

Paleoceanography and Paleoclimatology



RESEARCH ARTICLE

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Key Points:

- In Franklin Basin (Gulf of Maine), a basal radiocarbon age of 17.5 ± 0.2 cal ky BP provides a minimum-limiting age for deglaciation
- Oxygen isotopes combined with magnesium to calcium ratios in benthic foraminifera yield a new deglacial seawater oxygen isotope record
- Seawater oxygen isotope depletion during Heinrich Stadial 1 suggests increased meltwater input, likely driven by enhanced summer melt

Supporting Information:

Supporting Information may be found in the online version of this article.

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Intensified Meltwater Influx During Heinrich Stadial 1 Recorded in the Gulf of Maine, USA

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Abstract Abrupt, millennial-scale climate events characterize the most recent deglaciation. However, their nature remains poorly understood. Previous work has shown enhanced glacial melt signals from the European/Scandinavian Ice Sheet and a rapid recession of the Laurentide Ice Sheet (LIS) in Eastern Maine during Heinrich Stadial 1 (HS-1). These observations challenge the idea that stadials were year-round cold excursions in the Northern Hemisphere. We aim to determine if European HS-1 melt signals are coeval with LIS meltwater in the Gulf of Maine (GOM) marine sediment record, which was ice-proximal during much of the last deglaciation. Here we present geochemical records ($\delta^{18}\text{O}$, Mg/Ca, calculated $\delta^{18}\text{O}_{\text{seawater}}$) developed using benthic foraminifera from a new GOM sediment core, EN-669-GGC-40, spanning ~ 17.5 – 11.5 cal ky BP. We find that the documented portion of HS-1 in the GOM can be divided into two phases: the earlier (HS-1b, ~ 17.5 – 15.9 cal ky BP) being characterized by large-amplitude variations of $\delta^{18}\text{O}_{\text{sw}}$, which we interpret as a series of intervals of enhanced meltwater runoff from the LIS. The later period (HS-1a, ~ 15.9 – 14.7 cal ky BP) is characterized by a rapid and sustained isotopic depletion of $\delta^{18}\text{O}_{\text{sw}}$ from 15.7 to 15.3 cal ky BP which we interpret as a persistent melting episode, consistent with rapid collapse of glacial ice in Eastern Maine. Our results support the hypothesis that HS-1 was characterized by warm(ing) summers and hypercold winters in the North Atlantic region (Denton et al., 2005, <https://doi.org/10.1016/j.quascirev.2004.12.002>, Denton et al., 2022, <https://doi.org/10.1016/j.quascirev.2022.107750>).

1. Introduction

What caused the millennial-scale climate changes of the last glaciation and last glacial termination remains an important question of paleoclimatology. The last termination, for example, was recorded by mean-annual air temperatures inferred from Greenland ice-core isotopes, featured the Heinrich Stadial-1 (HS-1; 17.8–14.7 cal ky BP) and Younger Dryas (YD; 12.9–11.7 cal ky BP) cold intervals, divided by the relatively warm Bølling-Allerød (B-A; 14.7–12.9) interval (chronology from NGRIP Community Members, 2004; WAIS Divide Community Members, 2014). However, paleoclimatic records of millennial-scale abrupt climate change derived from independent proxies have yielded conflicting signals (Denton et al., 2005). Whereas Greenland ice cores (e.g., GISP2 and NGRIP) record cooling during HS-1, some studies of North Atlantic marine sediments indicate an increase in ice melting activity during HS-1, suggesting warming (e.g., Boswell et al., 2019; Toucanne et al., 2015; Wickert et al., 2023). Likewise, on land, many Northern Hemisphere alpine glacier snowline reconstructions and exposure age chronologies indicate increased summertime ablation during the same interval (Bromley et al., 2026; Foreman et al., 2022, 2025; Putnam et al., 2023), as well as during the more recent YD episode (Bromley et al., 2014, 2023; Wittmeier et al., 2020).

To reconcile these observations, Denton et al. (2005, 2022) introduced the Seasonality/Heinrich Summers Hypothesis: rather than stadials being characterized by year-round cooling, they were instead episodes of intensified seasonality in the North Atlantic, with relatively warm (or warming) summers juxtaposed against extremely cold winters. The hypothesis suggests that warm summers would stimulate enhanced meltwater flux into the North Atlantic, and that the resulting lid of fresh and cold water would promote extensive sea ice cover in winter, leading to intensified, “continental” winters in the North Atlantic region (Denton et al., 2022). Mean-annual air temperatures registered in Greenland ice cores may therefore have been skewed toward the strong winter signal that would have masked the more muted summer signal (Buizert et al., 2014; Denton et al., 2005). On the other hand,

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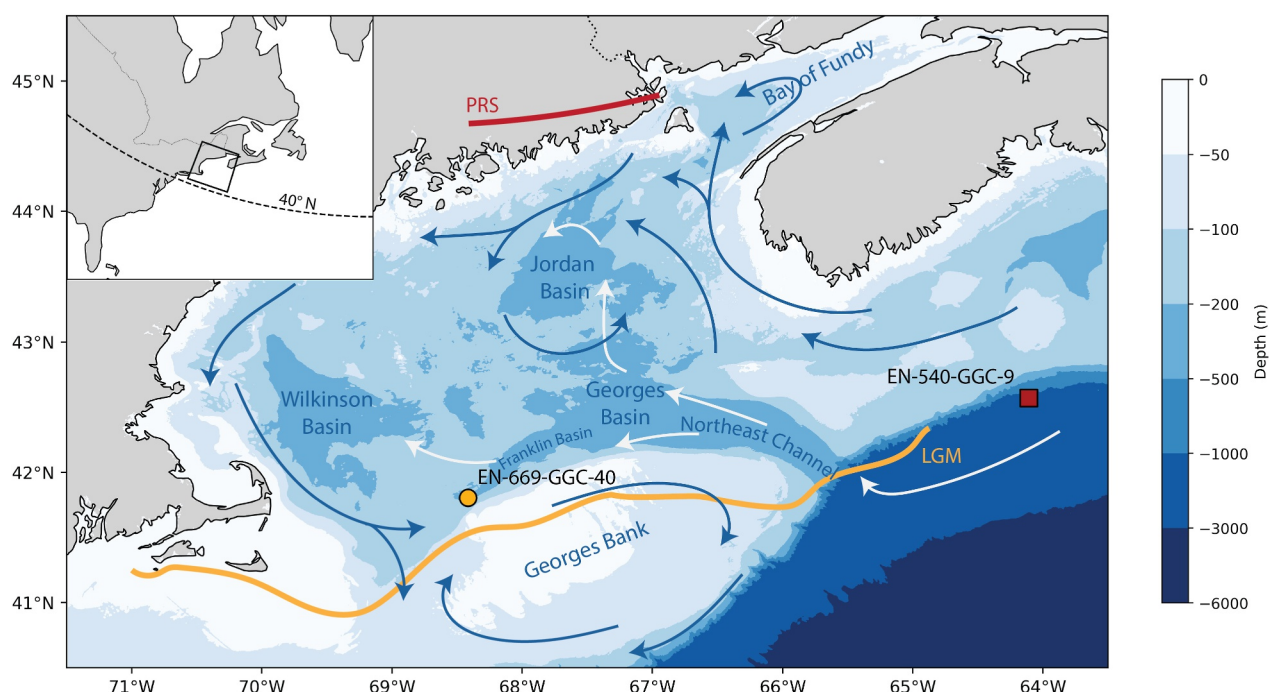


Figure 1. Bathymetric map of the Gulf of Maine region showing location of core site EN-669 GGC-40 (gold circle), EN-540-GGC-9 (red square), and relevant oceanographic features. White arrows represent summertime surface (upper 40 m) currents and blue arrows represent bottom currents, after Brooks, 1985; Pettigrew et al., 2005; Townsend et al., 2015. Franklin Basin is an arm of Georges Basin, connected to the greater Atlantic via the Northeast Channel. Gold line shows estimated LGM ice margin from Oldale and Uchupi (1970) and Schnitker et al. (2001), red line shows position of Pineo Ridge System (PRS; 15.3 ± 0.5 cal ky BP) from Hall et al. (2025). Bathymetric data are from GEBCO 2024 Global Grid (www.gebco.net).

alpine glacier snowlines reflect summer or ablation season temperatures (Oerlemans, 1994), which are associated with increased meltwater fluxes to the ocean. Therefore, it is possible that these records may not be in conflict, but rather are indicating episodes of intensified seasonality (Bromley et al., 2023; Putnam et al., 2023).

