

Response to Comment on “Mixed-Layer Deepening During Heinrich Events: A Multi-Planktonic Foraminiferal $\delta^{18}\text{O}$ Approach”

Harunur Rashid^{1,2*} and Edward A. Boyle²

Hillaire-Marcel and de Vernal question our hypothesis of the “homogenization of the upper water masses” of the North Atlantic during Heinrich events 0, 1, and 4. Here, we respond to their specific points and synthesize data from the southern Labrador Sea cores, the basic tenet of Hillaire-Marcel and de Vernal’s arguments.

We used oxygen isotopes from multiple planktonic foraminifera to show that substantial perturbation of the upper water masses of the North Atlantic occurred during Heinrich events (HEs) (1). Hillaire-Marcel and de Vernal (2) argue that mixed-layer deepening is implausible based on evidence of dense sea-ice cover and strong stratification of surface waters.

The minimum sedimentation rate of 20 to 28 cm/thousand years (ky) between 35 and 80 cm in Chain 82 Station 50 Core 20 (CHN82-20) is estimated by six ¹⁴C accelerator mass spectrometry (AMS) dates (1) consistent with the earlier anomalously high sedimentation rates estimated

by Keigwin and Lehman (3). Such a high sedimentation rate within the deglacial sequence covering HE0 and HE1 allowed us to distinguish various high-frequency climatic variabilities and minimize the effect of bioturbation. However, we acknowledge that the effects of bioturbation cannot be entirely ruled out for the older interval because of the lower sedimentation rate. Newly revised deglacial chronology of the Northern Hemisphere suggests that the meltwater pulse 1A (mwp1A) was anchored at 14 ky before the present (B.P.) (4). Furthermore, none of the circum-North Atlantic paleo records show any discernible lighter $\delta^{18}\text{O}$ signature for the pre-

sumed perturbation caused by the mwp1A. On the other hand, HE1 occurred between 15.4 and 16.4 ky B.P., nearly 2 ka before the mwp1A. Therefore, we argue against putting together HE1 and mwp1A signatures at sites CHN82-20 (2, 3) and CH69-K09 (5) and do not think it is appropriate or desirable.

We acknowledge the possibility that foraminifers identified as *N. pachyderma* dextral (d) could be represented by *N. incompta* during the warm periods. However, the discrimination between the *N. incompta* and *N. pachyderma* (d) can be achieved only by means of genetic analysis and is not possible for fossil material. Data from Bauch *et al.* (6) suggest that the $\delta^{18}\text{O}$ signature of *N. incompta* in the Holocene period is ~0.5 per mil (‰) lighter than that of *N. pachyderma* sinistral (s). This difference theoretically should have no major impact on our $\delta^{18}\text{O}$ record for *N. pachyderma* (d) if warmer intervals are represented by *N. incompta*. Furthermore, our inferences on water-column mixing remain valid with the replacement of *N. pachyderma* (d) by *N. incompta*, which is ecologically and isotopically similar to *G. bulloides*. In addition, the *N. pachyderma* (d) $\delta^{18}\text{O}$ data in core CHN82-20 are not flat all the time [see figure 3B in (1)] and

¹Byrd Polar Research Center, The Ohio State University, 1090 Carmack Road, Columbus, OH 43210–1002, USA.

²Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, 77 Massachusetts Avenue, Cambridge, MA 02139, USA.

