

## RESEARCH ARTICLE

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## Key Points:

- High-resolved Chl field is shown in the Madeira Basin from 1871 to 2008
- Strong decrease of the deep chl maximum is correlated to NAO and solar radiation
- Carbon uptake by subtropical oceans decreased by 700 Mt in 2008 compared to 1871

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Chlorophyll *a* reconstruction from in situ measurements:  
2. Marked carbon uptake decrease in the last century

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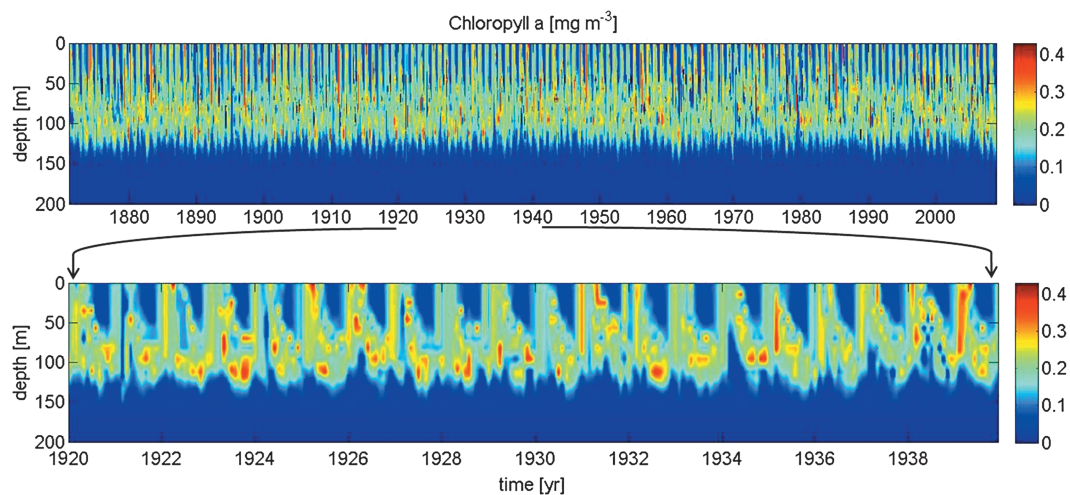
**Abstract** A chlorophyll *a* hindcast in the Madeira Basin from 1871 to 2008 was used to analyze the long-term variability in the oligotrophic, subtropical gyres in relation to the climate change of the last century. The deep chlorophyll maximum (DCM), as dominant pattern of the chlorophyll *a* field, showed a fast decrease in its strength in the 1940s. An absolute minimum was reached between 1967 and 1973 when no DCM established with a recovering to the end of the time series. Long-term variability of the DCM was related to the North Atlantic Oscillation with a time delay of 9 years. The marked decrease in the 1940s was correlated to the drop of the solar radiation in transition from early brightening to global dimming. Caused by the influence of the solar radiation and maybe related to increasing global temperatures in the last century, the integrated chlorophyll *a* concentration decreased by about 0.7 mg m<sup>-2</sup> in 2008 compared to 1871. The high-resolved chlorophyll *a* hindcast allowed an estimation of the carbon uptake by the ocean due to primary production in the euphotic zone. A rough calculation over the area of the global subtropical oceans showed 700 megaton less carbon uptake in 2008.

## 1. Introduction

Marine primary production by phytoplankton plays a key role in the global carbon uptake and is therefore an important factor influencing the global climate [Charlson et al., 1987; Murtugudde et al., 2002; Boyce et al., 2012]. Nearly half of the planetary primary production takes place in the euphotic zone, the uppermost water column of the global oceans [Behrenfeld et al., 2006; Boyce et al., 2010]. A large amount of carbon is subsequently transferred into the deep oceans, stored in sediments, and removed from the atmosphere [Field et al., 1998; Denman et al., 2007; Schwab et al., 2012].