The Laurentide Ice Sheet (LIS) was the largest of the Pleistocene ice sheets, with estimated ice volume responsible for ~ 70 m of global eustatic sea level change since the Last Glacial Maximum (LGM) (Dalton et al., 2020; Tarasov et al., 2012). Various southeastern ice streams of the LIS directly entered the North Atlantic Ocean (Dalton et al., 2020; Margold et al., 2018; Mayewski et al., 1981; Schnitker et al., 2001), meaning that large quantities of LIS-derived meltwater and icebergs could directly impact North Atlantic circulation and surface conditions (e.g., Broecker, 1994; Denton et al., 2005; You et al., 2023). Understanding LIS behavior in response to abrupt climate change events at a marine margin, and the role of the LIS in affecting local North Atlantic climate, are important for determining the nature and drivers of these events and intra-stadial variability.

The Gulf of Maine (GOM), on the continental shelf of the northeastern United States, has an average depth of 150 m (Oldale & Uchupi, 1970; Uchupi, 1968). It was glaciated by the LIS during the LGM (Hughes et al., 1985; Mayewski et al., 1981; Schnitker et al., 2001; Todd et al., 2016), and because of its southern location was among the first marine margins to experience glacial retreat during the late LGM (Hughes et al., 1985; Schnitker et al., 2001). Basal radiocarbon ages from the GOM region on benthic foraminifera indicate that deglaciation was underway as early as ~ 20 cal ky BP at the easternmost margins, and that the ice edge had receded through most of the present-day GOM by ~ 16 cal ky BP at the latest (Figure 1; Schnitker et al., 2001; Todd & Shaw, 2012). The ice edge reached the present-day GOM coastline at or before 15.3 ky BP, indicated by the chronology of the Pineo Ridge and other coastal moraine and delta systems (Hall et al., 2017; Kaplan, 1999, 2007; Schnitker et al., 2001; Todd & Shaw, 2012). This means that the GOM was LIS-proximal for the entirety of HS-1, and that meltwater from the LIS edge would very likely influence the composition of GOM waters.

Benthic foraminifera are nearly ubiquitous in GOM sediments dating back to deglaciation, whereas planktic foraminifera are notably absent before ~ 10 cal ky BP (Keigwin & Pilska, 2015; Schnitker et al., 2001). The two species of benthic foraminifera that dominate the deglacial GOM record are *Nonionellina labradorica*

(Dawson, 1860) and *Islandiella helenae* (Feyling-Hanssen & Buzas, 1976). Both are common Arctic/subarctic species whose abundances typically vary inversely (Kristjánsdóttir et al., 2007; Schnitker et al., 2001). *Islandiella helenae* appear to prefer more ice-marginal environments with lower salinities and cooler temperatures (Mudie et al., 1984; Polyak et al., 2002; Skirbekk et al., 2016), whereas *N. labradorica* tolerate more slope-like temperatures and salinities (Jennings et al., 2004; Mudie et al., 1984; Skirbekk et al., 2016). The isotopic composition of foraminiferal shells (or “tests”) reflects the isotopic composition and temperature of the seawater in which they grow, which can be strongly influenced by glacial meltwater (e.g., Ravelo & Hillaire-Marcel, 2007).

Using paired Mg/Ca and $\delta^{18}\text{O}$ of benthic foraminiferal tests, we reconstructed the oxygen isotope composition of seawater ($\delta^{18}\text{O}_{\text{sw}}$) in the Gulf of Maine for the duration of the most recent deglaciation to assess the timing of meltwater production relative to periods of warming or cooling in regional ice core records and terrestrial glacial records, thus generating new observations to bear on the Seasonality Hypothesis and the multi-phased nature of HS-1, and ultimately the response of ice sheets to abrupt climate change.

2. Methods

2.1. Core Collection and Shipboard Measurements

Marine sediment cores were collected on cruise EN669 aboard the R/V Endeavor in August 2021. A Giant Gravity Core, EN669-GGC-40 (41°48′12.8″N, 68°25′00.2″W, 214 m water depth), was collected from Franklin Basin in the eastern GOM (Figure 1). A Knudsen 3260 Chirp Echosounder was used with two transducers: (a) a high-frequency, primary, starboard-forward, 12 KHz EDO 323C conical beam transducer to track bathymetry, and (b) a low-frequency, 3.5 KHz array consisting of 12 Edgetech KT216 broadband transducers for sub-bottom profiling. Transducers were located at 5 m water depth (ship draft) and sound speed was assumed to be 1,500 m/s. Downcore multisensor-track (MST) data were also collected onboard in a temperature-controlled van using a GEOTEK MSCL 7.9 equipped with a Bartington Loop Sensor (MS2C) for measuring magnetic susceptibility. Marine sediment core EN-540-GGC-9 (42.57°N, 64.11°W, 1,111 m water depth), from the Scotian Slope, was taken on cruise EN540 (Chief Scientist Lloyd Keigwin) in 2014 aboard the R/V Endeavor (Figure 1).

2.2. Core Sampling

EN669-GGC-40 and EN-540-GGC-9 were split and sampled at the Woods Hole Oceanographic Institution (WHOI) Seafloor Samples Laboratory. Computed tomography (CT) scanning was performed on archival core halves using a GEOTEK RXCT scanner. Scan parameters included an X-ray voltage of 92 kV, an X-ray current of 240 μA , and a 0.125 mm Cu filter, which produced a scan with a resolution of 76 pixels per cm, or a 131.7 micron (μm) resolution. For EN-669-GGC-40, samples of ~ 10 cc volume were taken at ~ 200 -year intervals (every 2–3 cm) between 10 cm depth and 586 cm depth for a total of 248 samples of marine sediment.

2.3. Sediment Processing and Micropaleontology

The sediment samples were transported to the University of Maine, where they were wet-sieved at 63 μm using deionized water to separate coarse and fine fractions. The fine fraction was dried at room temperature and archived, and the coarse fraction was dried in an oven at low temperature ($<40^\circ\text{C}$) and subsequently dry-sieved through a 250 and 150 μm sieve stack. The coarse fractions were examined under a Nikon SMZ1270 Stereo Microscope to characterize the foraminifera present and select samples for analysis. Benthic species *Nonionellina labradorica* (Dawson, 1860) and *Islandiella helenae/norcrossi* (Feyling-Hanssen & Buzas, 1976, identified following morphology guidelines of Cage et al., 2021) were sorted into micropaleontology slides using a clean, fine-tip paintbrush. In order to check for signs of diagenesis (e.g., pitting, cracks, or crusts on primary calcite), Scanning Electron Microscope (SEM) photographs were taken of carbon-coated samples using a Tescan Vega II XMU tungsten filament Scanning Electron Microscope. SEM samples were not used for geochemical analyses. *Islandiella helenae* and *I. norcrossi* are similar but morphologically distinct species; however, they are combined in published faunal assemblages (e.g., Rytter et al., 2002) and trace element calibrations (e.g., Kristjánsdóttir et al., 2007; Skirbekk et al., 2016). We elected to combine *I. helenae* and *I. norcrossi* samples to maintain fidelity to the calibration curves and to achieve adequate sample weights for geochemical analysis.

Benthic species *Nonionellina labradorica* and *Islandiella helenae/norcrossi* appear throughout the deglacial record with a few monospecific gaps. The widespread occurrence of these species, combined with the existence of

Table 1
Chronological Information for Core EN-669-GGC-40

Sample name	Accession number	Depth (cm)	Material	^{14}C age	ΔR applied	MARINE20 calibrated age ($\mu \pm 1\sigma$)	Excluded from age model
GGC40-1	OS-169004	1–2	Benthic foraminifera (<i>Bolivina</i> sp.)	565 ± 65	-50 ± 62	Modern	
GGC40-2	OS-170707	32–33	Mixed benthic foraminifera	$2,230 \pm 35$	30 ± 61	$1,618 \pm 106$	
GGC40-3	OS-175723	110–111	Benthic foraminifera (<i>Bolivina</i> sp.)	$6,160 \pm 35$	221 ± 57	$6,154 \pm 101$	
GGC40-4	OS-175724	110–111	Bivalve shell (<i>A. subaequilatera</i>)	$6,990 \pm 30$	262 ± 57	$7,031 \pm 105$	X
GGC40-5	OS-170708	177–178	Mixed benthic foraminifera	$8,680 \pm 100$	344 ± 55	$8,704 \pm 165$	
GGC40-6	OS-175725	279–280	Benthic foraminifera (<i>Globobulimina</i> sp.)	$9,290 \pm 140$	373 ± 54	$9,430 \pm 205$	
GGC40-7	OS-179369	315–316	Mixed planktic foraminifera	$9,640 \pm 180$	390 ± 54	$9,860 \pm 249$	X
GGC40-8	OS-179370	315–316	Mixed benthic foraminifera	$10,050 \pm 80$	410 ± 54	$10,378 \pm 164$	
GGC40-9	OS-175726	371–372	Mixed Benthic foraminifera	$11,750 \pm 90$	493 ± 52	$12,605 \pm 119$	
GGC40-10	OS-175727	481–482	Barnacle shell	$14,600 \pm 85$	496 ± 52	$16,220 \pm 169$	
GGC40-11	OS-170709	582–583	Mixed benthic foraminifera	$15,300 \pm 270$	496 ± 52	$17,115 \pm 365$	
GGC40-12	OS-170710	583–593	Barnacle shell	$15,600 \pm 65$	496 ± 52	$17,472 \pm 166$	

