

1. Point-by-point response and relevant changes performed in the manuscript

In this section we inserted a point-by-point response to the Referees as well as Dr. Schmittner together with all relevant changes performed in the manuscript. To facilitate the review, we copied the Referees' as well as Dr. Schmittner's comments below in black and inserted our comments in green. The modifications performed to the manuscript are shown in italic green in quotes.

1.1 Responses to Referee #1

We thank Referee #1 for the constructive review of our manuscript.

This is an interesting manuscript presenting two high-resolution planktic $\delta^{13}\text{C}$ records from the Brazilian coast covering HS3 and HS2. It is worth publishing in *Climate of the Past* if the "Discussion" section is completely rewritten and therefore the interpretation of the results re-assessed.

The authors have not shown that the planktic $\delta^{13}\text{C}$ decrease measured in their core was due to stronger Southern Ocean upwelling. This is just an hypothesis.

Comment #1 - We agree that it is a hypothesis and have now treated it accordingly by rephrasing specific parts of the revised version of our manuscript.

Additionally considering this comment and the similar suggestions from Referee #2 and Dr. Schmittner, we also incorporated two additional hypotheses (see below) in the revised manuscript, in order to have a more thorough and balanced discussion:

Pages 10-12, lines 10-2: "*5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump*

Recent model experiments (e.g. Schmittner, 2005; Schmittner and Galbraith, 2008) have shown that AMOC slowdown events may cause a decrease in the global efficiency of the oceanic biological pump, being an important driver for the oceanic CO_2 outgassing within HS1 during the last deglaciation and possibly also during other HS, including HS3 and HS2.

NADW has low preformed nutrient waters because it is formed by nutrient depleted surface waters, where the biological pump has efficiently removed nutrients from surface waters (Marinov et al., 2008). AABW has high preformed nutrient waters because it is formed by nutrient-enriched Southern Ocean surface waters (nutrients have not being efficiently removed from surface waters). However, during weak AMOC two factors may alter the nutrient distribution and the global oceanic biological pump (Schmittner and Galbraith, 2008). First, the reduction in the NADW formation decreases the input of low preformed nutrient (high $\delta^{13}C_{DIC}$) waters to the ocean interior which becomes more dominated by high preformed nutrient (low $\delta^{13}C_{DIC}$) southern component waters (e.g., AABW). Second, the reduction of the Southern Ocean stratification induced by the decrease of salt input via NADW formation promotes the strengthening of the upwelling and subsequent sinking of high preformed nutrient (low $\delta^{13}C_{DIC}$) waters to the ocean interior, thus reducing the capacity of those unutilized nutrients to sequester carbon via the biological pump. The two factors acting in conjunction are thought to be responsible for the simulated weakening of the global efficiency of the biological pump, as well as for the increase in CO_{2atm} and decrease in $\delta^{13}CO_{2atm}$ (Schmittner and Galbraith, 2008; Schmittner and Lund, 2015; Hertzberg et al., 2016).

Schmittner and Lund (2015) show that the modeled weakening of the biological pump, induced by an AMOC slowdown, reduces the ability of the surface ocean to biologically sequester isotopically light organic carbon (^{12}C), producing a decrease in the surface ocean $\delta^{13}C_{DIC}$ and an increase of the intermediate ocean $\delta^{13}C_{DIC}$ (lower remineralization rate). For HS1, planktonic and benthic foraminiferal $\delta^{13}C$ records (Tessin and Lund, 2013; Lund et al., 2015; Curry and Oppo, 2005; Hertzberg et al., 2016) from the western South Atlantic (ca. 27°S) agree with the model output by showing a decrease in $\delta^{13}C$ in the upper water column (SACW) and an increase at intermediate water depths (AAIW). Thus, the weakening of the global oceanic biological pump and consequent negative anomaly of the $\delta^{13}C_{DIC}$ in the upper water column should be captured by the tests of planktonic foraminifera $\delta^{13}C$ during calcification (Spero and Lea, 1996; Bemis et al., 2000). The negative $\delta^{13}C_{DIC}$ during HS3 and HS2 revealed by our planktonic foraminifera provide the first observational evidence supporting the modeling results. Additionally this mechanism also provides a possible

*explanation for the larger negative $\delta^{13}\text{C}$ anomaly in *G. ruber w* (mixed layer dwelling) relative to the anomaly in *G. inflata* (permanent thermocline dwelling) (Fig. 4).*

It is noteworthy that the mechanism described in section 5.1, although based on a different driver for the decrease in $\delta^{13}\text{C}$, also suggests that the decreases in $\delta^{13}\text{C}$ of planktonic foraminifera from the South Atlantic would be carried by SACW (inherited from its precursor, SAMW) and thus both mechanisms (described in section 5.1 and here) are in this regard not mutually exclusive. However, the mechanism described in the present section goes against the assumption that weakening of the biological pump is related to stronger upwelling in the Southern Ocean, and that the Southern Ocean would be the source of the low $\delta^{13}\text{C}$ signal for the South Atlantic (Lund et al., 2015; Hertzberg et al., 2016).

5.3 Millennial-scale changes: the role of air–sea gas exchange

The $\delta^{13}\text{C}_{\text{DIC}}$ of the surface ocean can also be affected by air–sea gas exchange (Oppo and Fairbanks, 1989; Charles and Fairbanks, 1990; Lynch-Stieglitz et al., 1995). Although this process tends towards isotopic equilibrium, the CO_2 exchange between the ocean and the atmosphere does not lead to equilibrium because CO_2 uptake and emission will still occur in different regions and the movement and replacement of surface waters is faster than required for equilibration (Lynch-Stieglitz et al., 1995). Since the $\delta^{13}\text{CO}_{2\text{atm}}$ is lighter than $\delta^{13}\text{C}_{\text{DIC}}$, at areas of ocean $\text{CO}_{2\text{atm}}$ uptake (i.e., water mass formation regions) air–sea gas exchange has the potential to deplete $\delta^{13}\text{C}_{\text{DIC}}$ (Lynch-Stieglitz et al., 1995). Additionally, the isotopic fractionation is inversely correlated with temperature.

*Therefore, we cannot exclude the possibility that the likely decrease in $\delta^{13}\text{CO}_{2\text{atm}}$ during AMOC slowdown events (Eggleston et al. 2016) (e.g., HS3 and HS2) could have affected the $\delta^{13}\text{C}_{\text{DIC}}$ via air–sea gas exchange, especially in regions of water mass formation. The formation region of SACW is an area of ocean CO_2 uptake and may contribute to the $\delta^{13}\text{C}$ anomalies observed in our *G. inflata* record (Fig. 4b). Additionally, since the isotopic fractionation during air–sea gas exchange is temperature–dependent the weakening of the AMOC and subsequent warming of the upper subtropical South Atlantic (Barker et al. 2009; Chiessi et al. 2015) could have contributed to the observed $\delta^{13}\text{C}$ anomalies both in the *G. ruber w* and in the *G. inflata**

records (Fig. 4). However, the gradient is too small ($-0.1\text{‰ } \delta^{13}\text{C}$ per $^{\circ}\text{C}$, Broecker and Maier-Reimer, 1992) to explain the whole changes observed in our records. If temperature was the dominant driver, unrealistic changes between 5 and 13 $^{\circ}\text{C}$ would be required to explain the full amplitudes of the $\delta^{13}\text{C}$ variations.

The $\delta^{18}\text{O}$ records from *G. ruber w* and *G. inflata* from our core (Supplementary material Figure 1) should partially reflect changes in water temperature (ca. -0.22‰ per 1 $^{\circ}\text{C}$; e.g., Mulitza et al, 2003), but show no clear trends across HS3 and HS2. While temperature changes might be partially obscured in the foraminiferal $\delta^{18}\text{O}$ records by the influence of synchronous changes in seawater- $\delta^{18}\text{O}$, as has been hypothesized for the Holocene (Chiessi et al., 2014) and HS1 (Chiessi et al., 2015) in the western Atlantic, we consider it unlikely that temperature changes of the above magnitude would be completely masked.”

In the conclusions we also inserted the two additional hypotheses described above and removed the mention of the Southern Ocean as the main source of the CO_2 outgassed to the atmosphere during HS3 and HS2.

Page 13, lines 14-18: we added “(ii) Weakening of the global oceanic biological pump. A weak AMOC during HS3 and HS2 would promote an accumulation of ^{13}C -depleted CO_2 in the upper water column of the South Atlantic. This accumulation would result in a negative anomaly of the $\delta^{13}\text{C}_{\text{DIC}}$ (as well as of the $\delta^{13}\text{C}_{\text{CO}_2\text{atm}}$) that in turn would be captured by the tests of planktonic foraminifera at our core site. We further suggest that changes in air-sea gas exchange could have contributed to the decreases in $\delta^{13}\text{C}$ via both mechanisms.”.

Page 13, line 20: we removed “of the Southern Ocean”.

However, for the reasons stated below we still consider the possible role played by changes in Southern Ocean ventilation in producing our $\delta^{13}\text{C}$ anomalies:

- i. the mechanisms that invoke the Southern Ocean as one of the $\text{CO}_{2\text{atm}}$ -sources during the last deglaciation are based on reconstructions as well as conceptual and numerical models that, to our understanding, have not yet been negated (Toggweiler et al., 2006; Anderson et al., 2009; Sigman et al., 2010; Tschumi et al., 2011; Lee et al., 2011; Burke and Robinson, 2012);

- ii. model experiments that question the role of the Southern Ocean (e.g., Schmittner and Lund, 2015) were performed under preindustrial boundary conditions, which can significantly influence the results (e.g., Menviel et al., 2008), and do not constitute a firm negation of the Southern Ocean hypothesis.

Planktic $\delta^{13}\text{C}$ is influenced by several factors such as changes in oceanic circulation, mixing, SST, export production: : : (see Charles et al. 1993, Lynch-Stieglitz et al. 1995, Menviel et al. 2015: : :).

Comment #2 - We agree that planktonic foraminiferal $\delta^{13}\text{C}$ can be affected by several factors and that some of them were not thoroughly considered in the original version of our manuscript. To account for these different factors we have inserted two new sections (i.e., 5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump, and 5.3 Millennial-scale changes: the role of air-sea gas exchange) in our manuscript (please see our Comment #1 above) and clearly refer to the complexity of the proxy.

Additionally, by analyzing two different species of planktonic foraminifera, i.e., *G. ruber* w that is a symbiont-bearing species, and *G. inflata* that is a symbiont-facultative species (e.g., Hemleben et al., 1989), we have excluded the potential bias in our records related to changing symbiont activity (e.g., Spero and Lea, 1993; Spero et al., 1997; Bemis et al., 2000). Since the $\delta^{13}\text{C}$ record of both species show a similar pattern, changing symbiont activity can be disregarded as a potential bias to our results.

Page 4, lines 19-21: *“However, other factors such as calcification temperature, carbonate ion concentration, symbiont activity and air-sea gas exchange may also influence planktonic foraminiferal $\delta^{13}\text{C}$ (Lynch-Stieglitz et al., 1995; Spero and Lea, 1996, Spero et al., 1997; Bemis et al., 2000).”*

Page 7, lines 17-26: *“During HS, we expect warmer temperatures to have occurred in the upper water column of the western South Atlantic (Barker et al. 2009; Chiessi et al. 2015). This would trigger an increase in $\delta^{13}\text{C}$ values of the symbiont-bearing species investigated here if calcification temperature would dominate the $\delta^{13}\text{C}$ signal (Bemis et al., 2000), which is not the case (Fig. 4a). Additionally, given the lack of regional upper ocean reconstructions for carbonate ion concentration, we assume that increased $\text{CO}_{2\text{atm}}$ that is frequently associated with HS (Ahn and Brook, 2008; Ahn and Brook,*

2014) would have been accompanied by a decrease in sea surface carbonate ion concentration (Broecker and Peng, 1993). This would promote an increase in the $\delta^{13}\text{C}_{\text{DIC}}$ but our records show a negative $\delta^{13}\text{C}$ anomaly (Fig. 4). Furthermore, we analysed a symbiont-bearing and a facultative-symbiont species (i.e., *G. ruber w* and *G. inflata*, respectively) and both records show a similar pattern (Fig. 4) indicating that changes in symbiont activity can also be disregarded as a factor influencing our results (Spero et al., 1997; Bemis et al., 2000).”

An AMOC decrease will lead to significant surface ocean $\delta^{13}\text{C}$ changes. Changes in Southern Ocean upwelling are not the only solution.

Comment #3 - Agreed. As mentioned in our Comment #1 (please see above) we have revised our manuscript accordingly by incorporating two additional mechanisms (i.e., 5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump, and 5.3 Millennial-scale changes: the role of air-sea gas exchange) that could account for the $\delta^{13}\text{C}$ changes in our records.

In addition, it has been shown that during calcification temperature and carbonate ion content could have an impact on calcite $\delta^{13}\text{C}$ in planktic species (see Spero et al. 1997, Bemis et al. 2000: : :).

Comment #4 - Experiments with *O. universa* (a symbiont-bearing species) and *G. bulloides* (a non-symbiont species) suggest that under higher temperatures the first species presents higher $\delta^{13}\text{C}$ values whereas the second species presents lower $\delta^{13}\text{C}$ values (Bemis et al., 2000). These opposed $\delta^{13}\text{C}$ responses were related to the presence (i.e., *O. universa*) or absence (i.e., *G. bulloides*) of symbionts. We analyzed a symbiont-bearing and a facultative-symbiont species (i.e., *G. ruber w* and *G. inflata*, respectively), but both records showed a similar behavior. During HS, we would expect warmer temperatures in the upper water column of the western South Atlantic (Barker et al. 2009; Chiessi et al. 2015) and an increase in $\delta^{13}\text{C}$ values of our symbiont-bearing species if calcification temperature would dominate the $\delta^{13}\text{C}$ signal. Since *G. ruber w* $\delta^{13}\text{C}$ also showed a negative anomaly, we do not see the need to consider changes in calcification temperatures as a relevant driver for our $\delta^{13}\text{C}$ anomalies. We briefly refer to this issue in the revised version of our manuscript.

Page 7, lines 17-20: *“During HS, we expect warmer temperatures to have occurred in the upper water column of the western South Atlantic (Barker et al. 2009; Chiessi et al. 2015). This would trigger an increase in $\delta^{13}\text{C}$ values of the symbiont-bearing species investigated here if calcification temperature would dominate the $\delta^{13}\text{C}$ signal (Bemis et al., 2000), which is not the case (Fig. 4a).”*

Regarding the possible effect of changes in carbonate ion concentration, we are aware that changes in the seawater $[\text{CO}_3^{2-}]$ can impact planktonic foraminiferal $\delta^{13}\text{C}$ in an inverse way (i.e., higher (lower) $[\text{CO}_3^{2-}]$ decreases (increases) planktonic foraminiferal $\delta^{13}\text{C}$) (Spero et al., 1997). Given the lack of regional upper ocean reconstructions for $[\text{CO}_3^{2-}]$, we assumed that increased $\text{CO}_{2\text{atm}}$ would be accompanied by a decrease in sea surface $[\text{CO}_3^{2-}]$ (Broecker and Peng, 1993). This would promote an increase in the $\delta^{13}\text{C}_{\text{DIC}}$. HS are frequently associated with an increase in $\text{CO}_{2\text{atm}}$ (Ahn and Brook, 2008; Ahn and Brook, 2014) but our records show a negative $\delta^{13}\text{C}$ anomaly, suggesting that changes in $[\text{CO}_3^{2-}]$ are not the dominant driver of our $\delta^{13}\text{C}$ anomalies. We briefly refer to this issue in the revised version of our manuscript.

Page 7, lines 20-23: *“Additionally, given the lack of regional upper ocean reconstructions for carbonate ion concentration, we assume that increased $\text{CO}_{2\text{atm}}$ that is frequently associated with HS (Ahn and Brook, 2008; Ahn and Brook, 2014) would have been accompanied by a decrease in sea surface carbonate ion concentration (Broecker and Peng, 1993). This would promote an increase in the $\delta^{13}\text{C}_{\text{DIC}}$ but our records show a negative $\delta^{13}\text{C}$ anomaly (Fig. 4).”*

Moreover, the authors suggest that increased runoff from the Plata river drainage basin led to increased sediment rate. It is an interesting result, which nicely fits with a southward shift of the ITCZ during that time, however river runoff might potentially have a fairly low $\delta^{13}\text{C}$ signature, thus potentially also influencing surface $\delta^{13}\text{C}$ at the core location?

Comment #5 - Abrupt millennial-scale climate events of the last glacial period have been associated with increased precipitation over tropical and subtropical South America to the east of the Andes (e.g., Arz et al., 1998; Wang et al., 2007; Kanner et al., 2012; Strikis et al., 2015). During HS1, however, the millennial-scale signal of $\delta^{18}\text{O}_{\text{IVC-sw}}$ from the upper water column of our core site indicates an increase in salinity

(Chiessi et al., 2015). Thus, the upper water column of our core site was not affected by an increase in freshwater discharge from the Plata River at millennial-scale. Since the anomaly of precipitation during HS1 was stronger than during HS3 and HS2 in the Plata River drainage basin (Wang et al., 2007), it is unlikely that weaker precipitation anomalies in the Plata River drainage basin (i.e., HS3 and HS2) affected the upper water column of our core site more intensely than during HS1. This suggests that changes in the discharge of the Plata River drainage basin at millennial-scale are not a relevant driver of our $\delta^{13}\text{C}$ anomalies.

Pages 12-13, lines 27-3: *“Some aspects of the regional response to HS1 are useful to evaluate this possibility. During HS1, ice volume corrected seawater- $\delta^{18}\text{O}$ from the upper water column of our core site indicates an increase in salinity (Chiessi et al., 2015). Thus, despite of the increased terrigenous discharge, it seems that the upper water column of our core site was not affected by an increase in freshwater discharge from the Plata River during HS1. Since the precipitation anomaly of HS1 was stronger than that of HS3 and HS2 in the Plata River drainage basin (Wang et al., 2007), it is unlikely that weaker precipitation anomalies of HS3 and HS2 would have impacted the upper water column of our core site more intensely than during HS1. This suggests that changes in the discharge of the Plata River drainage basin at millennial-scale are not a relevant driver of our $\delta^{13}\text{C}$ decreases, and that the buoyant low salinity waters were advected elsewhere by winds, while terrigenous sediments were already too deep to be influenced by the wind.”*

Please note that: 1) the ice core data (Ahn and Brook 2014) do not support any atmospheric CO_2 increase during HS2 and HS3.