The subtropical oceans spread up to 40% of the planetary surface and their ecosystems contribute to 25% of the global primary production [Longhurst et al., 1995; Polovina et al., 2008; Schwab et al., 2012]. To estimate the primary production of the water column, measurements of chlorophyll *a* concentrations are commonly used as a proxy for the phytoplankton biomass [Cullen, 1982; Uitz et al., 2006]. In the subtropical oceans during the bloom in late winter and early spring, the primary production takes place mostly near the surface as an answer to enhanced wind mixing causing new nutrient import from beneath the nitracline which is deeper than 100 m in this region [Cullen, 1982]. After depletion of nutrients at the surface, the phytoplankton descends to the edge of the euphotic zone in approximately 80–100 m water depth and forms the deep chlorophyll maximum (DCM) [Teira et al., 2005]. The formation of the DCM depends on two main factors: the transport of nutrients upward through the nutricline by turbulent diffusion as well as the amount of light from above [Hodges and Rudnick, 2004]. Therefore, the upper and lower boundaries of the DCM are found between the lowest required concentration of nutrients and the deeper edge of the euphotic zone. Due to formation of the DCM, considerable carbon uptake takes place in the subtropical gyres even in summer and autumn while the surface chlorophyll *a* is depleted. This important contribution to the global carbon pump results in 50% of the global ocean carbon export to the deep oceans and sediments within the subtropical gyres [Schwab et al., 2012].

The sensitivity of phytoplankton growth to long-term changes in nutrient and light supply is still not understood and needs to be investigated for robust projections of the carbon export by the oceans in changing environments in the 21st century due to global temperature increase. Because no in situ measured time series of chlorophyll *a* over several decades exist, the analysis of past primary production depends on recalculations of the chlorophyll *a* distribution. In order to approach this task, we present an analysis of the long-term variability of the DCM using a chlorophyll *a* hindcast in the Madeira Basin for more than 100 years to identify the forcing mechanisms of primary production in low-light conditions like found



**Figure 1.** (top) The chlorophyll *a* hindcast from 1871 to 2008 in the Madeira Basin. (bottom) The chlorophyll *a* hindcast is shown from 1920 to 1939 indicating the transition from the strong DCM in the 1920s to the weaker DCM in the late 1930s.

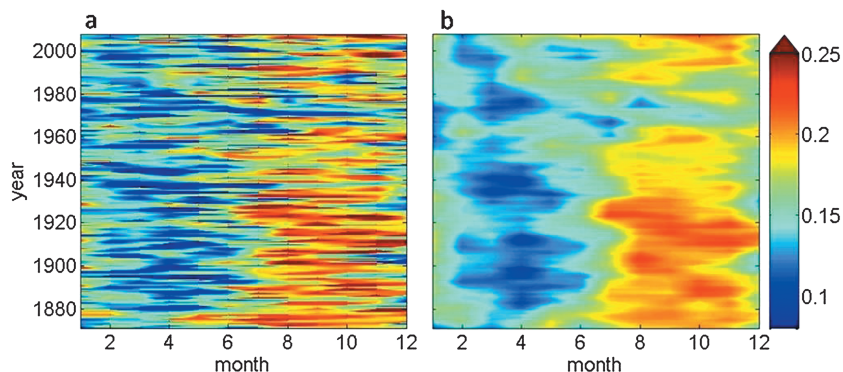
in the DCM. The strength of solar radiation is, apart from the recorded variability of sea surface temperature in the last century [Denman *et al.*, 2007; Levitus *et al.*, 2005, 2009], also subject to high multidecadal variability [Ohmura, 2006; Wild, 2009] and may influence the strength of the DCM. In context of evaluated decrease of primary production in the last century due to global warming [Behrenfeld *et al.*, 2006; Martinez *et al.*, 2009; Boyce *et al.*, 2010, 2011; Mackas, 2011; Rykaczewski and Dunne, 2011; McQuatters-Gollop *et al.*, 2011], an estimation of the carbon export in the subtropical gyres over more than one century aims to determine the possible effects of global climate change.

