ *et al.*, 2005), studied the seasonality of Chl-a and they found that the primary bloom did indeed occur in boreal summer, the strengthening of zonal winds and also noted the existence of a secondary late-season bloom.

The large pool of cooler SSTs that develops each spring just south of the equator, in the eastern EA, is called the Atlantic cold tongue (ACT). In May–June, the ACT develops rapidly when the Intertropical Convergence Zone (ITCZ) shifts northward over the eastern Atlantic, associated to the intensification of the southeast winds, and reaches its peak in July–August (CANIAUX *et al.*, 2011; OKUMURA; XIE, 2004). The ACT is reduced due to the weakening of the southerly cross–equatorial winds from September onward (DEPPENMEIER; HAARSMA; HAZELEGER, 2016). The southeast trade winds transport the coastal surface waters offshore, via the Ekman transport this induces the upwelling of cold and nutrient-rich subsurface waters (e.g. LUTJEHARMS; MEEUWIS, 1987) starting a phytoplankton bloom that consumes nitrate down to the depth where light becomes limiting (PÉREZ *et al.*, 2005). When the trade winds are strong, from May to August, this Ekman transport of surface waters produces a deepening of the thermocline toward the west, resulting in a warm pool as well as a deeper mixed layer depth (MLD), in the western part of the basin.

At the same time, the thermocline deepening in the western basin and the increase of local winds in the eastern region is similar to the Bjerknes feedback (SIGNORINI *et al.*, 1999). Several authors have described the activity of Bjerknes feedback in the EA (e.g. DEPPENMEIER; HAARSMA; HAZELEGER, 2016; KEENLYSIDE; LATIF, 2007; LÜBBECKE; MCPHADEN, 2013, 2017; ZEBIAK, 1993).

At interannual timescales, a recent study of BOYCE et al. (2014) reveal that average chlorophyll concentrations have declined across most of the global ocean area over the past century.

POLO et al. (2015) characterized the interannual variability of the Tropical Atlantic by two main modes: the zonal mode or Atlantic Niño and the meridional mode (CARTON *et al.*, 1996; KUSHNIR *et al.*, 2006; RUIZ-BARRADAS; CARTON; NIGAM, 2000; XIE, S. P.; CARTON, 2004b; ZEBIAK, 1993). These main modes are associated to the interannual variations of SST in EA, (CARTON *et al.*, 1996; KUSHNIR *et al.*, 2006; RUIZ-BARRADAS; CARTON; NIGAM, 2000; XIE, S. P.; CARTON, 2004b; ZEBIAK, 1993). The meridional mode of the Atlantic is associated with the surface wind speed, SST and evaporation (CHANG; JI; LI, 1997; XIE, S.-P., 1999). The zonal mode is represented by the Atlantic Niño as a result of the air-sea interaction, in which the equatorial thermocline movement and the ocean wave propagation are of extreme importance (ZHU; HUANG; WU, 2012). This mechanism is defined as Bjerknes feedback (CARTON; HUANG, 1993; HANDOH; BIGG, 2000; HUANG *et al.*, 1997; XIE, S. P.; CARTON, 2004b; ZEBIAK, 1993). The events associated with strong positive and negative interannual anomalies are considered as Atlantic El Niño and La Niña, respectively (BURLS *et al.*, 2011; MARIN *et al.*, 2009). These anomalies, which are similar to the Pacific El Niño, have been previously reported (BINET; GOBERT; MALOUEKI, 2001; HISARD, 1980; SHANNON *et al.*, 1986).

The Atlantic Niño/Niña can be characterized through the ATL3 index, defined as the anomalies of averaged SST in the region between 20 º W – 0 ºE and 3ºS – 3ºN (ZEBIAK, 1993). The different timescales of SST variability in the tropics have a strong effect on the marine ecosystem (LÜBBECKE *et al.*, 2014), and therefore a strong impact on the balance of life in the oceans. The understanding of the mechanisms that control the SST is of high importance improving their predictability

(LÜBBECKE; MCPHADEN, 2017), and also for a better management of the ecosystems.

Several authors have highlighted the phytoplankton biomass response to climate changes (DUNSTAN *et al.*, 2018; TROLLE *et al.*, 2014; WINDER; SOMMER, 2012). Understanding the chlorophyll variability will improve predictions of how climate change will affect the carbon uptake of the ocean (IRWIN; FINKEL, 2008).

In this paper, because the ocean circulation is predominantly zonal, we will focus on the impact of the equatorial Atlantic zonal mode on the Chl-a distribution.

Material and Methods

Datasets

We used 20 years (January 1998 to December 2017) of monthly sea surface temperature (SST) and mixed layer depth (MLD) data from Global Ocean Data Assimilation System (GODAS), provided by the NOAA/OAR/ESRL PSD (BEHRINGER *et al.*, 2016), Boulder, Colorado, USA, available from http://www.esrl.noaa.gov/psd/, and with a spatial resolution of 1º by 1º, enhanced to 1/3º in latitude. GODAS has 40 levels with a 10-meter resolution in the upper 200 meters. For the same period, we analyzed monthly zonal and meridional components of the wind, 0.25º x 0.25º, from ERA-Interim reanalysis, produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), and http://www.ecmwf.int/. We also examined monthly sea surface chlorophyll concentration from Sea-viewing Wide Field-of-view Sensor (SeaWiFS) Ocean Color Data, 9km resolution, from January 1998 to December 2002, and the dataset from Moderate Resolution Imaging Spectroradiometer (MODIS), also with 9km resolution, from January 2003 to December 2017, and provided by Ocean Biology Processing Group (OBPG)/NASA, available at https://oceancolor.gsfc.nasa.gov/.

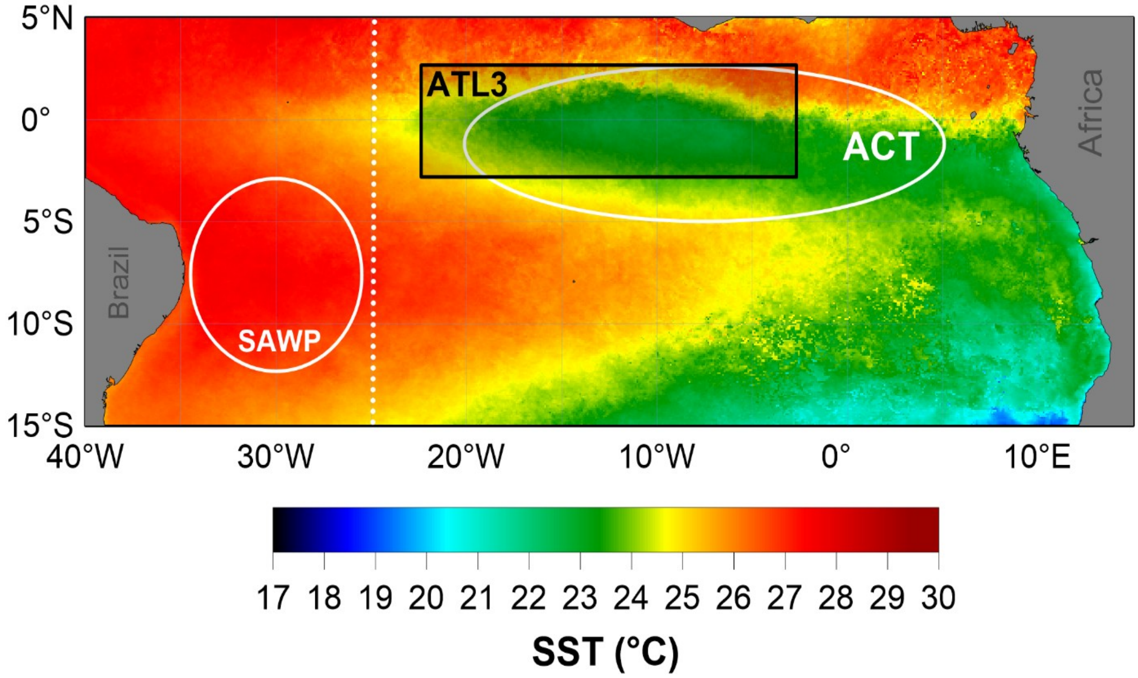
Analysis

The analysis domain is limited by 5ºN - 15ºS and 40ºW – 20ºE. The monthly Chl-a climatologies were calculated by the monthly average of SeaWiFS and MODIS data, thus obtaining the seasonal patterns of the Chl-a in the EA. The Chl-a monthly anomalies were calculating by removing the monthly mean climatology. Monthly wind climatologies were obtained from the monthly ERA-Interim data, including monthly anomalies for the zonal and meridional components, and surface wind speed.

Monthly climatologies of SST and MLD were constructed to characterize their seasonal patterns in the EA. Interannual patterns were observed by calculations of monthly anomalies. The GODAS MLD is based on potential density and can be defined as the depth where the buoyancy difference with the surface level is equal to 0.03 $kg.m^{-3}$.

The western part of the basin, near the northeast coast of Brazil, known as the South Atlantic Warm Pool (SAWP) (CINTRA *et al.*, 2015; HOUNSOU-GBO *et al.*, 2015a) and the eastern part, characterized by upwelling, known as Atlantic Cold Tongue (ACT) (SCHOTT *et al.*, 2004). With the purpose of analyzing the impact of the Atlantic zonal mode on the Chl-a anomalies of the eastern and western sectors of EA, we construct a timeseries of average of Chl-a anomalies in a latitudinal section at 25ºW (FIGURE 2, dashed line). This timeseries is used to analyze possible Chl-a transport to the western equatorial Atlantic, which is considered an oligotrophic area. We divide the EA in two areas, at west (SAWP) and east (ACT) of 25ºW, based on the distinct zonal pattern of SST (FIGURE 2). The box used for calculating the ATL3 index (ZEBIAK, 1993), through of detrended monthly SST anomalies in the region, characterizing the years of Atlantic Niño (positive anomalies) or Niña (negative anomalies). The ATL3 index was calculated considering the period in which the SST anomalies averaged over the ATL3 region exceed the standard deviation of the timeseries multiplied by 0.7 (-0.7). As defined by LÜBBECKE et al. (2010), the Atlantic Niño or Niña events were chosen when the anomalies occur at least three successive months.

FIGURE 2: Study domain, covering the area 5ºN - 15ºS and 40ºW – 20ºE. The color map is the SST averaged from June to August 2009. The dashed white line defines the limit between the influence of the South Atlantic Warm Pool (white line) at west and, the Atlantic Cold Tongue (white line) at east of the basin. The ATL3 index box covers the area 20 º W – 0 ºE and 3ºS – 3ºN (black line).



Source: Own authorship

In order to identify interannual impacts of the Atlantic zonal mode on the Chl-a concentration, we calculated monthly anomalies and constructed annual composites for years of Atlantic Niño and Niña. To understand the pattern of the winds in these years we also calculate composites of monthly wind anomalies.