numerous well-established Mg/Ca and $\delta^{18}\text{O}_{\text{sw}}$ records derived from them (e.g., Barrientos et al., 2018; Farmer et al., 2011; Kristjansdóttir et al., 2007; Skirbekk et al., 2016), led us to select them as the basis for constructing our record. In modern oceans, both species are prolific in Arctic Seas and North Atlantic fjords (Rytter et al., 2002; Skirbekk et al., 2016). We found both species in modern coretop samples, although the predominant *Islandiella* sp. in modern samples is *I. norcrossi* rather than *I. helenae*, which dominates deglacial sediments. Both *Islandiella* sp. are also much smaller ($<200 \mu\text{m}$ diameter) in the modern and Holocene portion of the core, whereas deglacial *Islandiella* sp. are typically $>250 \mu\text{m}$ in diameter. Test size and morphology of *N. labradorica* remain consistent throughout the core.

2.4. ^{14}C Dating and Age Model Development

For EN-669-GGC-40, intervals for sampling were selected based on initial dates derived from the top and bottom of the core, with the aim of obtaining at least one radiocarbon date per 2,000 calendar years. Carbonate material derived from bivalve shells, barnacle shells, or benthic foraminifera was cleaned via sonication in ultrapure water, dried at low temperature ($<40^\circ\text{C}$), and sent to WHOI-NOSAMS for ^{14}C analyses. The uppermost sample (1–2 cm core depth) was processed using a gas interface system (GIS) connected to an Ionplus Mini Carbon Dating System (MICADAS). All other samples were acidified in 100% phosphoric acid and analyzed on a Continuous-Flow AMS System (Roberts et al., 2007). Age-depth models were generated using a total of 12 radiocarbon dates, calibrated to MARINE20 via OxCal 4.2 (Heaton et al., 2020; Bronk Ramsey, 2009; Table 1). At depths where more than one radiocarbon date was available from multiple materials, dates from benthic foraminifera were preferentially selected for the age model in order to match the material of isotope analyses. A bivalve shell fragment at 110–111 cm yielded a radiocarbon age ~ 830 years older than benthic foraminifera from the same sample. This large age offset and the shell's broken edges suggest that this bivalve specimen may have been transported downslope and was therefore omitted from the age model. One age from planktic foraminifera was used to assess age offsets between contemporaneous benthic foraminifera (Figure S1 in Supporting Information S1).

A local correction to the global average Marine Reservoir Age (ΔR) was calculated for each sample. Due to changes in circulation, sea-ice coverage, and local relative sea level throughout the deglaciation, use of a single, static ΔR value is unsatisfactory (Heaton et al., 2023). For samples older than 12.8 ^{14}C ky BP, a ΔR value of 496 ± 52 years was applied, based on paired marine-terrestrial material from the glaciomarine Presumpscot Formation described in Thompson et al. (2011) and calibrated for use in MARINE20 using the *deltar* application (Reimer & Reimer, 2017). For samples with a ^{14}C age younger than 12.8 ky BP, a ΔR was calculated by linearly interpolating between a modern GOM ΔR value, -83 ± 70 years (Lower-Spies et al., 2020), and the ΔR value of Thompson et al. (2011) at 12.8 ^{14}C ky BP. Specific reservoir corrections for the samples in this age range can be found in the Supporting Information (Table 1).

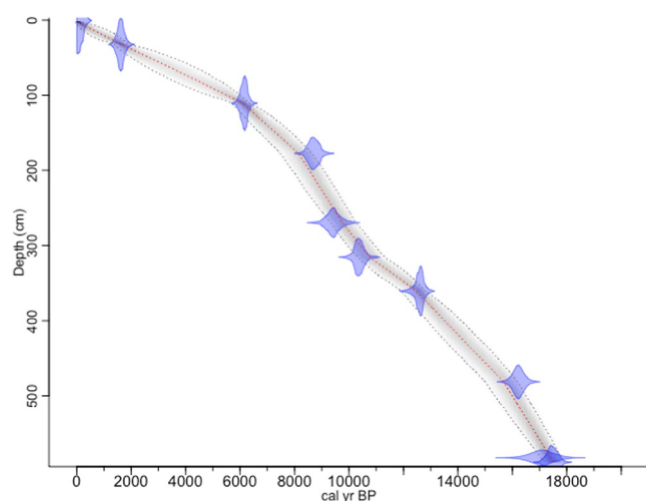


Figure 2. Age-depth model for core EN-669 GGC-40 using *rbacon* application (Blaauw & Christen, 2011). Blue areas represent ^{14}C dated material and associated uncertainties. Red dashed line indicates the modeled mean age, and gray dashed lines to the left and right of the red line indicate the minimum and maximum modeled age within the 95% confidence interval. Dates are provided in Table 1 and *rbacon* run parameters can be found in Table S1 in Supporting Information S1.

In the deep North Atlantic, reconstructed reservoir ages (R) for early deglaciation are large (in excess of 2,000 years, e.g. Stern & Lisiecki, 2013; Skinner et al., 2019). However, these reconstructions are based almost entirely on cores which are from >1,500 m water depth. Thus, they are much more likely to reflect larger reservoir ages. The GOM, which has a maximum depth of ~350 m and interacts primarily with shelf and shallow slope water masses, is unlikely to have experienced such drastic reservoir ages as those deep Atlantic sites, as it is more frequently mixed and equilibrated with the atmosphere. Nevertheless, we considered alternative reservoir age corrections (including static ΔR scenarios) (see Figure S1 in Supporting Information S1). Absolute maximum differences for these scenarios ranged $\pm \sim 0.7$ cal ky different from our final age model, but mean differences were ± 0.3 cal ky for the deglacial portion of the core. Ultimately, lacking a more contemporaneous and local measure of reservoir age, we favor the interpolated ΔR described above as the best estimation because it is based on proximal reservoir constraints, and thus is more likely to reflect the shallower, more coastal location of our EN-669-GGC-40 as opposed to deeper North Atlantic values.

To construct an age model for core EN-669-GGC-40, we employed the *rbacon* application to model, using Bayesian statistics, the downcore age curve, with attendant 95% confidence intervals, and uncertainty envelopes (Blaauw & Christen, 2011). The age-depth model is given in Figure 2 and Table 1.

2.5. Stable Isotope Analyses

Nonionellina labradorica and *I. helenae/norcrossi* were chosen from the coarse fraction of each sample with sufficient material for sampling (>20 μg). A total of 108 samples were chosen from EN669-GGC-40. Each sample contained 1–5 benthic foraminifers. The samples were bathed and sonicated in OptimaTM-grade methanol and allowed to dry completely before being shipped to the Iowa State Stable Isotope Laboratory, where they were acidified and analyzed for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ using a Thermo Scientific Delta V Plus mass spectrometer in continuous flow mode connected to a Gas Bench with a CombiPAL autosampler. Reference standards (NBS-18, IAEA 603) were used for isotopic corrections (Hut, 1987). The analytical uncertainties for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ varied depending on time of analysis, ranging from 0.04 to 0.10 and 0.05–0.11 ‰, respectively. These values represent the root-mean-square error, which combines the deviation of the standards used throughout each run with the correction error. Raw benthic foraminiferal calcite $\delta^{18}\text{O}$ values ($\delta^{18}\text{O}_{\text{bf}}$) were corrected for species-specific vital effects using the following disequilibrium corrections: 0.90 for *N. labradorica* (Wharton et al., 2024), and 0.289 for *I. helenae/norcrossi* (Kristjánssdóttir et al., 2007). Unless otherwise specified, $\delta^{18}\text{O}_{\text{bf}}$ values in text refer to these vital effect-corrected values. Samples with $\delta^{18}\text{O}_{\text{bf}}$ values in excess of 7‰ VPDB were excluded in order to avoid diagenetically altered signals. Raw values for each sample are available (Figure 3; see Statement of Data Availability). Twelve sample depths were analyzed for stable isotopes of both species to compare co-occurring stable isotope values. Before vital effect correction, the average offset between *N. labradorica* and *I. helenae/norcrossi* in these samples is -1.16‰ and -0.45‰ for $\delta^{13}\text{C}_{\text{bf}}$ and $\delta^{18}\text{O}_{\text{bf}}$, respectively.

