Key:
Black= Reviewers' comments
Blue= Authors' responses
Green = Modified text in the manuscript
Please note that all line numbers pertain to the first version of the manuscript.

We thank Reviewer 2 for providing helpful comments on our manuscript. Please see below our responses to these comments.

The manuscript by Bengston et al. seeks to document the d13C of the LIG ocean for comparison to the mid-Holocene. Using published datasets, the authors calculate the average d13C for the LIG and Holocene and find that the LIG in certain areas was more 13C-depleted, by ~0.2 per mil. Given that atmospheric d13C was lower during the LIG, differences in air-sea gas exchange cannot be invoked to account for the oceanic discrepancy. Instead, the authors suggest the light LIG reflects a long-term imbalance between weathering and burial of carbon.

Strengths

The background section is a comprehensive review of the LIG literature that nicely summaries the keys aspects of LIG climate.

The authors assembled an impressive array of time series and evaluated potential biases associated with the averaging techniques. While it would always be useful to have more d13C data, especially in the volumetrically dominant Pacific, they make a compelling initial case that oceanic d13C in certain oceanic regions during the LIG was lighter than during the Holocene. The authors explicitly acknowledge the paucity of data in the Indian and Pacific Oceans, and work to address the issue by focusing on a few areas with relatively high density of d13C records. In doing so, they are able to demonstrate, at least in certain regions, that there is a statistically different mean d13C during the LIG.

Weaknesses

The treatment of AMOC differences between the Holocene and LIG is underdeveloped. While there is evidence of short-term AMOC changes during the LIG that do not occur during the Holocene (e.g. Galassen et al., 2014), there are several other records from the North Atlantic that suggest the first half of the LIG had lower d13C values, which may reflect a weaker AMOC (see records summarized in Hodell et al., 2009, EPSL, 288, 10-19).

Thank you for drawing this to our attention. Please note that we have responded to this concern below when addressing the comment on our chosen time period within the LIG.

The age models used in the compilation are taken from published records. Given that most of the cores are from Lisiecki and Stern (2016), this shouldn't be a major issue because LS16 uses a consistent tuning method. However, the records in Oliver et al. (2010) and the other papers may use slightly different approaches. It would therefore be very useful to apply the methodology from LS16 to all of the cores in the presented compilation to eliminate potential age model biases. In lieu of such an effort, the authors could show how well the various d18O records during MIS 5d, 5e, and 6 align with the LS16 stack as evidence that age model offsets are not a major concern.

We thank the Reviewer for this comment. We have accordingly checked the d18O data from the other sources. There were indeed small dating offsets between some of the additional cores and the LS16 aligned data. We have adjusted these age models to align with the geographically closest LS16 stacks. We now provide a plot of the data before and after the adjustment in Fig S1. We have also updated our d13C analysis accordingly, noting that there were only small changes in our results due to the relatively small portion of the dataset that was affected. The following was added to the manuscript to L127:

In order to align all of the records, adjustments to the age models of cores from Oliver et al. (2010) and the four additional cores (CH69-K09, MD95-2042, MD03-2664 and ODP 1063) were made by aligning the d18O minima during the LIG to corresponding d18O minima of the nearest LS16 stack. The d18O data before and after the alignment is given in Fig. S1.

The other primary weakness is the limited number of records for the Pacific (18 LIG, 19 Holocene) and Indian Oceans (4 LIG, 7 Holocene). Given that the Pacific and Indian Oceans combined have ~3x the volume of the Atlantic, and therefore >3x the DIC, the paucity of data coverage in the Pacific and Indian Oceans is the greatest source of uncertainty for the mean oceanic d13C estimate. The addition of only a handful of Pacific records with slightly more positive d13C values could alter the conclusion that the mean oceanic d13C during the LIG was less than the Holocene.

Additionally, a non-trivial proportion of the Pacific records appear to come from relatively shallow locations, creating another source of potential bias. Here it would be useful to show not only the spatial coverage, as in Figure 1, but also a figure showing the depth coverage in zonal sections through the three major ocean basins. The authors address the depth dependency in Figure 3, where they calculate mean values based only those cores deeper than 2500 m. They also note that the volume weighted regional values are based on cores deeper than 1000 m. For the reader to get a better sense of the data coverage vs. depth, however, it would be very helpful a figure with the zonal sections or a figure showing the eight regions used to estimate the regional values, with core locations superimposed.