Comment #6 - We cannot agree with this point. The best resolved $\text{CO}_{2\text{atm}}$ record available (Ahn and Brook, 2014) shows an increase in CO_2 during HS3 and HS2 that is highlighted by red arrows in the original publication (please see Fig. 1c from Ahn and Brook (2014)). The slight offset between the increase in $\text{CO}_{2\text{atm}}$ and the HS3 and HS2 intervals used in our manuscript can be accommodated by age model uncertainties.

2) D^{13}CO_2 can't really be used due to poor resolution and most likely issues with age model.

Comment #7 - We agree that the resolution of the $\delta^{13}\text{CO}_{2\text{atm}}$ curve is still suboptimal to resolve changes during HS2, and we have removed this curve from Fig. 5 of the revised version of our manuscript. We are aware of Eggleston et al.'s (2016) new $\delta^{13}\text{CO}_{2\text{atm}}$ data. However, despite their reference to a reduction in $\delta^{13}\text{CO}_{2\text{atm}}$ during HS2, their record also lacks the necessary temporal resolution to allow an appropriate comparison to our records.

Page 9, lines 30-31: Despite the low temporal resolution, “Eggleston et al.’s (2016) Antarctic $\delta^{13}\text{CO}_{2\text{atm}}$ record shows a decrease during HS2.”

3) The link between changes in southern hemispheric westerlies and AMOC changes is still poorly documented.

Comment #8 - Despite the existence of some open issues, several publications (e.g., Lee et al., 2003; Anderson et al., 2009; Toggweiler et al., 2006; Tschumi et al., 2008; d’Orgeville et al., 2010; Laurantou et al., 2010; Sigman et al., 2010; Tschumi et al., 2011; Lee et al., 2011; Voigt et al., 2015) describe links between changes in AMOC and the Southern Hemisphere westerlies.

4) the opal flux in the Southern Ocean (Anderson et al. 2009) does not increase during HS2 and HS3.

Comment #9 - We are aware that Anderson et al.’s (2009) opal flux data (a proxy for the strength of Southern Ocean upwelling) did not change during the HS3 and HS2. However, this could be a temporal resolution issue. In core TN057-13-4PC, opal measurements show a mean temporal resolution of ca. 310 yr for HS1, but ca. 1550 yr for HS2 (this core does not reach HS3). In core E27-23, the mean temporal resolution of HS1 measurements is ca. 686 yr, but ca. 1550 yr for HS2, and ca. 1850 yr for HS3. HS1 opal measurements in core NBP9802-6PC have a mean temporal resolution of ca. 1715 yr, while HS2 measurements have a mean temporal resolution of ca. 3100 yr (this core does not reach HS3). Additionally, Anderson et al. (2009) suggest that such mechanism may be a common feature for other HS of the last glacial period.

As such the whole discussion section, as well as conclusion and abstract need to be rewritten.

Comment #10 - To account for the main topics raised by Referee #1 (but also Referee #2 and Dr. Schmittner) we have revised the necessary sections of our manuscript accordingly (please see above).

1.2 Responses to Referee #2

We thank Referee #2 for the constructive review of our manuscript.

Review of "Glacial d13C decreases in the western South Atlantic forced by millennial changes in Southern Ocean ventilation" by Campos et al.

The authors use the d13C of planktonic foraminifera to infer variability in the d13C of DIC in surface/thermocline waters in the SW Atlantic. One strength of the paper is the comprehensive review of the modern oceanographic setting. Another is that the planktonic time series are based on a sediment core with very high sedimentation rates, allowing for high resolution reconstruction of surface ocean d13C during Heinrich Stadial 2 and 3. It is also important that the authors used two different species to reconstruct d13C (one surface dwelling and another thermocline dwelling) to account for potential biases in habitat and vital effects that could overprint changes in the d13C of DIC. Given the high resolution and replicated nature of the record, the authors clearly show that this part of the Southwest Atlantic underwent significant changes in d13C during Heinrich Stadial 2 and 3.

The primary weakness of the paper is the interpretation of the d13C and sedimentation rate results. The authors are quick to assume that their d13C records reflect the input of light carbon from the Southern Ocean and neglect other possible explanations that are well documented in the published literature.

Comment #11 - We agree that other mechanisms could have also contributed to our records. Considering this comment and similar suggestions from Referee #1 and Dr. Schmittner, we have incorporated two additional hypotheses (see below) in the revised version of our manuscript, in order to have a more thorough and balanced discussion:

Pages 10-12, lines 10-2: *"5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump"*

Recent model experiments (e.g. Schmittner, 2005; Schmittner and Galbraith, 2008) have shown that AMOC slowdown events may cause a decrease in the global efficiency of the oceanic biological pump, being an important driver for the oceanic CO₂ outgassing within HS1 during the last deglaciation and possibly also during other HS, including HS3 and HS2.

NADW has low preformed nutrient waters because it is formed by nutrient depleted surface waters, where the biological pump has efficiently removed nutrients from surface waters (Marinov et al., 2008). AABW has high preformed nutrient waters because it is formed by nutrient-enriched Southern Ocean surface waters (nutrients have not being efficiently removed from surface waters). However, during weak AMOC two factors may alter the nutrient distribution and the global oceanic biological pump (Schmittner and Galbraith, 2008). First, the reduction in the NADW formation decreases the input of low preformed nutrient (high $\delta^{13}\text{C}_{\text{DIC}}$) waters to the ocean interior which becomes more dominated by high preformed nutrient (low $\delta^{13}\text{C}_{\text{DIC}}$) southern component waters (e.g., AABW). Second, the reduction of the Southern Ocean stratification induced by the decrease of salt input via NADW formation promotes the strengthening of the upwelling and subsequent sinking of high preformed nutrient (low $\delta^{13}\text{C}_{\text{DIC}}$) waters to the ocean interior, thus reducing the capacity of those unutilized nutrients to sequester carbon via the biological pump. The two factors acting in conjunction are thought to be responsible for the simulated weakening of the global efficiency of the biological pump, as well as for the increase in CO_{2atm} and decrease in $\delta^{13}\text{CO}_{2\text{atm}}$ (Schmittner and Galbraith, 2008; Schmittner and Lund, 2015; Hertzberg et al., 2016).

Schmittner and Lund (2015) show that the modeled weakening of the biological pump, induced by an AMOC slowdown, reduces the ability of the surface ocean to biologically sequester isotopically light organic carbon (¹²C), producing a decrease in the surface ocean $\delta^{13}\text{C}_{\text{DIC}}$ and an increase of the intermediate ocean $\delta^{13}\text{C}_{\text{DIC}}$ (lower remineralization rate). For HS1, planktonic and benthic foraminiferal $\delta^{13}\text{C}$ records (Tessin and Lund, 2013; Lund et al., 2015; Curry and Oppo, 2005; Hertzberg et al., 2016) from the western South Atlantic (ca. 27°S) agree with the model output by showing a decrease in $\delta^{13}\text{C}$ in the upper water column (SACW) and an increase at intermediate water depths (AAIW). Thus, the weakening of the global oceanic biological

*pump and consequent negative anomaly of the $\delta^{13}\text{C}_{\text{DIC}}$ in the upper water column should be captured by the tests of planktonic foraminifera $\delta^{13}\text{C}$ during calcification (Spero and Lea, 1996; Bemis et al., 2000). The negative $\delta^{13}\text{C}_{\text{DIC}}$ during HS3 and HS2 revealed by our planktonic foraminifera provide the first observational evidence supporting the modeling results. Additionally this mechanism also provides a possible explanation for the larger negative $\delta^{13}\text{C}$ anomaly in *G. ruber* w (mixed layer dwelling) relative to the anomaly in *G. inflata* (permanent thermocline dwelling) (Fig. 4).*

It is noteworthy that the mechanism described in section 5.1, although based on a different driver for the decrease in $\delta^{13}\text{C}$, also suggests that the decreases in $\delta^{13}\text{C}$ of planktonic foraminifera from the South Atlantic would be carried by SACW (inherited from its precursor, SAMW) and thus both mechanisms (described in section 5.1 and here) are in this regard not mutually exclusive. However, the mechanism described in the present section goes against the assumption that weakening of the biological pump is related to stronger upwelling in the Southern Ocean, and that the Southern Ocean would be the source of the low $\delta^{13}\text{C}$ signal for the South Atlantic (Lund et al., 2015; Hertzberg et al., 2016).

5.3 Millennial-scale changes: the role of air–sea gas exchange

The $\delta^{13}\text{C}_{\text{DIC}}$ of the surface ocean can also be affected by air–sea gas exchange (Oppo and Fairbanks, 1989; Charles and Fairbanks, 1990; Lynch-Stieglitz et al., 1995). Although this process tends towards isotopic equilibrium, the CO_2 exchange between the ocean and the atmosphere does not lead to equilibrium because CO_2 uptake and emission will still occur in different regions and the movement and replacement of surface waters is faster than required for equilibration (Lynch-Stieglitz et al., 1995). Since the $\delta^{13}\text{CO}_{2\text{atm}}$ is lighter than $\delta^{13}\text{C}_{\text{DIC}}$, at areas of ocean $\text{CO}_{2\text{atm}}$ uptake (i.e., water mass formation regions) air–sea gas exchange has the potential to deplete $\delta^{13}\text{C}_{\text{DIC}}$ (Lynch-Stieglitz et al., 1995). Additionally, the isotopic fractionation is inversely correlated with temperature.

Therefore, we cannot exclude the possibility that the likely decrease in $\delta^{13}\text{CO}_{2\text{atm}}$ during AMOC slowdown events (Eggleston et al. 2016) (e.g., HS3 and HS2) could have affected the $\delta^{13}\text{C}_{\text{DIC}}$ via air–sea gas exchange, especially in regions of water mass formation. The formation region of SACW is an area of ocean CO_2 uptake and may

contribute to the $\delta^{13}\text{C}$ anomalies observed in our *G. inflata* record (Fig. 4b). Additionally, since the isotopic fractionation during air–sea gas exchange is temperature–dependent the weakening of the AMOC and subsequent warming of the upper subtropical South Atlantic (Barker et al. 2009; Chiessi et al. 2015) could have contributed to the observed $\delta^{13}\text{C}$ anomalies both in the *G. ruber w* and in the *G. inflata* records (Fig. 4). However, the gradient is too small (-0.1‰ $\delta^{13}\text{C}$ per $^{\circ}\text{C}$, Broecker and Maier-Reimer, 1992) to explain the whole changes observed in our records. If temperature was the dominant driver, unrealistic changes between 5 and 13 $^{\circ}\text{C}$ would be required to explain the full amplitudes of the $\delta^{13}\text{C}$ variations.

The $\delta^{18}\text{O}$ records from *G. ruber w* and *G. inflata* from our core (Supplementary material Figure 1) should partially reflect changes in water temperature (ca. -0.22‰ per 1 $^{\circ}\text{C}$; e.g., Mulitza et al, 2003), but show no clear trends across HS3 and HS2. While temperature changes might be partially obscured in the foraminiferal $\delta^{18}\text{O}$ records by the influence of synchronous changes in seawater- $\delta^{18}\text{O}$, as has been hypothesized for the Holocene (Chiessi et al., 2014) and HS1 (Chiessi et al., 2015) in the western Atlantic, we consider it unlikely that temperature changes of the above magnitude would be completely masked.”

In the conclusions we also inserted the two additional hypotheses described above and removed the mention of the Southern Ocean as the main source of the CO_2 outgassed to the atmosphere during HS3 and HS2.

Page 13, lines 14-18: we added “(ii) Weakening of the global oceanic biological pump. A weak AMOC during HS3 and HS2 would promote an accumulation of ^{13}C –depleted CO_2 in the upper water column of the South Atlantic. This accumulation would result in a negative anomaly of the $\delta^{13}\text{C}_{\text{DIC}}$ (as well as of the $\delta^{13}\text{C}_{\text{CO}_{2\text{atm}}}$) that in turn would be captured by the tests of planktonic foraminifera at our core site. We further suggest that changes in air–sea gas exchange could have contributed to the decreases in $\delta^{13}\text{C}$ via both mechanisms.”.

Page 13, line 20: we removed “of the Southern Ocean”.

However, for the reasons stated below we still consider the possible role played by changes in Southern Ocean ventilation in producing our $\delta^{13}\text{C}$ anomalies:

- iii. the mechanisms that invoke the Southern Ocean as one of the CO_{2atm}-sources during the last deglaciation are based on reconstructions as well as conceptual and numerical models that, to our understanding, have not yet been negated (Toggweiler et al., 2006; Anderson et al., 2009; Sigman et al., 2010; Tschumi et al., 2011; Lee et al., 2011; Burke and Robinson, 2012);
- iv. model experiments that question the role of the Southern Ocean (e.g., Schmittner and Lund, 2015) were performed under preindustrial boundary conditions, which can significantly influence the results (e.g., Menviel et al., 2008), and do not constitute a firm negation of the Southern Ocean hypothesis.

The authors also invoke speculative connections between rainfall and sedimentation rate in the core as support for their climate interpretation, despite disagreement between the sedimentation rate and planktonic d13C patterns.

Comment #12 - We agree that the second HS2 peak in sedimentation rate and the respective planktonic $\delta^{13}\text{C}$ anomaly are not perfectly aligned in time. However, this apparent offset can be exclusively due to (i) the occurrence of ^{14}C plateaus during HS (e.g., Sarnthein et al., 2007; Franke et al., 2008), and/or (ii) the discretized way our age model was produced in relation to the “continuous” $\delta^{13}\text{C}$ measurements.

Finally, the authors neglect to mention much of the work at the Brazil Margin spanning the last deglaciation (including Heinrich Stadial 1) that is both relevant to their work and inconsistent with a Southern Ocean driver.

Comment #13 - We have included relevant publications from the Brazil Margin (e.g., Tessin and Lund, 2013; Lund et al. 2015) in the revised version of our manuscript together with a thorough and more balanced discussion (please see our Comment #11 above). However, the hypothesis that the Southern Ocean acts as a possible CO_{2atm}-source during HS has, to the best of our knowledge, not yet been proven incorrect and we still see the need to discuss it together with other possible explanations (please see our Comment #11 above).

Major issues:

While the prevailing view is that Southern Ocean outgassing drove surface ocean d13C anomalies during the last deglaciation, the authors also need to reference to Andreas Schmittner’s work that shows that weakening of the AMOC can alter the preformed

nutrient budget of the global ocean and therefore the efficiency of the biological pump. Weakening of the biopump would preferentially leave light carbon in the surface and create negative $\delta^{13}\text{C}$ anomalies in multiple ocean basins (e.g. Schmittner, 2005; Schmittner and Galbraith, 2008).

Also, detailed reconstructions from the Brazil Margin show that benthic $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ changed late in the deglaciation, suggesting that the abyssal ocean was an unlikely source for the surface ocean anomalies during HS1 (Lund et al., 2015). Additional published records from the SE Atlantic shows a similar pattern, with a late deglacial response in the abyssal records (Waelbroeck et al., 2011; Roberts et al., 2016).

Comment #14 – Considering that the role of the weakening of the biological pump was also raised by Referee #1 and Dr. Schmittner, we have incorporated this hypothesis in the revised version of our manuscript (please see our Comment #11 above).

Our records complement the Brazil Margin depth transect (e.g., Tessin and Lund, 2013; Lund et al., 2015; Hertzberg et al., 2016) with upper water column high temporal resolution data that also agree with the model output of Schmittner and Lund (2015) by suggesting a decrease in $\delta^{13}\text{C}_{\text{DIC}}$ at the upper water column of the western South Atlantic.

In the revised version of our manuscript, we suggest that two processes may have contributed to the negative $\delta^{13}\text{C}$ anomalies in our records during HS3 and HS2: (i) strengthening of Southern Ocean deep water ventilation and (ii) weakening of the biological pump. Additionally, we suggest that air–sea gas exchange could have contributed to the observed $\delta^{13}\text{C}$ decreases.

Furthermore, intermediate depth $\delta^{13}\text{C}$ reconstructions from the Brazil Margin (Oppo and Horowitz, 2000; Curry and Oppo, 2005; Lund et al., 2015) imply that AAIW and SAMW were not carriers of a light carbon signal during HS1, which is inconsistent with the mechanism invoked by Campos et al. While the authors did a commendable job summarizing the literature pertaining the modern oceanographic setting, the introductory material includes surprisingly little information from previously published work along the Brazil Margin. As a result, the introductory material is incomplete and the lack of context limits the interpretation in the discussion section.

Comment #15 - To account to this additional mechanism in the revised manuscript we have inserted a new section describing the AMOC-induced weakening of the biological pump (please see our Comment #11 above).

The other main issue with the data interpretation is that the continental hydroclimate response isn't well supported by the sedimentation rate data. While sedimentation rates peak early in HS2, at about the same time as a planktonic $\delta^{13}\text{C}$ minimum, there is a second peak in sedimentation rate after HS2 that corresponds to a broad maximum in $\delta^{13}\text{C}$. Furthermore, there is no peak in sedimentation rate during either HS3 or HS1. Taken as a whole, the sedimentation rate data therefore do not support a continental hydroclimate connection, which perhaps isn't surprising given the many factors that can influence sedimentation rates at a single core site.

Comment #16 - Please see our Comment #12 above. Additionally, although HS3 seems not to be related to a positive anomaly in sedimentation rate at our core site, HS1 and the Younger Dryas are both related to positive anomalies (of smaller amplitude during the Younger Dryas if compared to the anomalies during HS2 and HS1). The relatively high sea-level during HS3 and the Younger Dryas if compared to the intervening period (Waelbroeck et al., 2002) may have dampened a more significant anomaly in sedimentation rates during HS3 and the Younger Dryas (e.g., Lantzsich et al., 2014). This discussion has not been inserted in the revised manuscript because we have focused in HS3 and HS2.