## 2. Material and Methods

Due to a lack of continuous in situ measurements of the DCM in the subtropical regions, we applied a method to reconstruct a chlorophyll *a* field in the Madeira Basin in the northeastern part of the oligotrophic North Atlantic Subtropical Gyre (NAST) from 1871 to 2008 (Figure 1). This method uses in situ measurements of temperature and nitrate, and their correlation is adapted to a modeled temperature field resulting in a nitrate field. In a next step, in situ chlorophyll *a* measurements are fitted to both, in situ temperature and nitrate measurements and this relation is finally used to calculate the chlorophyll *a* field. The in situ measurements were taken from the Azores Frontal Region (29°N–38.5°N and 17°W–27°W) [Fründt and Waniek, 2012], a part of the Madeira Basin. The method and the in situ measurements are described in detail in Fründt *et al.* [2015]. As required modeled temperature field, we used the output of the simple ocean data assimilation parallel ocean program (SODA POP) [Carton *et al.*, 2005; Carton and Giese, 2008], version 2.1.6 in its entire length from 1871 to 2008, resulting in a chlorophyll *a* hindcast of the same length (Figure 1). The data set by SODA POP is forced by 20CRv2 winds of the Twentieth Century Reanalysis Project [Compo *et al.*, 2011].

Empirical orthogonal function (EOF) analysis [von Storch and Zwiers, 1999] was performed using annual mean profiles of the standardized chlorophyll *a* concentration. Furthermore, wavelet coherences and phase differences were computed by applying the complex Morlet wavelet ( $\omega_0 = 6$ ) as mother wavelet to determine the periods on which two time series behave synchronously [Torrence and Compo, 1998; Grinsted *et al.*, 2004].

The impact of the dominant large-scale atmospheric patterns was determined by using the winter North Atlantic Oscillation (NAO) index from December to March [Hurrell, 1995]. Analogous analyses were done with further modes in the Atlantic: the Atlantic Multidecadal Oscillation (AMO) [Enfield *et al.*, 2001], the Arctic Oscillation (AO) [Higgins *et al.*, 2000], the Atlantic Meridional Mode (AMM) [Chiang and Vimont, 2004], the East Atlantic Pattern (EAP) [Barnston and Livezey, 1987], and the North Atlantic Subpolar Gyre Index (SPG) [Hátún *et al.*, 2005]. To determine the impact of global teleconnections, additionally the Pacific Decadal



**Figure 2.** Hovmöller diagram of the mean chlorophyll *a* concentration ( $\text{mg m}^{-3}$ ) at depths of the DCM between 90 and 130 m from 1871 to 2008: (a) the unfiltered annual values and (b) a running mean over 10 years.

Oscillation (PDO), [Trenberth and Hurrell, 1994], and the El Niño–Southern Oscillation (ENSO), [Bradley et al., 1987] indices were used.

As solar radiation flux at surface, we used the data provided by NOAA-Cooperative Institute for Research in Environmental Sciences (CIRES) Twentieth Century Reanalysis Version II in the Azores Frontal Region in monthly resolution (calculated from 3-hourly values) from 1871 to 2008 [Compo et al., 2006]. A mean value from August to November for each year and, enhancing the multidecadal variability of the time series [Gulev et al., 2013], a running mean over 10 years is calculated (Figure 5). The same filtering method is applied to the chlorophyll *a* hindcast.