We agree with the Reviewer that it would be a useful addition to the manuscript to have a figure of the data presented zonally. For this reason, we have added a new figure to the supporting materials (Fig S2) and refer to this in L137:

The spatial distribution of the database for the Holocene and the LIG is shown in Fig. 2 and the depth distribution of each ocean basin is shown in Fig. S2.

We have added the following sentences to L203 and refer to Fig. S2:

We also note that the average depths of cores from the Pacific Ocean (LIG: 2,711 m, Holocene: 2,131 m) and Indian Ocean (LIG: 2,383 m, Holocene: 2,303 m) are shallower than that of the Atlantic Ocean (LIG: 3,531 m, Holocene: 3,157 m; Fig. S2). However, as the vertical gradient below 2,000 m depth in the Pacific Ocean is small (e.g. Eide et al., 2017), this might not significantly impact our results.

The other main weakness of the paper is the focus on the late LIG, which is motivated by the desire to avoid the lighter d13C observed in the early portion of many early LIG records. The authors note that their focus on the late LIG is to avoid low d13C values associated with the penultimate deglaciation, which is a reasonable consideration. However, many of these light d13C values occur well within MIS 5e as defined by the oxygen isotope stratigraphy in the associated cores (see for example the records in Hodell et al., 2009). Focusing on the late LIG for comparison to the Holocene makes sense for the mean d13C comparison, but it biases the Atlantic LIG records towards heavier d13C values, which therefore minimizes any differences in d13C that are related to AMOC variability. In other words, it is very likely that the authors are missing differences in the AMOC between the LIG and Holocene by focusing on the late LIG in the Atlantic d13C records.

Thank you for this suggestion. We have now added an analysis of d13C for a slightly earlier period (128-123 ka BP) to the new manuscript. Figure 4 shows the data distribution across 3 LIG time periods, as well as their median, first and third quartiles. Figure S3 compares the latitude-depth d13C distributions in the Atlantic basin during the early and late LIG. Figure 3 also shows the d13C time-evolution in the Northeast, Equatorial and southeast Atlantic from 130 to 118 ka. Our analysis suggests that there was indeed a difference in the volume weighted mean d13C between the early LIG (128-123 ka BP) and our time slice (125-120 ka BP), even though this difference is small (0.06 permil). We did not find a significant difference in NADW extent though additional studies are needed to fully resolve this. We would like to stress that we would not expect centennial-scale AMOC slowdown events to be detectable in this analysis because of age model uncertainties and the overall length of the time considered in this analysis (5 ka). We have added the following to the discussion L273:

A statistical reconstruction of the early LIG (128–123 ka BP) δ13C compared to our 125–120 ka BP reconstruction does not reveal a significant difference in either the NADW core depth or NADW extent as indicated by the meridional d13C gradients (Fig. S3). The volume weighted average d13C during the early LIG is 0.06 permil lighter than during the LIG period considered here (125-120 ka BP). Since both time slices (128–123 ka BP and 125–120 ka BP) are 5 ka averages and include dating uncertainties of ~2 ka, it is not possible to resolve potential centennial-scale oceanic circulation changes (e.g. Galaasen et al., 2014b; Tzedakiset al., 2018).

Additional points:

Title: As ‘interglacial’ is an adjective, please consider using instead ‘interglacial period’ or ‘interglacial interval’.

We have adjusted the title to be:

Lower oceanic d13C during the Last Interglacial Period compared to the Holocene

Figure 1: Given the issues associated with scaling dD to temperature, it would be helpful to use only dD for the y-axis here, with some explanation of how dD scales to temperature with modern spatial relationships. Alternatively, consider including dD on one of the y-axes so the reader understands the source of the temperature estimate. Please also include error estimates for the SST record so it is clear what part of the temperature signal is statistically meaningful. (On the positive side, the comparison of the LIG and Holocene on the same x and y axes is very useful for showing the clear difference in CO2 history between the two intervals.)

Thank you for the suggestion. We have added dD on a secondary y-axis along with the estimated temperature at this Antarctic site. We have also provided details on the age model and the method by which temperature was calculated from dD.

We have changed the SST data presented, now showing two cores, one from the Iberian Margin, and the other from the North Atlantic. The data from the North Atlantic is now presented with the standard deviation.