Finally, the manuscript would benefit from inclusion of the planktonic $\delta^{18}\text{O}$ data. If there is a clear hydroclimate signal at this site, it could appear in the $\delta^{18}\text{O}$ data as intervals of unusually low $\delta^{18}\text{O}$. The $\delta^{18}\text{O}$ data would also be informative for assessing the influence of air-sea gas exchange effects on the $\delta^{13}\text{C}$ signal. The $\delta^{13}\text{C}$ of surface ocean DIC is influenced by temperature dependent gas exchange, with a relationship of -0.1 per mil per deg C of warming (Broecker and Maier-Reimer, 1992; Lynch-Stieglitz et al., 1995). While such an effect is unlikely to explain the full magnitude of the $\delta^{13}\text{C}$ signals, it should be included in the discussion. Model results suggest that weakening of the AMOC warms the upper subtropical South Atlantic by 2-3 deg C (Marcotte et al., 2011) which would account for up to 0.3 per mil of the planktonic $\delta^{13}\text{C}$ anomalies.

Comment #17 - A similar topic was also raised by Referee #1. Please refer to our Comment #5 to Referee #1. Additionally, a section expanding the discussion on the possible role of air–sea gas exchange is included in the revised version of our manuscript. This mechanism is described in our Comment #11 above.

Regarding $\delta^{18}\text{O}$ data we inserted this data in Supplementary material (Figure 1). It is noteworthy that our foraminiferal $\delta^{18}\text{O}$ data do not present clear trends during HS3 and HS2. The $\delta^{18}\text{O}$ of planktonic foraminifera is markedly influenced by the temperature and seawater– $\delta^{18}\text{O}$ of the sea water where the tests calcified (e.g., Rohling and Cooke, 1999). The relationship between temperature and $\delta^{18}\text{O}$ of planktonic foraminifera is inversely proportional, and the relationship between seawater– $\delta^{18}\text{O}$ and $\delta^{18}\text{O}$ of planktonic foraminifera is directly proportional. Therefore, higher temperatures decrease foraminiferal $\delta^{18}\text{O}$, whereas higher seawater– $\delta^{18}\text{O}$ increases foraminiferal $\delta^{18}\text{O}$ (e.g., Rohling and Cooke, 1999).

For periods and regions of the global ocean where an increase (decrease) in temperature is associated to a decrease (increase) in seawater– $\delta^{18}\text{O}$, the $\delta^{18}\text{O}$ of planktonic foraminifera will clearly record such environmental changes (e.g., Guilderson and Pak, 2007). However, for those cases where temperature and seawater– $\delta^{18}\text{O}$ increase (decrease) simultaneously, these environmental changes may not be clearly recorded in foraminiferal $\delta^{18}\text{O}$. The last scenario seems to be the case for the upper water column under the influence of the Brazil Current for centennial-scale events of the Holocene (Chiessi et al., 2014), as well as for HS1 (Chiessi et al., 2015), and possibly also for HS3 and HS2 (given the lack of regional SST reconstructions for HS3 and HS2).

Pages 11-12, lines 29-2: *“The $\delta^{18}\text{O}$ records from *G. ruber w* and *G. inflata* from our core (Supplementary material Figure 1) should partially reflect changes in water temperature (ca. -0.22% per $1\text{ }^{\circ}\text{C}$; e.g., Mulitza et al, 2003), but show no clear trends across HS3 and HS2. While temperature changes might be partially obscured in the foraminiferal $\delta^{18}\text{O}$ records by the influence of synchronous changes in seawater– $\delta^{18}\text{O}$, as has been hypothesized for the Holocene (Chiessi et al., 2014) and HS1 (Chiessi et al., 2015) in the western Atlantic, we consider it unlikely that temperature changes of the above magnitude would be completely masked.”*

Minor issues:

Page 6, Line 12: Here the authors should consider the possibility that weakening of the biological pump could have produced $\delta^{13}\text{C}$ anomalies not only locally, but on a larger spatial scales.

Comment #18 - Agreed. We have changed the revised version of our manuscript accordingly, but have not inserted the suggested change at this passage. Rather, we inserted the suggested change to pages 10-11, lines 10-11 (please see our Comment #11 above).

Page 6, Line 20: It is important to mention that published mode and intermediate water records from the Brazil Margin either show a positive $\delta^{13}\text{C}$ anomaly during HS1 or a delayed negative $\delta^{13}\text{C}$ response that is inconsistent the light carbon being transported northward via mode and intermediate water.

Comment #19 - Agreed (please see our Comment #11 above). We have revised the manuscript accordingly. Additionally, we also incorporate two more hypotheses (i.e., 5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump, and 5.3 Millennial-scale changes: the role of air-sea gas exchange) to the revised version of our manuscript, in order to have a more thorough and balanced discussion.

Page 7, Line 5: The authors seem to take it as a given that Southern Hemisphere westerlies drives greater upwelling of deep waters when it actuality there is limited data to support such a link. Please qualify these types of statements to reflect the limited constraints that exist.

Comment #20 – Agreed (please see our Comment #8 to Referee #1). We have revised our manuscript accordingly.

Page 7, Line 9: See comments above that published intermediate depth records from the Brazil Margin don't show clear negative carbon isotope anomalies during HS1, which is relevant to whether a similar process occurred during HS2 and HS3.

Comment #21 - Agreed (please see our Comment #11 above). We have revised our manuscript accordingly. Additionally we also incorporate two more hypotheses (i.e., 5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump, and 5.3 Millennial-scale changes: the role of air-sea gas exchange) to the revised version of our manuscript, in order to have a more thorough and balanced discussion.

Figure 4: If the authors are going to pursue the sedimentation rate-hydroclimate link, then the sedimentation rates should also be included in this figure for comparison to the $\delta^{13}\text{C}$. The $\delta^{18}\text{O}$ records should also be included to assess the potential impact of sea surface temperature on $\delta^{13}\text{C}$.

Comment #22 - Agreed. We have inserted the sedimentation rate record in Figure 5 of the revised version of our manuscript. Regarding the $\delta^{18}\text{O}$ data, please refer to our Comment #5 to Referee #1 and our Comment #17 above.

Figure 5: There are several time series in this plot that aren't essential for the discussion, such as the Iberian Margin SST record, the EDML $\delta^{18}\text{O}$ time series, and the Siple Dome CO_2 .

Comment #23 - We agree that removing the $\delta^{18}\text{O}$ EDML curve (EPICA Community Members, 2006) would not harm the main focus of our manuscript. In addition, we have also removed the Taylor Dome $\delta^{13}\text{CO}_{2\text{atm}}$ curve (Smith et al., 1999) for the reason mentioned in our Comment #7 to Referee #1.

However, we would prefer not to delete the following curves: (i) the Iberian Margin SST curve (Bard, 2002) because it shows a “w-structure” during HS2 that gives support to the proposed “w-structure” of HS2; and (ii) the Siple Dome $\text{CO}_{2\text{atm}}$ (Ahn and Brook, 2014) due to its fundamental importance to our manuscript (please see our Comment #6 to Referee #1).

1.3 Response to Dr. Schmittner

We thank Dr. Schmittner for his constructive comment on our manuscript. To facilitate the discussion, we have copied his comment below in black and inserted our response in green.

I suggest to the authors to consider our relevant recent modeling work, which suggests a different mechanism for the CO_2 increase during H-events. As shown in Schmittner and Galbraith (2008, *Nature*, 456, 373-376, doi:10.1038/nature07531) an AMOC shutdown causes a decrease of the efficiency of the biological pump, which leads to an increase in atmospheric CO_2 consistent in both amplitude and rate-of-change with ice core observations. Schmittner and Lund (2015; *Climate of the Past*, 11, 135-152,

doi:10.5194/cp-11-135-2015) show that this leads to a decrease of surface ocean (and atmospheric) $\delta^{13}\text{C}$ that is particularly strong (more than 0.5 permil) in the South Atlantic (their Fig. 5G).

Comment #1 - A similar observation was also raised by Referee #1 and Referee #2. Please refer to our Comment #1 to Referee #1 or Comment #11 to Referee #2.

2. Additional relevant changes performed in the manuscript

After implementing the changes suggested by the Referees and by Dr. Schmittner, we identified the need to refine specific passages of the manuscript. Those relevant changes are shown below in italic green in quotes.

2.1 Changes performed in the title

Page 1, lines 1-2: the title was changed to “ $\delta^{13}\text{C}$ decreases in the upper western South Atlantic during Heinrich Stadials 3 and 2”.

2.2 Relevant changes performed in the abstract

Page 1, lines 12-13: we added “($\text{CO}_{2\text{atm}}$) and decreases of its stable carbon isotopic ratios ($\delta^{13}\text{C}$), i.e., $\delta^{13}\text{CO}_{2\text{atm}}$ ”.

Page 1, line 13: we exchanged the term “Southern Ocean” by “ocean”.

Page 1, lines 16-17: we removed “intensification in Southern Ocean deep water ventilation presumably associated with a”.

Page 1, lines 18-21: we added “and the consequent increase (decrease) in $\text{CO}_{2\text{atm}}$ ($\delta^{13}\text{CO}_{2\text{atm}}$). We hypothesise two mechanisms that could account for the decreases observed in our records, namely strengthening of Southern Ocean deep water ventilation and weakening of the biological pump. Additionally, we suggest that air–sea gas exchange could have contributed to the observed $\delta^{13}\text{C}$ decreases.”.

Page 1, lines 21-22: we removed “*After reaching the upper water column of the Southern Ocean, the $\delta^{13}\text{C}$ depletion would be transferred equatorward via central and thermocline waters.*”.

Page 1, line 24: we removed “*from the Southern Ocean*”.

Page 1, line 29-30: we exchanged the term “*Southern Ocean*” by “*Atlantic Meridional Overturning Circulation*”.

2.3 Relevant changes performed in the introduction

Page 2, lines 10-11: we added “*This increase in $\text{CO}_{2\text{atm}}$ was accompanied by a decrease of its stable carbon isotopic composition ($\delta^{13}\text{CO}_{2\text{atm}}$) (Eggleston et al., 2016).*”.

Page 2, lines 11-12: we added “*and the associated $\delta^{13}\text{CO}_{2\text{atm}}$ decrease*”

Page 2, lines 20-26: we added “*However, model experiments (e.g., Schmittner and Galbraith, 2008; Schmittner and Lund, 2015) and records from different ocean basins (e.g., Tessin and Lund, 2013; Lund et al., 2015; Curry and Oppo, 2005; Hertzberg et al., 2016) suggest that the increase in $\text{CO}_{2\text{atm}}$ and decrease in $\delta^{13}\text{CO}_{2\text{atm}}$ during HS1 are instead related to the weakening of the global oceanic biological pump and the consequent accumulation of ^{13}C -depleted CO_2 in the upper water column. Such anomalously low- $\delta^{13}\text{CO}_2$ would then be outgassed to the atmosphere. Since isotopic fractionation between reservoirs is temperature-dependent the air-sea gas exchange during HS1 could have additionally modified $\delta^{13}\text{CO}_{2\text{atm}}$ (Lynch-Stieglitz et al., 1995).*”.

Pages 2-3, lines 27-1: we removed “*The CDW forms from mixing of North Atlantic Deep Water (NADW), Indian Deep Water (IDW) and Pacific Deep Water (PDW) and upwells to the south of the Antarctic Polar Front driven by the prevailing westerly winds (Marshall and Speer, 2012). Therefore, the $\delta^{13}\text{C}$ signal of CDW (ca. 0.4‰) (Kroopnick, 1985) lies between that of NADW (ca. 1‰) (Kroopnick, 1985) and IDW/PDW (ca. 0.2 to -0.2‰) (Kroopnick, 1985) (Oppo and Fairbanks, 1987; Charles and Fairbanks, 1992). During periods of weak AMOC the influence of NADW on the Southern Ocean is reduced (Charles and Fairbanks, 1992), and the $\delta^{13}\text{C}$ of CDW should decrease since this water mass would have a relatively larger contribution from*

low- $\delta^{13}\text{C}$ IDW and PDW. Thus, the strengthening in the Southern Ocean upwelling would bring old, respired ^{13}C -depleted waters to the surface that would then be transferred equatorward via central and thermocline waters (Spero and Lea, 2002).”.

Page 3, lines 1-2: we added *“The reduction of the upper water column $\delta^{13}\text{C}$ caused by one or both of the”.*

Page 3, lines 2-3: we removed *“(i.e., a weak AMOC leading to a strengthening of the Southern Ocean upwelling, a $\delta^{13}\text{C}$ depletion and an increased outgassing of CO_2 to the atmosphere)”.*

Page 3, lines 11-12: we removed *“most likely caused by millennial-scale strengthening of Southern Ocean upwelling”.*

2.4 Relevant changes performed in the regional settings

Page 3, line 24: we added *“In the study area”.*

Page 3, lines 24-25: we added *“and Antarctic Intermediate Water (AAIW)”.*

Page 4, lines 12-13: we added *“the dissolved inorganic carbon $\delta^{13}\text{C}$ ($\delta^{13}\text{C}_{\text{DIC}}$) in”.*

Page 4, line 16: we added *“In the southwest South Atlantic”.*

2.5 Relevant changes performed in the materials and methods

Page 5, lines 19-22: we removed *“Default parameter settings were used, except for mem.mean (set to 0.4) and acc.shape (set to 0.5). Ages are modelled as drawn from a t-distribution, with 9 degrees of freedom (t.a=9, t.b=10). 1,000 age-depth realizations were used to estimate mean age and 95 % confidence intervals at 0.5 cm resolution (Fig. 3).”.*

Page 5, lines 22-25: we added *“With the exception of mem.mean (set to 0.4) and acc.shape (set to 0.5), default parameters were used. Radiocarbon ages were assumed to be t-distributed with 9 degrees of freedom (t.a.=9, t.b=10). Mean ages and 95% error*

margins were estimated from 10,000 downcore age-depth realizations at 0.5 cm resolution (Fig. 3).”.

Page 5, lines 24: in order to improve our age model we increased the number of age-depth realizations modeled within the software Bacon 2.2 from “1,000” to “10,000”.

2.6 Relevant changes performed in the results

Page 6, line 7: the section title was changed to “*Age model and sedimentation rates*”.

Page 6, line 8: we exchanged “32.6” by “32”.

Page 6, line 8: we exchanged “5.7” by “6”.

Page 6, line 9: we exchanged “3.8” by “5”.

Page 6, line 9: we exchanged “111” by “91”.

Page 6, lines 11-12: we removed “*and received special attention due to the higher sedimentation rate which provides increased temporal resolution*”.

Page 6, line 14: the section title was changed to “*Stable carbon isotope values of G. ruber and G. inflata*”.

Page 6, line 15: we exchanged “32.6” by “32”.

Page 6, line 15: we exchanged “28.5” by “28.2”.

Page 6, line 16: we exchanged “24.8” by “24.9”.

Page 6, line 17: we exchanged “27” by “27.2”.

Page 6, line 18: we exchanged “30.6” by “29.3”.

Page 6, line 18: we exchanged “30.4” by “29.1”.

Page 6, line 23: we exchanged “32.5” by “31.5”.

Page 6, line 23: we exchanged “30.6” by “29.3”.

Page 6, line 23: we exchanged “29.8” by “28.8”.

Page 6, line 23: we exchanged “28.3” by “28”.

2.7 Relevant changes performed in the discussion

Page 7, lines 4-12: we added “Concomitantly, a weak AMOC was described based on $^{231}\text{Pa}/^{230}\text{Th}$ records from the Bermuda Rise (ODP Site 1063, 33.7° N, 57.6° W) (Lippold et al., 2009) (Fig. 5d). Both events are also marked by pulses of ice-rafted debris (IRD) (MD99-2331, 42.2° N, 9.7° W) (Eynaud et al., 2009) and by decreases in SST (SU8118 and MD952042, 37.5° N, 10.1° W) (Bard, 2000) in the north–eastern North Atlantic (Iberian Margin). The Greenland GISP2 ice core (72.6° N, 38.5° W) shows synchronous increases in Ca^{+2} , indicating changes in atmospheric circulation over Greenland (Mayewski et al., 1997) (Fig. 5a, b, c). It is noteworthy that the four records (i.e., Fig. 5a, b, c, d) mentioned above also show a “w–structure” during HS2, similar to the one shown in our $\delta^{13}\text{C}$ records. The IRD (Eynaud et al., 2009) and Ca^{+2} (Mayewski et al., 1997) records also show a “w–structure” similar to ours during HS3.”

Page 7, line 13: we added “local”.

Page 7, lines 26-29: we added “We propose two primary mechanisms to explain our $\delta^{13}\text{C}$ decreases: (i) changes in the strength of Southern Ocean deep water ventilation (detailed in section 5.1), and (ii) the weakening of the global oceanic biological pump (detailed in section 5.2). Additionally, air–sea gas exchange may have acted as a secondary factor contributing to our $\delta^{13}\text{C}$ decreases (detailed in section 5.3).”.

Pages 7-8, lines 30-6: we removed “A pervasive feature of planktonic foraminiferal $\delta^{13}\text{C}$ records in the Indo–Pacific Ocean (Spero and Lea, 2002), Southern Ocean (Ninnemann and Charles, 1997), and South Atlantic Ocean (Oppo and Fairbanks, 1989) is a negative excursion during HS1. Ninnemann and Charles (1997) suggested that the source for this signal is in the Southern Ocean. They further proposed that the anomaly is related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the Southern Ocean via SAMW and/or AAIW. A low–density type of SAMW actually contributes to SACW that spreads into the South Atlantic (Stramma and England, 1999). Additionally, AAIW also influences SACW through vigorous eddy mixing at the Brazil/Malvinas Confluence

(Piola and Georgi, 1982). Thus, SACW represents a potential conduit for the $\delta^{13}\text{C}$ signal from the sub-Antarctic region to the subtropical South Atlantic (Fig. 1 and 2). Therefore, we propose that the negative excursions in our $\delta^{13}\text{C}$ records are related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the subantarctic zone to the western South Atlantic via central and thermocline waters.”.