The time series of chlorophyll *a* content over more than the twentieth century allowed estimation of the change in carbon uptake by the oceans due to changes in the amount of primary production. The difference in carbon uptake ( $\Delta\text{CU}$ ) in the subtropical gyres between 2008 and 1871 was calculated using

$$\Delta\text{CU} = \Delta_{\text{CHL}} \cdot R \cdot A \cdot \mu \tag{1}$$

with  $\Delta_{\text{CHL}}$  the difference in integrated chlorophyll *a* content of the water column from surface to 465 m depth between 2008 and 1871,  $R$  = carbon/chlorophyll  $a$  = 50 as a weighted mean between the ratio of 25 in DCMs [Eppley, 1968; Cullen, 1982] lasting all year and of 150–250 near the surface [Beers et al., 1975; Cullen, 1982] contributing only during the bloom, the area of the subtropical gyres  $A = 2.04 \cdot 10^8 \text{ km}^2$  [Polovina et al., 2008] and the growth rate of phytoplankton  $\mu = 0.26 \text{ day}^{-1}$  in the NAST [Marañón, 2005].

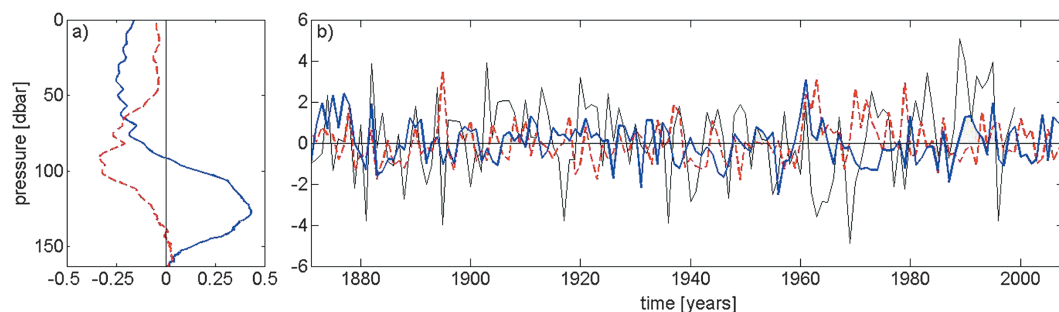
### 3. Results

In the second half of the year between August and November, the DCM in a depth of about 90 to 130 m was most distinct with values typical for the oligotrophic waters in the NAST of 0.25–0.3  $\text{mg m}^{-3}$  (Figure 2a). Contrary to the surface, where chlorophyll *a* was completely depleted after the bloom due to exhausted nutrients, low chlorophyll *a* concentrations in the DCM were still found between December and July in most of the years. Besides the interannual variability, the time series of 138 years of chlorophyll *a* concentration revealed a variability of the strength of the DCM on time scales of several decades (Figure 2b). Beginning in the nineteenth century, the chlorophyll *a* concentration continuously increased until the 1930s where a marked DCM lasted over 4 months.

Afterward, the chlorophyll *a* concentration rapidly dropped in the 1940s to 1960s (Table 1). In these more than 20 years, the chlorophyll *a* values remained nearly constant on the lowest level since 1871. A further decrease of the chlorophyll *a* concentration led to the absolute minimum between 1967 and 1973 when no

**Table 1.** Mean Chlorophyll *a* Concentrations in the DCM (90–130 m)  $\text{CHL}_{\text{DCM}}$  From August to November for Selected Time Periods

Time Period	$\text{CHL}_{\text{DCM}}$ ( $\text{mg m}^{-2}$ )
1920–1939	$0.20 \pm 0.04$
1940–1966	$0.183 \pm 0.025$
1967–1973	$0.15 \pm 0.05$
1974–2008	$0.18 \pm 0.04$



**Figure 3.** (a) The first (blue line) and second EOF (red dashed line) and (b) the time coefficients of the first EOF (blue line) and second EOF (red dashed line). Note that due to calculation with the standardized chlorophyll *a* field, they have no units. Additionally, the winter NAO index (black line) with a time lag of 9 years is given.

DCM was established, which held for the entire upper 100 m. This exceptional situation was followed by a steady intensification until the last years of our time series. However, the chlorophyll *a* concentration did not reach the high values from the 1920s to 1930s (Table 1).