2.6. Trace Element Analyses

Nonionellina labradorica and *I. norcrossi/helenae* were chosen from the coarse fraction material of each sample depth ($n = 1\text{--}5$ specimens per sample). A total of 119 samples were selected for analysis from EN669-GGC-40. The selected foraminifera were cleaned and dissolved in acid-leached vials following the methods of Boyle and Keigwin (1985) and Rosenthal et al. (1997), forgoing the crushing step due to test fragility as it was observed that sonication was sufficient to break the tests and fully expose the chamber interiors. Cleaned samples were dissolved in 0.065 N OptimaTM-grade HNO_3 in acid-leached 0.5 mL Eppendorf snap-cap vials and then diluted to a total volume of 400 μL using 0.5 N OptimaTM-grade HNO_3 . A suite of major and trace elements was measured, including magnesium to calcium ratios (Mg/Ca), on a Thermo Scientific Element XR ICP-MS at the University of Maine Climate Change Institute. ICP-MS runs were conducted using a quartz cyclonic spray chamber with

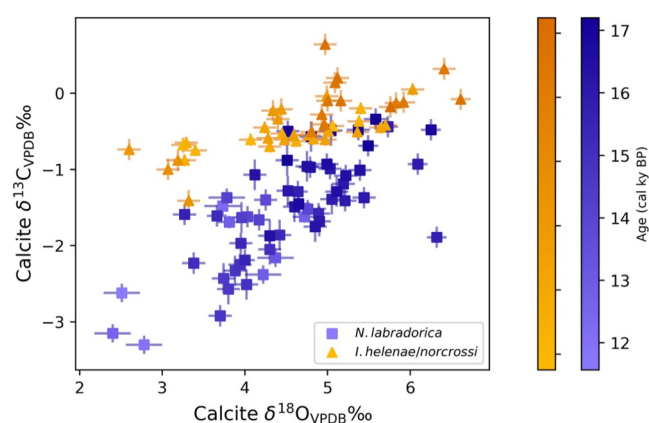


Figure 3. Cross plot of raw (uncorrected for vital effect) $\delta^{18}\text{O}_{\text{br}}$ versus $\delta^{13}\text{C}_{\text{br}}$. Darker colors reflect older ages and lighter colors reflect younger ages, in order to illustrate the deglacial baseline shift in stable isotopes. Error bars reflect 2-sigma analytical uncertainties of $2 \times 0.07\text{--}0.11\text{‰}$ VPDB for $\delta^{18}\text{O}_{\text{br}}$ and $0.04\text{--}0.06\text{‰}$ VPDB for $\delta^{13}\text{C}_{\text{br}}$.

additional gas inlet, PFA self-aspirating nebulizer, and platinum cones. Samples were run with a probe uptake rate of $50\text{ }\mu\text{L}/\text{min}$, sample gas flow of $\sim 0.8\text{ L}/\text{min}$, additional gas of $\sim 0.2\text{ L}/\text{min}$ and RF power of 1,250.

Four linear calibration standards encompassing the typical ranges of foraminiferal trace element to calcium ratios were run at the beginning and end of each analytical session (Marchitto, 2006). A set of 5 matrix standards with identical trace element-to-calcium ratios but varying calcium concentrations ($0.5\text{--}4\text{ mM}$) was also run to enable quantification and correction of any influences of solution calcium concentration on measured values of element ratios, also known as matrix effects (Rosenthal et al., 1997; Standish et al., 2019). Long-term analytical precision for Mg/Ca , estimated as % RSD of repeated analysis of a suite of standards spread across multiple days of analysis was 2.6%. The Instrument Detection Limit (IDL), defined as the minimum amount of analyte that can be determined from zero with 99% confidence by an instrument (Thomas, 2013), is $0.6\text{ mmol}/\text{mol}$ for Mg/Ca (Table S3 in Supporting Information S1). Twenty-one of the 119 samples (18%) were marked as outliers after the process of analysis due to insufficient sample size ($<0.3\text{ mM}$ Ca in solution). Though in pelagic depositional environments, high silicate contamination is defined as $>0.1\text{ mol}/\text{mol}$ Fe/Mg

(Barker et al., 2003; Bryan & Marchitto, 2008), it has been suggested that at marginal sites with high siliciclastic sedimentation regimes, Fe may be a primary calcite signal rather than an indicator of diagenesis (Cook & Hendy, 2024). Relationships between “contaminant” signals (Al/Ca , Fe/Ca , Mn/Ca) and Mg/Ca in our samples are weak but statistically significant ($R^2 = 0.03\text{--}0.16$, $p \leq 0.02$; see Figures S2 and S3 in Supporting Information S1). This indicates that while contaminant ratios covary modestly with Mg/Ca , they account for only a small fraction of the total Mg/Ca variance. We therefore infer that test Mg/Ca primarily reflects temperature, and elect to include samples with Fe/Mg ratios greater than $0.1\text{ mol}/\text{mol}$, as there is no robust or consistent evidence to indicate contamination from an Fe-bearing phase. Samples in the upper 5% of values for Al/Ca , Fe/Ca , and Mn/Ca were flagged on plots, but still included in our interpretation for the reasons outlined above.

Geochemical results were translated into paleotemperatures using established Mg/Ca calibration curves: Barrientos et al. (2018) for *N. labradorica*, and Kristjánsdóttir et al. (2007) for *I. norcrossi/helenae*. $\delta^{18}\text{O}$ of seawater was calculated using the low-light equation of Bemis et al. (1998). Prediction intervals were calculated for calibration equations following the method of Allen et al. (2020) using calculated standard errors for data sets: $\pm 0.76^\circ\text{C}$ for Barrientos et al. (2018), $\pm 0.63^\circ\text{C}$ for Kristjánsdóttir et al. (2007), and $\pm 0.7^\circ\text{C}$ ($\pm 0.15\text{‰}$ VSMOW) for Bemis et al. (1998). Calculated 95% prediction intervals are $\pm 1.33^\circ\text{C}$ for Barrientos et al. (2018), $\pm 1.11^\circ\text{C}$ for Kristjánsdóttir et al. (2007), and $\pm 0.29\text{‰}$ VSMOW for Bemis et al. (1998).

3. Results and Discussion

3.1. Chronology and Implication for the Timing of Deglaciation

Results of ^{14}C dating are given in Table 1. Ten ^{14}C dates on marine micro and macrofaunal shells underpin the age-depth model for core EN-669-GGC-40 in Franklin Basin of the GOM. A ^{14}C date of $17.5 \pm 0.2\text{ cal ky BP}$ of a barnacle shell recovered from the lowermost core-cutter sediment of EN-669-GGC-40 affords a minimum-limiting age for the onset of deglaciation of Franklin Basin. This barnacle age, and ^{14}C ages of two samples (mixed benthic foraminifera and another barnacle, respectively; Table 1) at depths above the lowermost sample, all afford ages that are within the HS-1 interval during the last glacial termination. The lowermost age indicates that ice recession was underway in the GOM, to allow for sedimentation in Franklin Basin at least by $17.5 \pm 0.2\text{ cal ky BP}$, in early HS-1. The presence of sediment packages extending below the base of our core in the seismic profile (Figure S4 in Supporting Information S1) suggests the retreat had begun prior to 17.5 cal ky BP . We consider this to be direct, physical evidence that deglaciation and intensified melt of the ice-sheet margin was underway during HS-1. This age is in agreement with compiled terrestrial ^{14}C and ^{10}Be chronologies, which show that deglaciation of the Maine coast, located well inboard of Franklin Basin, was underway by at least $15.9 \pm 0.2\text{ cal ky BP}$, and possibly as early as $17.2 \pm 0.2\text{ cal ky BP}$ (see figure 11 of Hall et al., 2025).

3.2. Sedimentological Results

The sediment of EN-669-GGC-40 is comprised of two main sedimentological units, partitioned at 431 cm core depth, or 14.35 ± 0.7 cal ky BP, that align with two distinct seismic stratigraphic units (Figure S4 in Supporting Information S1). The lowermost unit is a red-brown silty clay with abundant sand and gravel, and the uppermost unit is a green-gray marine clay. This two-unit sequence has been documented elsewhere in the GOM (e.g., Bacchus, 1993; Keigwin & Pilskaln, 2015; Schnitker et al., 2001) and is sometimes described as the transition from a “distal” glaciomarine to a fully marine sedimentation regime. Figure 4d shows the downcore magnetic susceptibility (MS). The lower, red-brown unit is over six times more magnetically susceptible than the upper, green-gray unit. Elevated MS is due, at least in part, to elevated coarse sediment in the lower unit, including heavy, magnetic minerals, which can indicate glacial or ice-rafted transport in a marine setting. The coarse sediment percent weight (defined as $>63 \mu\text{m}$ diameter) averages 4.3% for the upper unit and 30.6% for the lower unit. The coarse sediment that prevails in the lowermost unit is primarily ice-rafted debris (IRD), which is typically characterized by highly variable composition and grain maturity, resulting from its origin as unsorted glacial debris melting out of icebergs onto the sea floor (Schnitker et al., 2001). Large, rounded clasts are visible in the CT scans of the core's lower unit (Figure S4 in Supporting Information S1). The transition away from ice-rafted material in the core at ~ 14.5 cal ky BP matches the proposed timing that the LIS retreated inland of the marine limit in Maine (Thompson & Borns, 1985), and further supports the interpretation that this change represents a relatively rapid shift in sedimentation regime from glaciomarine sedimentation to fully marine, open-ocean sedimentation.

