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Interactive comment on "Western Mediterranean hydro-climatic consequences of Holocene iceberg advances (Bond events)" *by* Christoph Zielhofer et al.

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Dear Editor, Anonymous Referee #1 gave as important suggestions about our proxy data set. We will consider his comments in the revised version of our manuscript.

The following paragraphs are preliminary replies to the referee's comments:

Issue 1) "The authors have published the interpretation of the 18O record in Zielhofer et al. (2017). They interpret the 18O record as a proxy for winter precipitation, which is based on a multi-proxy approach with a charcoal, and cedar pollen abundance records. Although, I can follow their line of arguments that cedar trees need enough moisture

and the charcoal record may represent fire activity, I don't see a clear correlation between the charcoal record and the cedar pollen abundance."

In the first version of the CP manuscript, we applied a multi-proxy interpretation that is based on two scales. Orbital scale and multi-centennial to millennial scale.

At orbital scale δ18O values increase from approx. +3 to +5 ‰ in the Early Holocene to values from +5 to +7 ‰ in the Late Holocene. As the most straightforward scenario for a subhumid, closed basin (Roberts et al., 2008) this implies a decrease of the precipitation/evaporation ratio with generally more arid conditions towards the Late Holocene. This corresponds with Sidi Ali diatom, TOC and carbonate records that indicate in average higher lake levels during the Early Holocene and lower levels at later stages (Zielhofer et al., 2017).

These findings seem to contradict the Cedrus pollen record from Lake Sidi Ali (Campbell et al., 2017) that shows a low or even missing occurrence of cedars during the Early Holocene, indicating reduced moisture availability at that time. However, reduced moisture availability for the cool-preferring cedar seems to be the result of enhanced summer heat during the Early Holocene. Due to their shallow roots, cedars are vulnerable to summer heat in contrast to the warm tolerant and deep-rooting evergreen oaks that dominate the Sidi Ali pollen record during the Early Holocene (Campbell et al., 2017). In this context, we attest summer temperature-driven drought stress and not winter precipitation as the limiting factor for the long-term trend of Holocene cedar occurrence in the Middle Atlas. This inference is in good agreement with the orbital-forced summer insolation maximum during the Early Holocene (Berger 1978) and the chironomid-based summer temperature reconstructions that indicate enhanced Mediterranean summer temperatures during the Early Holocene as well (Samartin et al., 2017).

Following our interpretation, at orbital scale our proxies show reduced winter precipitation (high δ18O) and (summer) cool conditions (high amount of Cedrus pollen) during the Late Holocene and enhanced winter precipitation (low δ18O) and enhanced summer heat (low Cedrus pollen amounts) during the Early Holocene. We think that this is currently the best interpretation but alternatives may exist. Furthermore, we worked out that enhanced δ18O values might be also the result of a specific origin and seasonality of the precipitation-bearing air masses. We assume that Late Holocene precipitation from springtime Mediterranean cyclones reveal higher δ18O values than Atlantic winter rains (Zielhofer et al. 2017).

However, we have to agree with Referee #1 that the orbital pattern between Cedrus and Sidi Ali δ18O is not clearly visible in the comparison of the Cedrus record with our δ18O record at multi-centennial and millennial time scales (here see Fig. 2 D and G) . In the revised manuscript, we will discuss this issue and will argue that the Cedrus record is influenced by summer heat stress and that summer heat might be predominantly forced by the subtropical high and not by North Atlantic air masses. This is visible in the "in phase" pattern between subtropical summer SST and our Cedrus record (Figure 1). We argue that reduced summer heat ("cooling at Sidi Ali") can be in phase with reduced winter rainfall at Sidi Ali (e.g. 8.2 and 10.2 ka) but that there are also indications for out-of-phase pattern (e.g. 1.2, 7.3, or 9.3 ka). Hence, this out-of-phase pattern might be influenced by different forcing mechanisms for summer cooling (sub-tropical high) and winter rain (North Atlantic winter cyclones). In the revised manuscript we will follow this line of argument that might explain the weak matching between Cedrus and δ18O at multi-centennial to millennial time scales. In the revised manuscript, we will argue that the Sidi Ali δ18O record is "in phase" with Bond's HSG record to support the idea of a teleconnection between Western Mediterranean winter precipitation and North Atlantic cooling. As Cedrus might reflect predominantly summer cooling we might not be able to detect North Atlantic (winter) cooling from our own record directly.