The DCM as dominant pattern of the chlorophyll *a* field over the entire time period was manifested in the first, explaining 45.4%, and second EOF, explaining 19.3% (Figures 3a and 3b). The integrated chlorophyll *a* content of our time series (mean value  $24 \pm 5 \text{ mg m}^{-2}$ ) showed a decrease of  $\Delta_{\text{CHL}} = -0.72 \text{ mg m}^{-2}$  from 1871 to 2008.

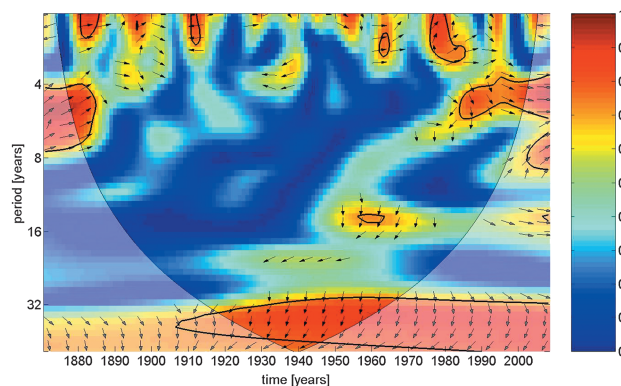
#### 4. Discussion

##### 4.1. Long-Term Variability of the DCM

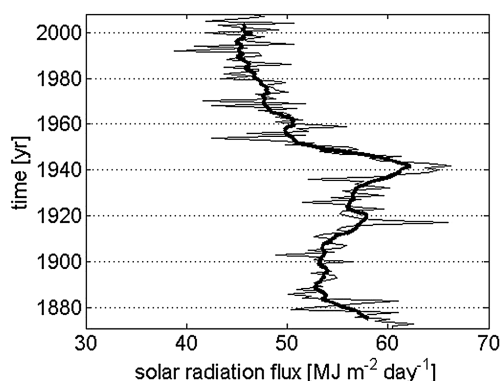
Fründt *et al.* [2015] showed with a shorter time series from 1980 to 2008, by applying EOF analysis, a correlation between the DCM and the NAO on a period of roughly 5 years. Application of the wavelet coherence between the time coefficient of the first EOF (Figure 3b) and the NAO showed that the relationship between NAO and the DCM found between 1980 and 2008 was less contributed to the long-term variability of the DCM since this relationship had not been established before the 1980s (Figure 4). This was maybe linked to the regime shift 1976–1981 when the Atlantic Multidecadal Oscillation (AMO) as well as the NAO changed from a negative to a more positive trend. This shift was related with a change from a colder to warmer period associated with variations in the wind field over the North Atlantic

[Kerr, 2000; Dima and Lohmann, 2007; Dippner *et al.*, 2012] influencing mixed layer depths and nutrient supply into the euphotic zone from below.

On periods in range of several years, no continuous relationship over the entire time series could be identified. On decadal scale, the DCM was significantly correlated with the NAO since 1910 on periods of 30–40 years with a consistent phase difference of  $90^\circ$ , i.e., a time lag of approximately 7 to 10 years (Figure 4). The calculation of the correlation coefficients  $\rho$  between the NAO and the time coefficient of the first EOF showed significant results at a time lag of 0 years ( $\rho = 0.17$ ,  $p = 0.041$ ) and of 9 years ( $\rho = 0.21$ ,  $p = 0.016$ , Figure 3b) which corresponded to the phase difference as seen in the wavelet coherence. The direct correlation with



**Figure 4.** Wavelet coherence and phase between the winter NAO index and the time coefficient of the first EOF. Areas bordered by the black solid lines are within the 95% confidence level using the red noise model. The arrows indicate the phase difference ( $0^\circ$  is given by a horizontal arrow pointing to the right, and an arrow pointing vertically upward means the time coefficient lags the NAO by  $90^\circ$ ). The U-shaped curve shows limits of the cone of influence (COI). All signals outside COI are statistically not certain.