Notably, the elevated MS that occurs throughout HS-1 shows a distinct two-phased structure, with a period of low MS partitioning the phases at 15.8 ± 0.6 cal ky BP (Figure 4d). If we take MS to primarily reflect concentration of glaciogenic sediment/IRD at our core site, our two phases of IRD deposition align with those identified at a site in the North Atlantic IRD belt (Hole U1308A) by Hodell et al. (2017). They found two separate peaks of IRD (identified by Ca/Sr) within HS-1, the earlier centered at 16.2 cal ky BP with a 2-sigma range of 17.1 to 15.5 cal ky BP, and the later centered at 15.1 cal ky BP with a 2-sigma range of 15.9 to 14.3 cal ky BP (Hodell et al., 2017). The second peak at Hole U1308A is smaller in magnitude than the first, and the lithology of the IRD switches from primarily detrital carbonate and no hematite-stained grains during the older peak to very little detrital carbonate and more hematite-stained grains during the younger peak (Hodell et al., 2017), which has been observed in other records on the Laurentian Fan (Gil et al., 2015) and Iberian Margin (Bard et al., 2000). Hematite-stained quartz grains were qualitatively observed throughout the coarse sediment found in core EN669-GGC-40, during both the early and later MS maxima.

Other regional IRD records, including from just outside the GOM (Zhao et al., 2019), and the Grand Banks area (Bukar et al., 2025), show a similar double-peaked structure during HS-1 (Figure S5 in Supporting Information S1). These records show a late HS-1 peak ~ 0.5 cal ky earlier, but within age model uncertainty, than the peaks of this study and Hodell et al. (2017).

3.3. Bottom Water Temperature Reconstruction

Bottom water temperatures (BWTs) reconstructed from the Mg/Ca (mmol/mol) of benthic foraminiferal calcite yielded temperatures ranging from $-1.3 \pm 1.4^\circ\text{C}$ to $8.8 \pm 1.4^\circ\text{C}$ during deglaciation (Figure 4b). BWTs exhibited consistent values, on average, between the early and later phase of HS-1; an average temperature of $2.4 \pm 1.5^\circ\text{C}$ reconstructed from 17.5 to 14.9 cal ky BP. This is followed by a warming of 2.9°C that rose steadily throughout the Bølling-Allerød Interstadial (B-A). Late B-A foraminifers (13.9–12.9 cal ky BP) yielded an average temperature of $5.3 \pm 1.4^\circ\text{C}$. BWTs were more variable during the YD compared to earlier intervals, but slightly cooler on average ($4.5 \pm 1.4^\circ\text{C}$).

These reconstructed BWTs for the last deglaciation are $2\text{--}3^\circ\text{C}$ warmer on average than other estimates of GOM seawater temperatures during that time ($0\text{--}2^\circ\text{C}$), which are derived from ice-marginal bivalves (Dorion et al., 2001; Kreutz, 1994). However, our BWT reconstruction is consistent with a deeper, more offshore and ice-distal location than those reconstructions from Eastern Maine. Therefore, we consider our benthic foraminiferal records to be more representative of mixed external slope and glacial water sources, whereas shallower, ice-proximal settings may reflect a bias toward glacial melt and terrestrial runoff alone. The warmest BWTs in our reconstruction, occurring in the late B-A and mid-YD, fall within the lower range of modern temperatures

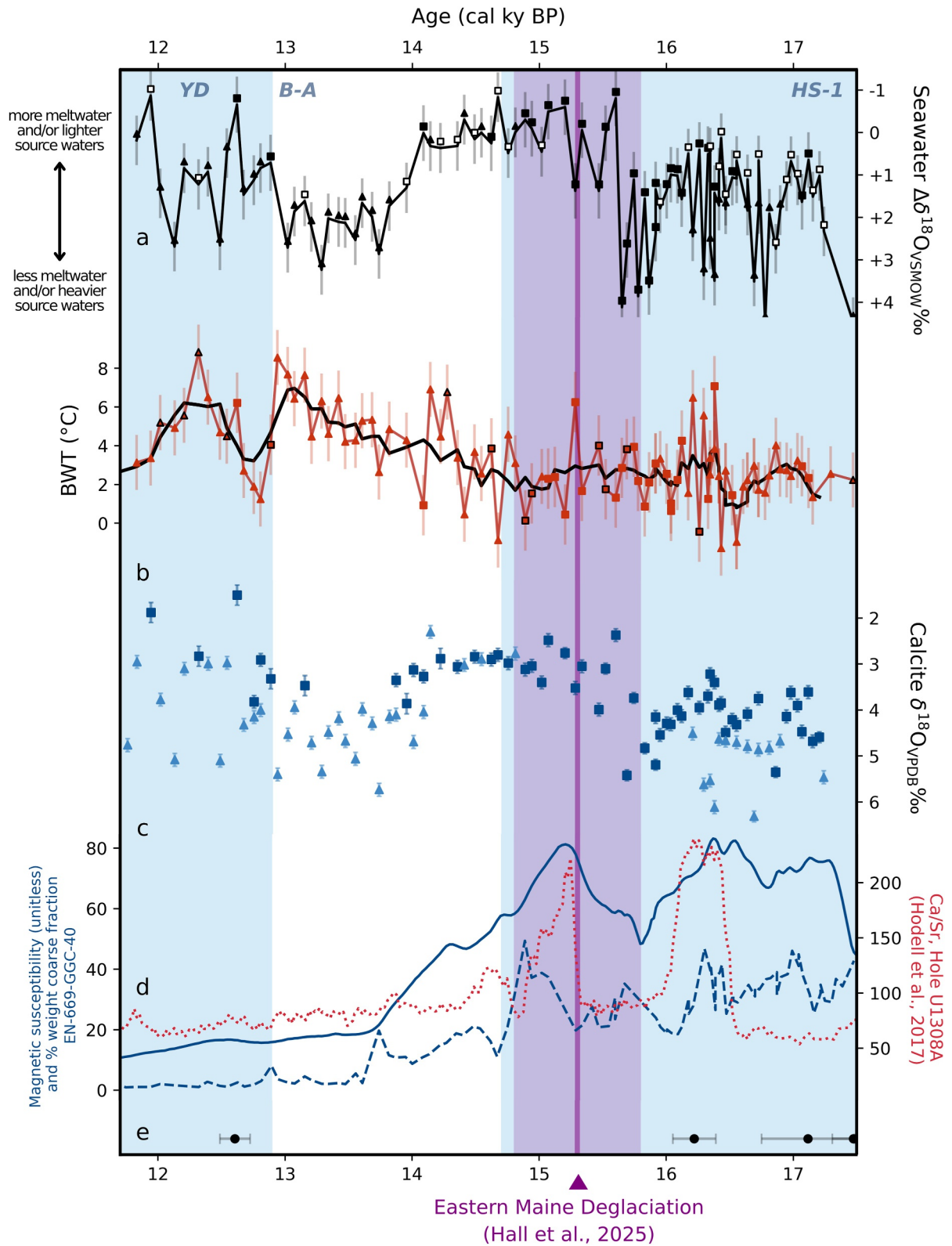


Figure 4.

recorded from a buoy in Jordan Basin, GOM, which varies from 8°C to 10°C seasonally at 250 m depth (Townsend et al., 2023).

Relative sea level (RSL) in eastern Maine was nearly 100 m higher than present immediately following deglaciation but dropped rapidly to a position ~60 m below present before rising more slowly to its current position (Baril et al., 2023; Kelley et al., 2010; Thompson et al., 1989). Therefore, we expect that our site was affected to some extent by changes in global sea level as well as local isostatic adjustment. Considering that our core site in Franklin Basin is located ~300 km outboard of coastal sites from where sea-level reconstructions were derived, and therefore also farther from the thickest ice and greatest crustal depression (Baril et al., 2023), those coastal RSL fluctuations likely represent a maximum estimate of the possible RSL change in Franklin Basin. If our site followed the same RSL pattern as eastern Maine, it may have experienced a maximum water depth of up to 244 m (+40 m from modern RSL) during the highstand in late HS-1 (Baril et al., 2023). Franklin Basin may have experienced a change of up to –80 m between 15 and 12.5 cal ky BP, reaching a minimum depth of 174 m (Baril et al., 2023). The adjacent Georges Bank was subaerially exposed during this period, which is indicated by the presence of terrestrial peat that has been dated to this interval (Emery et al., 1965). Therefore, we consider that a component of BWT warming registered during the B-A may be related to shoaling due to glacio-isostatic rebound.