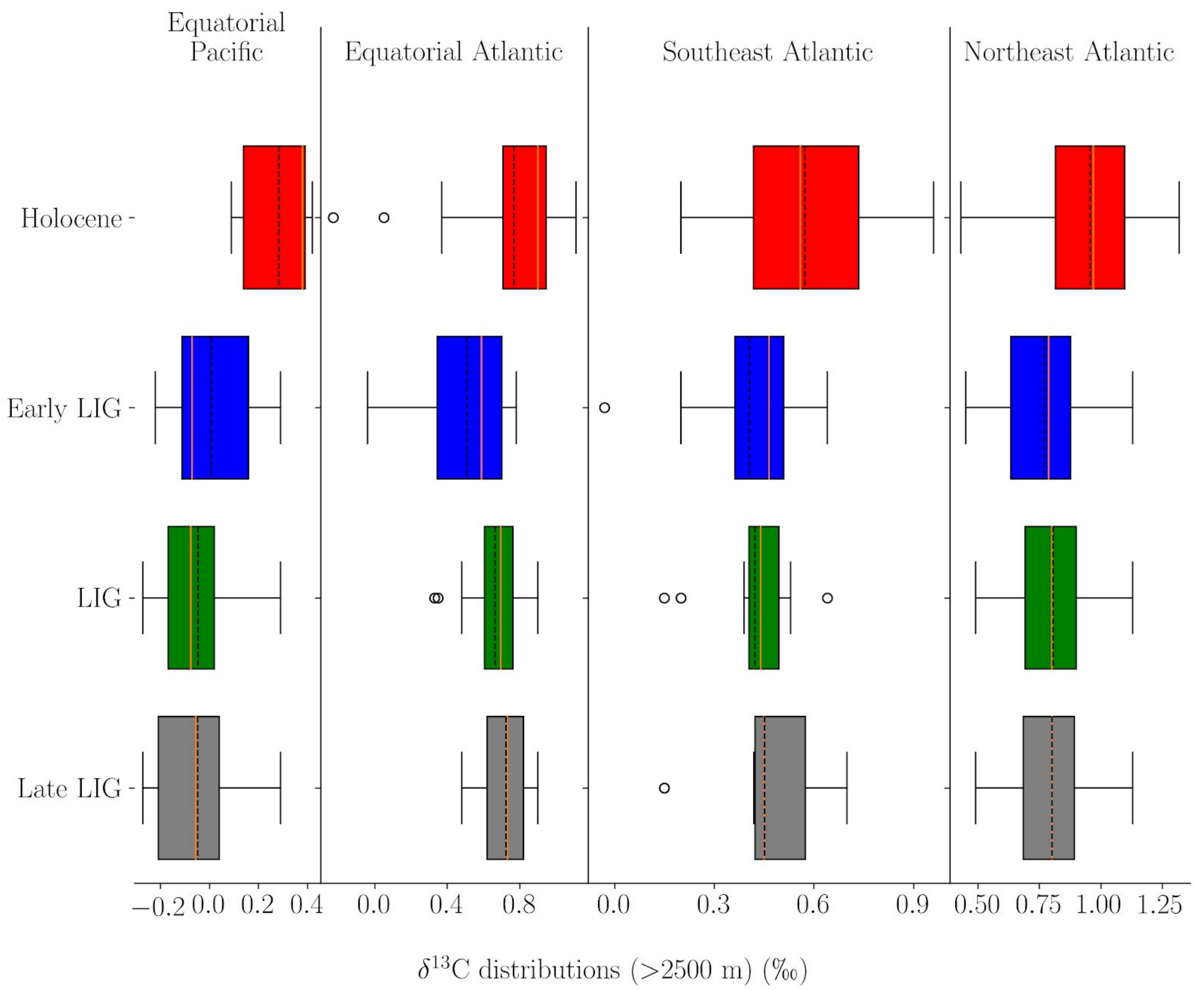
Figure 2: Please specify the confidence limit associated with the whiskers. Also note the statistical range noted by the colored boxes (box and whisker diagrams aren't particularly common in the paleo literature).

We have added the following details to the figure caption (now Figure 3):

Lower end of the box indicates quartile 1 (Q1) and the upper end indicates quartile 3 (Q3). Orange vertical lines show the median and dotted vertical lines show the mean. The whiskers indicate the lower and upper fences of the data calculated as Q1-1.5*(Q3-Q1) and Q3+1.5*(Q3-Q1), respectively, and the clear circles are outliers.