Page 8, lines 7-8: the section title was changed to “*Millennial-scale changes: AMOC-induced strengthening of Southern Ocean deep water ventilation*”.

Page 8, lines 9-20: we added “*A negative excursion during HS1 was described in planktonic foraminiferal $\delta^{13}\text{C}$ records from the Indo-Pacific Ocean (Spero and Lea, 2002), Southern Ocean (Ninnemann and Charles, 1997), and South Atlantic Ocean (Oppo and Fairbanks, 1989). Ninnemann and Charles (1997) suggested that the source for this signal was the Southern Ocean.*

In the Southern Ocean CDW forms from mixing of NADW, Indian Deep Water (IDW) and Pacific Deep Water (PDW) and upwells to the south of the Antarctic Polar Front driven by the prevailing westerly winds (Marshall and Speer, 2012). Therefore, the $\delta^{13}\text{C}$ signature of CDW (ca. 0.4‰) (Kroopnick, 1985) lies between that of NADW (ca. 1‰) (Kroopnick, 1985) and IDW/PDW (ca. 0.2 to -0.2‰) (Kroopnick, 1985) (Oppo and Fairbanks, 1987; Charles and Fairbanks, 1992). During periods of weak AMOC the inflow of NADW to the Southern Ocean is reduced (Charles and Fairbanks, 1992), and the $\delta^{13}\text{C}$ of CDW should decrease since the latter would have a relatively larger contribution from low- $\delta^{13}\text{C}$ IDW and PDW (Spero and Lea, 2002).”.

Page 8, lines 23-31: we removed “*Concomitantly to the periods of low $\delta^{13}\text{C}$ in our records a weak AMOC was described based on $^{231}\text{Pa}/^{230}\text{Th}$ records from the Bermuda Rise (ODP Site 1063, 33.7° N, 57.6° W) (Lippold et al., 2009) that characterize HS3 and HS2 (Fig. 5d). Both events are also marked by pulses of ice-rafted debris (IRD) (MD99-2331, 42.2° N, 9.7° W) (Eynaud et al., 2009) and by decreases in SST (SU8118 and MD952042, 37.5° N, 10.1° W) (Bard, 2002) in the north-eastern North Atlantic (Iberian Margin). The Greenland GISP2 ice core (72.6° N, 38.5° W) shows synchronous increases in Ca^{+2} , indicating changes in atmospheric circulation over Greenland (Mayewski et al., 1997) (Fig. 5a, b, c). It is worth to note that the four records (i.e., Fig. 5a, b, c, d) mentioned above also show a w-structure during HS2,*

similar to the one shown in our $\delta^{13}\text{C}$ records. The IRD (Eynaud et al., 2009) and Ca^{+2} (Mayewski et al., 1997) records also show a w-structure similar to ours during HS3.”.

Page 9, lines 5-7: we added *“A recent model experiment (Bauska et al., 2016) corroborates this hypothesis by showing that stronger Southern Ocean upwelling would promote a weakening of the biological pump in the Southern Ocean.”.*

Page 9, lines 11-13: we added *“Actually, a low-density type of SAMW contributes to SACW that spreads into the South Atlantic (Stramma and England, 1999). Additionally, AAIW also influences SACW through vigorous eddy mixing at the Brazil/Malvinas Confluence (Piola and Georgi, 1982).”.*

Page 9, lines 15-17: we removed *“However, we cannot exclude the possibility that the upwelled low- $\delta^{13}\text{C}$ respired CO_2 could have been first outgassed from the Southern Ocean, and then re-dissolved into the ocean via air-sea exchanges at the formation regions of SACW and TW, eventually reaching the upper water column at our core site.”.*

Page 9, lines 23-26: we removed *“Millennial-scale changes of the Southern Ocean temperature and deep water ventilation also led to the increase in $\text{CO}_{2\text{atm}}$ (Spero and Lea, 2002; Ahn and Brook, 2008; Ahn and Brook, 2014; Gottschalk et al., 2015). During HS3 and HS2, positive peaks in $\text{CO}_{2\text{atm}}$ (Siple Dome, 81.7°S , 148.8°W) (Ahn and Brook, 2014) (Fig. 5k) were described to be synchronous to circum-Antarctic warming (EPICA Community Members, 2006) (Fig. 5i), and most likely have originated from”.*

Page 9, lines 29-30: we added *“and the consequent increase in $\text{CO}_{2\text{atm}}$ (Siple Dome, 81.7°S , 148.8°W) (Ahn and Brook, 2014) (Fig. 5j)”.*

Page 9, lines 31-32: we removed *“even the $\delta^{13}\text{C}$ record of $\text{CO}_{2\text{atm}}$ at Taylor Dome (77.8°S , 158.7°E) on Antarctica shows a decrease during HS2 (Smith et al., 1999) (Fig. 5j).”.*

Page 10, lines 2-3: we removed *“An alternative explanation relates to a possible time lag between the weakening of the AMOC and the increase in $\text{CO}_{2\text{atm}}$ (Ahn and Brook, 2014)”.*

Page 10, lines 4-5: we added “*Therefore, the negative excursions in our $\delta^{13}\text{C}$ records could be related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the subantarctic zone to the western South Atlantic via central and thermocline waters.*”.

Page 10, lines 6-9: we removed “*Thus, our records are consistent with the hypothesis that the increase in $\text{CO}_{2\text{atm}}$ during abrupt millennial-scale climate change events of the last glacial period is originated by ocean processes (Smith et al., 1999; Ahn and Brook, 2008; Bereiter et al., 2012; Ahn and Brook, 2014) and is most likely related to a weak AMOC and associated strengthened Southern Ocean upwelling.*”.

Page 12, line 3: the section title was changed to “*Changes in continental climate*”.

Page 13, lines 3-5: we removed “*However, we discard this possibility because the expected signal of stronger local primary productivity on planktonic foraminiferal $\delta^{13}\text{C}$ would be opposite to the one observed at GeoB6212-1 (Mulitza et al., 1999).*”.

2.9 Relevant changes performed in the acknowledgements

Page 13, lines 28-1: we added “*We thank two anonymous reviewers and A. Schmittner for constructive comments that greatly improved this manuscript.*”.

2.10 Relevant changes performed in table 1

Page 22, table line 1: we exchanged “5694” by “5929”.

Page 22, table line 2: we exchanged “10112” by “10047”.

Page 22, table line 3: we exchanged “11345” by “11395”.

Page 22, table line 4: we exchanged “14374” by “14342”.

Page 22, table line 5: we exchanged “15179” by “15247”.

Page 22, table line 6: we exchanged “18616” by “18620”.

Page 22, table line 7: we exchanged “20731” by “20748”.

Page 22, table line 8: we exchanged “21843” by “21834”.

Page 22, table line 9: we exchanged “23511” by “23537”.

Page 22, table line 10: we exchanged “25583” by “25587”.

Page 22, table line 11: we exchanged “26164” by “26184”.

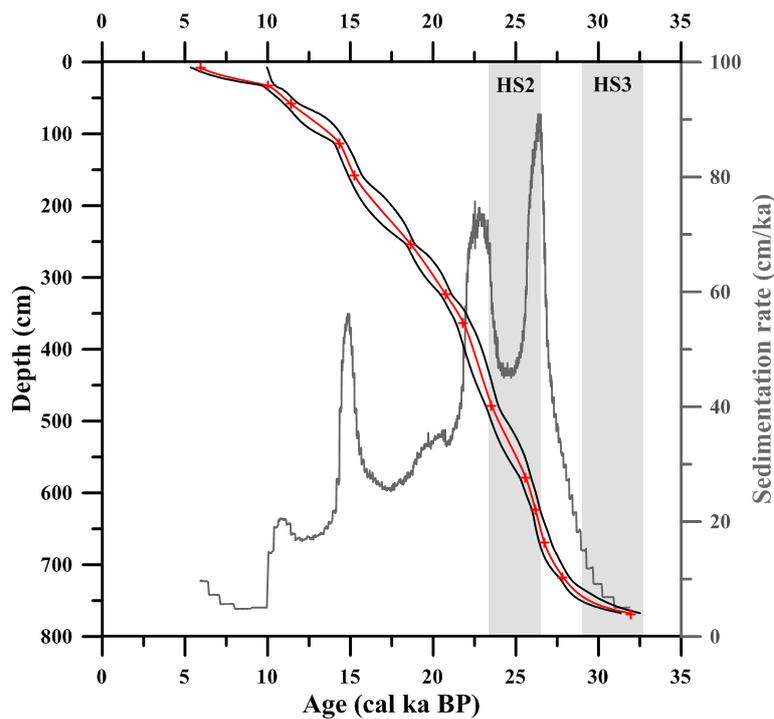
Page 22, table line 12: we exchanged “26645” by “26711”.

Page 22, table line 13: we exchanged “27894” by “27850”.

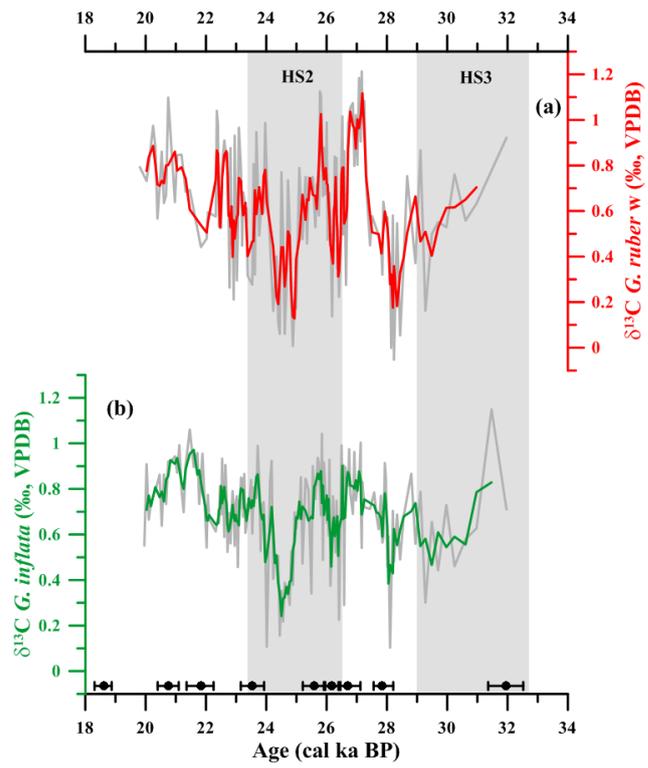
Page 22, table line 14: we exchanged “32583” by “31966”.

2.11 Relevant changes performed in the figures

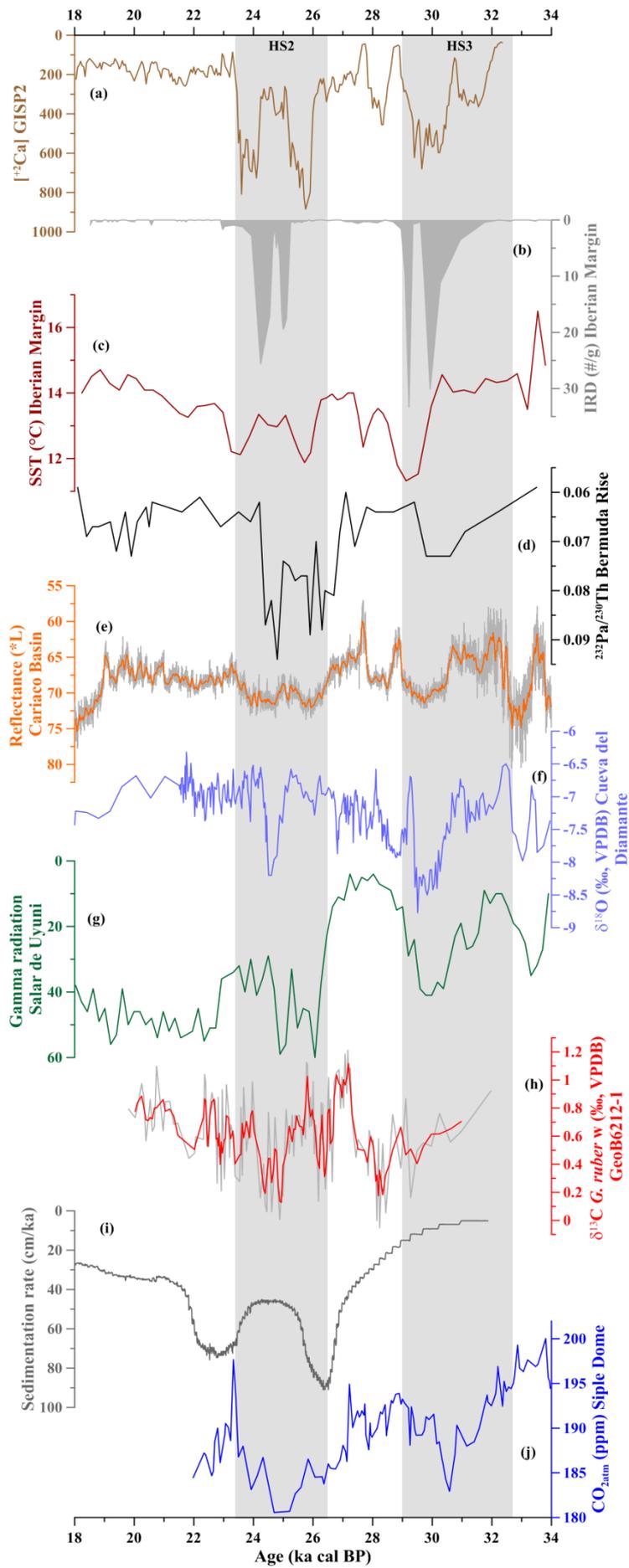
Page 26: we exchanged Figure 3 by



Page 28: we exchanged Figure 4 by



Page 29: we exchanged Figure 5 by



Page 30, lines 16-17: we added “*sedimentation rates from marine sediment core GeoB6212-1 collected in the western South Atlantic at 32.4° S, 50.1° W (this study)*”.

Page 30, lines 17-20: we removed “*atmospheric temperature over Antarctica indicated by EPICA Dronning Maud Land (EPICA Community Members, 2006) $\delta^{18}O$ (higher values indicate warmer conditions) plotted versus its original chronology at 75° S, 0° E; (j) $\delta^{13}C$ of atmospheric CO_2 ($\delta^{13}CO_{2atm}$) from Taylor Dome (lower values indicate increased input to the atmosphere of $\delta^{13}C$ -depleted respired CO_2) at 77.8° S, 158.7° E (Smith et al., 1999); (k)*”.

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3. Marked up version of the manuscript

In this section we inserted the marked up revised version of the manuscript. The changes performed to the manuscript are marked up in purple and green.

$\delta^{13}\text{C}$ decreases in the upper western South Atlantic during Heinrich Stadials 3 and 2

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Abstract. Abrupt millennial-scale climate change events of the last deglaciation (i.e., Heinrich Stadial 1 and the Younger Dryas) were accompanied by marked increases in atmospheric CO_2 ($\text{CO}_{2\text{atm}}$) and decreases of its stable carbon isotopic ratios ($\delta^{13}\text{C}$), i.e., $\delta^{13}\text{CO}_{2\text{atm}}$, presumably ~~originated by~~ due to outgassing from the ~~ocean~~ Southern Ocean. However, information on the preceding Heinrich Stadials during the last glacial period is scarce. Here we present ~~stable carbon isotopic data~~ ($\delta^{13}\text{C}$) records from two species of planktonic foraminifera from the western South Atlantic that reveal major decreases (up to 1‰) during Heinrich Stadials 3 and 2. These $\delta^{13}\text{C}$ decreases are most likely related to millennial-scale periods of ~~intensification in Southern Ocean deep water ventilation presumably associated with a weakening of the~~ Atlantic meridional overturning circulation ~~and the consequent increase (decrease) in $\text{CO}_{2\text{atm}}$ ($\delta^{13}\text{CO}_{2\text{atm}}$).~~ We hypothesise two mechanisms that could account for the decreases observed in our records, namely strengthening of Southern Ocean deep water ventilation and weakening of the biological pump. Additionally, we suggest that air-sea gas exchange could have contributed to the observed $\delta^{13}\text{C}$ decreases. After reaching the upper water column of the Southern Ocean, the $\delta^{13}\text{C}$ depletion would be transferred equatorward via central and thermocline waters. Together with other lines of evidence, Our data are consistent with the hypothesis that the CO_2 added to the atmosphere during abrupt millennial-scale climate change events ~~of~~during the last glacial period also originated in the ocean and reached the atmosphere by outgassing ~~from the Southern Ocean~~. The temporal evolution of $\delta^{13}\text{C}$ during Heinrich Stadials 3 and 2 in our records is characterized by two relative minima separated by a relative maximum. This “w-structure” is also found in North Atlantic and South American records, ~~giving us confidence~~ further suggesting that such structure is a pervasive feature of Heinrich Stadial 2 and, possibly, also Heinrich Stadial 3.

30 **Keywords:** Planktonic Foraminifera. Stable Carbon Isotopes. Heinrich Stadials. ~~Southern Ocean~~ Atlantic meridional overturning circulation.