Results and discussion

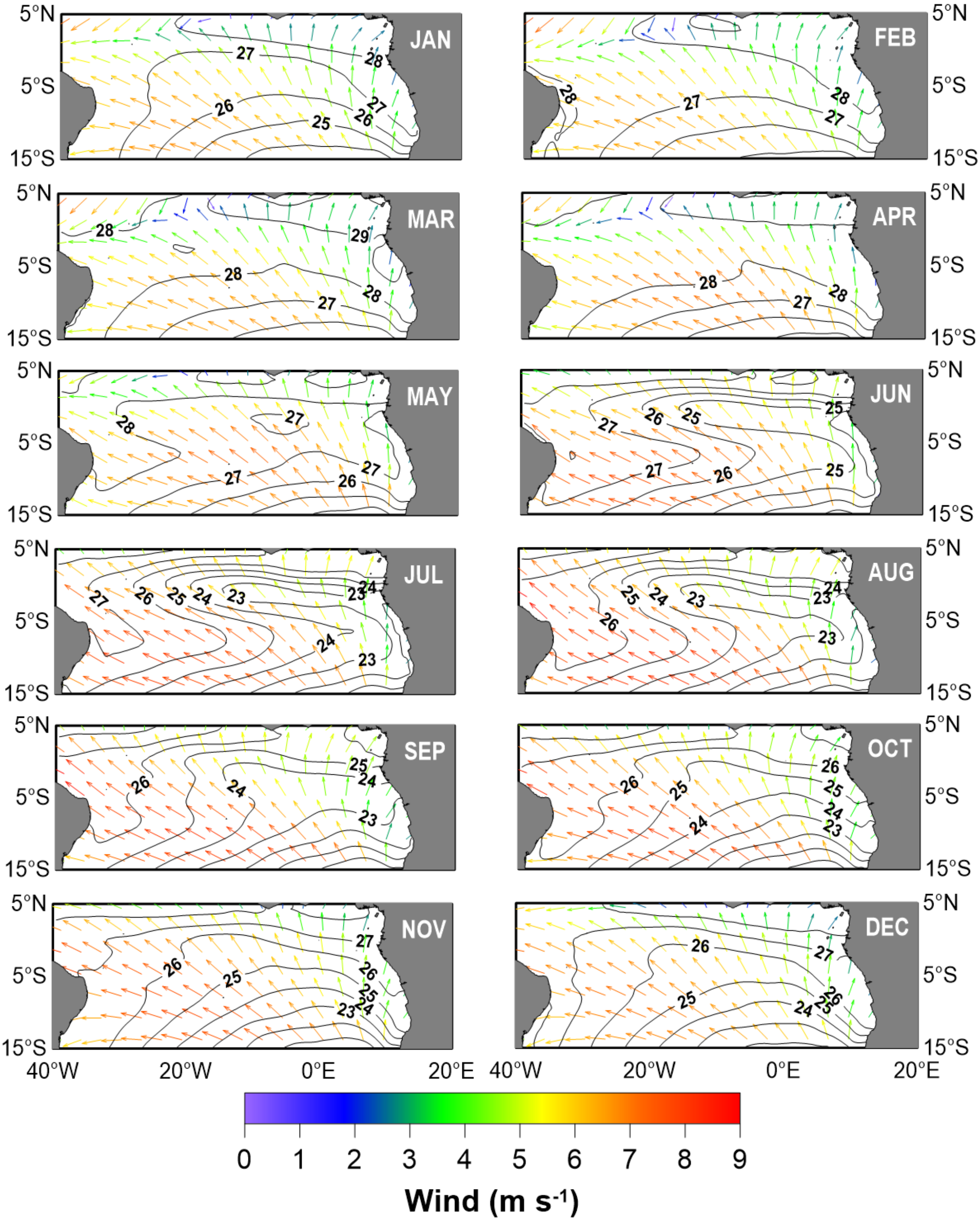
The SST and wind fields in EA are connected by the variability of the South Atlantic subtropical high-pressure system, or South Atlantic anticyclone (SAA). The responses of the wind anomalies to the SAA affect the SST anomalies in the eastern equatorial Atlantic (LÜBBECKE; MCPHADEN, 2017). FIGURE 3 shows the seasonal variability of SST and the wind speed.

The highest SSTs are found between November and April, mainly in the ACT, and from February to May in the western SAWP. The most intense SST zonal gradient is present in June, July and August. Considering our climatology, from 1998 to 2017, the weakest east-west SST gradient is in January to April, as such as the easterly surface winds. From October to January the SST zonal gradient changes to a meridional one. Two wind seasons are associated to these patterns, with the strongest winds from May to October, predominantly from Southeast and the weakest winds from November to March, from east, in the western side of the Atlantic basin. In the eastern EA, the winds are predominantly from the South throughout the year.

The intense zonal SST gradient from June to August (FIGURE 3) is known as the ACT. According to CANIAUX et al. (2011) and LÜBBECKE et al., (2010) the ACT starts its development between March and June, in response to the SST spatial seasonality and depends on the intensification of the SAA.

During the strong winds season, the MLD deepens in the western basin (RUGG; FOLTZ; PEREZ, 2016). At the same time in the eastern basin, the strong wind intensifies the upwelling, as well more productivity (GARCÍA-REYES *et al.*, 2015), which leads to more nutrients and hence more productivity. As a result, Chl-a concentration is higher in the eastern basin from May to October (FIGURE 4). From November to April the mean Chl-a concentration is 0.26 mg m^{-3}, from May to October the mean concentration over the EA is 0.38 mg m^{-3}, which corresponds an increase of 46%.

FIGURE 3: Monthly mean climatology of SST (black line) and winds (colored arrows corresponding to wind speed) constructed over the period from 1998 to 2017.

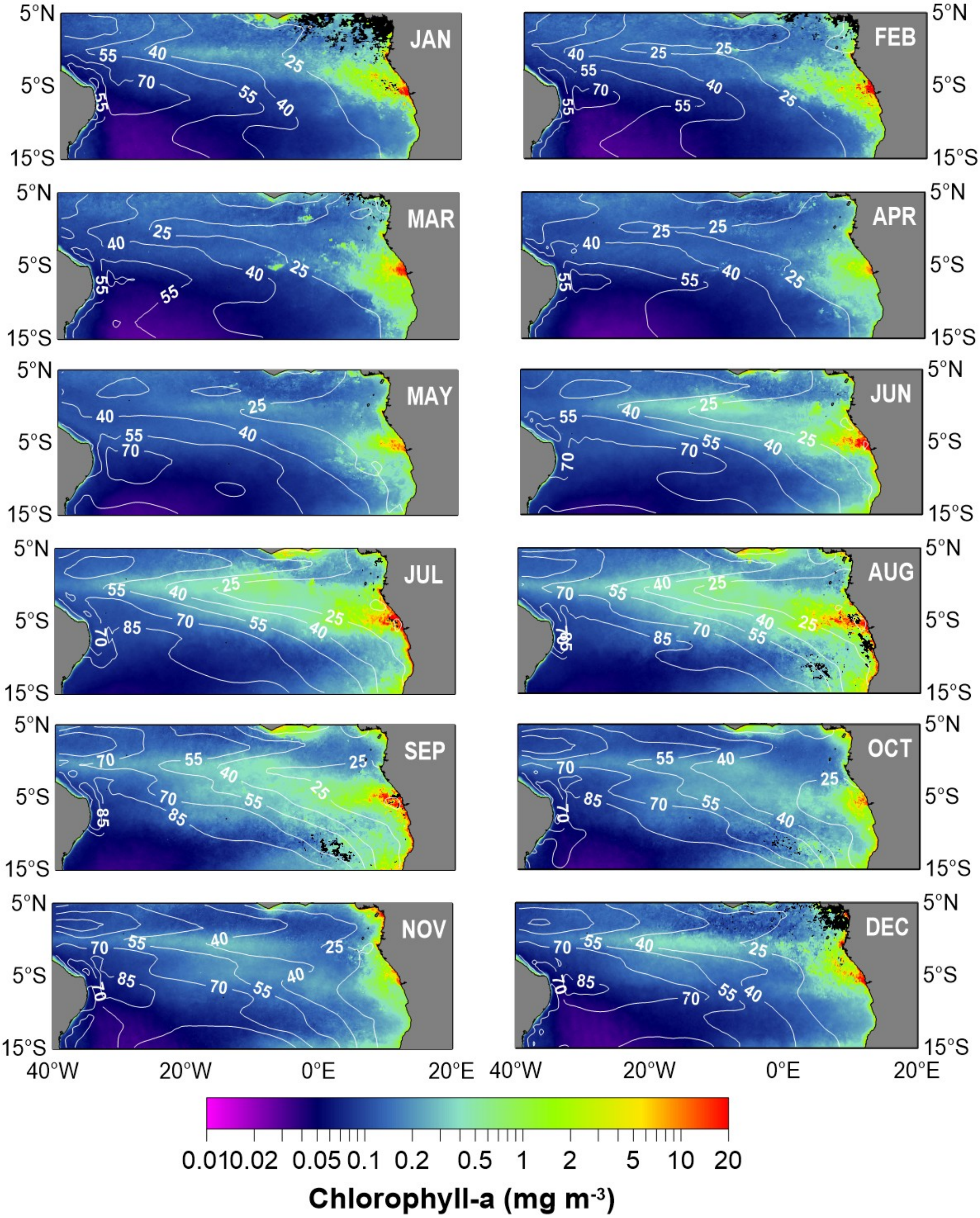


Source: Own authorship.

Despite the continuity of the ACT throughout the year, the westward transport of strongest Chl-a occurs from June to September. This westward transport is associated to a period of high SST (FIGURE 3) and a strong MLD zonal gradient.

The Chl-a follows an axis along the cold tongue (FIGURE 4), covering part of the western EA, known as an oligotrophic zone, with low Chl-a concentration, as identified by VEDERNIKOV et al. (2007).

FIGURE 4: Monthly mean climatology of Chlorophyll-a (mg m-3) and MLD (in white line) over the period from 1998 to 2017.

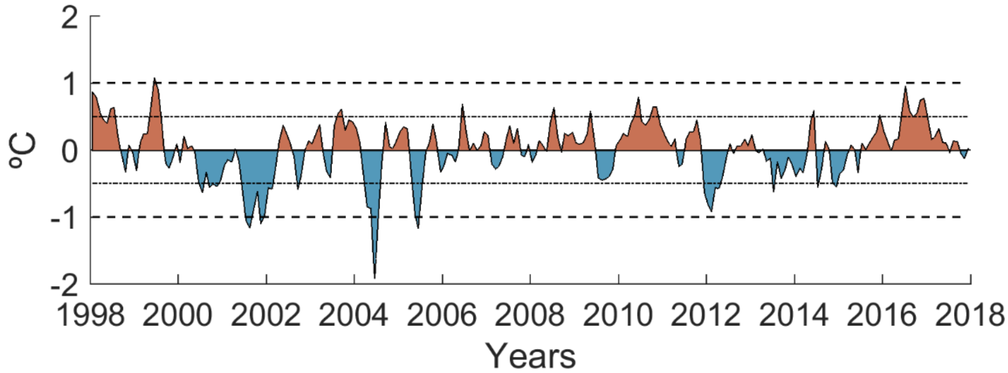


Source: Own authorship.

At the same time, the Congo River plume reaches its maximum extend and Chl-a concentration from May to September, coinciding with the development of the Chl-a maximum offshore (PÉREZ *et al.*, 2005) They found the same pattern of seasonality, with mean values of Chl-a of 0.72 mg m^{-3} at 10ºW and 0.1 mg m^{-3} at 25ºW.

To analyze the interannual variability of the Atlantic zonal mode, we constructed the ATL3 index from 1998 to 2017 (FIGURE 5). In this work we follow the ATL3 index defined in LÜBBECKE et al. (2010). The Atlantic Niño (Niña) events are identified as the periods when the SST anomalies exceed the standard deviation of the timeseries multiplied by 0.7 (-0.7), for at least 3 successive months, in the ATL3 area. Based on this definition the Atlantic Niño years are 1998, 1999, 2003, 2007, 2008, 2010, 2016 and 2017. Atlantic Niña years are 2000, 2004, 2005, 2012 and 2015.