Nonthermal effects on Mg/Ca, like the Mg/Ca of seawater, should also be considered as potential drivers of the recorded change (e.g., Haynes et al., 2023). Magnesium concentrations in glacial meltwater are typically lower than that of seawater (Brown, 2002), and thus a large influx of meltwater could temporarily drive GOM seawater Mg/Ca to lower values. However, the magnitude of this effect on GOM foraminiferal calcite, and thus paleo-temperature reconstructions, was estimated to be relatively small (~–0.5°C for a meltwater influx of 350 cubic kilometers, smaller than the ~1°C combined uncertainty (2 σ) of analytical and calibration uncertainties).

3.4. Stable Isotope Record and Meltwater Reconstruction

Measured raw benthic foraminiferal calcite $\delta^{18}\text{O}$ values ($\delta^{18}\text{O}_{\text{bf}}$) and calculated change in $\delta^{18}\text{O}$ of seawater ($\delta^{18}\text{O}_{\text{sw}}$) from modern Franklin Basin bottom waters (referred to as $\Delta\delta^{18}\text{O}_{\text{sw}}$, with a modern reference value of 0.156‰ VSMOW, from Houghton & Fairbanks, 2001), are shown in Figures 4a and 4c, respectively. The deglacial GOM is almost always more isotopically enriched throughout the record than its modern deep source waters, which range from 0.4‰ to –1.09‰ VSMOW (Houghton & Fairbanks, 2001; Khatiwala et al., 1999). This is due at least in part to whole-ocean $\delta^{18}\text{O}$ depletion that occurred during the last glacial period due to the preferential sequestration of isotopically light water from the ocean into land-based ice (Shackleton, 1967). GOM $\delta^{18}\text{O}_{\text{bf}}$ values for the older phase of HS-1b (>16.5 cal ky BP) are intermittently in agreement with compilations of LGM-aged $\delta^{18}\text{O}$ from benthic foraminifera around the Atlantic (e.g., Oppo et al., 2018), though slightly more enriched than may be expected for a shallower, more coastal site. In the Labrador Sea, $\delta^{18}\text{O}_{\text{bf}}$ of benthic foraminifera *C. wuellerstorfi* ranges from ~5‰ VPDB during early deglaciation to ~2.5‰ VPDB after deglaciation (Rasmussen et al., 2003), and recent work on LGM-aged *N. labradorica* from Gulf Stream-influenced Cape Hatteras yielded average values around 4‰ VPDB (Wharton et al., 2024). During deglaciation, Franklin Basin likely still received some Labrador-derived slope water and mixed Gulf Stream-derivative Warm Slope Water via the Northeast Channel, as it does today as a function of its geometry and proximity to the Slope Sea (Townsend et al., 2015, 2023). We interpret the occurrence of relatively enriched values in HS-1b as reflecting communication with external slope water masses during deglaciation. Similarly, within HS-1b, there are large-amplitude

Figure 4. Stacked plot of geochemical and sedimentological results: (a) Calculated change in $\delta^{18}\text{O}_{\text{sw}}$ (± 0.6 ‰ VSMOW) from modern bottom water values in Franklin Basin reported by Houghton & Fairbanks (2001), where black squares represent all sample data were from *N. labradorica*, black triangles represent all sample data were from *I. helenae/norcrossi*, and white squares with black outlines represent samples where *N. labradorica* was used for $\delta^{18}\text{O}$ while *I. helenae/norcrossi* was used for Mg/Ca, (b) Calculated Bottom Water Temperature (error $\pm 1.5^\circ\text{C}$), where triangles represent paleotemperatures from *I. helenae/norcrossi*, and squares represent *N. labradorica*. Markers outlined in black indicate the flagged samples described in Section 2.6, and black line represents a 5-point moving average, (c) $\delta^{18}\text{O}_{\text{bf}}$ of *N. labradorica* (dark blue squares) and *I. helenae/norcrossi* (light blue triangles). Error bars reflect two-sigma analytical uncertainties of 0.07–0.11‰ VPDB for $\delta^{18}\text{O}_{\text{bf}}$, (d) Downcore multisensor-track (MST) magnetic susceptibility data for core EN-669-GGC-40 (solid dark blue line, this study), downcore percent dry weight coarse fraction (>63 μm) (dashed dark blue line, this study), and Ca/Sr XRF data from Hole U1308A in the North Atlantic IRD belt (dotted red line), from Hodell et al. (2017), and (e) calibrated radiocarbon ages and associated uncertainties (see Table 1). Vertical light blue bars represent the Younger Dryas (12.9–11.7 cal ky BP) and HS-1 (18.9–14.7 cal ky BP). Purple bar and vertical line at 15.3 ± 0.5 cal ky BP represent the mean interval of eastern Maine deglaciation according to recent surface-exposure age dating (Hall et al., 2025).

excursions (in excess of 2 per mil) between enriched and depleted $\delta^{18}\text{O}$ values. This variability is consistent with intervals of intensified melt interspersed with intervals of melt reduction during early HS-1.

A distinct shift in isotopic enrichment in the GOM record occurs mid-way through HS-1, at the HS-1a/HS-1b boundary. The transition between HS-1b and HS-1a occurs at ~ 16.0 cal ky BP. A period of intense isotopic enrichment, 16 ± 0.6 – 15.7 ± 0.6 cal ky BP, precedes a depletion event, which occurs between 15.7 ± 0.6 and 15.3 ± 0.6 cal ky BP. The average $\Delta\delta^{18}\text{O}_{\text{sw}}$ for the thousand years preceding this transition is 1.5‰ VSMOW; after the transition, it is -0.2‰ VSMOW: a -1.7‰ depletion in a 500-year period. This depletion is followed by a reversal beginning at 14.3 ± 0.6 cal ky BP ($+1.6\text{‰}$ VSMOW for the thousand-year average from 14.3 to 13.3 cal ky BP as compared to the thousand years preceding it). This rapid and sustained isotopic depletion of GOM bottom waters during HS-1 was most likely driven by a large influx of isotopically depleted meltwater from the adjacent LIS, rather than by a shift in the proportion of oceanographic inputs, which were all relatively enriched at the time (Oppo et al., 2018). We interpret the enriched state from 15.9 to 15.7 cal ky BP to be a brief cessation or reduction of meltwater input. This preceded the breakup of ice and influx of meltwater from the LIS beginning around 15.7 cal ky BP. This interpretation is bolstered by the $\delta^{13}\text{C}_{\text{bf}}$ record, which shows a gradual depletion in carbon isotopes throughout both phases of HS-1, but partitioned by a brief increase in enrichment from around 16.4 to 15.9 cal ky BP, aligning with the phase of enriched oxygen isotopes that separates HS-1b and HS-1a. This could indicate that shifts in circulation outside the GOM preceded the largescale thinning and recession of the LIS in Eastern Maine (15.8–14.8 cal ky BP; Hall et al., 2025). Subsequent enrichment in both $\delta^{18}\text{O}_{\text{sw}}$ and $\delta^{13}\text{C}_{\text{bf}}$ during the B-A may indicate an abatement in LIS meltwater production or increasing distance of the site from the ice edge and the increased influence of enriched external slope waters in Franklin Basin bottom water composition.

3.5. Paleoenvironmental Indicators

Both species of interest, *Islandiella helenae/norcrossi* and *Nonionellina labradorica*, were present throughout the core, except where *N. labradorica* is absent in the earliest part of the record, and where *I. helenae/norcrossi* is absent during the first half of HS-1a (Figure 5b). *Islandiella helenae* is typically considered an indicator of sea ice and commonly considered a distal glaciomarine species (Cage et al., 2021; Lloyd, 2006; Rytter et al., 2002; Seidenkrantz et al., 2021). *Nonionellina labradorica* is typically indicative of more “Atlantic” conditions, meaning warmer, more saline, and more productive environments than *I. helenae/norcrossi* (Jernas, 2012; Skirbekk et al., 2016). Most critically, the absence of *N. labradorica* in the earliest part of the record (17.5–17.0 ka) indicates ice-proximal, cold, or brackish conditions. It follows that the absence of *I. helenae/norcrossi* preceding and into the transition to HS-1a, from 16.4 to 15.4 cal ky BP, may suggest a rapid recession of more ice-proximal conditions in Franklin Basin at that time, temporarily destabilizing the preferred habitat of *I. helenae/norcrossi*. The preference of *Nonionellina labradorica* for environments with high organic flux and primary productivity may indicate that in this interval, the GOM was experiencing intense seasonal phytoplankton blooms (Schmidt et al., 2022).