Figure 3: Note that the choice of averaging interval for the LIG has a non-trivial influence on the mean d13C for the equatorial and SE Atlantic. Given that the chosen interval of 125-120 ka is somewhat arbitrary, it is necessary to more fully explore the sensitivity of the findings to the choice of time interval. For example, if the defined LIG interval were 124-120 ka, several light d13C points would be excluded, resulting in a higher mean LIG d13C. If the lighter points are excluded, like those earlier in the LIG, what is the resulting mean LIG d13C for the equatorial and SE Atlantic? Is it statistically different than the Holocene d13C?

Thank you for raising these questions. We have now included box plots (moved from Fig. 2 to a new Figure, now Figure 3) which explore the sensitivity of the anomaly to the LIG time period considered within 128 ka BP and 118 ka BP:



We have added the following accompanying paragraph to Section 3.1:

We also compare the distribution of d13C for cores deeper than 2,500 m for three overlapping periods within the LIG (early LIG: 128--123 ka BP; LIG: 125--120 ka BP; late LIG: 123--118 ka BP). The results for the four regions are shown in Fig. 4. The statistical characteristics do not show much variation between the LIG and late LIG d13C distributions. In the equatorial Pacific, the difference between the early LIG and the Holocene is smaller than between LIG and Holocene, but this is countered with a larger difference in the equatorial Atlantic between early LIG and Holocene. The spread in the data is generally larger during the Holocene than during the other time periods which might be due to the greater number of measurements during the Holocene. The spread of data during the early LIG is slightly larger than during the LIG and late LIG in the equatorial and southeast Atlantic. The equatorial Atlantic is the only region which displays significantly more points with lower d13C during the early LIG. Overall, these distributions do not suggest that the LIG-Holocene anomalies that we have determined would be significantly impacted by slight variations in the selected time window. We perform an analysis of variance (ANOVA) on each region and post hoc tests on the data. We find that the Holocene data is significantly different from the three LIG periods in the northeast Atlantic, the southeast Atlantic and the equatorial Pacific, while the three periods within the LIG are not significantly different from each other for any of the regions.

Additionally, we have improved our explanation of our selected time periods in the manuscript:

We then define the time periods within the LIG and the Holocene to perform our analyses. For the Holocene, as most of the available data is dated prior to 2 ka BP, we define the end of our Holocene time period as 2 ka BP. To capture as much of the Holocene data as possible, we include data back to 7 ka BP, ensuring that we do not include instability associated with the 8.2 kiloyear event (Alley et a., 2005; Thomas et al., 2007). This provides a time span of 5 ka of data that we will consider for our analysis of the Holocene.

For the LIG, we seek to avoid data associated with the end of the penultimate deglaciation, which is characterised by a benthic d13C increase in the Atlantic until ~128 ka BP (Govin et al., (2015); Menviel et al. (2019); Oliver et al. (2010), Fig. 3). In addition, a millennial-scale event has been identified in the North Atlantic between ~127 and 126 ka BP (Galaasen et al., 2014, Tzedakis et al., 2018). Considering the typical dating uncertainties associated with the LIG data (2 ka), we thus decide to start our LIG time period at 125 ka BP. To ensure that the two time periods are of same length (5 ka BP), we define the LIG period for our analysis to be 125-120 ka BP. We note that our definition should also avoid data associated with the glacial inception (Govin et al. (2015); Past Interglacial Working Group of PAGES, 2016). We verify that the LIG time period has sufficient data across the four selected regions, noting that the highest density of data falls within the 125-120 ka BP time period---particularly in the equatorial Atlantic and southeast Atlantic (Fig. 3b, c).

Line 85: While remineralization contributes to the lowering of NADW d13C as waters flow toward the Southern Ocean, the residence time of NADW is quite short in the Atlantic, minimizing the influence of remineralization. Mixing with 13C-depleted UCDW and AABW also contributes to the deep South Atlantic being 13C-depleted relative to the North Atlantic.

It is true that the remineralisation is not the only mechanism responsible for the decrease in d13C of NADW. We have rephrased the sentence to read:

Along its path through the Atlantic basin interior, organic matter remineralisation and mixing with southern source waters lowers $\delta^{13}C$, with $\delta^{13}C$ values of ~0.5 ‰ in the deep Southern Ocean.

Line 87: This sentence is written in such a way to give the impression that the use of d13C as a circulation proxy is a recent phenomenon. But in the following sentence, there are citations of classic papers where d13C was used for exactly this purpose. Please clarify.