FIGURE 5: ATL3 index constructed over the period from 1998 to 2017.



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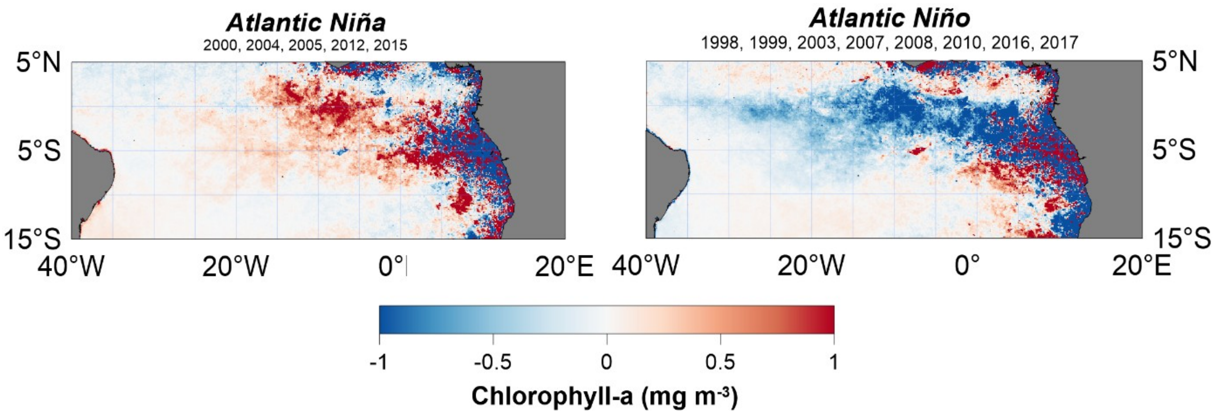
Composites of surface winds and Chl-a for Atlantic Niño and Niña years were performed (FIGURE 6). The FIGURE 7 shows the composites for surface winds anomalies, highlighting distinct patterns in intensity and direction. In Atlantic Niña years the eastern EA presents stronger predominant southeasterly winds and weaker northerly winds in the western EA. In Atlantic Niño years an opposite pattern occurs and, the winds in the eastern basin converge to the African coast.

Similar patterns of winds in Atlantic Niño years was found by NNAMCHI et al. (2015), who show maps of monthly anomalies of SST, heat fluxes and winds regressed on the Atlantic Niño index. Six to two months before the peak phase is reached, the predominant winds are from the Northwest. When the peak phase is reached, the associated westerly and northerly wind anomalies decline and, the loss of heat flux is maximum. After that, the SST decreases, leading to the decay of an

Atlantic Niño event. At the same time, a southerly component of the wind starts to act in the western basin.

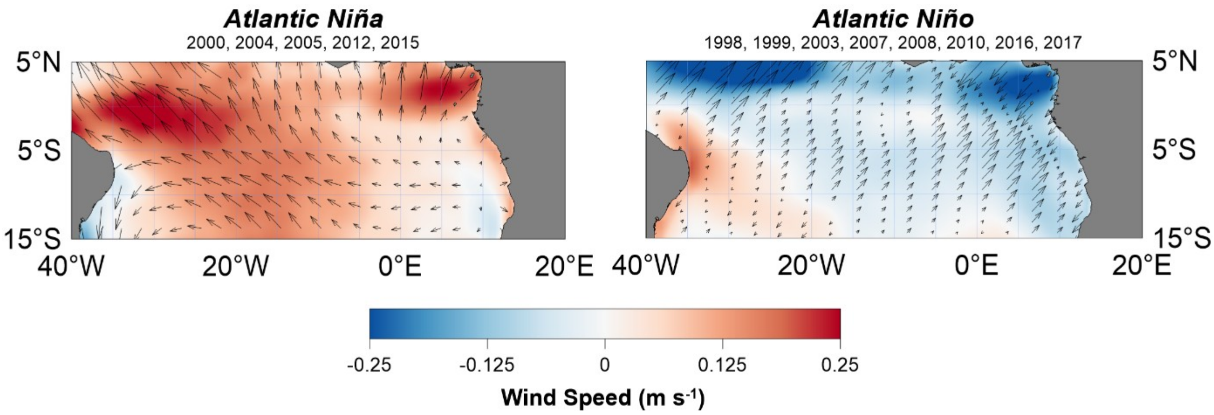
FIGURE 6: Composites of Chlorophyll-a concentration anomalies during Atlantic Niña and Niño years.



Source: Own authorship.

According to NNAMCHI et al., (2016) the Atlantic Niño is not purely an equatorial phenomenon as the SST anomalies extend toward the southeastern Atlantic Ocean. (NNAMCHI; LI; ANYADIKE, 2011) argue that this near-equatorial warming pattern may be associated with cooling of similar magnitude in the southwestern Atlantic during June to August, phenomenon named as the South Atlantic Ocean dipole (SAOD). The negative phase of the SAOD is characterized by an opposite pattern, with cooling anomalies in the Atlantic Niño region and warming over the southwestern Atlantic.

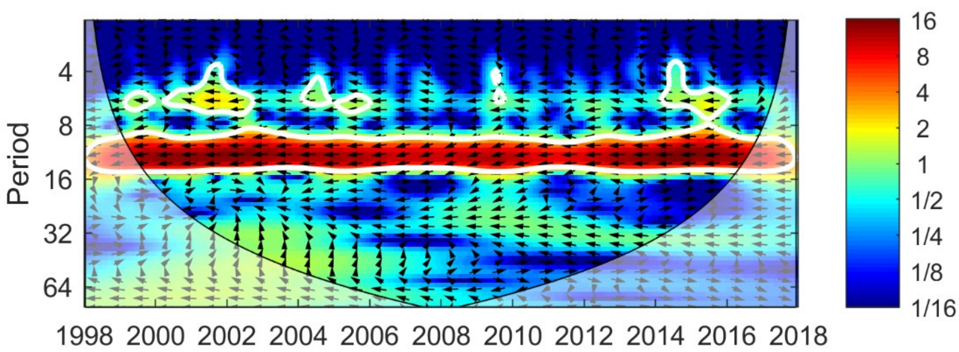
FIGURE 7: Composites of wind anomalies during Atlantic Niña and Niño years.



Source: Own authorship.

FIGURE 8 shows coupled structures are evident at 6-month periodicity during specific years, as in 2000, 2004, 2005, 2012 and 2015. These years are identified as Atlantic Niña events (FIGURE 5).

FIGURE 8: Cross-wavelet between ATL3 index and the averaged Chl-a concentration at 25 º W. The white line means 95% confidence level.



Source: Own authorship.

With the purpose of analyzing the westward propagation of the Chl-a concentration to the oligotrophic part of the EA, we extract a timeseries of the average of Chl-a anomalies in the section at 25ºW. Considering that in the eastern EA, during Atlantic Niño and Niña years, the winds are predominantly southeastward and northwestward, respectively, we analyze the possible influences of the ATL3 in the Chl-a. Through the Cross-wavelet analysis between ATL3 signal and Chl-a anomaly timeseries at 25ºW we identify the main coupled covariance between both variables. The annual signal presents the strongest covariance, which take place at the same time of year, evidencing the coupling mode between the ATL3 and Chl-a (FIGURE 8). The blocked out “white” margins indicate the “Cone of Influence” (COI), where edge artifacts become important (GRINSTED; MOORE; JEVREJEVA, 2004). The color scale for indicates maxima covariance (red) and minima (blue).

The arrows to the left indicate that both signals are coupled but in anti-phase. This means that, when the ATL3 is positive, during Niño years, the Chl-a is negative and during Niña years, when the ATL3 is negative, the Chl-a is positive in the western EA. This relation is caused by the wind patterns in the years of Atlantic Niño/Niña. The southeastward winds in the eastern EA during Niño events push the waters toward the coast, favoring water piling up and weakening of the upwelling. Instead, during Niña years the inverse winds in this area favor the westward Ekman transport and, consequently, the upwelling intensification.

Conclusions

We analyzed 20 years of SST, winds, MLD and Chl-a to understand the impact of the predominant Atlantic mode of variability on the Chl-a distribution in the EA.
During this period, we found predominant Atlantic Niño years, which contribute to reduce the Chl-a in the EA. On the contrary, during Atlantic Niña years the Chl-a concentration increases in the basin, and the westward transport of Chl-a is stronger. This transport supplies Chl-a to the western oligotrophic region.

These results highlight the importance of monitoring the Chl-a in the equatorial Atlantic through a climatic index, as the Atlantic Niño signal, which is the main mode of interannual variability in this region. It is necessary to understand the processes that control the Chl-a variability due to its importance on the marine ecosystems. Indeed, the biomass is the food supply to higher trophic level and an increase of the frequency of Atlantic Niño may affect the fisheries and the ecological balance of the equatorial Atlantic.

5 IMPACTS OF THE SOUTHWEST TROPICAL ATLANTIC ON THE EXTREMES OF PRECIPITATION AND FLOOD IN PERNAMBUCO

The second manuscript is the final report of the BIC-1843-1.08/17, which is a program from FACEPE, and was developed during the years 2017 and 2018. The main objective of this work is to understand the relationship between the MLD in the SWTA and the precipitation in the ENEB. This work is described as follows:

Introduction

The Northeast Brazil (NEB), 47°W-35°W and 18°S-1°S (RAO; DE LIMA; FRANCHITO, 1993) is characterized basically as a semiarid climate and can be divided into different climatic zones with a seasonal variability and year-on-year of the rainfall regime (HASTENRATH, 2012; RAO; DE LIMA; FRANCHITO, 1993). It is also known for large droughts and floods that are associated with global and regional climate phenomena.

The State of Pernambuco is part of the climatic region, knowing as the East of the Brazilian Northeast (ENEB), which presents a rainfall period between March and August, with maximum rainfall centered in the months of May and June, and an annual average above 1500 mm of precipitation. The western boundary of the Southwest Tropical Atlantic (SWTA), near the northeast coast, influences the climate of the ENEB, as the water vapor generated in this region, that is carried to the continent by the southeast trade winds (LUIZ DO VALE SILVA *et al.*, 2018). (HOUNSOU-GBO et al. (2015) showed a positive response of rainfall anomalies in the ENEB region in relation to SWTA sea surface temperature (SST) anomalies.

In past years, several programs have conducted studies in the tropical Atlantic (TA), for its important role in climate change and, to understand the mechanisms that cause climatic anomalies on the Brazilian Northeast (ANDREOLI; KAYANO, 2007). Other studies have already shown the relationship between SST in the TA and the climate in the NEB (FOLLAND; PALMER; PARKER, 1986; MOURA, ANTONIO D.; SHUKLA, 1981).

The SST's seasonal variability in the TA is similar to the pattern in the Pacific (DING; WANG, 2005), however the dynamics and interaction are different. In the TA, the standard variability does not only occur on the equator, but it extends through mid-latitudes, thus may change the SST and the wind regime in both hemispheres (DING; WANG, 2005; NOBRE *et al.*, 1996; PHILANDER; PACANOWSKI, 1986; XIE, S. P.; CARTON, 2004a).