**Figure 5.** The solar radiation flux at surface in the Azores Frontal Area from 1871 to 2008 (provided by NOAA-CIRES Twentieth Century Reanalysis Version II). Shown is the mean value between August to November (thin line) during which months the DCM is mostly established, and a running mean over 10 years (bold line).

AMM, EAP, SPG, PDO, and ENSO) gave no significant results considering phase differences. Potentially, the regime shift in the North Atlantic 1957/58 with impact on the sea surface temperatures in the midlatitudes [Yasanaka and Hanawa, 2002] supported the decline in the strength of the DCM. However, the influence of this regime shift on the phytoplankton community in the subtropical regions was not studied yet.

The globally longest time series of in situ measured chlorophyll *a* were recorded at Bermuda Atlantic Time-series Study and Hawaii Ocean Time series between 1989 and 2007 and show an increase of primary production in this period [Saba *et al.*, 2010; Chavez *et al.*, 2011]. These measured trends support our results of the chlorophyll *a* hindcast in which an increase of the chlorophyll *a* concentration occurred from 1973 to 2008 (Figure 2).

Besides the hydrographic and atmospheric influences affecting the amount of nutrient supply into the DCM, its formation depended highly on the incoming solar radiation (Figure 5). The planetary increase of the solar radiation in the first half of the twentieth century, especially in the 1930s and 1940s is known as the “early brightening” [Ohmura, 2006; Wild, 2009]. Afterward, mostly caused by an increase of aerosols, clouds and their interaction, the “global dimming” reduced the solar radiation flux at the sea surface. On a planetary scale, changes in cloudiness and atmospheric transmission terminated the dimming in the 1980s followed by a “brightening” until nowadays [Wild *et al.*, 2005; Wild, 2009]. Most pronounced in the Madeira Basin was the moment from early brightening to global dimming in the 1940s (Figure 5). The transition from brightening to dimming caused a reduction of the DCM strength in the 1970s (Figure 2b). With the start of the second brightening, the DCM intensified again. The direct connection of the strength of the DCM and the solar radiation flux was manifested in a significant correlation coefficient of 0.25 (*p* value 0.004) calculated with the annual mean chlorophyll *a* in 90–130 m. Hence, the intensity of the solar radiation was a dominant factor explaining the long-term variability of the chlorophyll *a* concentration in the deeper levels of the euphotic zone where the phytoplankton was adapted to low-light conditions. Whereas the light-saturated upper levels of the euphotic zone were not affected, smallest changes in the light intensity had high impact on the primary production in the DCM.

#### 4.2. Effect of Long-Term Trends of Primary Production on the Ocean’s Carbon Uptake

It is worth considering which impact did the decrease of the DCM as dominant feature of the chlorophyll *a* field in the subtropical gyres [Pérez *et al.*, 2006] have for the carbon uptake by the oceans in the last century. In this region, the surface chlorophyll *a* was depleted after the bloom for almost half a year, making the DCM an important fraction of the integrated chlorophyll *a* content. Our method allowed a calculation of the carbon uptake for the entire year with lower error bars due to a more precise estimation of the integrated chlorophyll *a* over time and depth. The decrease of the integrated chlorophyll *a* was on the one hand caused by the long-term variability of the solar radiation flux and its impact on the DCM. On the other hand, changes in the growth conditions like varying SST and associated stratification of the water

NAO, i.e., no time lag, was diminished compared to the value found in Fründt *et al.* [2015] when only the last 28 years were regarded, because a continuous correlation over the entire length was not found. The time lag of 9 years might be related to the transmission of NAO-induced oceanic signals by slow Rossby waves with speeds of around  $1 \text{ cm s}^{-1}$  as reported by Cipollini *et al.* [1997].