Sorry that this sentence was misleading. L86-87 has been changed to:

The tight relationship between the water masses' apparent oxygen utilisation, nutrient content and d13C allows d13C to be used as a water mass ventilation tracer (e.g. Boyle and Keigwin, 1987; Curry and Oppo, 2005; Duplessy et al., 1988; Eide et al., 2017).

Lines 285-305: The authors suggest that the -0.2 per mil difference in mean oceanic d13C during the LIG may have been due to less organic carbon in the land biosphere. Unfortunately, there is no effort to estimate how much land carbon would be required to create the d13C anomaly. While this would assume a closed atmosphere-biosphere-ocean system, making this assumption explicit would then allow for informed speculation on the likely sources of terrestrial carbon. The estimate of terrestrial carbon loss could then be compared to various reservoirs (e.g. peats) to assess whether they are likely sources.

A mass balance calculation would imply that the system is closed. Given that the LIG and Holocene are more than 100,000 years apart, the closed system approximation is associated with uncertainties that are too large to be included in the main part of the manuscript. Nevertheless, we now include the d13C mass balance calculation in the supplementary materials. And reference it in the following paragraph in the revised discussion:

An alternative explanation for the anomaly is a change in the terrestrial carbon storage, which has a typical signature of approximately -37 to -20 ‰ for C3 derived plant material (Kohn, 2010) and -13 ‰ for C4 derived plant material (Basu et al., 2015). The total land carbon content at the LIG is poorly constrained. Proxies generally suggest extensive vegetation during the LIG compared to the Holocene (CAPE, 2006; Govin et al., 2015; Larrasoaña et al., 2013; Muhs et al., 2001; Tarasov et al., 2005; de Vernal and Hillaire-Marcel, 2008), which would imply a greater land carbon store. However, other terrestrial carbon stores including peatlands and permafrost may also have differed during the LIG compared to the Holocene. With an estimated ~550 Gt C stored in peats today (mean δ 13 305 C ~-28 ‰, Dioumaeva et al. (2002); Novák et al. (1999)) and ~1,000 Gt C in the active layer in permafrost, which may have been partially thawed during the LIG (Reyes et al., 2010; Schuur et al., 2015; Stapel et al., 2018), less carbon stored in peat and permafrost at the LIG could have led to a lower total land carbon store compared to the Holocene. However, it is not possible to infer this total land carbon change from the oceanic and atmospheric δ13C anomalies because it cannot be assumed that the mass of carbon and 13C is preserved within the ocean-atmosphere-land biosphere system on glacial-interglacial timescales.

There is indeed continuous exchange of carbon and 13C between the lithosphere and the coupled ocean, atmosphere and land biosphere carbon reservoirs. Isotopic perturbations associated with changes in the terrestrial biosphere are communicated to the burial fluxes of organic carbon and CaCO3 and are therefore removed on multi-millennial time scales (Jeltsch-Thömmes et al., 2019; Jeltsch-Thömmes and Joos, 2020). Nevertheless, when hypothetically neglecting any exchange with the lithosphere, we find that the change in terrestrial carbon needed to explain the difference in δ13C would be in the order of 295±44 Gt C less during the LIG than the Holocene (Text S1).

Lines 309-311: The idea about long-term imbalance between weathering and burial of carbon needs to be explained more thoroughly. How would these processes create the difference in LIG and Holocene d13C of DIC? The cited paper by Jeltsch-Thommes and Joos (2020) is a modeling study that evaluates the influence a large pulse of carbon introduced to the atmosphere, assuming that the carbon comes from the terrestrial biosphere. The simulations suggest that that oceanic d13C responds quickly to the addition of 500 Gt of terrestrial organic carbon, creating an oceanic anomaly of ~ -0.2 per mil within about 500 years. The d13C anomaly persists for 10 kyr, before slowly returning to its initial value after approximately 100 kyr (due to removal of light carbon through biogenic sedimentation). Are the authors suggesting that such a process could explain the apparent difference between LIG and Holocene d13C?

We agree with the Reviewer that the discussion on exchanges with the lithosphere as cause of the d13C anomaly was not clear. We have revised major parts of the discussion to better explore why we believe this mechanism is critical to understanding the anomaly.