The interannual variability of rainfall in the ENEB is also associated with anomalies in the position and intensity of the Intertropical Convergence Zone (ITCZ), caused by positive anomalies in the SST in the South Atlantic, according to MOURA, and SHUKLA, (1981) and (NOBRE, 1993). The ITCZ also causes rainfall maxima from February to April (KOUSKY, 1979; RODRIGUES et al., 2011) in the western, northern and central parts of the NEB. However, the main atmospheric phenomenon that causes intense rainfall in the ENEB are the easterly waves, which propagate from the African continent towards the coast of Brazil. These phenomena are associated with instabilities in the atmospheric pressure field by approximately 1500 m, altering vorticity at these levels and causing convergence of ocean moisture to high levels (LUIZ DO VALE SILVA *et al.*, 2018). These atmospheric instabilities arrive on the coast of the ENEB causing extreme rainfall events. This rainfall variability in the ENEB is mainly controlled by the SST and the TA's mixed layer depth (MLD).

The MLD plays a key role in the variability of precipitation in the ENEB. The NEB presents a large spatial and temporal variability of rainfall, as a result of different atmospheric forcings (CHAVES; ALBUQUERQUE CAVALCANTI, 2001; HASTENRATH, 2012; KOUSKY, 1979; RAO; DE LIMA; FRANCHITO, 1993), it is of extreme importance to study and understand the oceanic and atmospheric phenomena, which are associated with the rainfall patterns at the ENEB. According to TALLEY et al. (2011), observations of the MLD and the understanding of how it develops seasonally and at different time scales are important for modeling and understanding the climate.

The MLD responds to turbulent processes in the upper oceans and, is responsible for the transfer of mass, momentum and energy, and acting as energy source of the most dynamics movements in all the oceans. The MLD thickness defines the heat content and mechanical inertia of the ocean superficial layer, that interacting directly with the atmosphere. Estimating the MLD with precision and objectivity is crucial for understanding the dynamics of the oceans and climate change (CHU *et al.*, 2011).

The present study analyzes *in situ* data, from ARGO floats (ROEMMICH *et al.*, 2009), and reanalysis data, from ORAS4 and ERA-Interim, to achieve how the MLD is associated with the rainfall regime of coastal region of the ENEB. The depth of the thermocline and the pycnocline and the seasonal variation of both were also analyzed. The rainfall data were obtained from meteorological stations of the Pernambuco State Agency for Climate and Water (APAC) during the years 2000 to 2011.

Objective

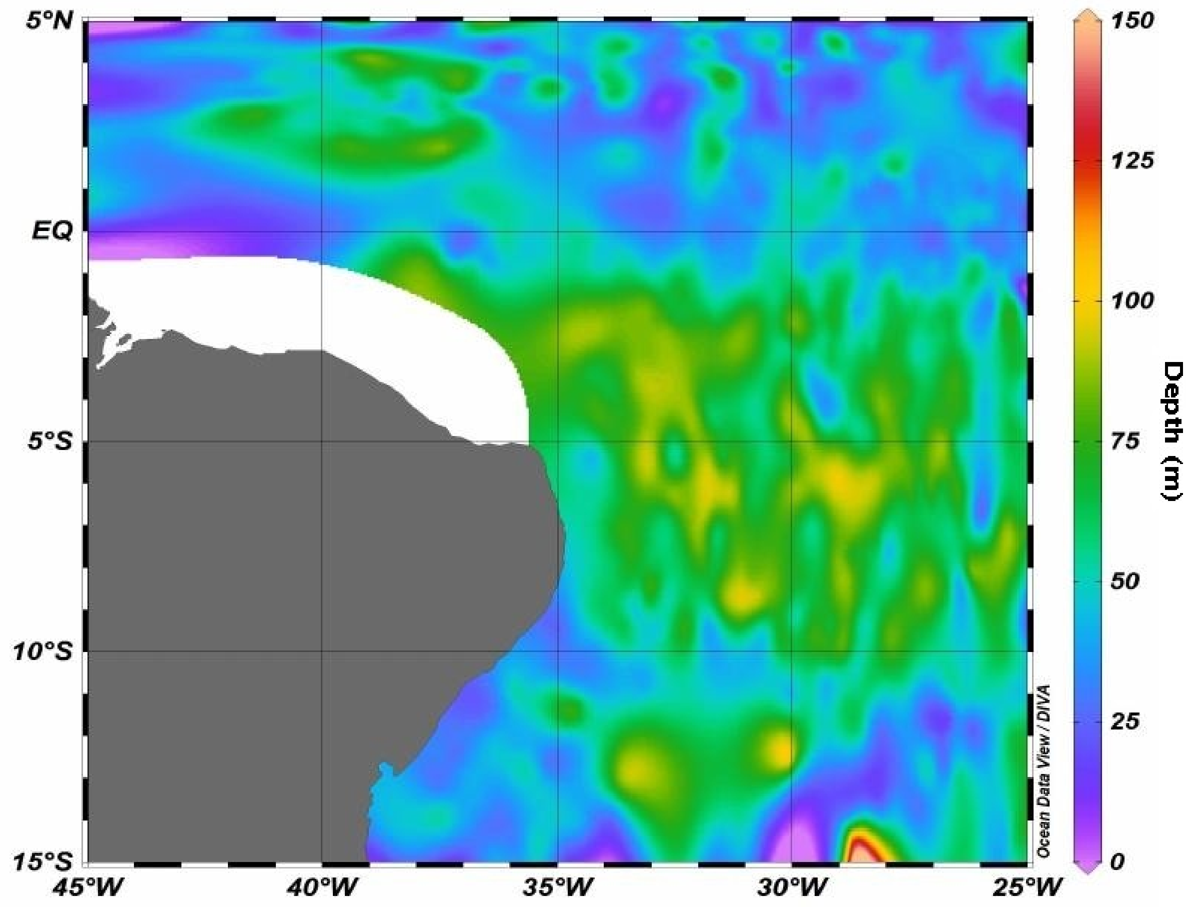
The main objective of this work is to evaluate the impact of the MLD in the SWTA in the rainfall on the ENEB. To achieve this intent, specific goals were performed: 1) a quality control in the oceanographic database of the ARGO floats; 2) a determination of the MLD based on the most commonly methods that have been proposed in the previous studies; 3) an identification of the most appropriate MLD's method for Southwest tropical Atlantic (SWTA); 4) an estimation and analysis of the spatial-temporal variability of MLD; 5) an identification of the relation between the MLD and the meteorological variables over the ENEB.

Material and Methods

Study Area

The present study investigated the Southwest tropical Atlantic (SWTA) region (FIGURE 9), between 5º N - 15º S and 45º W - 25º W, in the near of the Brazilian Northeast, during the years of 2000 to 2011. This study investigated MLD in the SWTA at the limits of this area were based on the classification of SPALDING et al. (2007) in Marine Ecoregions of the World: The Bioregionalization of Coastal and Shelf Areas. The analysis was realized with monthly average data for a better climatological analysis of the region.

FIGURE 9: Study Area. Quarterly means (September, October and November), from the depth of the 2001 MLD to 2011.



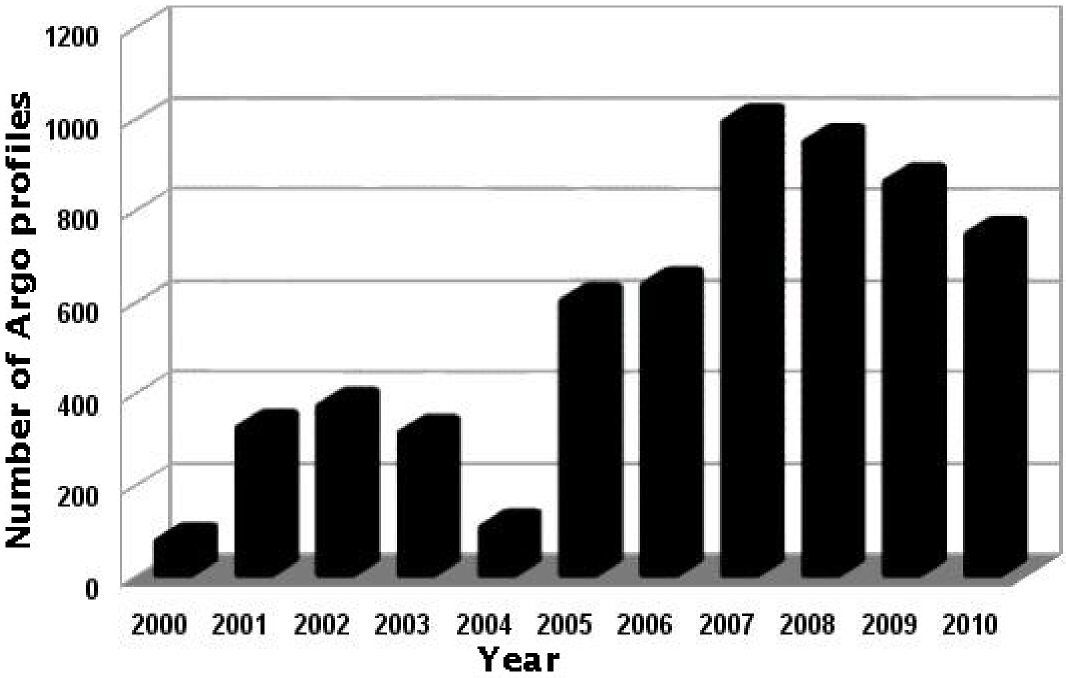
Source: Own authorship.

From the surface to the bottom of the ocean in the study region, it is possible to detect large masses of water, which are characterized by their specific temperature and salinity and organisms found within them. An important distinction between the MLD and the deepest part of the ocean is that the first one responds quickly to changes in atmospheric forcing. However, in the deep ocean, because of the very slow circulation and high heat capacity of the water, the deeper layer responds more slowly to changes in surface conditions. Therefore, the MLD has diurnal, seasonal and interannual variations, whereas in the deep ocean the changes occur in interannual, decadal and centennial time scales (MARSHALL; PLUMB, 2008).

Database

In order to calculate the MLD, from the *in situ* data, were used the salinity and temperature profiles of the ARGO floats and data collected from CTDs in the cruise performed in the region, with this database was possible to evaluate and to determinate which MLD's calculation method was more effective and consistent with the region. The ARGO profile numbers collected during the period can be observed in FIGURE 10.

FIGURE 10: Bar Graffic - number of profiles (temperature and salinity) of the ARGO floats after quality control.



Source: Own authorship.

Monthly data from the Ocean Reanalysis System, version 4 (ORAS4) (BALMASEDA; MOGENSEN; WEAVER, 2013), were used to understand the seasonal variability of the MLD and SST, considering that the number of profiles collected by the floats after quality control was much lower before 2005, from this year onwards the data became more frequent and with a better quality. Monthly climatologies of MLD and SST and anomalies were performed for the study area.

Were analyzed monthly zonal and meridional components of the wind, with a resolution of 0.5º x 0.5º, from ERA-Interim reanalysis, produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), and http://www.ecmwf.int/

(BERRISFORD et al., 2009), and also monthly precipitation data were obtained from 132 meteorological stations, along the coast of the State of Pernambuco, produced by the Pernambuco State Agency for Water and Climate (APAC). All data were collected for the period of 11 years (January 2000 to December 2010), and monthly climatology and anomalies were performed for the winds and precipitation.

Through computational scripts, a quality control of the data was performed, as was select the profile with the lowest number of errors, during the gathering of the Argo floats. The data chosen after the quality control process were stored according to the year, month and local.