Due to uncertainties in age models (Sidi Ali vs. Bond record), we are not able to provide significant correlations between HSG and Sidi Ali δ18O. Here, we argument more carefully now as suggested by referee #1. We avoid the term "correlation" and use the terms "in phase pattern" and "out of phase". Further, we apply lowpass filters (pro-

gramme PAST) that reduce the centennial-scale variabilities of the proxy records (see Fig. 2). In the revised manuscript, therefore, we argue more carefully. The blue and orange bars in Fig. 2 indicate in phase pattern of the filtered HSG and δ18O records. Further, we agree with referee #1 that the charcoal record is a difficult proxy for the interpretation of landscape dynamics at multi-centennial time scales. There might be no consistent matching between the charcoal and vegetation record. The charcoal can be dominated by individual fire events and these do not appear systematically linked to the fluctuations in the Cedrus pollen abundance. The charcoal is a complex proxy – influenced by climate but also by fuel availability, and the relative importance of these two factors seems to shift over the Holocene (Campbell et al. 2017). Therefore, we will not use the charcoal record for multi-centennial proxy interpretation in the revised manuscript.

Issue 2) "The authors clearly state that lake Sidi Ali is a closed basin lake where the Precipitation-Evaporation balance (P-E) plays an essential role in controlling the oxygen isotope composition. This is evident from the highly elevated present day 18O values of the water that range from 0 to +4 ‰ whereas the surrounding karst springs and streams range from -6 to -9 ‰ (Zielhofer et al., 2017). The lake shows a huge range in surface area varying from 2 to 2.8 km2 (Zielhofer et al., 2017) due to varying P-E balance on interannual/decadal timescales. This is extremely likely visible in the 18O of the water. This can be controlled by both evaporation during the dry season, and by replenishment during the winter season, but not only through winter precipitation."

We fully agree with Referee #1 that the lake is strongly affected by evaporation. This was clearly stated in the first version of the manuscript (page 7 line 15). Nevertheless, we tried to work out in our published manuscripts (Zielhofer et al. 2017 and Campbell et al. 2017) that the δ18O record at Sidi Ali is a complex proxy and that the most convincing line of argument results in the variability of winter rains. We would like to mention here that we already worked out a convincing scenario for the early Holocene millennial fluctuations (Zielhofer et al. 2017).

Issue 3) "In order to show that the ostracod 18O variability represents the 18O of the water the authors calculate the theoretical calcite 18O values based on the present-day water 18O and the isotope fractionation factor from Friedman and O'Neill (1977). Why not using the much more recent isotope fractionation factor from Kim and O'Neill (1997)? Please show a range of possible temperatures that can be calculated taking into account different isotope fractionation factors."

We followed this suggestion and used the equation by Kim and O'Neil (1997) to calculate δ18O values for theoretical calcite based on measured water temperatures and the stable isotope composition of modern lake and spring and stream waters nearby. We corrected the resulting values in the text and changed the reference.

Further, we calculated additional δ18O values (Table 1) for carbonate precipitated from Sidi Ali in equilibrium with host water at specific temperature scenarios and depths using the equation by Kim and O'Neil (1997) as suggested by the reviewer.

The lowest water temperature at the lake bottom in September 2012 was 7.7°C. Here, we have a δ18O value (water, SMOW) of 1.21 ‰ for the depth of 30 m. Temperatures are quite stable around this depth. Using the Kim and O'Neil (1997) equation results in a δ18O value (carbonate, VPDB) of 2.55 ‰(The earlier used equation of Friedman and O'Neil (1977) had resulted in a value of 3.05 ‰ which is not so very different.)

The value of 2.55 ‰ is well within the range of data for the Holocene ostracod shells, which is -1.1 to 8.1 ‰ (min - max). In table 1, we combined measured and assumed temperature scenarios with today's δ18O values (water, SMOW). For example, we had measured the highest δ18O value with 2.58 ‰ (water, SMOW) at 5 m water depth in September 2012. Measured temperatures were highest in surface waters and were 19.6°C at maximum. With the exception of one δ18O (carb) value for assumed 25°C that is slightly lower than our measured range for ostracod shells, all calculated theoretical δ18O values (carbonate, VPDB) lie between 0.19 and 4.78 ‰ (Table 1) and are

in the range of our measured δ18O values for Holocene ostracod shells at Sidi Ali.

However, we need to keep in mind that ostracod shells are not precipitated in isotopic equilibrium but often show a vital offset of 1 to 2 ‰ due to the metabolism of the animals (von Grafenstein et al., 1999). Therefore, δ18O values of ostracod calcite are usually 1 to 2 ‰ higher than the assumed values for inorganic calcite. With this in mind, all calculated values in table 1 are within the range of our Holocene ostracod shell data and all these combinations are realistic and cannot be ruled out. The wider range of δ18O from ostracod shells indicates that Holocene lake water δ18O was sometimes higher and sometimes lower than today's δ18O (water) values.