Vertical migration of infaunal benthic foraminifera can cause changes in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of test material independently from changes in seawater chemistry (e.g., Hoogakker et al., 2024). During HS-1a, the $\delta^{18}\text{O}_{\text{bf}}$ of *N. labradorica* was more depleted by about -1‰ VPDB on average than in the millennium preceding it (Figure 4a). If this were due to a migratory shift upward in *N. labradorica* microhabitat responding to limited bottom water oxygenation, the $\delta^{13}\text{C}_{\text{bf}}$ of the same samples should be more enriched (Hoogakker et al., 2024; Jorissen et al., 1995; Van der Zwaan et al., 1999). However, the opposite is observed: during late HS-1, $\delta^{13}\text{C}_{\text{bf}}$ becomes more depleted by nearly -2‰ VPDB (Figure 5a), implying either a shift in bottom water characteristics or a downward migration, assuming a typical $\delta^{13}\text{C}$ porewater curve (e.g., Hoogakker et al., 2024). It may be interpreted in tandem with the foraminiferal presence/absence results to indicate that during HS-1a, the GOM may have been experiencing intense seasonal productivity events, perhaps due to warming summers (e.g., Denton et al., 2022) affecting the sea surface temperature combined with influx of glacial meltwater, which is associated with driving large phytoplankton blooms in modern environments (e.g., Arrigo et al., 2017). These large blooms would drive a high flux of isotopically light organic material to the seafloor and, upon respiration, the porewater in which the infaunal *N. labradorica* lived (Hesse et al., 2014; Hoogakker et al., 2024).

Another possible explanation is that increased delivery of terrigenous material to the bottom waters of the GOM via turbid meltwater or ice-rafting resulted in increased delivery of isotopically depleted terrigenous organic

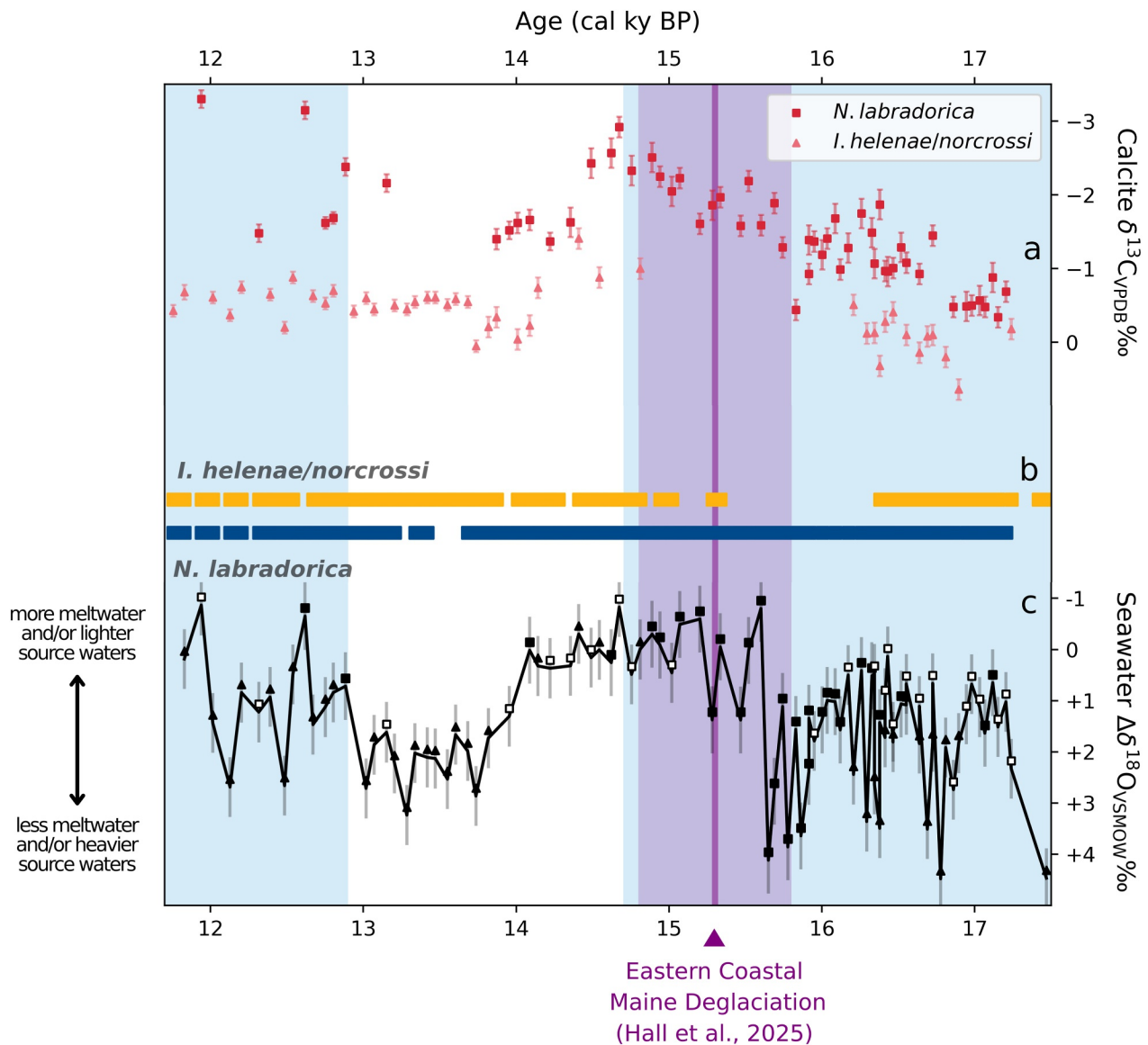


Figure 5. (a) $\delta^{13}\text{C}_{\text{br}}$ for *N. labradorica* and *I. helenae/norcrossi*. Error bars represent two sigma analytical error of $2 \times 0.04\text{--}0.06\text{‰}$ VPDB. (b) Presence/absence intervals for *N. labradorica* (dark blue) and *I. helenae/norcrossi* (gold), where gaps in the line indicate periods of absence (c) calculated $\Delta\delta^{18}\text{O}_{\text{sw}}$, as described in Figure 4 caption, where black squares represent all sample data were from *N. labradorica*, black triangles represent all sample data were from *I. helenae/norcrossi*, and white squares with black outlines represent samples where *N. labradorica* was used for $\delta^{18}\text{O}$ while *I. helenae/norcrossi* was used for Mg/Ca. Vertical light blue bars represent the Younger Dryas and HS-1, and purple bar represents the interval of eastern Maine deglaciation from Hall et al. (2025).

matter. However, this explanation fails to justify the steady depletion that occurs over thousands of years. If this were the primary driver of the $\delta^{13}\text{C}_{\text{br}}$ depletion, one would expect the record to show more sudden, episodic events associated with large pulses of terrigenous material delivery.

Changes in RSL may also affect the $\delta^{18}\text{O}$ record. Today, the most isotopically enriched water in the GOM is a mixture of Gulf-Stream-influenced slope waters that enters at depth via the Northeast Channel (Houghton & Fairbanks, 2001). Assuming this basic regime applies throughout the record, a shoaling or lowstand of RSL could result in less isotopically enriched Atlantic waters entering GOM basins due to a shallower Northeast Channel. This would lead to increased influence of shelf and coastal water masses, which are generally more isotopically depleted (e.g., Whitney et al., 2020). It is reasonable to assume that coastal waters would also have been more depleted during deglaciation due to the persistent influence of high-latitude runoff and (intermittently) local/regional meltwater inputs. During highstands, increased influence of enriched Atlantic waters could have