1 Introduction

Heinrich Stadials (HSs) are abrupt millennial-scale climate change events marked by an anti-phase interhemispheric temperature pattern which is usually termed the bipolar seesaw (Broecker, 1998). One widely accepted mechanism for the bipolar seesaw is related to changes in the strength of the Atlantic meridional overturning circulation (AMOC), likely caused by fresh water input into the high latitudes of the North Atlantic (Mix et al., 1986; Crowley, 1992; Stocker, 1998). During HSs, a weak AMOC occurred simultaneously ~~to~~with cooling in the high latitudes of the surface North Atlantic (Sachs and Lehman, 1999; Bard et al., 2000), warming of the surface South Atlantic (Barker et al., 2009; Chiessi et al., 2015), a southward migration of the Intertropical Convergence Zone (ITCZ) (Arz et al., 1998; Deplazes et al., 2013), strengthening of the South American monsoon system (SAMS) (Cruz et al., 2006; Kanner et al., 2012), and an increase in atmospheric CO₂ (CO_{2atm}) (Ahn and Brook, 2008; Ahn and Brook, 2014). This increase in CO_{2atm} was accompanied by a decrease of its stable carbon isotopic composition ($\delta^{13}\text{CO}_{2atm}$) (Eggleston et al., 2016). It has been suggested that the origin of the CO_{2atm} rise and the associated $\delta^{13}\text{CO}_{2atm}$ decrease was ocean-sourced (Schmittner and Galbraith, 2008; Anderson et al., 2009; Denton et al., 2010; Mariotti et al., 2016; Eggleston et al., 2016; Hertzberg et al., 2016). The occurrence of stable carbon isotope ($\delta^{13}\text{C}$) minima during HS1 (last deglaciation) in planktonic foraminiferal records from the Indo-Pacific Ocean, Southern Ocean, and South Atlantic Ocean (Oppo and Fairbanks, 1989; Ninnemann and Charles, 1997; Mulitza et al., 1999; Spero and Lea, 2002) suggests that the signal ~~should have~~ originated from the ocean region most directly connected to all major oceanic basins, i.e., the Southern Ocean (Ninnemann and Charles, 1997). Under a weak AMOC, wind-driven upwelling of the Circumpolar Deep Water (CDW) in the Southern Ocean would become stronger, ~~breaking up~~reducing the stratification of the Southern Ocean, and ~~enhance~~ing outgassing of low- $\delta^{13}\text{C}$ CO₂ to the atmosphere (Anderson et al., 2009; Denton et al., 2010; Tschumi et al., 2011; Bauska et al. 2016). However, model experiments (e.g., Schmittner and Galbraith, 2008; Schmittner and Lund, 2015) and records from different ocean basins (e.g., Tessin and Lund, 2013; Lund et al., 2015; Curry and Oppo, 2005; Hertzberg et al., 2016) suggest that the increase in CO_{2atm} and decrease in $\delta^{13}\text{CO}_{2atm}$ during HS1 are instead related to the weakening of the global oceanic biological pump and the consequent accumulation of ^{13}C -depleted CO₂ in the upper water column. Such anomalously low- $\delta^{13}\text{C}$ CO₂ would then be outgassed to the atmosphere. Since isotopic fractionation between reservoirs is temperature-dependent the air-sea gas exchange during HS1 could have additionally modified $\delta^{13}\text{CO}_{2atm}$ (Lynch-Stieglitz et al., 1995).

~~The CDW forms from mixing of North Atlantic Deep Water (NADW), Indian Deep Water (IDW) and Pacific Deep Water (PDW) and upwells to the south of the Antarctic Polar Front driven by the prevailing westerly winds (Marshall and Speer, 2012). Therefore, the $\delta^{13}\text{C}$ signal of CDW (ca. 0.4‰) (Kroopnick, 1985) lies between that of NADW (ca. 1‰) (Kroopnick, 1985) and IDW/PDW (ca. 0.2 to -0.2‰) (Kroopnick, 1985) (Oppo and Fairbanks, 1987; Charles and Fairbanks, 1992). During periods of weak AMOC the influence of NADW on the Southern Ocean is reduced (Charles and Fairbanks, 1992), and the $\delta^{13}\text{C}$ of CDW should decrease since this water mass would have a relatively larger contribution from low $\delta^{13}\text{C}$ IDW and PDW. Thus, the strengthening in the Southern Ocean upwelling would bring old, respired ^{13}C depleted waters to the~~

surface that would then be transferred equatorward via central and thermocline waters (Spero and Lea, 2002). This ~~The reduction of the upper water column $\delta^{13}\text{C}$ caused by one or both of the mechanisms (i.e., a weak AMOC leading to a strengthening of the Southern Ocean upwelling, a $\delta^{13}\text{C}$ depletion and an increased outgassing of CO_2 to the atmosphere) described above~~ may be a common feature of ~~other HSs (Anderson et al., 2009) as well~~ but, so far, planktonic foraminiferal $\delta^{13}\text{C}$ records corroborating this ~~assumption process often only cover the are mostly available for~~ abrupt millennial-scale climate change events of the last deglaciation (i.e., HS1 and the Younger Dryas) while ~~high temporal resolution information on the HS~~ of the last glacial period is still scarce (Oppo and Fairbanks, 1989; Ninnemann and Charles, 1997; Spero and Lea, 2002; Hertzberg et al., 2016). Here we ~~addresses investigate~~ this issue for HS3 and HS2 using planktonic foraminiferal (*Globigerinoides ruber* white (*G. ruber* w) and *Globorotalia inflata* (*G. inflata*)) $\delta^{13}\text{C}$ data from a high temporal resolution marine sediment core (GeoB6212-1), collected near 32° S off south-eastern South America (SESA). Our data suggest that HS3 and HS2 were ~~also~~ marked by significant $\delta^{13}\text{C}$ decreases in ~~the upper water column most likely caused by millennial-scale strengthening of Southern Ocean upwelling.~~

2 Regional setting

The upper water column of the study area is dominated by the southward flowing Brazil Current (BC) ~~that forms~~, the western branch of the South Atlantic subtropical gyre. The BC is one of the weakest western boundary currents in the world ocean (Peterson and Stramma, 1991) carrying warm, saline and nutrient-depleted subtropical waters southward (Olson et al., 1988). The BC originates near 10-15° S from the bifurcation of the Southern South Equatorial Current as it approaches the western slope of the Brazil Basin (Stramma et al., 1990; Peterson and Stramma, 1991). Around 38° S the BC encounters the northward flowing Malvinas Current (MC) ~~(i.e., Brazil/Malvinas Confluence)~~, where the opposing flows turn south-east and flow offshore, ~~the so-called Brazil/Malvinas Confluence~~. The ~~offshore region~~ ~~Brazil/Malvinas Confluence~~ is characterized by intense mesoscale variability. After collision and considerable mixing the warm-salty BC fractions flow eastward as the South Atlantic Current (Olson et al., 1988; Peterson and Stramma, 1991), while the majority of the cold fresh MC waters veer southeastward to rejoin the Antarctic Circumpolar Current.

~~In the study area,~~ the BC transports Tropical Water (TW), ~~and~~ South Atlantic Central Water (SACW) ~~and Antarctic Intermediate Water (AAIW)~~. TW occupies the mixed layer, i.e., the upper ca. 100 m of the water column, with a mean temperature of 20 °C and mean salinity of 36 psu (Tsuchiya et al., 1994). TW originates in the tropics-subtropics transition region by subduction, creating a subsurface salinity maximum capping the central waters (Mémery et al., 2000; Tomczak and Godfrey, 2003) (Fig. 1).

SACW occupies the permanent thermocline from ca. 100 to 500 m water depth. Its temperature ranges from 6 to 20 °C and its salinity spans from 34.6 to 36 psu (Mémery et al., 2000). Two types of SACW have been identified (Stramma et al., 2003). The low-density type of SACW which is mainly found in the South Atlantic subtropical gyre is formed by subduction of a low-density type of Subantarctic Mode Water (SAMW) along the southern edge of the gyre (Gordon, 1981; Stramma

and England, 1999). ~~The~~A denser variety of SACW originates in the South Indian Ocean and is brought into the South Atlantic by the Agulhas Current (Sprintall and Tomczak, 1993) (Fig. 1).

Just below the permanent thermocline, ~~Antarctic Intermediate Water (AAIW)~~ occupies the water column from ca. 500 to 1200 m water depth (Stramma and England, 1999). AAIW is characterized as a cold and low salinity water mass (Piola and Georgi, 1982; Tomczak and Godfrey, 2003). Around the southern tip of South America, AAIW originates by subduction of cold and fresh Antarctic Surface Water across the Antarctic Polar Front, and by contribution of a dense type of SAMW that originates from deep winter convection in the Subantarctic Zone (Molinelli, 1981; Naveira Garabato et al., 2009). AAIW is advected eastward through the Drake Passage by the Antarctic Circumpolar Current and turns northwards with the MC into the South Atlantic (Piola and Gordon, 1989). Since AAIW circulation follows the anticyclonic flow of the subtropical gyre the majority of the northward flow at mid-latitude occurs in the eastern basin (McCartney, 1977; Stramma and England, 1999; Tomczak and Godfrey, 2003). However, intense mixing in the Brazil/Malvinas Confluence also leads to direct northward ~~flow influence~~ in the western South Atlantic that can, to some extent-, influence the dissolved inorganic carbon $\delta^{13}\text{C}$ ($\delta^{13}\text{C}_{\text{DIC}}$) in the formation region of SACW (e.g., Piola and Georgi, 1982) (Fig. 1 and 2).

In the modern South Atlantic, the distribution of ~~dissolved inorganic carbon $\delta^{13}\text{C}$ ($\delta^{13}\text{C}_{\text{DIC}}$)~~ allows the identification of its major water masses. TW and SACW show high $\delta^{13}\text{C}_{\text{DIC}}$ values of ca. 2‰. AAIW presents $\delta^{13}\text{C}_{\text{DIC}}$ values of ca. 0.7‰. NADW derives from the North Atlantic and shows $\delta^{13}\text{C}_{\text{DIC}}$ values of ca. 1‰. ~~The~~In the southwest South Atlantic NADW ~~layer is sandwiched between surrounded by~~ Upper and Lower CDW which present $\delta^{13}\text{C}_{\text{DIC}}$ values of ca. 0.4‰ (Kroopnick, 1985). Since planktonic foraminiferal $\delta^{13}\text{C}$ reflects the $\delta^{13}\text{C}_{\text{DIC}}$ of the ambient seawater, we use it as a proxy for the past oceanic carbon system (Mulitza et al., 1999; Spero, 1992). However, other factors such as calcification temperature, carbonate ion concentration, symbiont activity and air-sea gas exchange may also influence planktonic foraminiferal $\delta^{13}\text{C}$ (Lynch-Stieglitz et al., 1995; Spero and Lea, 1996, Spero et al., 1997; Bemis et al., 2000).

Changes in upper ocean properties and circulation patterns are also closely associated with changes in the atmospheric circulation. Positive sea surface temperature (SST) anomalies in the western South Atlantic, likely associated to changes in the strength of the AMOC (Knight et al., 2005), have been correlated with positive anomalies in the strength of the SAMS and, consequently, with the increase of precipitation over SESA (Chaves and Nobre, 2004). The SAMS and its main components – the ITCZ, the South Atlantic Convergence Zone (SACZ), and the South American Low Level Jet (SALLJ) – are the main atmospheric drivers of the hydroclimate of tropical and subtropical SESA to the east of the Andes (Garreaud et al., 2009). The ITCZ is a global convective belt in the equatorial region, and the SACZ is an elongate NW-SE convective belt that originates in the Amazon Basin and extends southeastward above the northern portion of SESA and the adjacent subtropical South Atlantic. The SALLJ is a NW-SE humidity flux from the west Amazon Basin to the subtropical region of SESA (Zhou and Lau, 1998; Carvalho et al., 2004; ~~Schneider et al., 2014~~). This southward water vapour flux is a crucial source of precipitation to the Plata River drainage basin (Berbery and Barros, 2002), which is a source of continental borne sediments to our core site.

3 Materials and methods

3.1 Marine sediment core

We investigated sediment core GeoB6212-1 (32.41° S, 50.06° W, 1010 m water depth, 790 cm core length) (Schulz et al., 2001) collected from the continental slope off SE SAAS where the upper water column is under the influence of the BC, and thus the TW and SACW (Fig. 1). This gravity core was raised at the Rio Grande Cone, a major sedimentary feature in the western Argentine Basin. As our study focuses here are in HS3 and HS2, we analysed the section from the bottom of the core (768 cm core depth; ca. 33 cal ka BP) up to 290 cm core depth (ca. 20 cal ka BP). Visual core inspection provided evidence for the presence of sand lenses at 330 and 368 cm core depth (Schulz et al., 2001; Wefer et al., 2001). Therefore we did not sample these depths. The section of interest of GeoB6212-1 was sampled every 2.5 cm with syringes of 10 cm³. All samples were wet sieved, oven-dried at 50 °C and the fraction larger than 150 µm was stored in glass vials for subsequent analyses.

3.2 Age model

The age model of core GeoB6212-1 is based on 14 AMS radiocarbon ages from planktonic foraminifera (Table 1, Fig. 3). For each sample, we hand-picked under a binocular microscope around 10 mg of planktonic foraminifera shells from the sediment fraction larger than 150 µm. Samples were analysed at the Poznan Radiocarbon Laboratory, Poland, and at the Beta Analytic Radiocarbon Dating Laboratory, USA (Table 1). All radiocarbon ages were calibrated with the calibration curve IntCal13 (Reimer et al., 2013) with the software Bacon 2.2 (Blaauw and Christen, 2011). A marine reservoir correction of 400 years was applied (Bard, 1988). All ages are reported as calibrated years before present (cal a BP; present is 1950 AD). To construct the age model we used Bayesian statistics in the software Bacon 2.2 (Blaauw and Christen, 2011). ~~Default parameter settings were used, except for mem.mean (set to 0.4) and acc.shape (set to 0.5). Ages are modelled as drawn from a t distribution, with 9 degrees of freedom (t.a=9, t.b=10). 1,000 age depth realizations were used to estimate mean age and 95 % confidence intervals at 0.5 cm resolution (Fig. 3). With the exception of mem.mean (set to 0.4) and acc.shape (set to 0.5), default parameters were used. Radiocarbon ages were assumed to be t-distributed with 9 degrees of freedom (t.a=9, t.b=10). Mean ages and 95% error margins were estimated from 10,000 downcore age-depth realizations at 0.5 cm resolution (Fig. 3).~~

3.3 Stable carbon isotope analyses

Around 10 tests of *G. ruber* ~~white~~ *sensu stricto* (Wang, 2000) within the size range 250-350 µm and 8 tests of *G. inflata* non-encrusted with 3 chamber in the final whorl (Groeneveld and Chiessi, 2011) within the size range 315-400 µm were hand-picked under a binocular microscope every 2.5 cm from 290 to 768 cm core depth. While the first species records the conditions at the top of the mixed layer (down to ca. 30 m) (Chiessi et al., 2007; Wang, 2000), the second species records the conditions at the permanent thermocline (ca. 350-400 m) (Groeneveld and Chiessi, 2011), allowing the reconstruction of the

$\delta^{13}\text{C}$ signal of the TW and the SACW, respectively. The $\delta^{13}\text{C}$ analyses were performed on a Finnigan MAT 252 mass spectrometer equipped with an automatic carbonate preparation device at the MARUM – Centre for Marine Environmental Sciences, University of Bremen, Germany. Isotopic results are reported in the usual delta–notation relative to the Vienna Peedee belemnite (~~VPDB~~). Data were calibrated against the house standard (Solnhofen limestone), itself calibrated against the NBS19 standard. The standard deviation of the laboratory standard was lower than 0.05‰ for the measuring period.

4 Results

4.1 Age model and sedimentation rates

Our age model covers the period between 32.6 and 5.76 cal ka BP (Table 1, Fig. 3). Sedimentation rates change markedly during this time interval with values ranging from 3.85 to 11191 cm ka⁻¹. Three main peaks in sedimentation rate were identified at ca. 26, 23 and 15 and one minor peak at 11 cal ka BP. The two oldest sedimentation peaks occur within our period of interest (i.e., from ca. 33 until 20 cal ka BP), ~~and received special attention due to the higher sedimentation rate which provides increased temporal resolution~~ (Fig. 3). The mean temporal resolution of our $\delta^{13}\text{C}$ records is ca. 90 yr with values ranging from 28 ~~to~~ 195 yrs.

4.2 Stable carbon isotopes values of *G. ruber* and *G. inflata* analyses

The *G. ruber* $\delta^{13}\text{C}$ record shows two long–term decreases, from ca. 32.6 to 28.25 cal ka BP with an amplitude of ca. 1‰, and from ca. 26.5 to 24.98 cal ka BP also with an amplitude of ca. 1‰ (Fig. 4a). These two negative long–term trends are separated from each other by an abrupt increase of ca. 1.3‰ ending at ca. 27.2 cal ka BP. Both long–term decreases ~~sing slopes~~ were interrupted by brief positive excursions, one from 29.330.6 to 29.130.4 cal ka BP with an amplitude of ca. 0.7‰ and another from ca. 26.2 to 25.8 cal ka BP with an amplitude of ca. 1‰. After the second long–term decrease, the $\delta^{13}\text{C}$ values of *G. ruber* w varied around 0.7‰. Both long–term negative excursions determine a pattern we refer to as “w–structure”.

The *G. inflata* $\delta^{13}\text{C}$ record shows four negative excursions departing from a baseline of ca. 0.8‰ (Fig. 4b). The first occurs from ca. 31.52.5 to 29.330.6 cal ka BP with an amplitude of ca. 0.5‰, the second from ca. 28.89.8 to 28.3 cal ka BP with the same amplitude, the third from ca. 26.5 to 26.4 cal ka BP with an amplitude of ca. 0.8‰, and the forth from ca. 25.8 to 24.4 cal ka BP with an amplitude of ca. 0.9‰. Also in the $\delta^{13}\text{C}$ record from *G. inflata* two “w–structures” are present and are defined by the previously described negative excursions.

The “w–structures” ~~for both species~~ as well as the $\delta^{13}\text{C}$ minima are synchronous for both species (Fig. 4).