Analysis

The theoretical definition of MLD is quite diverse, since it is arbitrary and can be obtained by different methods (DE BOYER MONTÉGUT *et al.*, 2004; HOLTE; TALLEY, 2009). In this work it was evaluated different threshold methods, according to TABLE 1, using 6521 profiles of the ARGO floats during the years of 2000 and 2011. To calculate the MLD it was applied different threshold method with finite difference criteria from a reference value in the near of the surface.

TABLE 1: Different threshold methods with specific criteria for calculating the depth of MLD. With the respective authors, depth of reference, profiles and method for choosing the criterion.

Authors and study area	Profiles	MLD – Threshold method	Ref. Depth	Selection of the criterion	Shortcut
***Weller and Pueddeman* (1996) – North Pacific**	Individual	$\Delta\sigma_\theta$ = 0.03 $Kg.m^{-3}$	10	Arbitrary	M1
***De Boyer Montegut et al.* (2004) - Global**	Individual	ΔT = 0.2 ° C	10	*In situ* data from several profiles	M2
***Obata et al.* (1996) - Global**	Average	ΔT = 0.5 ° C	0	Arbitrary	M3
***Kara et al.* (2006) - Global**	Average	ΔT = 0.8 ° C	10	Statistical comparison between several oceanographic station	M4
***Suga et al.* (2004) – North Pacific**	Individual	$\Delta\sigma_\theta$ = 0.125 $Kg.m^{-3}$	10	Arbitrary	M5
***Monterey and Levitus* (1997) - Global**	Average	$\Delta\sigma_\theta$ = 0.125 $Kg.m^{-3}$	0	Characteristic of the North Subtropical Atlantic Modal Water	M6

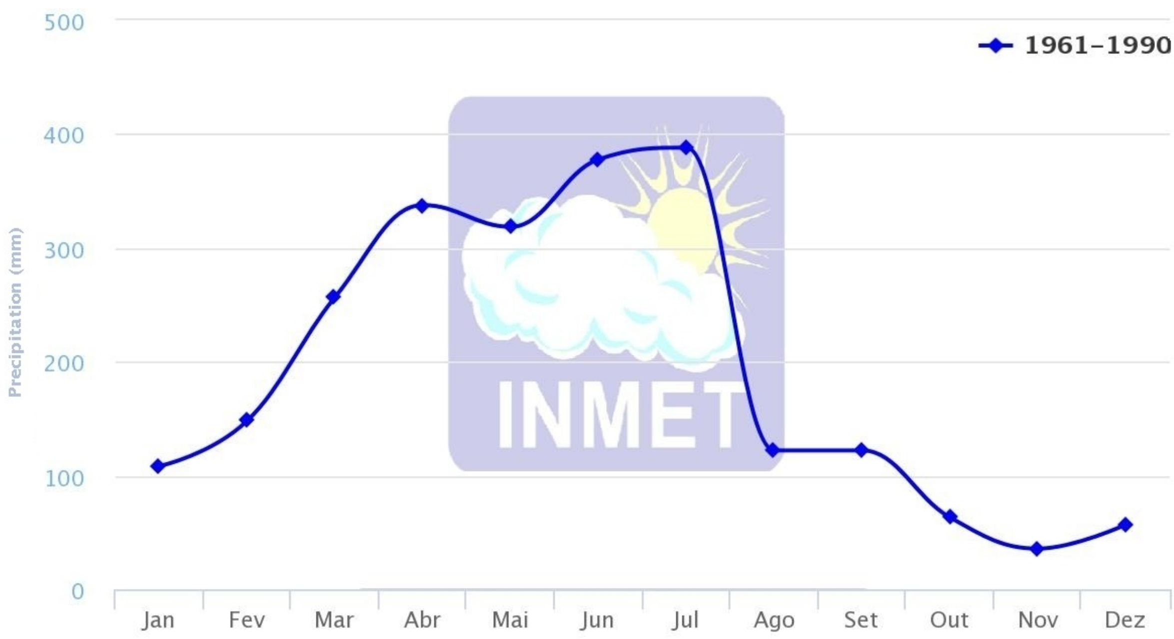
Source: Own authorship.

For the threshold method is required a fixed reference depth and a finite difference value from the density and/or temperature at the fixed depth determined. An example was the method chosen by DE BOYER MONTÉGUT et al., (2004), which fixed the depth at 10 meters below the surface of the sea, to avoid influence of the strong surface diurnal cycle of the ocean. The temperature criterion is defined as ΔT = 0.2 ° C absolute difference, from the surface, and the calculation of the MLD was from the following equation:

MLD_DT002 = depth where ($\theta = \theta_{10m} \pm 0.2$ ° C), Where θ is the potential temperature. For better results using data from ARGO floats, we used the hybrid method, according to HOLTE and TALLEY (2009), in which all the methods of the TABLE 1 were calculated for each profile of temperature and salinity.

With data from CTD, obtained through oceanographic cruises from the Department of Hydrography and Navigation (DHN) of the Brazilian Navy, it was analyzed each of the methods in different seasons of the year, to choose the best method for the region of study.

FIGURE 11: Cumulative Precipitation (mm) - Climatology (1961-1990) - Recife / PE, for theNational Institute of Meteorology (INMET)



Source: INMET, 2018.

To select the best method for the region, it was used the limits of the top the thermocline and pycnocline, inferring that the MLD, because it is a homogeneous and well a mixed superficial layer, would coincide with the depth of one of the two, and always be above of them. The thermocline is characterized by a strong temperature gradient, whereas the pycnocline is characterized by a strong density gradient.

According to MIGNOT et al. (2007), the top of the thermocline depth was defined as the depth where the temperature changes $\Delta T = 0.2$ ° C, relative to the temperature at 10 meters in the water column. The surface layer of the ocean above the thermocline is warmer when it is compared with o the ocean below. However, this layer above the top of the thermocline is not necessarily a homogeneous or isothermal layer, due to the potential salinity effects on the density (DE BOYER MONTÉGUT *et al.*, 2004; MIGNOT, J.; DE BOYER MONTÉGUT; TOMCZAK, 2009; MIGNOT, JULIETTE *et al.*, 2007).

Bearing in mind the potential temperature, the potential density does not increase uniformly with depth, both have a similar pattern in the vertical structure, there is a shallow upper layer of nearly uniform density, followed by a layer in which the density rapidly increases with depth, termed pycnocline, analogous to the thermocline (TALLEY *et al.*, 2011).

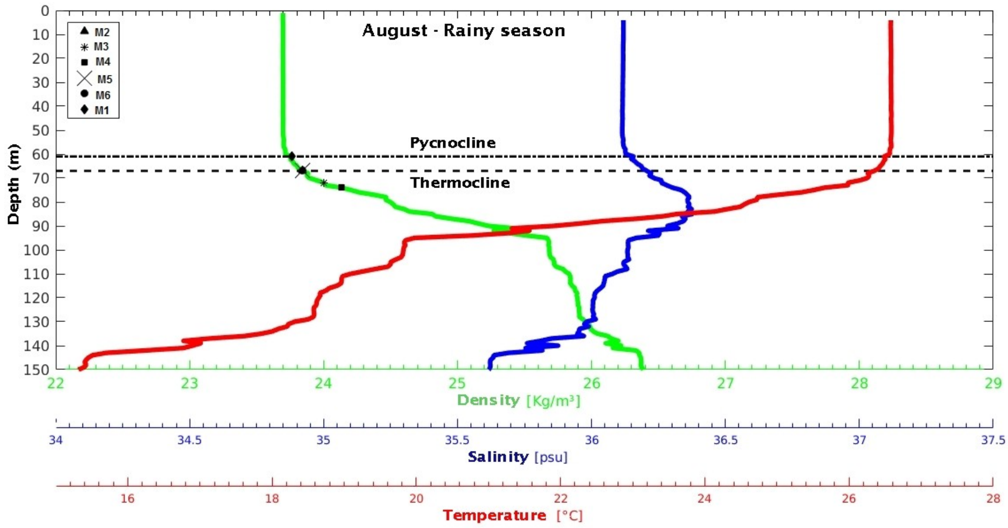
The statistical analyzes were performed with monthly averages within the seasonality of the dry and rainy seasons. The determination of the dry and rainy seasons was based on the precipitation climatology data of the Metropolitan Region of Recife (PE), according to the National Institute of Meteorology (INMET), between 1961 and 1990 (FIGURE 11). After construct the monthly climatology, it was defined the rainy season between March and August and the dry season from September to February.

Results and discussion

The first stage of the work was to define which method would be used to determine the MLD, we compare the different mixed layer depths obtained through several methods, using the ARGO float profiles and four CTD stations, and analyze the behavior of MLD in parallel to the depth of the top of thermocline and pycnocline.

The figures 12, 13, 14 and 15 show different CTD profiles. The figures 12 and 13 the data were carried out during the rainy season, while 14 and 15 were during the dry season. Regardless of the season, M1 showed to be more coherent for the determination of the MLD, since it was always between the depth limits of the pycnocline and thermocline, which are known depths due to its high gradient, means that a homogeneous layer should be above these gradients.

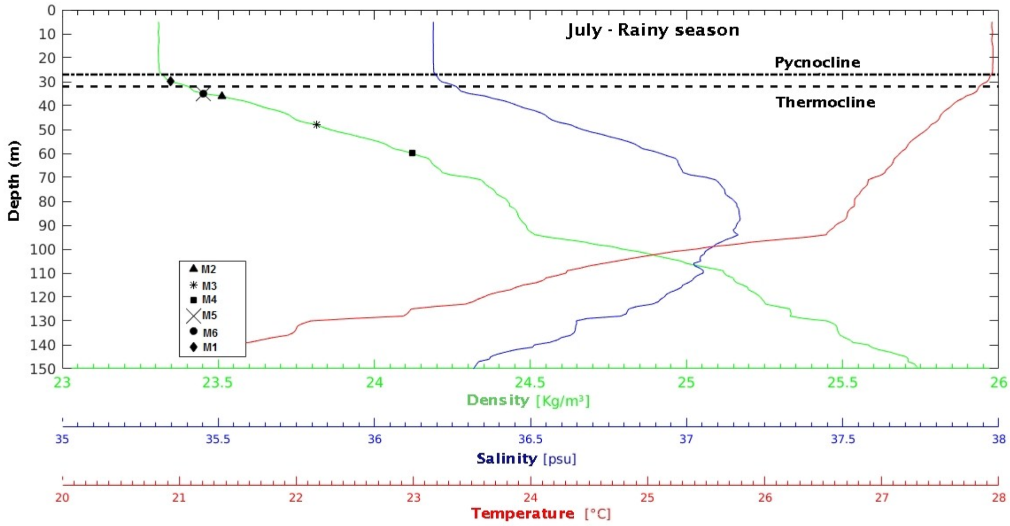
FIGURE 12: CTD profile of oceanographic cruise by DHN (August 2014 03ºS and 33ºW). Profiles of salinity, temperature and density were plotted, as well as depths: of the MLD according to each method, the thermocline and the pycnocline.



Source: Own authorship.

