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Comment on “The deglaciation over the Laurentide Fan: History of diatoms, IRD, ice and fresh water”

Gil et al. (2015) present new and existing data in core OCE326 GGC14 from the Laurentian Fan off of the Laurentian Channel. In their study, they criticize data published from the same core by Obbink et al. (2010) and from a nearby core by Carlson et al. (2007). One major conclusion of Gil et al. (2015) is that “The YD [Younger Dryas cold event; 12.9–11.7 ka; Rasmussen et al., 2006] cannot be regarded as an event triggered by a fresh water input through the Laurentian Channel since only a weak brief input nearly 1000 yrs after its onset is recorded.” Here we show that their study misrepresents existing data, and ignores their own data and its uncertainty, which leads to different conclusions than in Gil et al. (2015).

Gil et al. (2015) suggest that the Mg/Ca calcification temperature (CT) record of Obbink et al. (2010) during Heinrich Event 1 is confounded by other factors than temperature, based on their conclusion of a warming trend during this event in the Obbink et al. (2010) CT data. First, the warming trend during Heinrich Event 1 is only significant, within uncertainty of Mg/Ca–CT measurements, at ~16.5 ka on the age model in their Fig. 3I, which Gil et al. (2015) note as a reason to reject this record across this period of time. However, while not plotted in the Gil et al. (2015) Fig. 3J (the record is mentioned on page 61 as extending back to ~18.1 ka), their own percent *Neogloboquadrina pachyderma* sinistral data show a major reduction in *N. pachyderma* (s) percent at the same core depth as the only significant CT warm event (Fig. 1B at ~16 ka on the original Keigwin et al. (2005) age model; depth is 376–378 cm in GGC14). Second, the Gil et al. (2015) diatom record shows warming in their diatom factor scores at the same time (their Fig. 3D). Third, all of the Obbink et al. (2010) data (outside the one data point) are below 5 °C. This temperature level is close to saturation for *N. pachyderma* (s) concentration (e.g., Hilbrecht, 1996) and so the Obbink et al. (2010) data are in good agreement with the faunal estimates. As such, the Gil et al. (2015) basis for disregarding this part of the Obbink et al. (2010) record seems unwarranted. This warm event off of the Laurentian Channel in the middle of Heinrich Event 1 is coincident with the break up of ice in the Laurentian Channel (Shaw et al., 2006). Gil et al. (2015) misinterpret this Laurentian Channel ice retreat as happening at ~14 ka (page 64). This ~14 ka is in uncalibrated ^{14}C years; when calibrated, the actual timing of this ice retreat presented in Shaw et al. (2006) is at 16–17 ka.

There is good agreement between the δ^{18}O of seawater (sw) record from Obbink et al. (2010) and that calculated from the Gil et al. (2015) percent *N. pachyderma* (s) record converted to temperature (Fig. 1B) (Hilbrecht, 1996; Govin et al., 2012) and the Keigwin et al. (2005) *N. pachyderma* (s) δ^{18}O record. We show this in Fig. 1C on the original Keigwin et al. (2005) age model; the different age model does not change this discussion as the data are from same core. The global change in ocean δ^{18}O is removed from the Keigwin et al. (2005) *N. pachyderma* (s) δ^{18}O record (Fig. 1A; Clark and Mix, 2002; note the minor affect this correction has on the original δ^{18}O record). $\delta^{18}O_{sw}$ is then calculated from Bemis et al. (1998) to account for temperature changes. The Obbink et al. (2010) Mg/Ca record is now converted to CT using the Kozdon et al. (2009) relationship that accounts for low temperatures and drifting sea ice (i.e., addressing the Gil et al. (2015) criticism on their page 62). This new CT calculation only slightly changes the CT record from that of Obbink et al. (2010), and is well within the CT uncertainty (Mashiotta et al., 1999; Kozdon et al., 2009). In the resulting records, the data overlap within the one-sigma $\delta^{18}O_{sw}$ uncertainty of 0.2–0.3 per mil (e.g., Obbink et al., 2010; Govin et al., 2012) (note that there are no data at ~15.6 ka in the Obbink et al. (2010) record). As such, these two different approaches to calculating $\delta^{18}O_{sw}$ support the original record of Obbink et al. (2010).