The parts of the discussion exploring the possible mechanisms for the d13C anomaly now read:

Explanations for the 0.2 ‰ lower δ13C anomaly in the ocean may include a redistribution between the ocean-atmosphere system. Such a redistribution can result from a change in end-member values (Fig. 7). As fractionation during air-sea gas exchange is temperature dependent, globally higher SSTs at the LIG could lead to a lower oceanic δ13C. However, the effect of this is likely small (Brovkin et al., 2002) and this would also lead to a higher atmospheric δ13CO2 at the LIG, which is inconsistent with Antarctic ice core measurements that suggest an anomaly of -0.3 ‰ (Schneider et al., 2013). Lower nutrient utilisation in the North Atlantic would decrease surface ocean δ13C and thus the δ13C end-members. However, this would also imply that less organic carbon would be remineralised at depth. Therefore, it is unlikely that the lower average oceanic mean δ13C results from a change in end-members through lower surface ocean nutrient utilisation. Currently, there is still a lack of constraints on nutrient utilisation in these end-member regions during the LIG compared to the Holocene. Therefore, the lower δ13C in the ocean-atmosphere system cannot be explained by a simple redistribution of δ13C between the atmosphere and the ocean.

An alternative explanation for the anomaly is a change in the terrestrial carbon storage, which has a typical signature of approximately -37 to -20 ‰ for C3 derived plant material (Kohn, 2010) and -13 ‰ for C4 derived plant material (Basu et al., 2015). The total land carbon content at the LIG is poorly constrained. Proxies generally suggest extensive vegetation during the LIG compared to the Holocene (CAPE, 2006; Govin et al., 2015; Larrasoaña et al., 2013; Muhs et al., 2001; Tarasov et al., 2005; de Vernal and Hillaire-Marcel, 2008), which would imply a greater land carbon store. However, other terrestrial carbon stores including peatlands and permafrost may also have differed during the LIG compared to the Holocene. With an estimated ~550 Gt C stored in peats today (mean δ 13 305 C ~-28 ‰, Dioumaeva et al. (2002); Novák et al. (1999)) and ~1,000 Gt C in the active layer in permafrost, which may have been partially thawed during the LIG (Reyes et al., 2010; Schuur et al., 2015; Stapel et al., 2018), less carbon stored in peat and permafrost at the LIG could have led to a lower total land carbon store compared to the Holocene. However, it is not possible to infer this total land carbon change from the oceanic and atmospheric δ13C anomalies because it cannot be assumed that the mass of carbon and 13C is preserved within the ocean-atmosphere-land biosphere system on glacial-interglacial timescales.

There is indeed continuous exchange of carbon and 13C between the lithosphere and the coupled ocean, atmosphere and land biosphere carbon reservoirs. Isotopic perturbations associated with changes in the terrestrial biosphere are communicated to the burial fluxes of organic carbon and CaCO3 and are therefore removed on multi-millennial time scales (Jeltsch-Thömmes et al., 2019; Jeltsch-Thömmes and Joos, 2020). Nevertheless, when hypothetically neglecting any exchange with the lithosphere, we find that the change in terrestrial carbon needed to explain the difference in δ13C would be in the order of 295±44 Gt C less during the LIG than the Holocene (Text S1).

In addition, due to the warmer conditions at the LIG than during the Holocene, there could have been a release of methane clathrates which would have added isotopically light carbon

($\delta^{13}C$: ~-47 ‰) to the ocean-atmosphere system. However, available evidence suggests that geological CH_4 sources are rather small (Bock et al., 2017; Hmiel et al., 2020; Petrenko et al., 2017; Saunois et al., 2020) making this explanation unlikely, although we cannot completely exclude the possibility that the geological CH_4 source was larger at the LIG than the Holocene. Similarly, since the $\delta^{13}C$ value of CO_2 from volcanic outgassing is close to zero (Brovkin et al., 2016) and modelling suggests volcanic outgassing likely only had a minor impact on $\delta^{13}CO_2$ (Roth and Joos, 2012), it is unlikely that volcanic outgassing of CO_2 played a significant role in influencing the mean oceanic $\delta^{13}C$.

While we are not in the position to firmly pinpoint the exact mechanism, the LIG-Holocene differences in the isotopic signal of both the atmosphere and ocean were most likely due to a long-term imbalance between the isotopic fluxes to and from the lithosphere, including the net burial (or redissolution) of organic carbon and $CaCO_3$ in deep-sea sediments, and changes in shallow water sedimentation and coral reef formation (Jeltsch-Thömmes and Joos, 2020).