Further, the significantly wider range of Holocene δ18O values from ostracods (-1.1 to 8.1 ‰ shows that solely water temperature changes cannot explain past δ18O variability but changes in the precipitation/evaporation ratio must be considered as well.

Issue 4) "One aspect that has not been discussed is the role of changing water temperatures on ostracod 18O. Particularly during the Early Holocene where the authors argue that there are less cedar trees due to heat stress. There are four phases where there is a clear correlation with the 18O at 10.2, 8.2, 6.0, and 5.2 (I do not see a coherence at 7.3), why are these phases not interpreted as cooler summers? Cooler water temperatures may also result in heavier calcite 18O, and could provide a different interpretation that is consistent with the cedar pollen abundance record. This may also be in line with 'Atlantic cooling'".

Generally, this is exactly what we said. Increased Cedrus pollen indicate cooler summers (Page 8 line 10). However, multi-centennial changes in our δ18O signal are quite large (more than 2 ‰ e.g. 8.2 ka) and the effect of temperature-dependent stable isotope fractionation during the formation of carbonate in water is not large enough for explaining these large changes. The core location is quite deep, and non-marine ostracods are (with very few exceptions) all benthic. Temperatures in modern Sidi Ali approach 8°C beneath the thermocline at 10-14 m. If temperatures were even as low as 4°C (surely not colder at lake floor because of the density maximum of water at 4°C), δ18O values could have been not higher than ca. 1 ‰ (see table 1) due to the temperature change. Early Holocene δ18O fluctuations are often much larger in the record and water temperature alone was surely not driving these fluctuations. Even the slighter variabilities of Late Holocene δ18O values cannot be inferred from temperature-dependent stable isotope fractionation because the Sidi Ali curve shows lower values during Atlantic cooling.

Issue 5) "I do see a possible correlation with the HSG record from Bond et al. for the Early Holocene for the positive 18O peaks around 11.4, 10.2, 8.2. However, for the peaks at 9.3, 7.3, 6.7, 6.0, and 5.2 the variation in the 18O is either very small or the timing is not comparable to the Bond-events. The timing might be due to age-model uncertainties. But in its present form, I'm unable to assess whether the Bond events and the positive peaks in 18O are within error of the age model or not, because the age uncertainties are not indicated in Fig. 2. This is definitely a must."

The age model and age uncertainties were submitted as supplementary online material in the first version of the manuscript. However, we add the error bars in the newly compiled figure (see Fig. 2). Further, we add lowpass filters for a better visualisation of "in phase" and "anti-phase" pattern between HSG and δ18O. The filtered records (500yr lowpass filter) show a good match in multi-centennial variability.

Issue 6) "The 25-point running correlation calculated between the 18O and the Bond record shows correlation that barely reach 0.3, is this significant? Can you draw a line that indicates the 95% confidence level? I'm aware that age model uncertainties should also be taken into account, so this can be discussed."

We checked significance levels. Attained values above 0.4 and below approx. -0.4 are significant (95% confidence level). However, we will remove the 25-point correlation in the revised version of the manuscript and will argue more carefully ("in phase" vs. "out of phase").

Issue 7) "During the Late Holocene the authors try to link peaks at 4.6, 4.2, 3.2, 2.7 to peaks in HSG. I truly think that this is very hard to see, because the variation in 18O is very small."

We add lowpass filters for a better visualisation of in phase and anti-phase pattern between δ18O and HSG records (see Fig. 2). The filtered records might provide a better visualisation.

Issue 8) "The paper shows no figure with a comparison with regional records to test their interpretation of the ostracod 18O record, for example the pollen record from MD95-2043 (Fletcher et al., 2013) should be included. Furthermore, if the authors are correct and their 18O record represents winter precipitation, then a figure with a comparison with NAO records is necessary."

We add a NAO record in the attached figure (Fig. 2 H; Olsen et al., 2012). However, the comparison is limited due to the high variability/higher resolution of the NAO record and age uncertainties. However, both records show similar pattern during the LIA and MCA, for example.

We have doubts about the value of showing MD95-2043 pollen record for comparison. The aim of this paper is to highlight a shift in phasing between δ18O and the Bond record, and to suggest an explanation for that. For example, during the early Holocene (high summer orbital insolation, residual ice sheets), the Bond Events were associated with strong latitudinal temperature gradient and intensification of the westerly flow and weak penetration of winter rains into the W Mediterranean. During the Late Holocene (weak summer insolation, no ice sheets, modern ocean configuration), the Bond events may be more associated with ocean current changes around the dynamics of the ocean gyres and a similar-to-present linking of cold subpolar Atlantic & NAO-like negative pattern leading to increased rainfall. The MD95-2043 shows some similarities for the early Holocene but shows a slow changing millennial behaviour for the Late Holocene that does not really help support or refute the ideas about centennial variability at Sidi

Ali.