Gil et al. (2015) discuss the Bølling warm period (14.7–14.1 ka; Rasmussen et al., 2006) and Meltwater Pulse 1A, although it is not clear why they reference Carlson et al. (2007) in this context (page 64), as that study did not deal with these time periods. Carlson (2009) calculated the maximum amount of meltwater that could be contributed to Meltwater Pulse 1A through the Laurentian Channel (0.045–0.049 Sverdrups, 10^6 m^3 s^{-1}) using the Keigwin et al. (2005) *N. pachyderma* (s) δ^{18}O record and an older estimate of the timing of Meltwater Pulse 1A of 14.7–14.2 ka. Deschamps et al. (2012) now document Meltwater Pulse 1A as occurring between 14.6 and 14.3 ka (gray bar in Fig. 1). Obbink et al. (2010) also showed that this was a maximum estimate and that meltwater input more likely decreased during Meltwater Pulse 1A (Fig. 1C), which is further evident in the Gil et al. (2015) calculated $\delta^{18}O_{sw}$ record (Fig. 1C).

Gil et al. (2015) conclude that freshwater was not discharged through the Laurentian Channel during the Younger Dryas cold event and refer to Levac et al. (2015) in support of this conclusion. We note, however, that the title of Levac et al. (2015) is “Evidence for meltwater drainage via the St. Lawrence River Valley in marine

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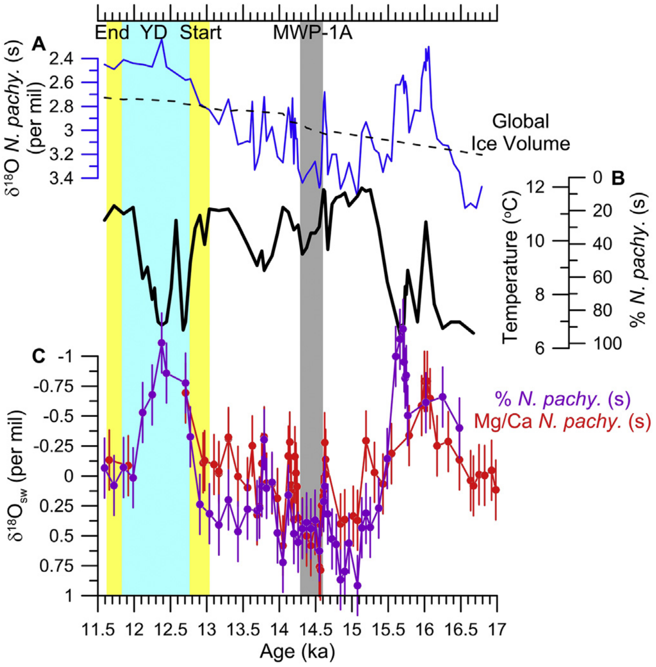

Fig. 1. Data from GGC14 on the age model of Keigwin et al. (2005). (A) *N. pachyderma* (s) δ^{18}O record (blue; Keigwin et al., 2005) with the global mean δ^{18}O record (dashed black; Clark and Mix, 2002). (B) Percent *N. pachyderma* (s) (Gil et al., 2015) converted to temperature following Govin et al. (2012); this relationship holds for sea surface to 200 m depth temperatures (Hilbrecht, 1996). (C) $\delta^{18}O_{sw}$ records relative to their mean using Mg/Ca–CT (red) and percent *N. pachyderma* (s) temperature (purple). Gray bar is Meltwater Pulse 1A (MPW-1A) (Deschamps et al., 2012). Light blue bar is the Younger Dryas with the yellow bars showing the range in its onset and termination according the Greenland ice-core records (Rasmussen et al., 2006). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

cores from the Laurentian Channel at the time of the Younger Dryas", and as implied by their title, Levac et al. do, in fact, conclude that freshwater was discharged through this outlet during the Younger Dryas, consistent with previous findings (e.g., Broecker et al., 1989; Licciardi et al., 1999; Clark et al., 2001; Brand and McCarthy, 2005; Rayburn et al., 2005, 2011; Carlson et al., 2007; Katz et al., 2011; Carlson and Clark, 2012; Cronin et al., 2012). The evidence that Gil et al. (2015) present to support their conclusion is the percent of low salinity diatoms. Although they note that species <2% are too rare for analysis (page 59), their diatom record used to support their claim of no freshwater discharge fluctuates around 2% (always less than 3%), and thus may not be a robust recorder of salinity changes.