*To whom correspondence should be addressed. E-mail: rashid.29@osu.edu

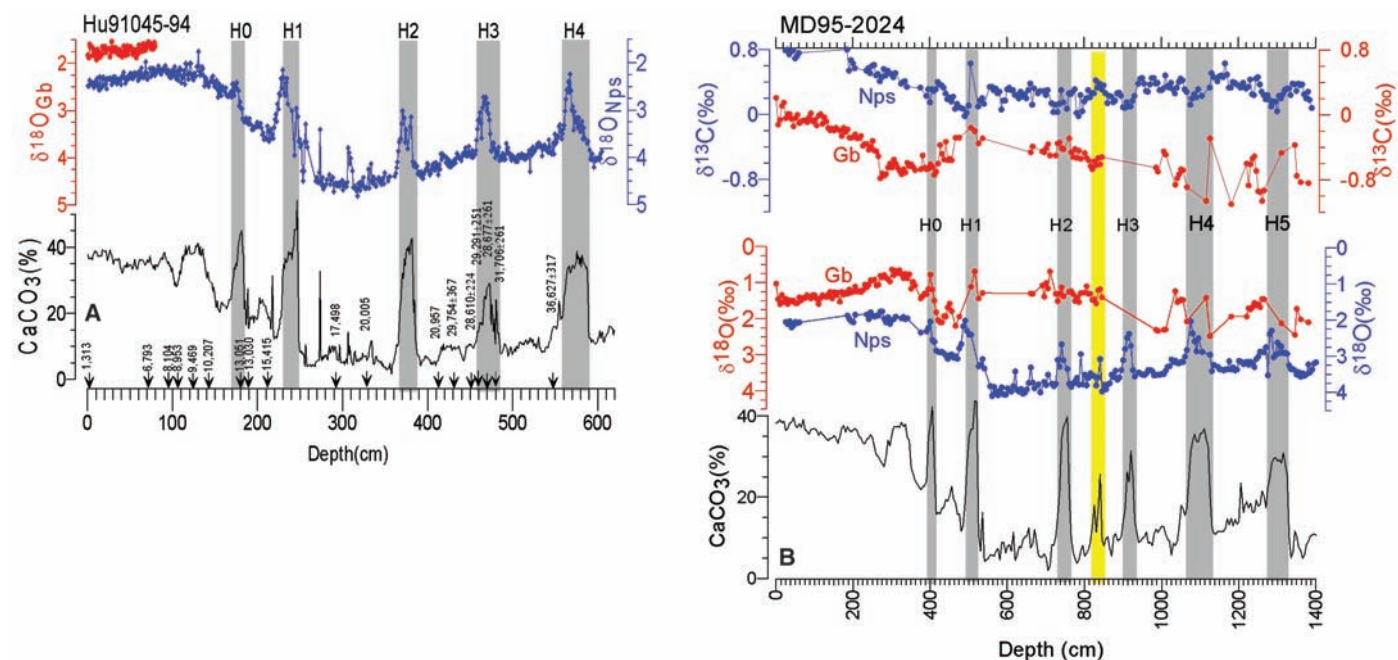


Fig. 1. (A) Downcore carbonate (% of CaCO_3) and $\delta^{18}\text{O}$ in *N. pachyderma* (s) (Nps) and *G. bulloides* (Gb) are plotted as a function of depth (cm) in core HU91045-94 retrieved at 3448 m water depth from the Orphan Knoll, southern Labrador Sea. Downward arrows denote calibrated ¹⁴C AMS dates

[data from (7, 8)], and $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data in Gb are not available beyond 82 cm. **(B)** Carbonate (% of CaCO_3), $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in Nps and Gb are plotted versus depth (cm) from core MD95-2024 [see (7) for details]. Note difference in x axes in core MD95-2024 compared with HU91045-94.

show 0.5‰ enrichment before HE1, not 3‰ as reported by Hillaire-Marcel and de Vernal (2).

Hillaire-Marcel and de Vernal (2) state that our hypothesis of the deepening of the mixed layer during HEs 0, 1, and 4 was originally put forward by Hillaire-Marcel and Bilodeau (7) based on the $\delta^{18}\text{O}$ in *N. pachyderma* (s) and *G. bulloides* from the southern Labrador Sea cores HU91045-94 and MD95-2024. Further investigation led these authors to abandon the mixed-layer deepening hypothesis because it appeared not plausible based on the evidence for dense sea-ice cover (inferred from the dinocyst assemblages) (8) and strong stratification of water masses (from the $\delta^{18}\text{O}$ in *N. pachyderma* (s) and *G. bulloides*). To shed light on these arguments and in response to the critiques, we plotted data from cores HU91045-94 and MD95-2024 (Fig. 1). The $\delta^{18}\text{O}$ data show systematic enrichment in core HU91045-94 compared with MD95-2024, both from the same site, regardless of the species (e.g., $\delta^{18}\text{O}$ of *N. pachyderma* (s) are $\sim 0.6\text{‰}$ heavier between HE3 and HE4 and $\sim 0.2\text{‰}$ heavier in early Holocene in core HU91045-94 compared with MD95-2024) (7, 8). The two carbonate records are also different at some times (e.g., in the early Holocene and just below HE2). We would like to point out that Hillaire-Marcel and Bilodeau (7) used terms such as “upper and halo-thermocline,” which meant the upper few hundred meters of the water column, and we assume that this translates to the mixed layer and upper thermocline. Instead of using the “homogenization of upper water masses,” the authors used stratification of the upper and halo-thermocline layers to explain $\delta^{18}\text{O}$ data in *N. pachyderma* (s) and *G. bulloides*, thus differentiating depth habitats for these species, as shown in Fig. 1B. Therefore, it is incorrect to suggest that these authors first hypothesized the “homogenization of the upper water masses” in their earlier publications (7, 8).