5 Discussion

The synchronous “w-structures” present in the $\delta^{13}\text{C}$ records of both planktonic foraminiferal species analysed here occur in ~~coeval resonance~~ with the millennial-scale events HS3 and HS2 (Sarnthein et al., 2001; Goni and Harrison, 2010) (Fig. 4). ~~Concomitantly, to the periods of low $\delta^{13}\text{C}$ in our records a weak AMOC was described based on $^{231}\text{Pa}/^{230}\text{Th}$ records from the Bermuda Rise (ODP Site 1063, 33.7° N, 57.6° W) (Lippold et al., 2009) that characterize HS3 and HS2 (Fig. 5d). Both events are also marked by pulses of ice-rafted debris (IRD) (MD99-2331, 42.2° N, 9.7° W) (Eynaud et al., 2009) and by decreases in SST (SU8118 and MD952042, 37.5° N, 10.1° W) (Bard, 2000) in the north-eastern North Atlantic (Iberian Margin). The Greenland GISP2 ice core (72.6° N, 38.5° W) shows synchronous increases in Ca^{+2} , indicating changes in atmospheric circulation over Greenland (Mayewski et al., 1997) (Fig. 5a, b, c). It is noteworthy worth to note that the four records (i.e., Fig. 5a, b, c, d) mentioned above also show a “w-structure” during HS2, similar to the one shown in our $\delta^{13}\text{C}$ records. The IRD (Eynaud et al., 2009) and Ca^{+2} (Mayewski et al., 1997) records also show a “w-structure” similar to ours during HS3.~~

Based on modern conditions, we expect our core site not to be significantly influenced by changes in the local nutrient content of the upper water column since the region is dominated by the oligotrophic BC, characteristic of western boundary currents, and is far from upwelling cells (Brandini et al., 2000). Thus, it is unlikely that changes in our $\delta^{13}\text{C}$ records are associated with local productive events driven by nutrient-cycle processes (Mulitza et al., 1999).

~~During HS, we expect warmer temperatures to have occurred in the upper water column of the western South Atlantic (Barker et al. 2009; Chiessi et al. 2015). This would trigger an increase in $\delta^{13}\text{C}$ values of the symbiont-bearing species investigated here if calcification temperature would dominate the $\delta^{13}\text{C}$ signal (Bemis et al., 2000), which is not the case (Fig. 4a). Additionally, given the lack of regional upper ocean reconstructions for carbonate ion concentration, we assume that increased $\text{CO}_{2\text{atm}}$ that is frequently associated with HS (Ahn and Brook, 2008; Ahn and Brook, 2014) would have been accompanied by a decrease in sea surface carbonate ion concentration (Broecker and Peng, 1993). This would promote an increase in the $\delta^{13}\text{C}_{\text{DIC}}$ but our records show a negative $\delta^{13}\text{C}$ anomaly (Fig. 4). Furthermore, we analysed a symbiont-bearing and a facultative-symbiont species (i.e., *G. ruber w* and *G. inflata*, respectively) and both records show a similar pattern (Fig. 4) indicating that changes in symbiont activity can also be disregarded as a factor influencing our results (Spero et al., 1997; Bemis et al., 2000). We propose two primary mechanisms to explain our $\delta^{13}\text{C}$ decreases: (i) changes in the strength of Southern Ocean deep water ventilation (detailed in section 5.1), and (ii) the weakening of the global oceanic biological pump (detailed in section 5.2). Additionally, air-sea gas exchange may have acted as a secondary factor contributing to our $\delta^{13}\text{C}$ decreases (detailed in section 5.3).~~

~~A pervasive feature of planktonic foraminiferal $\delta^{13}\text{C}$ records in the Indo-Pacific Ocean (Spero and Lea, 2002), Southern Ocean (Ninnemann and Charles, 1997), and South Atlantic Ocean (Oppo and Fairbanks, 1989) is a negative excursion during HS1. Ninnemann and Charles (1997) suggested that the source for this signal is in the Southern Ocean. They further proposed that the anomaly is related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the Southern Ocean via SAMW and/or~~

~~AAIW. A low density type of SAMW actually contributes to SACW that spreads into the South Atlantic (Stramma and England, 1999). Additionally, AAIW also influences SACW through vigorous eddy mixing at the Brazil/Malvinas Confluence (Piola and Georgi, 1982). Thus, SACW represents a potential conduit for the $\delta^{13}\text{C}$ signal from the sub-Antarctic region to the subtropical South Atlantic (Fig. 1 and 2). Therefore, we propose that the negative excursions in our $\delta^{13}\text{C}$ records are related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the subantarctic zone to the western South Atlantic via central and thermocline waters.~~

5.1 Millennial-scale changes: Atlantic Ocean, AMOC-induced strengthening of Southern Ocean deep water ventilation and $\text{CO}_{2\text{atm}}$

~~A negative excursion during HS1 was described in a pervasive feature of planktonic foraminiferal $\delta^{13}\text{C}$ records in from the Indo-Pacific Ocean (Spero and Lea, 2002), Southern Ocean (Ninnemann and Charles, 1997), and South Atlantic Ocean (Oppo and Fairbanks, 1989) is a negative excursion during HS1. Ninnemann and Charles (1997) suggested that the source for this signal is in was the Southern Ocean. They further proposed that the anomaly is related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the Southern Ocean via SAMW and/or AAIW.~~

~~In the Southern Ocean The CDW forms from mixing of North Atlantic Deep Water (NADW), Indian Deep Water (IDW) and Pacific Deep Water (PDW) and upwells to the south of the Antarctic Polar Front driven by the prevailing westerly winds (Marshall and Speer, 2012). Therefore, the $\delta^{13}\text{C}$ signature of CDW (ca. 0.4‰) (Kroopnick, 1985) lies between that of NADW (ca. 1‰) (Kroopnick, 1985) and IDW/PDW (ca. 0.2 to -0.2‰) (Kroopnick, 1985) (Oppo and Fairbanks, 1987; Charles and Fairbanks, 1992). During periods of weak AMOC the inflow influence of NADW onto the Southern Ocean is reduced (Charles and Fairbanks, 1992), and the $\delta^{13}\text{C}$ of CDW should decrease since this latter water mass would have a relatively larger contribution from low- $\delta^{13}\text{C}$ IDW and PDW. Thus, the strengthening in the Southern Ocean upwelling would bring old, respired ^{13}C depleted waters to the surface that would then be transferred equatorward via central and thermocline waters (Spero and Lea, 2002).~~

~~Concomitantly to the periods of low $\delta^{13}\text{C}$ in our records a weak AMOC was described based on ^{231}Pa / ^{230}Th records from the Bermuda Rise (ODP Site 1063, 33.7° N, 57.6° W) (Lippold et al., 2009) that characterize HS3 and HS2 (Fig. 5d). Both events are also marked by pulses of ice-rafted debris (IRD) (MD99-2331, 42.2° N, 9.7° W) (Eynaud et al., 2009) and by decreases in SST (SU8118 and MD952042, 37.5° N, 10.1° W) (Bard, 2002) in the north-eastern North Atlantic (Iberian Margin). The Greenland GISP2 ice core (72.6° N, 38.5° W) shows synchronous increases in Ca^{+2} , indicating changes in atmospheric circulation over Greenland (Mayewski et al., 1997) (Fig. 5a, b, c). It is worth to note that the four records (i.e., Fig. 5a, b, c, d) mentioned above also show a w structure during HS2, similar to the one shown in our $\delta^{13}\text{C}$ records. The IRD (Eynaud et al., 2009) and Ca^{+2} (Mayewski et al., 1997) records also show a w structure similar to ours during HS3. Additionally, during periods of The reduced AMOC would decrease the sub-tropical heat transport towards the north would decrease, leading to rising temperatures in the circum-Antarctic region (EDML, 75° S, 0° E, EPICA) (EPICA~~

Community Members, 2006) (Fig. 5i). ~~Consequently, Furthermore, during phases of weak AMOC~~ the Southern Hemisphere westerlies ~~would become~~ stronger and shift southward, strengthening CDW upwelling (Toggweiler et al., 2006; Anderson et al., 2009; Denton et al., 2010; Tschumi et al., 2011; Lee et al., 2011; Voigt et al., 2015). Increased upwelling would supply the ocean surface ~~of the Southern Ocean to the~~ south of the Antarctic Polar Front with ~~more~~ Si(OH)₄-rich, low- $\delta^{13}\text{C}$ waters ~~and with a higher concentration of Si(OH)₄~~ (Anderson et al., 2009; Hendry et al., 2012). ~~A recent model experiment (Bauska et al., 2016) corroborates this hypothesis by showing that stronger Southern Ocean upwelling would promote a weakening of the biological pump in the Southern Ocean.~~ Since upwelled CDW is hypothesized to be the dominant source of the upper and intermediate waters that leave the Southern Ocean (i.e., SAMW and AAIW) (Fig. 2), increased upwelling would transfer the low $\delta^{13}\text{C}$ signal as well as the positive Si(OH)₄ anomaly ~~to those waters northward into the adjacent subtropical gyres~~ (Oppo and Fairbanks, 1989; Ninnemann and Charles, 1997; Spero and Lea, 2002; Anderson et al., 2009; Hendry et al., 2012). ~~Actually, a low-density type of SAMW contributes to SACW that spreads into the South Atlantic (Stramma and England, 1999). Additionally, AAIW also influences SACW through vigorous eddy mixing at the Brazil/Malvinas Confluence (Piola and Georgi, 1982).~~ These signals would then propagate through the thermocline (i.e., SACW) of the South Atlantic, and be transferred to the mixed layer TW by vertical exchange process (i.e., TW) (Tomczak and Godfrey, 2003). ~~However, we cannot exclude the possibility that the upwelled low- $\delta^{13}\text{C}$ respired CO₂ could have been first outgassed from the Southern Ocean, and then re-dissolved into the ocean via air-sea exchanges at the formation regions of SACW and TW, eventually reaching the upper water column at our core site.~~

~~Hendry et al. (2012) describe H~~ higher seawater concentrations of Si(OH)₄ ~~were described through silicon isotope composition of in sponge spicules (benthic organisms) at intermediate water depths (i.e., 1048 m water depths) of the western South Atlantic (ca. 27° S) close to our core site during abrupt millennial-scale climate change events (Hendry et al., 2012), suggesting that the preformed signal from the Southern Ocean could indeed reach~~ed subtropical latitudes in the South Atlantic.

~~Millennial-scale changes of the Southern Ocean temperature and deep-water ventilation also led to the increase in CO_{2atm} (Spero and Lea, 2002; Ahn and Brook, 2008; Ahn and Brook, 2014; Gottschalk et al., 2015). During HS3 and HS2, positive peaks in CO_{2atm} (Siple Dome, 81.7° S, 148.8° W) (Ahn and Brook, 2014) (Fig. 5k) were described to be synchronous to circum-Antarctic warming (EPICA Community Members, 2006) (Fig. 5i), and most likely have originated from~~ ~~†~~The reduced stratification of the Southern Ocean and ~~consequent release~~ intensification of the upward transport of the remineralized carbon (¹²C-enriched CO₂) stored for a long period in deep waters (Anderson et al., 2009; Denton et al., 2010; Jaccard et al., 2016; Mariotti et al., 2016) ~~would increase CO_{2atm} (Siple Dome, 81.7° S, 148.8° W) (Ahn and Brook, 2014) (Fig. 5j).~~ Despite the low temporal resolution, ~~Eggleston et al.'s (2016) Antarctic $\delta^{13}\text{C}$ CO_{2atm} record shows a decrease during HS2, even the $\delta^{13}\text{C}$ record of CO_{2atm} at Taylor Dome (77.8° S, 158.7° E) on Antarctica shows a decrease during HS2 (Smith et al., 1999) (Fig. 5j).~~ However, the CO_{2atm} peaks occur ca. 1 ka later than the initiation of the $\delta^{13}\text{C}$ decrease in our records. Spero and Lea (2002) also observed a similar offset between the increase in CO_{2atm} and the decrease in Pacific Ocean

planktonic foraminifera $\delta^{13}\text{C}$ during HS1, and attributed this apparent offset to uncertainties in the age models of their records. ~~An alternative explanation relates to a possible time lag between the weakening of the AMOC and the increase in $\text{CO}_{2\text{atm}}$ (Ahn and Brook, 2014)~~

~~Therefore, the negative excursions in our $\delta^{13}\text{C}$ records could be related to the transfer of a preformed $\delta^{13}\text{C}$ signal from the subantarctic zone to the western South Atlantic via central and thermocline waters.~~

~~Thus, our records are consistent with the hypothesis that the increase in $\text{CO}_{2\text{atm}}$ during abrupt millennial-scale climate change events of the last glacial period is originated by ocean processes (Smith et al., 1999; Ahn and Brook, 2008; Bereiter et al., 2012; Ahn and Brook, 2014) and is most likely related to a weak AMOC and associated strengthened Southern Ocean upwelling.~~

5.2 Millennial-scale changes: AMOC-induced weakening of the biological pump

~~Recent model experiments (e.g. Schmittner, 2005; Schmittner and Galbraith, 2008) have shown that AMOC slowdown events may cause a decrease in the global efficiency of the oceanic biological pump, being an important driver for the oceanic CO_2 outgassing within HS1 during the last deglaciation and possibly also during other HS, including HS3 and HS2.~~

~~NADW has low preformed nutrient waters because it is formed by nutrient depleted surface waters, where the biological pump has efficiently removed nutrients from surface waters (Marinov et al., 2008). AABW has high preformed nutrient waters because it is formed by nutrient-enriched Southern Ocean surface waters (nutrients have not being efficiently removed from surface waters). However, during weak AMOC two factors may alter the nutrient distribution and the global oceanic biological pump (Schmittner and Galbraith, 2008). First, the reduction in the NADW formation decreases the input of low preformed nutrient (high $\delta^{13}\text{C}_{\text{DIC}}$) waters to the ocean interior which becomes more dominated by high preformed nutrient (low $\delta^{13}\text{C}_{\text{DIC}}$) southern component waters (e.g., AABW). Second, the reduction of the Southern Ocean stratification induced by the decrease of salt input via NADW formation promotes the strengthening of the upwelling and subsequent sinking of high preformed nutrient (low $\delta^{13}\text{C}_{\text{DIC}}$) waters to the ocean interior, thus reducing the capacity of those unutilized nutrients to sequester carbon via the biological pump. The two factors acting in conjunction are thought to be responsible for the simulated weakening of the global efficiency of the biological pump, as well as for the increase in $\text{CO}_{2\text{atm}}$ and decrease in $\delta^{13}\text{CO}_{2\text{atm}}$ (Schmittner and Galbraith, 2008; Schmittner and Lund, 2015; Hertzberg et al., 2016).~~

~~Schmittner and Lund (2015) show that the modeled weakening of the biological pump, induced by an AMOC slowdown, reduces the ability of the surface ocean to biologically sequester isotopically light organic carbon (^{12}C), producing a decrease in the surface ocean $\delta^{13}\text{C}_{\text{DIC}}$ and an increase of the intermediate ocean $\delta^{13}\text{C}_{\text{DIC}}$ (lower remineralization rate). For HS1, planktonic and benthic foraminiferal $\delta^{13}\text{C}$ records (Tessin and Lund, 2013; Lund et al., 2015; Curry and Oppo, 2005; Hertzberg et al., 2016) from the western South Atlantic (ca. 27°S) agree with the model output by showing a decrease in $\delta^{13}\text{C}$ in the upper water column (SACW) and an increase at intermediate water depths (AAIW). Thus, the weakening of the global oceanic biological pump and consequent negative anomaly of the $\delta^{13}\text{C}_{\text{DIC}}$ in the upper water column should be~~

captured by the tests of planktonic foraminifera $\delta^{13}\text{C}$ during calcification (Spero and Lea, 1996; Bemis et al., 2000). The negative $\delta^{13}\text{C}_{\text{DIC}}$ during HS3 and HS2 revealed by our planktonic foraminifera provide the first observational evidence supporting the modeling results. Additionally this mechanism also provides a possible explanation for the larger negative $\delta^{13}\text{C}$ anomaly in *G. ruber w* (mixed layer dwelling) relative to the anomaly in *G. inflata* (permanent thermocline dwelling) (Fig. 4).

It is noteworthy that the mechanism described in section 5.1, although based on a different driver for the decrease in $\delta^{13}\text{C}$, also suggests that the decreases in $\delta^{13}\text{C}$ of planktonic foraminifera from the South Atlantic would be carried by SACW (inherited from its precursor, SAMW) and thus both mechanisms (described in section 5.1 and here) are in this regard not mutually exclusive. However, the mechanism described in the present section goes against the assumption that weakening of the biological pump is related to stronger upwelling in the Southern Ocean, and that the Southern Ocean would be the source of the low $\delta^{13}\text{C}$ signal for the South Atlantic (Lund et al., 2015; Hertzberg et al., 2016).

5.3 Millennial-scale changes: the role of air-sea gas exchange

The $\delta^{13}\text{C}_{\text{DIC}}$ of the surface ocean can also be affected by air-sea gas exchange (Oppo and Fairbanks, 1989; Charles and Fairbanks, 1990; Lynch-Stieglitz et al., 1995). Although this process tends towards isotopic equilibrium, the CO_2 exchange between the ocean and the atmosphere does not lead to equilibrium because CO_2 uptake and emission will still occur in different regions and the movement and replacement of surface waters is faster than required for equilibration (Lynch-Stieglitz et al., 1995). Since the $\delta^{13}\text{CO}_{2\text{atm}}$ is lighter than $\delta^{13}\text{C}_{\text{DIC}}$, at areas of ocean $\text{CO}_{2\text{atm}}$ uptake (i.e., water mass formation regions) air-sea gas exchange has the potential to deplete $\delta^{13}\text{C}_{\text{DIC}}$ (Lynch-Stieglitz et al., 1995). Additionally, the isotopic fractionation is inversely correlated with temperature.

Therefore, we cannot exclude the possibility that the likely decrease in $\delta^{13}\text{CO}_{2\text{atm}}$ during AMOC slowdown events (Eggleston et al. 2016) (e.g., HS3 and HS2) could have affected the $\delta^{13}\text{C}_{\text{DIC}}$ via air-sea gas exchange, especially in regions of water mass formation. The formation region of SACW is an area of ocean CO_2 uptake and may contribute to the $\delta^{13}\text{C}$ anomalies observed in our *G. inflata* record (Fig. 4b). Additionally, since the isotopic fractionation during air-sea gas exchange is temperature-dependent the weakening of the AMOC and subsequent warming of the upper subtropical South Atlantic (Barker et al. 2009; Chiessi et al. 2015) could have contributed to the observed $\delta^{13}\text{C}$ anomalies both in the *G. ruber w* and in the *G. inflata* records (Fig. 4). However, the gradient is too small (-0.1‰ $\delta^{13}\text{C}$ per $^{\circ}\text{C}$, Broecker and Maier-Reimer, 1992) to explain the whole changes observed in our records. If temperature was the dominant driver, unrealistic changes between 5 and 13 $^{\circ}\text{C}$ would be required to explain the full amplitudes of the $\delta^{13}\text{C}$ variations.