Gil et al. (2015) state that "This interpretation [from the % diatom data], however, contradicts the results obtained from Mg/Ca on foraminifera (Carlson et al., 2007) that might have been strongly influenced by the development of sea-ice and the presence of icebergs (notably phase 1)." It is not clear what "Mg/Ca on foraminifera" results they are referring to. Carlson et al. (2007) used U/Ca and ^{87}Sr/^{86}Sr, with additional data from Mg/Ca calculated back to ocean Mg/Ca, in planktonic foraminifera from near the Laurentian Channel to show the addition of western plains Canadian freshwater (i.e., glacial Lake Agassiz) at the start of the Younger Dryas and modeled these changes to indicate an increase in freshwater discharge of ~0.06 Sverdrups above background runoff. Carlson et al. (2007) did not calculate $\delta^{18}O_{sw}$ from the Mg/Ca data. These records agree with δ^{13}C in *N. pachyderma* (s) in the same core (de Vernal et al., 1996) that suggest increased discharge of terrestrial-derived carbon. Carlson et al. (2007) furthermore placed more weight on the U/Ca and ^{87}Sr/^{86}Sr records than they did on the Mg/Ca and δ^{13}C records; these data should have been included in the discussion by Gil et al. (2015).

Although it is well established that changes in ocean temperature affect the δ^{18}O of *N. pachyderma* (s) (e.g., Wu and Hillaire-Marcel, 1994), Gil et al. (2015) ignore this fact despite documenting an ~70% increase in *N. pachyderma* (s) at the start of the Younger Dryas (Fig. 1B). This cooling of the near surface waters masks an ~1.4 per mil decrease in $\delta^{18}O_{sw}$ calculated using the Gil et al. (2015) data (Fig. 1C). Obbink et al. (2010) also show a similar decrease at the beginning of the Younger Dryas (Fig. 1C), while noting that further samples into the Younger Dryas were not analyzed because L. Keigwin informed the authors that he was working on these samples for Mg/Ca analysis (L. Keigwin, personal communication, 2007). This $\delta^{18}O_{sw}$ record calculated from the Gil et al. (2015) data agrees with the Gil et al. (2015) diatom factor 5 that shows "extremes" during the Younger Dryas (their Fig. 3E) that they suggest are indicative of "decreased surface salinity" (page 59). The Gil et al. (2015) $\delta^{18}O_{sw}$ record has an increase near to, but before, the ice-core-defined end of the Younger Dryas (Fig. 1). This finding is, however, in good agreement with the Carlson et al. (2007) U/Ca, ^{87}Sr/^{86}Sr and Mg/Ca data that show that freshwater runoff was routed away from the Laurentian Channel at ~12 ka and potentially to the Arctic Ocean, consistent with the accurate stratigraphic interpretations of Mackenzie River sedimentary sequences (Carlson and Clark, 2012) rather than those of Murton et al. (2010) that violate the geologic laws of hiatuses and superposition.

Gil et al. (2015) also do not take into account the evidence of large-volume freshwater discharge events directed to the Laurentian Channel from upstream sources before and during the Younger Dryas. Rayburn et al. (2011) demonstrated two large freshwater outflow events through the northern Champlain Valley that were routed to the Laurentian Channel between ~13.2 ka and ~12.9 ka. The age constraints were calculated using ^{14}C ages from terrestrially sourced macrofossils combined with varve records. The first event occurred with the final drainage of glacial Lake Vermont 13.2–13.1 ka (Rayburn et al., 2005, 2011). The second event occurred 13.1–12.9 ka, had a duration on the order of a century, and was sourced upstream of the Champlain Sea (Rayburn et al., 2011; Katz et al., 2011; Cronin et al., 2012). This event is evident in both benthic microfossil assemblage changes (Rayburn et al., 2011) and δ^{18}O decreases (Cronin et al., 2012), which are concurrent with the $\delta^{18}O_{sw}$ decrease observed in GGC14 (Fig. 1C) and the arrival of Lake Agassiz-runoff geochemical signatures off of the Laurentian Channel (Carlson et al., 2007). Brand and McCarthy (2005) show this δ^{18}O depletion in the Champlain Sea continued through much of the Younger Dryas, which is linked to eastward Lake Agassiz drainage by their ^{87}Sr/^{86}Sr records, similar to the record of Carlson et al. (2007).

We conclude that the Gil et al. (2015) statement that "The YD (Younger Dryas) cannot be regarded as an event triggered by a fresh water input through the Laurentian Channel..." is not supported by the existing data. Instead, these data are consistent with an increase in freshwater discharge through the Laurentian Channel at the start of and during the Younger Dryas. Moreover, the Obbink et al. (2010) records are in good agreement, within uncertainties, with the new Gil et al. (2015) data supporting this earlier study.

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