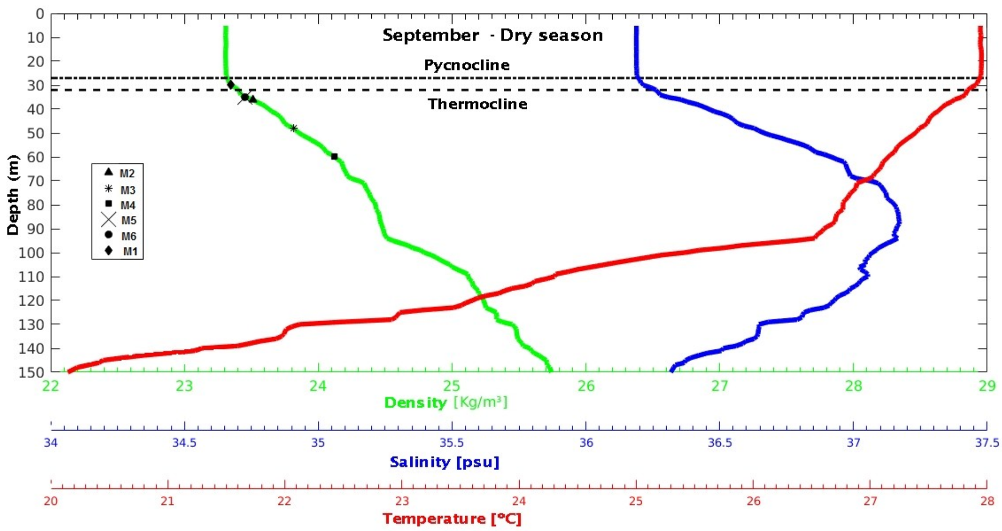
The pycnocline acts as a barrier of the vertical transport for the water and its property, because the pycnocline is very stable, a large amount of energy is necessary for the vertical displacement of particle at the boundary regions. Therefore, turbulence, which causes most of the mixture between different water masses, is less able to penetrate through the stable pycnocline than through less stable layers (TALLEY *et al.*, 2011).

FIGURE 13: CTD profile of oceanographic cruise by DHN (July 2010 06ºS and 34ºW). Profiles of salinity, temperature and density were plotted, as well as depths: of the MLD according to each method, the thermocline and the pycnocline



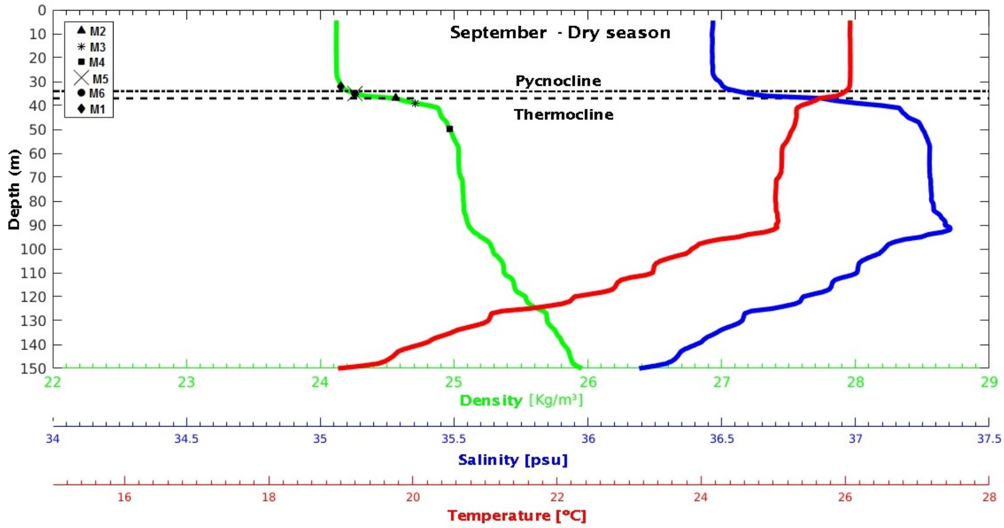
Source: Own authorship.

FIGURE 14: CTD profile of oceanographic cruise by DHN (September 2012 07ºS and 34ºW). Profiles of salinity, temperature and density were plotted, as well as depths: of the MLD according to each method, the thermocline and the pycnocline



Source: Own authorship.

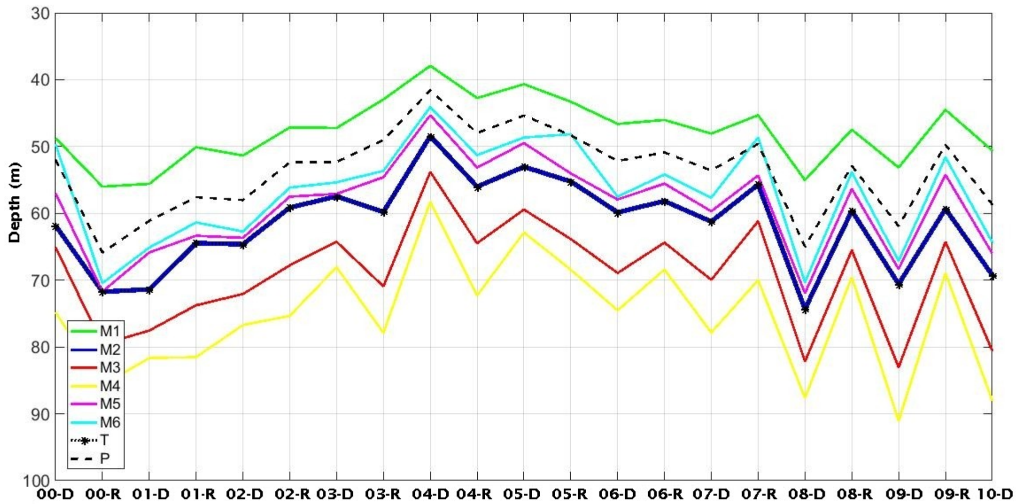
FIGURE 15: CTD profile of oceanographic cruise by DHN (September 2012 07ºS and 34ºW). Profiles of salinity, temperature and density were plotted, as well as depths: of the MLD according to each method, the thermocline and the pycnocline



Source: Own authorship.

The permanent thermocline, in the tropical regions, avoids the mixing of surface and deep waters (SPRINTALL; TOMCZAK, 1992). This buffering effect is also known for its influence on primary production, in the low latitude regions the sunlight is available throughout the year, the solar radiation is not a limiting factor for primary production (GRODSKY; CARTON; MCCLAIN, 2008) hence primary production mainly depends on the amount of nutrients available in the ocean mixed layer (LONGHURST, 1993), in this region productivity is limited because the thermocline prevents replenishment of nutrients from deeper waters (TRUJILLO; THURMAN, 2008).

FIGURE 16: MLD Variability by different methods, during dry and rainy periods between 2000 and 2010. (R) for rainy and (D) for dry. Depth and variability of thermocline and pycnocline can be observed.



Source: Own authorship.

Figure 16 shows the seasonal variability of MLD by the different methods (TABLE 1), thermocline and pycnocline between 2000 and 2010, with seasonal means by rainy (R) and dry (D) season. The same patterns was observed by the CTDs analysis, M1 is the most accurate method for estimating MLD. During the analyzed period, M5 and M6, both density-based methods, were above the thermocline; on the other hand M3 and M4, temperature-estimated methods were below the thermocline and pycnocline.

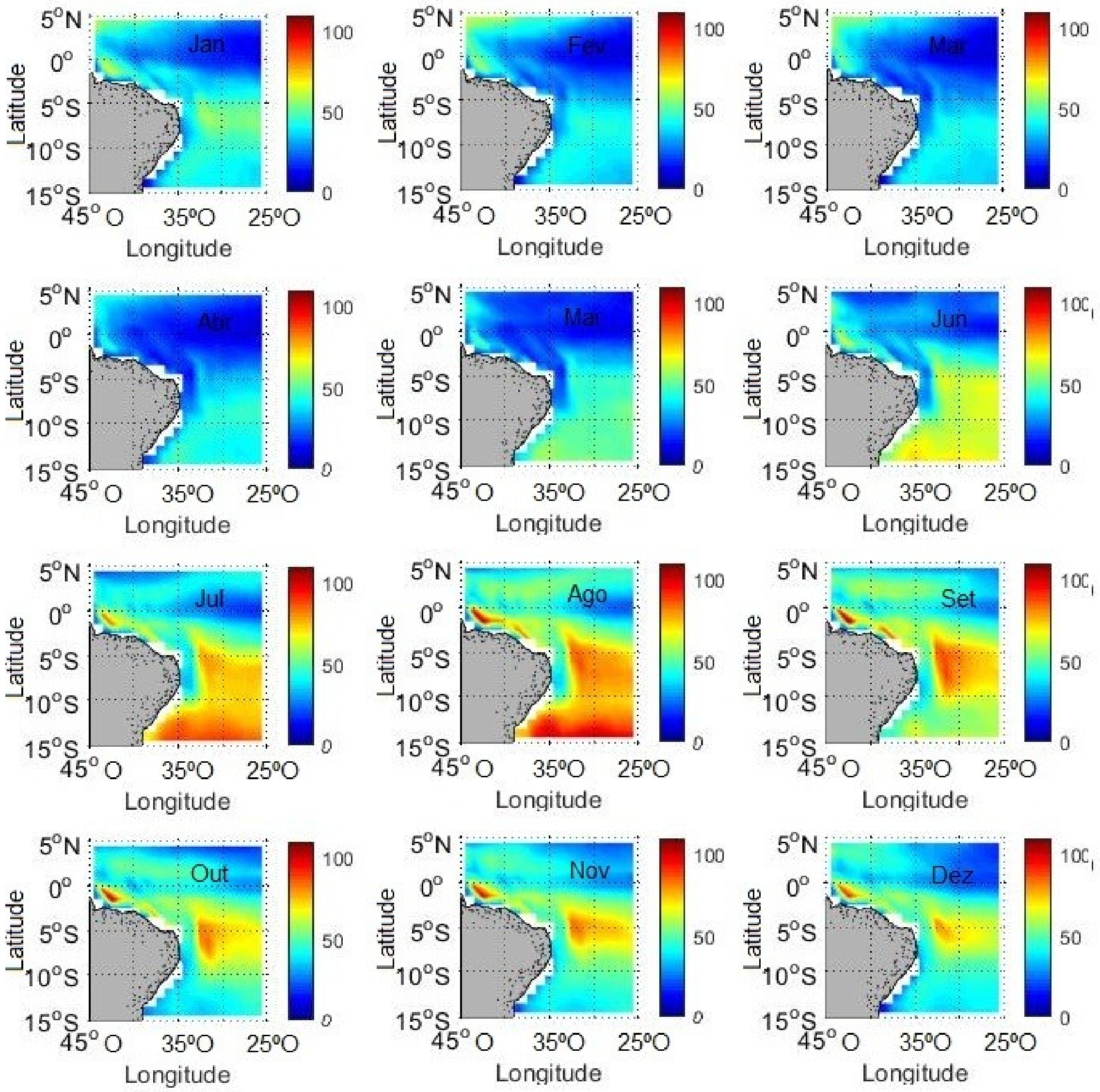
The M2 coincided throughout the period with the depth of the thermocline, this happens because the methodology for the calculation is almost the same, although a difference is observed in the CTD profiles, this is a consequence of the data gathering methodology of the floats ARGO, which collects information spaced every 20 meters, interpolating the temperature value in the profile.

The threshold methods are limited by their dependence on the values of temperature or density at the reference depth, and temperature-based threshold methods tend to overestimate the MLD (LUKAS; LINDSTROM, 1991; SPRINTALL; TOMCZAK, 1992), which is consistent with the results found in the analyzes of the study area.