The $\delta^{18}\text{O}$ records from *G. ruber w* and *G. inflata* from our core (Supplementary material Figure 1) should partially reflect changes in water temperature (ca. -0.22‰ per 1 $^{\circ}\text{C}$; e.g., Mulitza et al, 2003), but show no clear trends across HS3 and HS2. While temperature changes might be partially obscured in the foraminiferal $\delta^{18}\text{O}$ records by the influence of synchronous

changes in seawater- $\delta^{18}\text{O}$, as has been hypothesized for the Holocene (Chiessi et al., 2014) and HS1 (Chiessi et al., 2015) in the western Atlantic, we consider it unlikely that temperature changes of the above magnitude would be completely masked.

5.34 Changes in cContinental climate

Paleoclimate records from South America ~~have shown~~ indicate marked hydrological changes during abrupt millennial-scale climate events (Arz et al., 1998; Peterson et al., 2000; Baker et al., 2001; Cruz et al., 2006; Strikis et al., 2015). Reconstructions of the SAMS activity suggests ~~its~~ strengthening during HSs (Cruz et al., 2006; Kanner et al., 2012; Cheng et al., 2013). Changes in speleothem oxygen isotopic composition from the western Amazon Basin (NAR-C, Cueva del Diamante cave, northern Peru, 5.4° S, 77.3° W) (Cheng et al., 2013) as well as changes on gamma radiation records from the Bolivian Altiplano (Salar de Uyuni, 20.3° S, 67.5° W) (Baker et al., 2001) (Fig. 5f, g) ~~suggest~~ indicate increased precipitation during HS3 and HS2. ~~To the n~~North of the equator, a reflectance record from the Cariaco Basin (off northern Venezuela, MD03-2621, 10.7° N, 65° W) (Deplazes et al., 2013) suggests decreased precipitation during the same millennial-scale events (Fig. 5e). The opposite ~~behaviour of~~ precipitation variations at these sites reflects the interhemispheric anti-phased response of tropical precipitation during HSs (Wang et al., 2007; Cheng et al., 2013). ~~Importantly, d~~During HS3 and particularly HS2 the three above mentioned records (Fig. 5e, f, g) show a “w-structure” similar to the one observed in our $\delta^{13}\text{C}$ records. Strikis et al. (2015) reported a similar “w-structure” during HS1 related to two distinct hydrologic phases within HS1.

Periods of intensified SAMS would have strengthened the discharge from the Plata River drainage basin (Chiessi et al., 2009), increasing the delivery of terrigenous sediments to the Rio Grande Cone (Lantsch et al., 2014), our coring site. ~~We show~~ Our records present for the first time increased sedimentation rates during a HS off SESA. ~~Thus, the increased sedimentation rates during HS2 in our records and~~ corroborate the suggestion of Chiessi et al. (2009) during HS2. Furthermore, GeoB6212-1 sedimentation rates also show a “w-structure” during HS2 (Fig. 3), hinting for a sensitive response of the Plata River drainage basin to the increase in activity of the SAMS. The occurrence of a similar “w-structure” in North Atlantic records, in South American records and in our $\delta^{13}\text{C}$ and sedimentation rate records gives us confidence that such “w-structure” is indeed an ubiquitous feature of HS2, and possibly also HS3.

The increased continental runoff that led to increased delivery of terrigenous sediments to our core site could have also enhanced the nutrient availability and the local primary productivity, affecting our planktonic foraminiferal $\delta^{13}\text{C}$ records. Some aspects of the regional response to HS1 are useful to evaluate this possibility. During HS1, ice volume corrected seawater- $\delta^{18}\text{O}$ from the upper water column of our core site indicates an increase in salinity (Chiessi et al., 2015). Thus, despite of the increased terrigenous discharge, it seems that the upper water column of our core site was not affected by an increase in freshwater discharge from the Plata River during HS1. Since the precipitation anomaly of HS1 was stronger than that of HS3 and HS2 in the Plata River drainage basin (Wang et al., 2007), it is unlikely that weaker precipitation anomalies of HS3 and HS2 would have impacted the upper water column of our core site more intensely than during HS1. This

suggests that changes in the discharge of the Plata River drainage basin at millennial-scale are not a relevant driver of our $\delta^{13}\text{C}$ decreases, and that the buoyant low salinity waters were advected elsewhere by winds, while terrigenous sediments were already too deep to be influenced by the wind. However, we discard this possibility because the expected signal of stronger local primary productivity on planktonic foraminiferal $\delta^{13}\text{C}$ would be opposite to the one observed at GeoB6212-1 (Mulitza et al., 1999).

6 Conclusions

Our mixed layer and permanent thermocline $\delta^{13}\text{C}$ records from the western South Atlantic show in-phase millennial-scale decreases of up to 1‰ during the HS3 and HS2. We hypothesize that the ~~source of the~~ low $\delta^{13}\text{C}$ signal may be related to two ~~can be explained by~~ millennial-scale mechanisms. (i) Changes in the Southern Ocean deep water ventilation. A weak AMOC during HS3 and HS2 would produce stronger Southern Ocean upwelling that in turn, would supply the surface of the Southern Ocean with more low- $\delta^{13}\text{C}$ waters as well as promote increased outgassing of this old and low- $\delta^{13}\text{C}$ respired CO_2 . The low- $\delta^{13}\text{C}$ waters at the surface of the Southern Ocean would be subducted into the central and thermocline waters and transferred equatorward via the South Atlantic subtropical gyre circulation and southward along the western boundary towards our core site. (ii) Weakening of the global oceanic biological pump. A weak AMOC during HS3 and HS2 would promote an accumulation of ^{13}C -depleted CO_2 in the upper water column of the South Atlantic. This accumulation would result in a negative anomaly of the $\delta^{13}\text{C}_{\text{DIC}}$ (as well as of the $\delta^{13}\text{CO}_{2\text{atm}}$) that in turn would be captured by the tests of planktonic foraminifera at our core site. We further suggest that changes in air-sea gas exchange could have contributed to the decreases in $\delta^{13}\text{C}$ via both mechanisms. Together with other lines of evidence, our data are consistent with the hypothesis that the CO_2 added to the atmosphere during abrupt millennial-scale climate change events of the last glacial period originated in the ocean and reached the atmosphere by outgassing ~~of the Southern Ocean~~. Moreover, the occurrence of a similar “w-structure” during HS2 (and possibly HS3) in North Atlantic and South American records as well as in our planktonic foraminiferal $\delta^{13}\text{C}$ and sedimentation rate records gives us confidence that such “w-structure” is a pervasive feature that characterizes HS2 (and possibly HS3).

Data availability

The data reported here will be archived in Pangaea (www.pangaea.de).

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Table 1: Accelerator mass spectrometer radiocarbon dates and calibrated ages used to construct the age model of core GeoB6212-1.

Core depth (cm)	Lab ID	Species	Radiocarbon age $\pm 1\sigma$ error (a BP)	Calibrated ages (cal a BP)
8	Poz-47236*	<i>G. ruber</i>	5250 \pm 40	5929694
33	382580**	<i>G. ruber</i>	9440 \pm 40	10047442
58	Poz-47237*	<i>G. ruber</i>	10250 \pm 50	11395345
113	382581**	Planktonic forams	12870 \pm 50	14342374
158	Poz-47238*	Planktonic forams	13050 \pm 70	15247479
253	382582**	Planktonic forams	15750 \pm 50	18620646
323	382583**	Planktonic forams	17560 \pm 60	20748731
363	Poz-47239*	Planktonic forams	18610 \pm 140	21834843
478	382584**	Planktonic forams	19810 \pm 70	23537544
578	Poz-47240*	Planktonic forams	21750 \pm 150	25587583
623	382585**	Planktonic forams	22320 \pm 80	26184464
668	382586**	Planktonic forams	22480 \pm 80	26711645
717	424077**	Planktonic forams	24190 \pm 110	27850894
768	382587**	Planktonic forams	29520 \pm 160	319662583

* Poz: Poznan Radiocarbon Laboratory, Poznan, Poland.

** Beta Analytic Radiocarbon Dating Laboratory, Miami, USA.

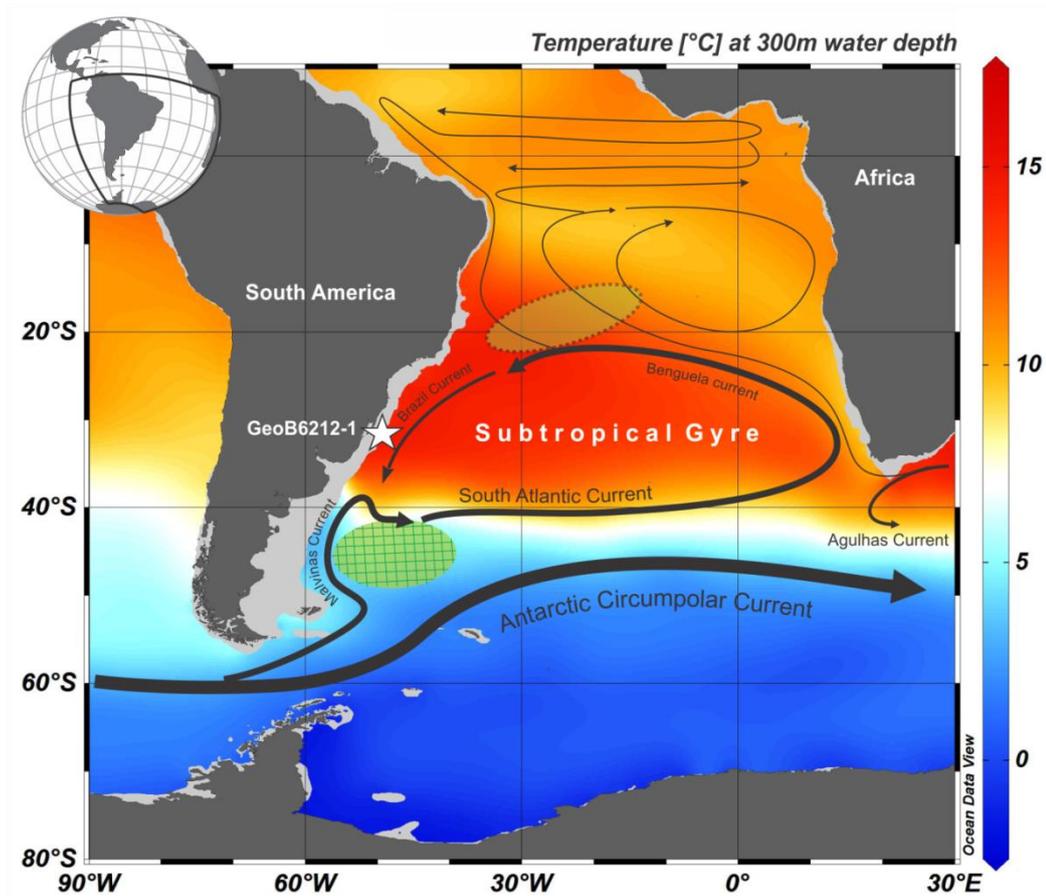


Figure 1: Schematic representation of the large-scale circulation of South Atlantic Central Water (SACW) (Stramma and England, 1999). The main SACW source region is depicted by the gridded green ellipse, and the main source region of tropical subsurface water (TW) is indicated by the dotted yellow ellipse. Mean annual temperature at 300 m water depth is shown by the colour shading (Locarnini et al., 2013) (<http://odv.awi.de>).

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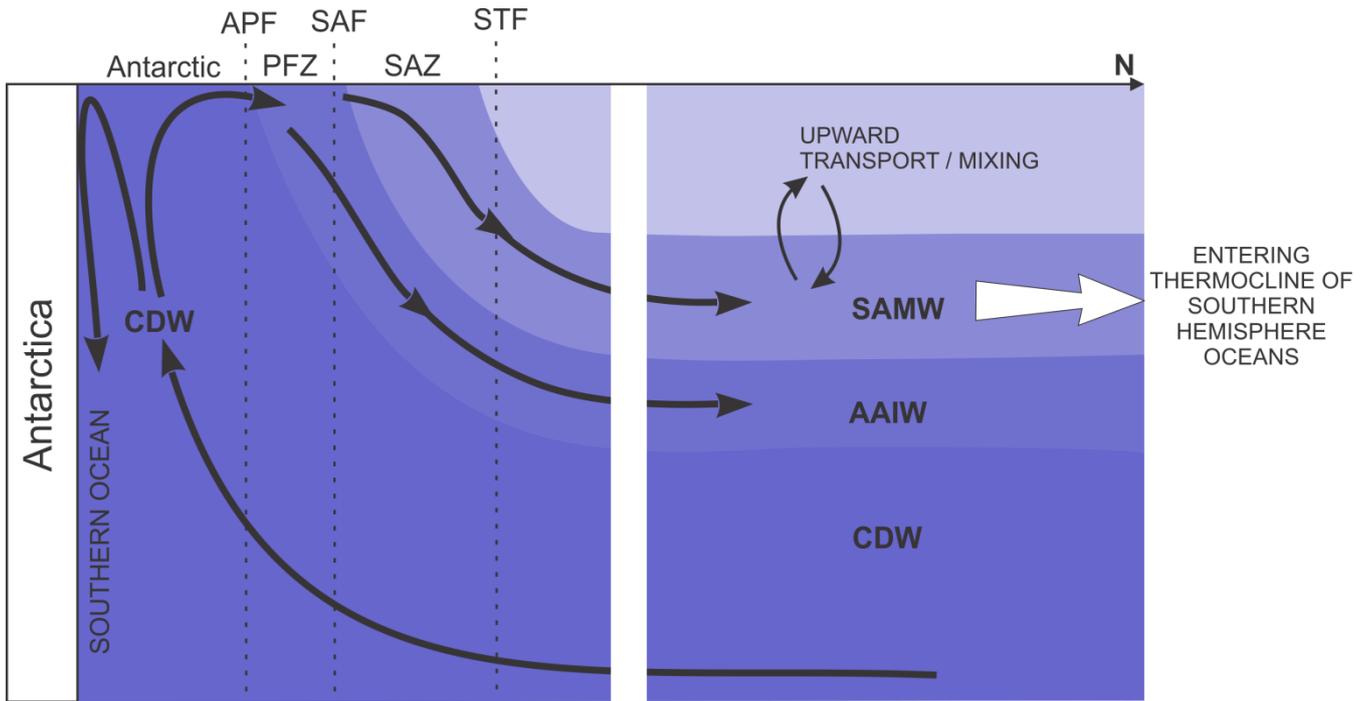
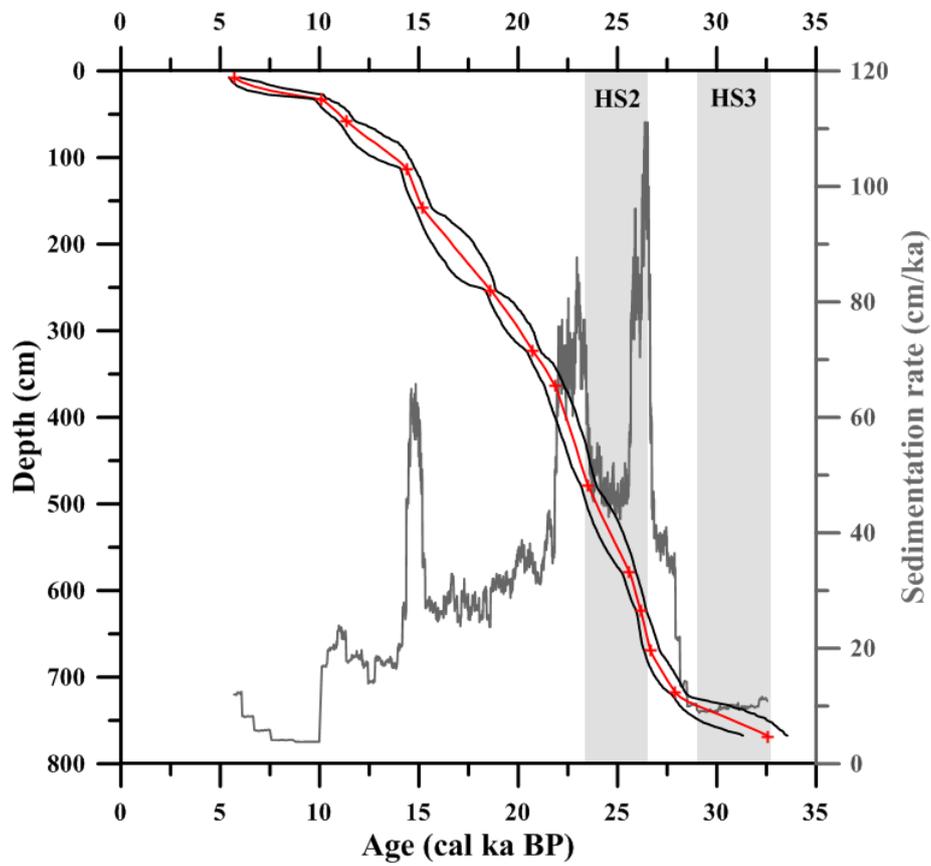


Figure 2: Schematic representation of ventilation and subduction of water masses in the Southern Ocean (modified after Anderson et al., 2009). Wind-driven upwelling south of the latitude of maximum westerlies brings Circumpolar Deep Water (CDW) to the surface and contributes to Antarctic Surface Waters. Antarctic Surface Waters represent the dominant source of the upper and intermediate waters that leave the Southern Ocean. Antarctic Intermediate Water (AAIW) originates by subduction of cold and fresh Antarctic Surface Waters across the Antarctic Polar Front (APF) and enters the South Atlantic mainly via the subtropical gyre. Subantarctic Mode Water (SAMW) originates from deep winter convection north of the Subantarctic Front (SAF). A low-density type of SAMW enters the thermocline of the Southern Hemisphere oceans along the southern edge of the subtropical gyres where it becomes part of central waters and contributes to ventilating the thermocline, while a denser type of SAMW formed in the eastern South Pacific is regarded as a precursor of the AAIW. The Polar Front Zone (PFZ) and Subantarctic Zone (SAZ) are the regions between the APF and SAF, and between the SAF and Subtropical Front (STF), respectively.