Regarding the rapid decrease of the chlorophyll *a* concentration in the 1940s (Figure 2), the NAO seemed not to be the dominant forcing mechanism. Even though the NAO was in a predominantly negative phase from the 1940s to the 1960s (Figure 3b), the negative trend of the NAO was not strong enough to explain the sudden decline in chlorophyll *a* concentration in the DCM. The analogous correlation analyses of the DCM with the further climate indices (AMO, AO,

column as well as mixing processes interfered or promoted nutrient supply. The long-lasting global warming of the upper water column has recently been reported by *Levitus et al.* [2009]. An increase of temperature led to an decrease of the globally integrated primary production over the entire last century [Boyce et al., 2010] due to higher stratification preventing nutrient import from below the euphotic zone [Bopp et al., 2001; Sarmiento et al., 2004; Polovina et al., 2008; Irwin and Oliver, 2009]. The study by Boyce et al. [2010] has been intensely discussed by Mackas [2011], Rykaczewski and Dunne [2011], McQuatters-Gollop et al. [2011], and Boyce et al. [2011] because the results were biased by combining different data sets of ocean transparency and in situ measured chlorophyll *a*. This difficulty was evaded using the method introduced in Fründt et al. [2015] resulting in an unbiased chlorophyll *a* hindcast. Assuming both the solar radiation flux and the warming of the world's ocean were similarly manifested in the subtropical oceans, it conceded an estimation of the global change of carbon uptake by the decrease of primary production in the last 138 years. Comparison of the first (1871) with the last year (2008) of our time series showed a reduction in carbon uptake by the oceans of  $\Delta\text{CU} = -700$  megaton (Mt)  $\text{yr}^{-1}$  in 2008.

The global carbon uptake by the oceans amounted for 92.2 Gt  $\text{yr}^{-1}$  in the 1990s [Denman et al., 2007]. The subtropical gyres contributed to 50% of this global carbon export [Emerson et al., 1997]. Therefore, the 700 Mt less carbon uptake in 2008 amounted to 1.5% in the subtropical gyres nowadays compared to the nineteenth century. Integrating over the entire time period from 1871 to 2008 and assuming a linear decrease of the carbon uptake in these 138 years, the oceans stored 48 Gt less carbon, i.e., half of the annual carbon uptake nowadays. The ocean's capability to uptake and storage carbon in intermediate and deep waters as well as in the sediment is considered as an important factor to reduce the  $\text{CO}_2$  concentration in the atmosphere and hence the temperature increase until the end of the 21st century [Denman et al., 2007]. Our back-of-the-envelope calculation above showed this capability decreased in the last century. Large parts of the subtropical gyres are oligotrophic. Especially these areas of the world's ocean for which we determined a reduced capacity of carbon uptake are extending [Polovina et al., 2008]. This may result in even less global carbon uptake in the 21st century, further backed up by the accelerated warming of the oceans in the last decades of the twentieth century [Levitus et al., 2009].

## 5. Conclusion

The analysis of more than 100 years of modeled chlorophyll *a* in the Madeira Basin showed the DCM as very sensitive to long-term changes in Northern Hemisphere atmospheric pattern and in the solar radiation flux. This chlorophyll *a* hindcast, spanning more than one century, offered the opportunity to highlight the direct influence of global dimming and brightening on the DCM for the first time. Identifying the influence of the NAO on periods of 30–40 years with a time delay of 9 years required such a long time series, but needs further studies to understand the variability of the underlying mechanisms. In the subtropical gyres, the DCM greatly contributes to the integrated primary production and is an important factor in the global carbon cycle. Assuming a linear trend of carbon uptake between 1871 and 2008, this led to a reduction in carbon uptake in the order of magnitude of half an annual carbon uptake today. Similar studies in other regions of the world's ocean are needed to allow a global estimation of the variability in carbon uptake over the last 100 years, especially before and during the period of increasing global temperatures. However, applying the method introduced by Fründt et al. [2015] requests unbiased time series of temperature, nitrate, and chlorophyll *a* to minimize systematical errors. These results will help to comprehend the sensitivity of the primary production to changes in atmospheric and oceanic conditions and contribute to more solid projections of subsequent global climate conditions.

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