Although M1 is more effective, or consistent with the real MLD, many profiles collected in the ocean present only temperature data and, that makes methods based on temperature more effective for long time-series studies (LORBACHER *et al.*, 2006), so M2 was also tested for correlation with the precipitation of the coast of Pernambuco.

Based on analyses we decided about the better MLD’s methods for the region, and whit the ORAS4 data for M1 and M2 for the period we performed a time-series analyses, in FIGURES 17 and 18 we have the monthly climatology for the MLD based on M1 and M2 respectively. Both figures show the seasonal variability observed in the analysis whit the *in situ* data, with the months of July and August having the thickest MLD in the region, and in the months of March and April, the MLD is shallower, on average.

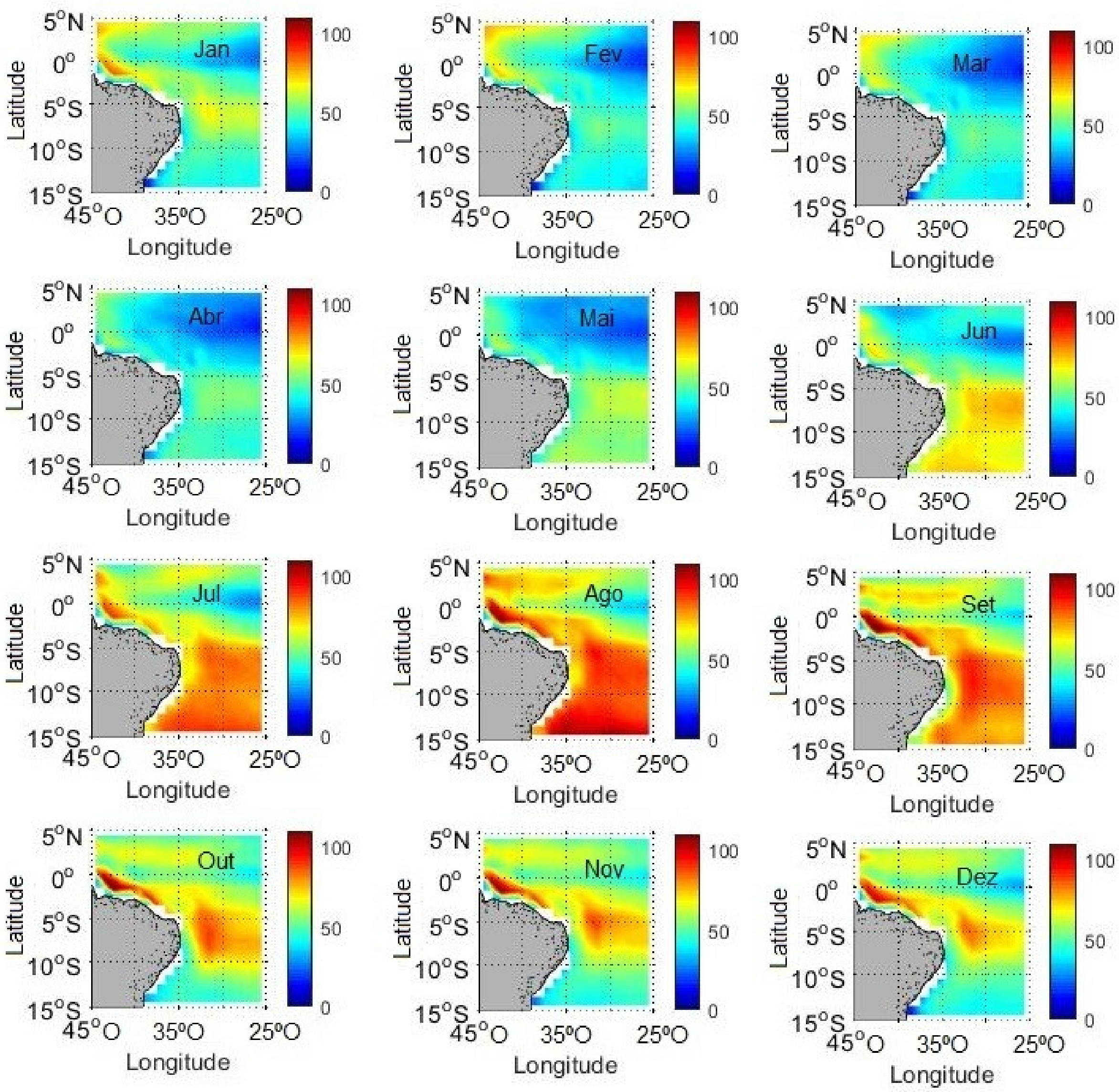
FIGURE 17: Monthly climatology (2000 to 2010) of the MLD, by the M1 method, in meters. ORAS4 database.



Source: Own authorship.

The results shows that the MLD presents a seasonality, thickest in the winter (or rainy season), about 60 to 80m, and decreases during the summer (or dry season), about 30 to 60m. TALLEY *et al.* (2011), have found the same patterns of MLD, where the authors verified that the MLD is thicker in the winter, since it accumulates months of cooling increasing the density and deepening the MLD through the wind patterns. This seasonal variability shows the ability of the MLD to act as an indicator of the ocean heat storage and, also a possible forcing of atmospheric instabilities on the western side of the TA. The ocean has a large heat capacity, which allows it to store the summer heat and, reaches the maxima of heat content in the months that coincide with the highs of the rainfall.

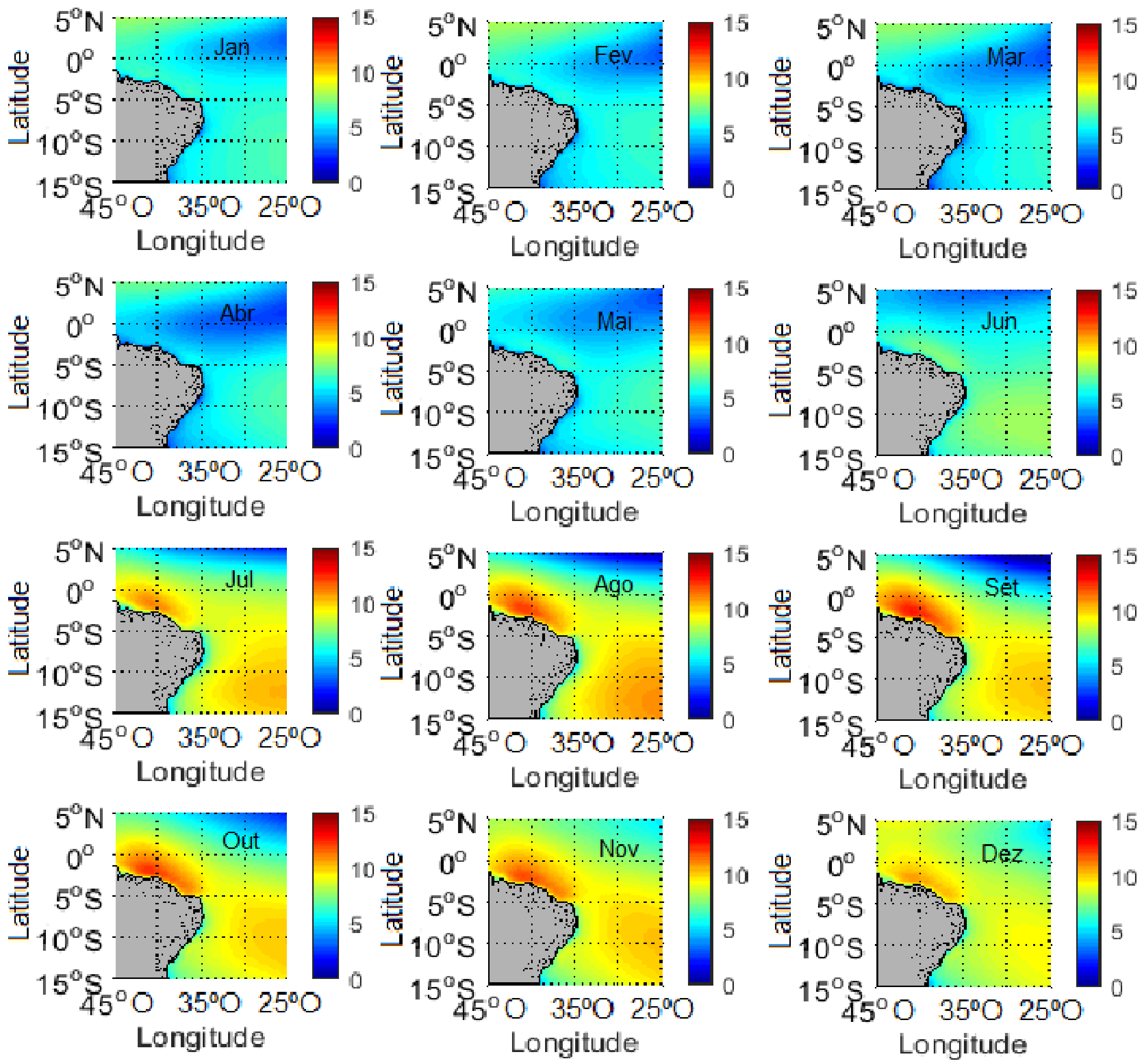
FIGURE 18: Monthly climatology (2000 to 2010) of the MLD, by the M2 method, in meters. ORAS4 database.



Source: Own authorship.

The variability of the MLD is associated whit some mechanisms, such as winds, solar radiation, precipitation and the ocean's circulation, the heat flux or fresh water incoming in the surfaces ocean decreases the density of the surface layer, resulting in a more stable stratified profile. If the winds are incapable to mix the surface layer, the final MLD is shallower than the initial MLD (TALLEY *et al.*, 2011), the months whit the thickest MLD are also associated with the greater intensity and speed of the wind over the ocean, the FIGURE 19 shows monthly climatology of the winds.

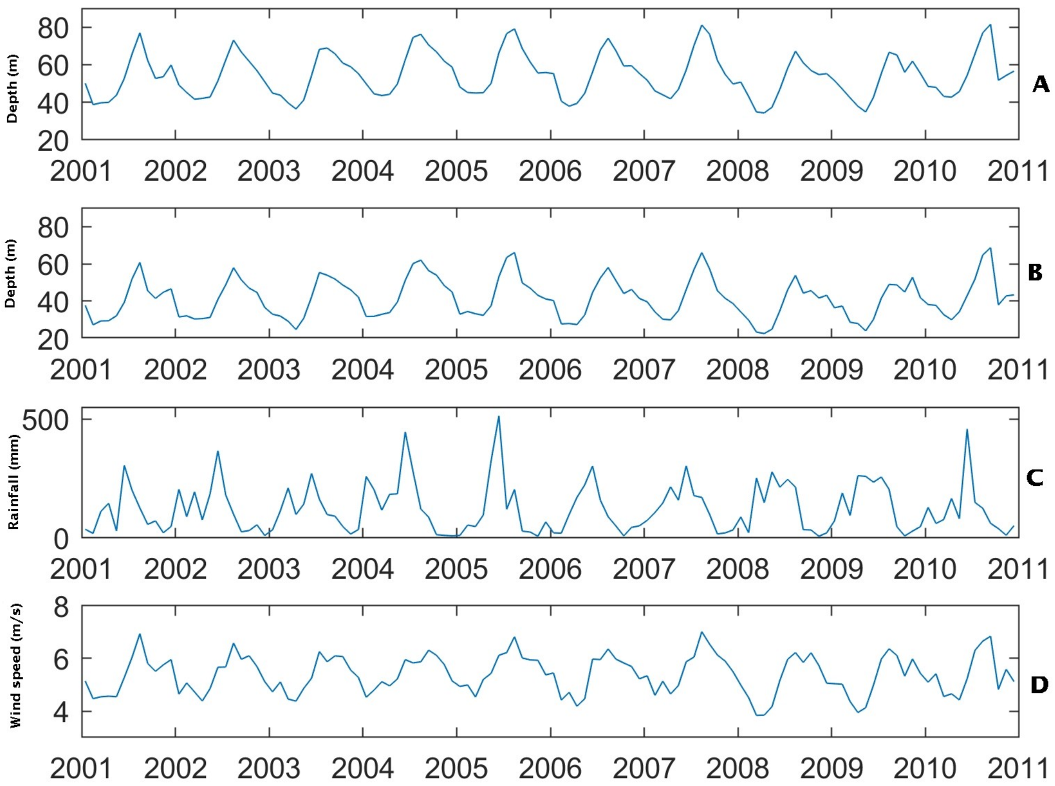
FIGURE 19: Monthly climatology (2000 to 2010) of wind speed, in meters per second. Data obtained by ERA-Interim, with resolution 0.5x0.5.