Kind regards Christoph Zielhofer, William Fletcher and Steffen Mischke

Table 1. Calculation of theoretical δ18O values for carbonates precipitated from Lake Sidi Ali in equilibrium with host water at specific temperatures using the equation by Kim and O'Neil (1997).

Figure 1: Sidi Ali Cedrus record, Sidi Ali δ18O and SST Hole 658C (1000 yr lowpass filter). The Cedrus record might show a good match with the subtropical SST record but not with winter rain (Sidi Ali δ18O).

Figure 2: Holocene North Atlantic ice-rafted debris record versus Western Mediterranean hydro-climatic record: A) Total solar irradiance (ΔTSI, Steinhilber et al., 2009); B) Holocene Bond events 0 to 8 derived from Bond et al. (1997, 2001); C) Ice-rafted debris (IRD) record based on hematite stained grains of stacked MC52 and VM29-191 cores from the subpolar North Atlantic (Bond et al. 2001), the black line through the Bond record shows results of lowpass filter (500 year) removing centennial variability; D) Improved Sidi Ali δ18O record from closely related species Fabaeformiscandona sp. and Candona sp (this study). The grey line represents the original data. The black line shows results of lowpass filter (500 year) removing centennial variability. Blue/red numbers and pale blue/orange bars indicate North Atlantic cooling events and wet/dry winters in the Western Mediterranean; E) Modelled ages with 2 sigma ranges (Fletcher et al., 2017); F) Summer insolation (65°N, June, Berger, 1978) (note reversed axis); G) Sidi Ali pollen record (Campell et al., 2017) with 500 year lowpass filter; H) Palaeo-NAO record (Olsen et al., 2012).

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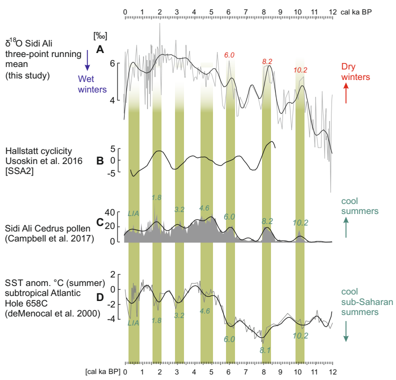

Fig. 1.

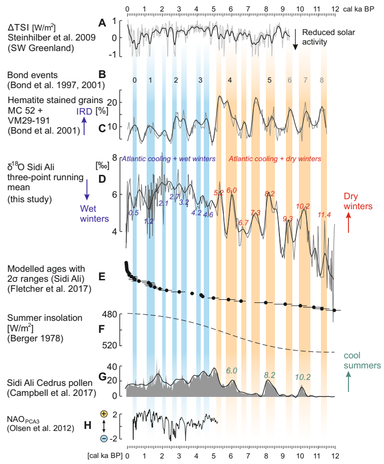

Fig. 2.

Table 1. Calculation of theoretical $\delta^{18}O$ values for carbonates precipitated from Lake Sidi Ali in equilibrium with host water at specific temperatures using the equation by Kim and O'Neil (1997).

Lake water depth	Temp.	$\delta^{18}O$ (water, SMOW)	Theoretical $\delta^{18}O$ (carb., VPDB)	Remarks
30 m	7.7 °C (Sep. 2012)	1.21 ‰ (Sep. 2012)	2.55 ‰	Calculation for actually measured water temperature and $\delta^{18}O$ (water) at 30 m
0 to 5 m	19.6 °C (Sep. 2012)	2.58 ‰ (Sep. 2012)	1.30 ‰	Calculation for maximum measured $\delta^{18}O$ (water) at 5 m depth and maximum surface water temperature (ca. 1 °C warmer than at 5 m)
30 m	4 °C	1.21 ‰ (Sep. 2012)	3.41 ‰	Calculation for assumed coldest water during colder times and present bottom water $\delta^{18}O$ (water)
30 m	4 °C	2.58 ‰	4.78 ‰	Calculation for assumed coldest water during colder times and more evaporated water (similar to surface water)
0 to 5 m	25 °C	1.21 ‰	-1.18 ‰	Calculation for assumed very warm (or shallow) conditions and present bottom water $\delta^{18}O$ (water)
0 to 5 m	25 °C	2.58 ‰	0.19 ‰	Calculation for assumed very warm (or shallow) conditions and today's maximum measured $\delta^{18}O$ (water) at 5 m depth

Fig. 3. Table 1