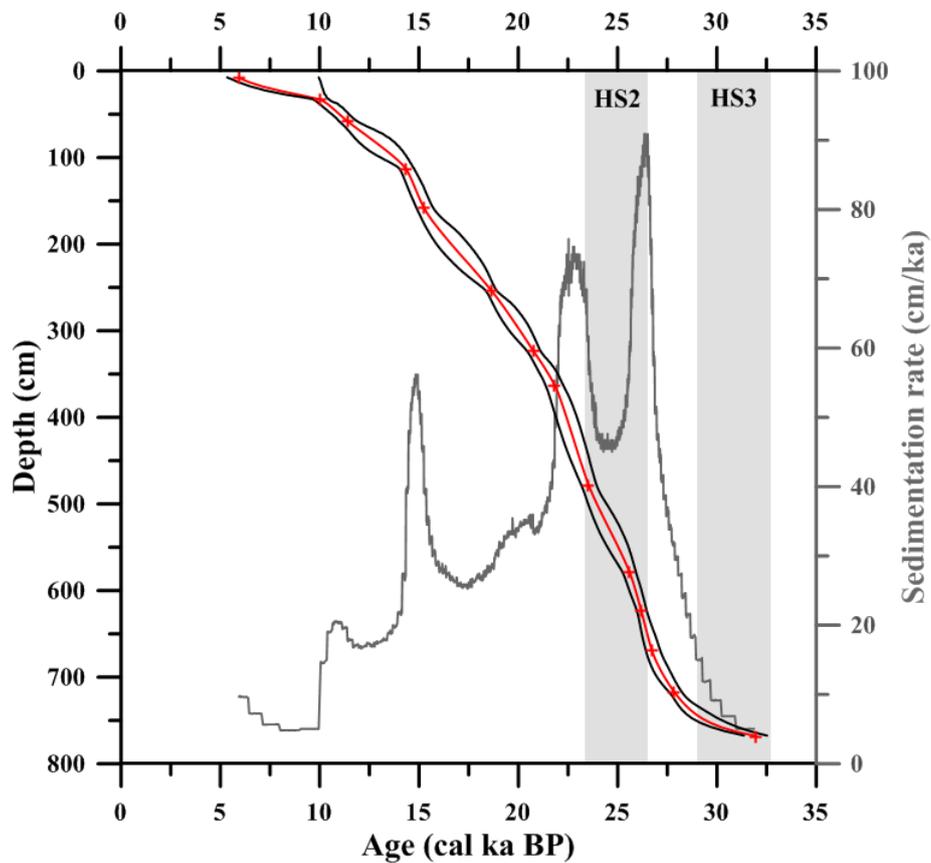
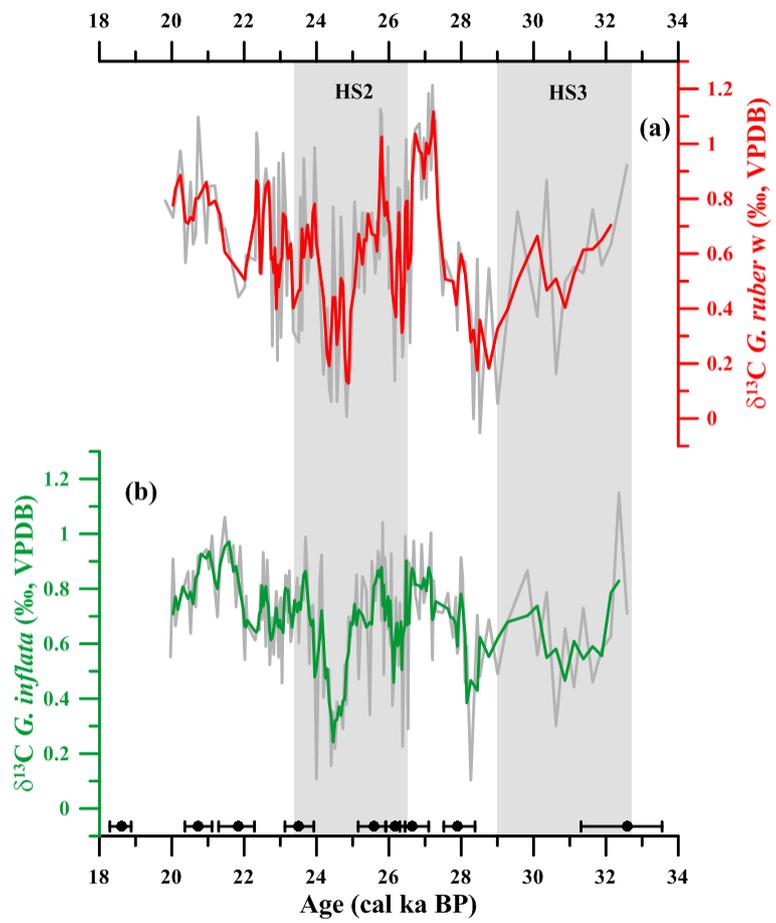


Figure 3: Age model (left hand side y-axis; red line and black enveloping curves) and sedimentation rates (right hand side y-axis; grey line) for marine sediment core GeoB6212-1 produced with the software Bacon 2.2 (Blaauw and Christen, 2011). For the age model, the red symbols show calibrated ages, the red line depicts mean ages and the upper (lower) black line depicts maximum (minimum) ages. Grey vertical bars show abrupt millennial-scale climate change events Heinrich Stadial 3 (HS3) and Heinrich Stadial 2 (HS2) (Goni and Harrison, 2010; Sarnthein et al., 2001).

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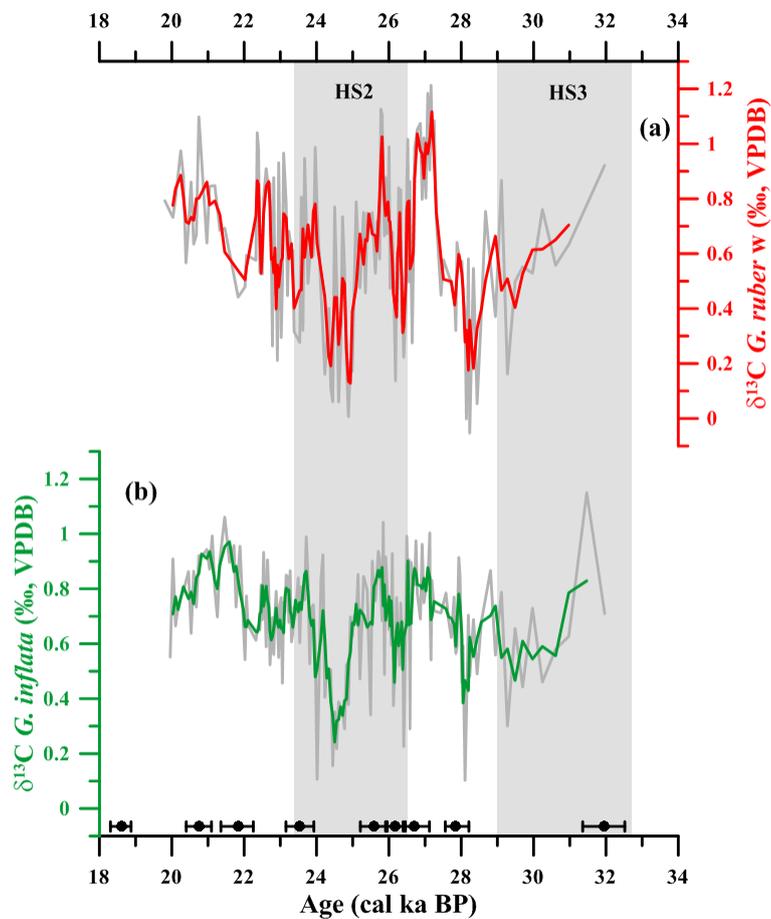


Figure 4: Stable carbon isotopic ($\delta^{13}\text{C}$) records from sediment core GeoB6212-1. (a) *Globigerinoides ruber white* (*G. ruber w*) $\delta^{13}\text{C}$ and (b) *Globorotalia inflata* (*G. inflata*) $\delta^{13}\text{C}$. Red and green lines represent three-point running averages for *G. ruber w* and *G. inflata*, respectively. Black symbols at the bottom of the panel depict calibrated ages. Grey vertical bars show abrupt millennial-scale climate change events Heinrich Stadial 3 (HS3) and Heinrich Stadial 2 (HS2) (Goni and Harrison, 2010; Sarnthein et al., 2001).

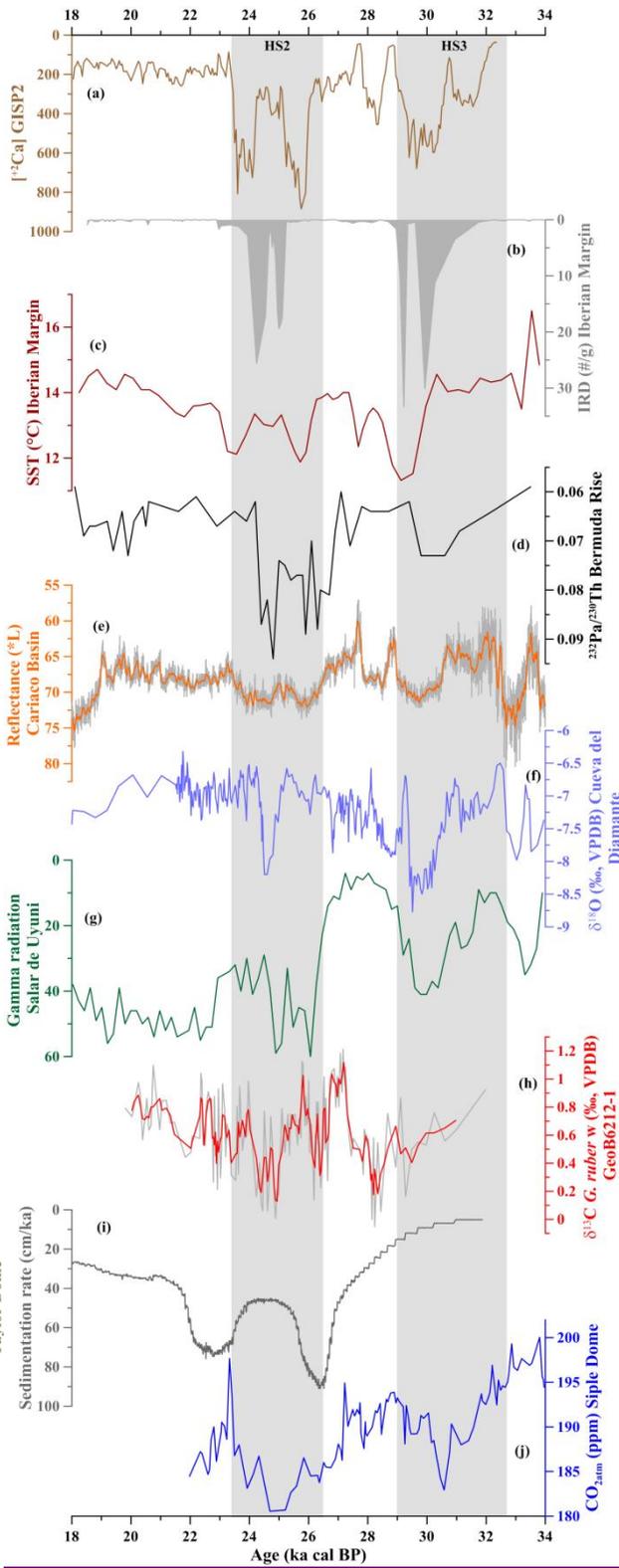
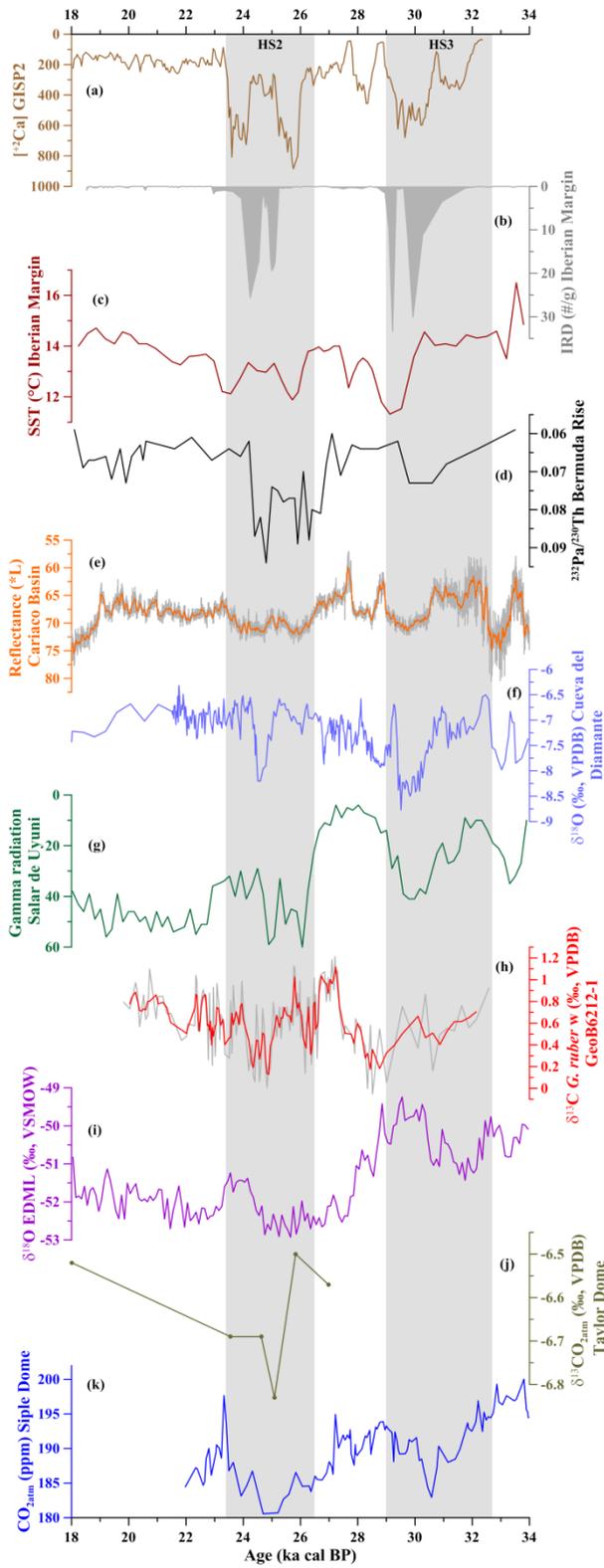


Figure 5: Proxy records from the western South Atlantic, western and eastern North Atlantic and tropical South America spanning Heinrich Stadial 3 (HS3) and Heinrich Stadial 2 (HS2) (Goni and Harrison, 2010; Sarnthein et al., 2001). (a) Changes in atmospheric circulation over Greenland derived from Greenland Ice Sheet Project 2 (GISP2) Ca^{+2} concentration (Mayewski et al., 1997) plotted versus the Greenland Ice Core Chronology 2005 (GICC05) (Andersen et al., 2006; Rasmussen et al., 2006) at 72.6°N , 38.5°W ; (b) Heinrich layers indicated by the presence of ice-rafted debris (IRD) from the Iberian Margin marine sediment core MD99-2331 at 42.2°N , 9.7°W (Eynaud et al., 2009); (c) sea surface temperature (SST) ($^\circ \text{C}$) changes from Iberian Margin marine sediment cores SU8118 and MD952042 at 37.5°N , 10.1°W (Bard, 2000~~2~~); (d) Atlantic meridional overturning circulation (AMOC) strength derived from Bermuda Rise sedimentary $^{231}\text{Pa}/^{230}\text{Th}$ ratio – ODP Site 1063 (higher values indicate a reduced AMOC) at 33.7°N , 57.6°W (Lippold et al., 2009); (e) position of the Intertropical Convergence Zone (ITCZ) indicated by reflectance (*L) (higher values indicate decreased precipitation) from the Cariaco Basin marine sediment core MD03-2621 at 10.7°N , 65°W (Deplazes et al., 2013) (orange line represents a 399-point running average); (f) strength of western Amazon precipitation indicated by the $\delta^{18}\text{O}$ from stalagmite NAR-C collected in the Cueva del Diamante Cave, western Amazon (more negative values indicate increased precipitation) at 5.4°S , 77.3°W (Cheng et al., 2013); (g) presence of paleolakes indicated by the natural gamma radiation from Bolivian Altiplano Salar de Uyuni (higher values indicate increased precipitation) at 20.3°S , 67.5°W (Baker et al., 2001); (h) *Globigerinoides ruber* white (*G. ruber* w) $\delta^{13}\text{C}$ from marine sediment core Geob6212-1 collected in the western South Atlantic at 32.4°S , 50.1°W (red line represents a three–point running average) (this study); (i) sedimentation rates from marine sediment core Geob6212-1 collected in the western South Atlantic at 32.4°S , 50.1°W (this study); atmospheric temperature over Antarctica indicated by EPICA Dronning Maud Land (EPICA Community Members, 2006) $\delta^{18}\text{O}$ (higher values indicate warmer conditions) plotted versus its original chronology at 75°S , 0°E ; (j) $\delta^{13}\text{C}$ of atmospheric CO_2 ($\delta^{13}\text{C}_{2\text{atm}}$) from Taylor Dome (lower values indicate increased input to the atmosphere of $\delta^{13}\text{C}$ -depleted respired CO_2) at 77.8°S , 158.7°E (Smith et al., 1999); (k) atmospheric CO_2 concentration (ppm) from ice core Siple Dome (Ahn and Brook, 2014) plotted versus the Greenland Ice Core Chronology 2005 (GICC05) (Svensson et al., 2008) at 81.7°S , 148.8°W .