Source: Own authorship.

The mean variability of the MLD, rainfall and wind speed can be seen in FIGURE 20, the seasonality of wind speed, rainfall and MLD is clear and, the maxima of wind speed and the thickest MLD occur at the rainy season. The first and second graphs of the figure do not show only the variability of the MLD, but they show that the temperature threshold (M2) overestimates the depth relative to the density method (M1), given that the potential density is not only a function of temperature, but also salinity and pressure. According to Webber (2003), in some places the density can be approximately constant in a surface layer.

FIGURE 20: Seasonal variability between 2001 and 2011. From top to bottom: MLD for M2(A) and M1(B), rainfall(C) and wind speed(D).

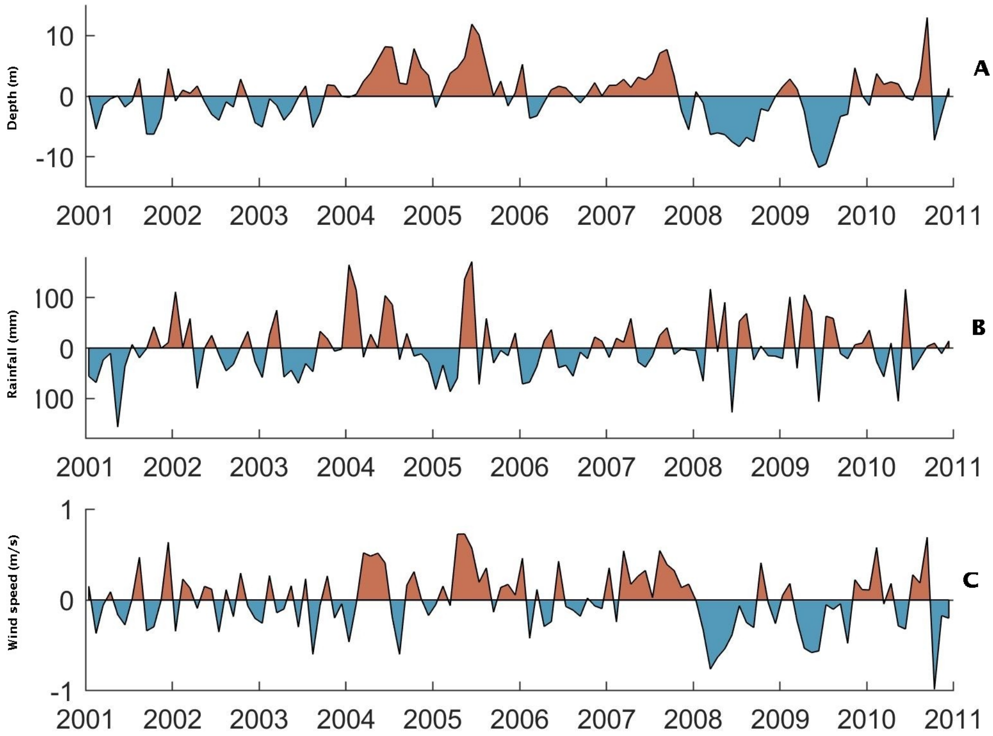


Source: Own authorship.

The monthly means highlights the variability of the MLD, which is correlated whit the rainfall regime and the winds, since the wind stress increases the mixing in the upper layer thus generating a thickest MLD and at the same time the wind carries the water vapor to the continent, allowing the formation of rainfall.

We analyzed the anomalies (FIGURES 21 and 22) and observed a relationship between MLD and the wind patterns, since positive anomalies observed in the MLD in the years 2004 and 2005 were also observed in the wind speed, during the same period there was a large variability between positive and negative rainfall anomalies. In 2009, both wind and MLD have negative anomalies, which were not observed in the rains.

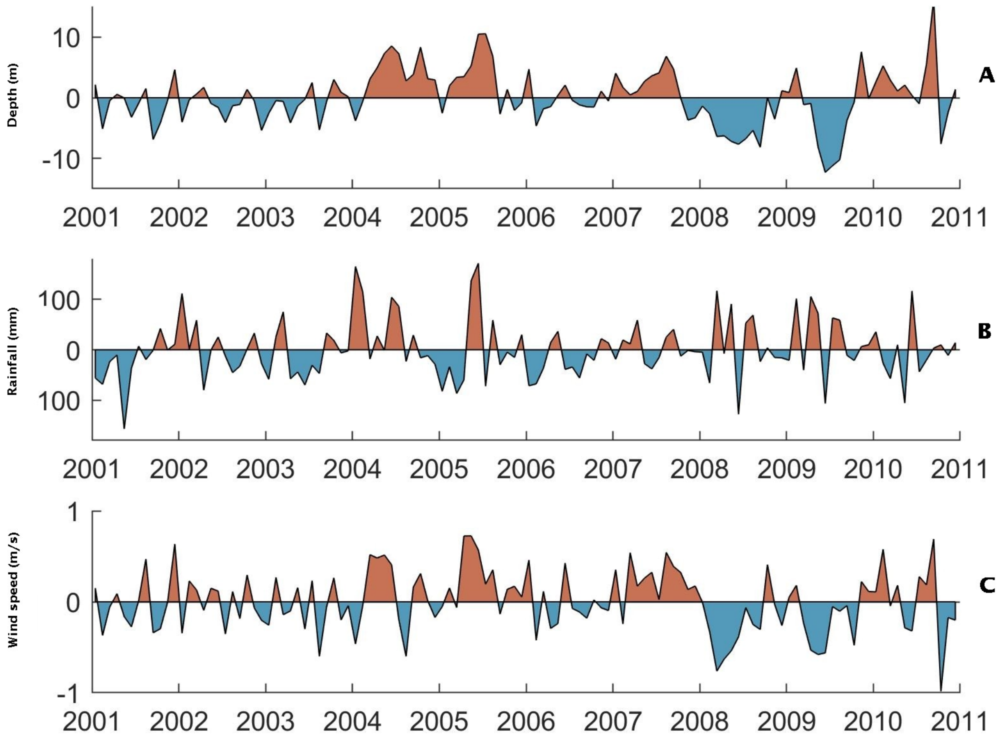
FIGURE 21: Monthly anomalies between 2001 and 2011: MLD by M1(A), rainfall(B) and wind speed(C). Data by ORAS4.



Source: Own authorship.

The direct relationship between MLD and the winds was already expected, however, despite following the same annual cycle (FIGURE 20), anomalies of MLD and rainfall do not respond in the same way, which raises a question about how both are related.

FIGURE 22: Monthly anomalies between 2001 and 2011: MLD by M2 (A), rainfall(B) and wind speed(C). Data by ORAS4.



Source: Own authorship.

As future work, it is clear that the study area should be subdivided into a smaller region closer to the coast of Pernambuco, considering that the selected area has different behavior in relation to SST, wind patterns and MLD. For example, the direction of the trade winds to the north of the equator, the difference in temperature and salinity of the near of large rivers, such as, the Amazon to the north and the São Francisco to the south. These characteristics may change the mean of MLF and, also may hide the interanuall effects and some anomalous behavior at the region.

Conclusions

The main objective of this work was the quantification of the time-space variability of the ocean mixed layer depth on the southwestern border of the tropical Atlantic and its influence on rainfall regimes in the state of Pernambuco. The analysis and characterization of the MLD in this region contributes to the studies of sea-air interaction that directly affect the climate in the eastern part of the Northeast of Brazil (ENEB).

The observational data used in this work allowed a better reliability of the results besides contributing to future studies with application of oceanic modeling and monitoring techniques.

The results allowed the identification of periods, that contributed to a deepening of MLD and that were identified in the rainy season.

The best method identified for the calculation of MLD was M1, based on the density gradient. The ORAS4 reanalysis data, based on this method, identified a relationship between MLD, rainfall and wind. The analyzes showed a better MLD response to wind patterns over the SWTA. The wind can then be an important parameter for the monitoring of MLD and consequently associated atmospheric instabilities.

In addition, this study contributes to the prediction and analysis of extreme events of precipitation events in the ENEB, which are the object of study in our laboratory, as applications for the use of coupled model of sea-air interaction.

6 CONCLUSION AND PERSPECTIVES

The oceans have a direct impact on society, acting as a thermal regulator, providing a well-being of life on the planet and as a food supplier through fishing. Understanding the air-sea interaction is essential for the public policies, which could make better decisions, with this information, in the management of regions such as Northeast Brazil, which suffer from extreme events of rain and drought, improving the human's life quality and, also for fisheries and environment management in the tropical Atlantic.

This work highlights the importance of studies and monitoring of the Equatorial Atlantic and its meteorological and oceanographic variables. In both results (section 4.1 and 4.2) there was an impact of these variables on the climate such as on the biota, mainly in the chlorophyll-a, essential for the maintenance of the marine food web. Continuous studies ensure a better understanding of thermodynamics in the region and how this directly affects the planet's hydrosphere, atmosphere, and biosphere.

This work contribute to a better understanding of the variability of the meteorological and oceanographic patterns in the Equatorial Atlantic and, its relationship with the climate and biota, specifically regarding to the chlorophyll-a (Chl-a) and Atlantic Niño/Nina events over the EA, and the rainfall patterns associated with the variability of the MLD. To understand spatial and temporal patterns of these variables, *in situ* and reanalysis data were used to perform statistical analyses, such as averages, anomalies, composites and wavelets.

These results highlight the importance of monitoring the Chl-a in the equatorial Atlantic through a climatic index, as the Atlantic Niño signal, which is the main mode of interannual variability in this region. It is necessary to understand the processes that control the Chl-a variability due to its importance on the marine ecosystems. Indeed, the biomass is the food supply to higher trophic level and an increase of the frequency of Atlantic Niño may affect the fisheries and the ecological balance of the equatorial Atlantic.

The observational data used in this work allowed a better reliability of the results besides contributing to future studies with application of oceanic modeling and monitoring techniques, these results allowed the identification of periods, that contributed to a deepening of MLD and that were identified in the rainy season.

In addition, this study contributes to the prediction and analysis of extreme events of precipitation events in the ENEB, which are the object of study in our laboratory, as applications for the use of coupled model of air-sea interaction.

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