Furthermore, the resolution of $\delta^{18}\text{O}$ data in *G. bulloides* compared with *N. pachyderma* (s) in HE1, HE3, HE4, and HE5 is poor (core MD95-2024) (Fig. 1B), with lighter $\delta^{18}\text{O}$ values in *N. pachyderma* (s) but no response of *G. bulloides* in HE2, and lighter $\delta^{18}\text{O}$ in *N. pachyderma* (s) but no *G. bulloides* $\delta^{18}\text{O}$ data in HE3. However,

a single $\delta^{18}\text{O}$ point in *G. bulloides* does coincide with the $\delta^{18}\text{O}$ in *N. pachyderma* (s) during HE4. Therefore, the MD95-2024 $\delta^{18}\text{O}$ data might suggest a variable response of *N. pachyderma* (s) and *G. bulloides* to Heinrich events. If the modern depth habitat zonation of *N. pachyderma* (s) and *G. bulloides* is applied, it suggests that the southern Labrador Sea received warmer water during the last glacial cycle, like today, if all other variables remain constant. Hillaire-Marcel and de Vernal (2) refer to “dense sea-ice cover and strong stratification of surface waters during most of the Heinrich events, including HE0,” citing their earlier works (8, 9). However, there are no sea-ice data during Heinrich events, including HE0, in references (8, 9) to argue one way or the other. Furthermore, it is difficult to imagine the extension of sea ice at 43°N during glacial times, while concomitantly cooling the water temperature to at least 4°C during the spring when *G. bulloides* blooms. This suggestion contradicts estimates of sea surface temperature (10, 11) near core CHN82-20.

We also question the interpretation of the $^{231}\text{Pa}/^{230}\text{Th}$ data of Gherardi *et al.* (12) from the northeast Atlantic to explain the salinity gradient in the northwest Atlantic because the Gherardi *et al.* data do not extend beyond 20 ky B.P. (i.e., beyond HE1). It is not clear whether the $^{231}\text{Pa}/^{230}\text{Th}$ can alone be used as a proxy for the variability of the meridional overturning circulation (MOC) because other factors such as the high rate of opal production (13) could have masked the particle fluxes and, by extension, that could affect the $^{231}\text{Pa}/^{230}\text{Th}$ signal (14). During HE1, the $^{231}\text{Pa}/^{230}\text{Th}$ data in (12) lead the Bermuda Rise $^{231}\text{Pa}/^{230}\text{Th}$ data by ~ 1.8 ky. Furthermore, from the Irish margin (farther north of Gherardi *et al.*'s core), Hall *et al.* (15) demonstrated that the intermediate waters were ventilating during HE1, in stark contrast to the “sluggish MOC” suggestion of Gherardi *et al.* Even if the MOC during HE1 was sluggish, this would in theory not contradict a stronger turbulent mixing of the upper ocean that would mainly be wind-driven. Therefore, we question the validity of speculation of “large salinity gradients” (2) based on the $^{231}\text{Pa}/^{230}\text{Th}$ data, which itself is contradictory.

We also do not understand the relevance of the Chukchi Sea of the Arctic Ocean data (16) to explain the salinity gradient in the North Atlantic.

Finally, we would like to pose a question regarding an order of magnitude increase of Na^+ and Ca^+ in GISP2 ice cores and their connection to the strength of wind stress to modulate the storminess and homogenization of water masses of the Northern Atlantic Ocean. Mayewski *et al.* (17) show a similar increase in Na^+ concentrations regardless of Heinrich or Dansgaard-Oeschger (D/O) ice-rafting events, which suggest a similar strength of storminess. We suggest that the “lack of enough meltwater” can be attributed to the weaker homogenization of the upper water masses during D/O ice-rafting events because no changes in storminess is observed. But there was not a shortage of meltwater during the Heinrich events in the open Atlantic Ocean. So the question is, “What does the wind-driven turbulent mixing do when the catabatic winds are stronger during the glacial time?”

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