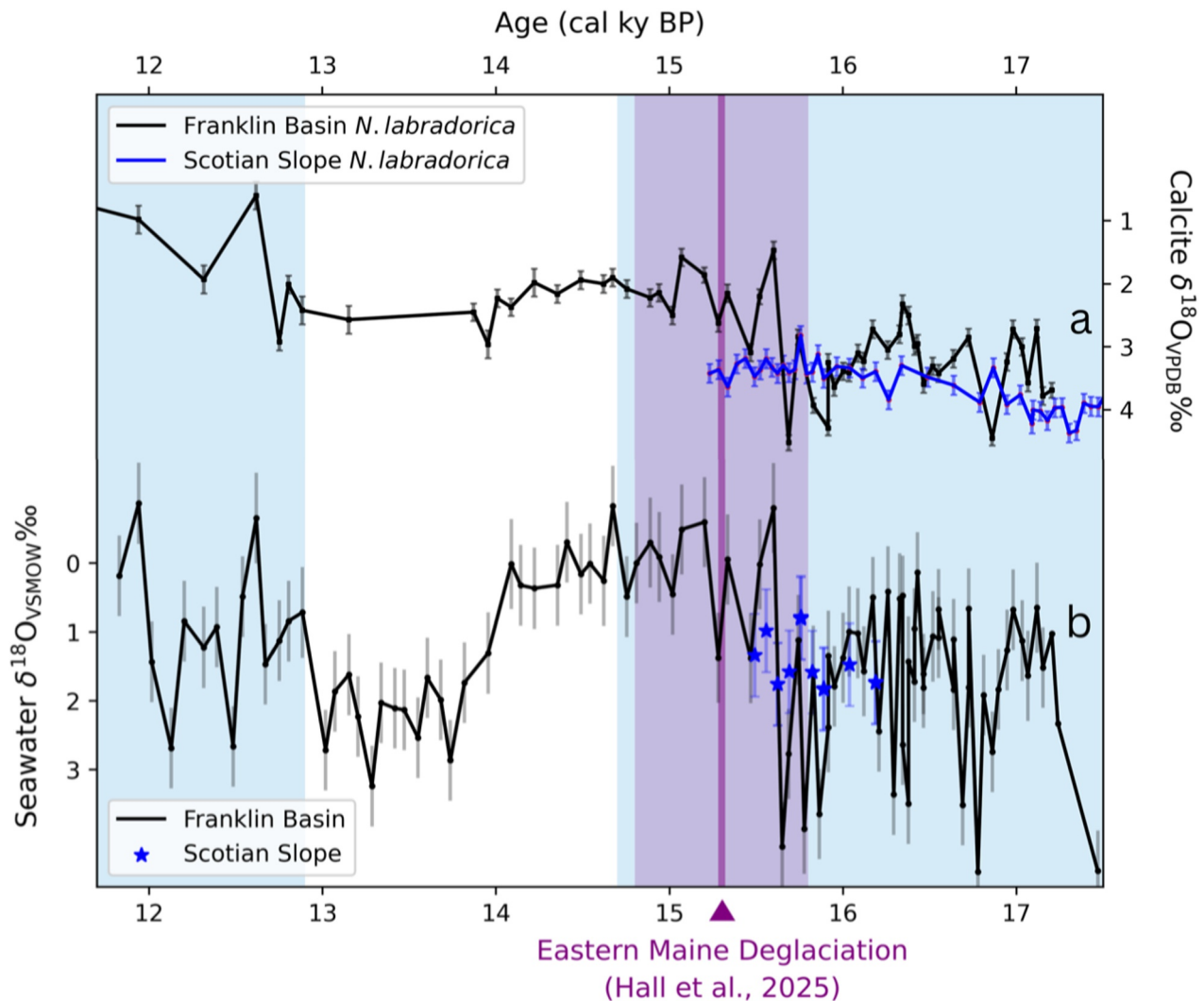


Figure 6. (a) Franklin Basin (EN-669-GGC-40) $\delta^{18}\text{O}_{\text{calcite}}$ compared to Scotian Slope core EN-540-GGC-9. Franklin Basin and Scotian Slope values reflect only *N. labradorica*, an infaunal benthic species. (b) GOM $\delta^{18}\text{O}_{\text{sw}}$ compared to Scotian Slope $\delta^{18}\text{O}_{\text{sw}}$ from core EN-540-GGC-9. Purple bar and vertical line at 15.3 ± 0.5 cal ky BP represents the mean interval of eastern Maine deglaciation according to recent surface exposure age dating (Hall et al., 2025).

enriched GOM waters. However, the entirety of our late HS-1 excursion is considered to have occurred under persistent “highstand” conditions (Kelley et al., 2010), and the large shoaling of RSL that occurs from 14 to 12 cal ky BP is accompanied by isotopic enrichment in our record, contrary to a depletion signal that would be expected if RSL were an important control on the isotopic record. Therefore, we infer that although RSL and its effects on basin geometry could have exerted a minor influence on our temperature record (Section 3.2), it is unlikely to be the primary driver of the isotopic excursions we observed.

3.6. Comparison to Scotian Slope Records

Foraminiferal $\delta^{18}\text{O}_{\text{bf}}$ from *N. labradorica* on the Scotian Slope (core EN-540 GGC-9, Table S4 in Supporting Information S1) shows a gradual -1‰ VPDB depletion during the entirety of HS-1 (Figure 6). This is a smaller shift than is documented in GOM foraminiferal calcite, and lacks the distinct two-phased structure that appears in the GOM record. When paired with a small number of Mg/Ca-derived BWTs, they yield an average $\delta^{18}\text{O}_{\text{sw}}$ in agreement with the GOM at that time.

If the observed isotopic depletion in the GOM were due primarily to transported Arctic meltwater, a record closer to those source waters should show a stronger depletion signal than a more distal, mixed signal, resembling the

magnitude of GOM $\delta^{18}\text{O}_{\text{sw}}$ depletion. However, the isotopic depletion that occurs within the GOM is stronger than immediately outside its main channel, through which Arctic-influenced slope waters arrive (Townsend et al., 2015). This suggests that although the isotopic depletion occurring in the GOM at this time may be amplified by its communication with higher latitude sourced waters, like Labrador Slope Water, external inputs cannot be responsible for the full magnitude of the observed GOM depletion signal. The isotopic depletion from distal meltwater sources becomes progressively muted via mixing with water masses more enriched in $\delta^{18}\text{O}$. Therefore, we argue that a portion (perhaps a majority) of the meltwater driving seawater $\delta^{18}\text{O}$ depletion in Franklin Basin must have been locally sourced from the LIS in the GOM drainage network.

Similarly, we suggest that isotopic depletion on the Scotian Shelf could indicate meltwater generated at the same time and by the same forcing mechanism as the signal in the GOM. However, this signal may be muted because of the increased water depth and distance away from the coast, allowing any meltwater discharged from land more time to mix with seawater. It may also represent meltwater from higher latitude ice sources that was routed southward along the Scotian Slope.

3.7. Comparison to Glacial Records and Climatic Implications

Surface-exposure dating of glacial erratics with ^{10}Be , together with ^{14}C dating of marine macrofossils in glaciomarine mud, indicate that deglaciation of the outermost islands of Maine was underway as early as 17.2 ± 0.2 cal ky BP, according to surface-exposure ages on Isle Au Haut. The eastern Maine mainland deglaciated rapidly between 15.8 and 14.8 cal ky BP (Borns et al., 2004; Hall et al., 2017, 2025; Thompson & Borns, 1985). This latter phase of deglaciation coincides with HS-1a, the interval of greatest isotopic depletion in the GOM. The interval 15.8–14.8 cal ky BP was, on average, more depleted by $0.6 \pm 0.6\text{‰}$ VSMOW than HS-1b. Therefore, our marine record supports the finding that significant LIS thinning, recession, and resulting release of meltwater occurred during HS-1. The dominance of *N. labradorica* over *I. helenae/norcrossi* during this same interval, paired with the depletion of $\delta^{13}\text{C}_{\text{br}}$, is also consistent with HS-1 warming of summer surface layers that respond most to atmospheric heating, as well as being highly productive during HS-1a (e.g., Denton et al., 2022).

Regionally, our finding that deglaciation proceeded and intensified meltwater release occurred during HS-1 is in agreement with the New England Varve Chronology of the Connecticut Valley (Ridge et al., 2012). Although meltwater from ice in the Connecticut Valley would not have drained into the GOM, it indicates that the southeastern margin of the LIS was actively deglaciating during HS-1 and out of phase with Greenland ice-core records.

Due to isostatic depression, portions of the ice edge were in direct contact with GOM/marine water until it retreated above sea level, likely around 14.5 cal ky BP (Borns et al., 2004; Dorion et al., 2001; Hall et al., 2017). Around this same time, Franklin Basin $\delta^{18}\text{O}_{\text{sw}}$ changes from becoming increasingly depleted to becoming increasingly enriched, and MS falls below all earlier levels. This change may be in response to the removal of ice from the marine system, and the increasing distance of the ice margin from the GOM and Franklin Basin. Overall, our geochemical record is also in good agreement with local terrestrial records of deglaciation in Maine.

We see no clear indication that the GOM experienced a mid to late B-A meltwater event akin to that suggested for the Hudson Valley in Donnelly et al. (2005). Meltwater routed through the Hudson Valley would discharge south of the GOM, and likely become too well-mixed with shelf and slope water masses during transit to be detectable within the GOM. Moreover, ice reconstructions (e.g., Hall et al., 2025) show that very little ice remained in areas likely to feed into the GOM by mid to late B-A. Finally, Franklin Basin $\delta^{18}\text{O}_{\text{sw}}$ signal shows a trend toward more enriched values during B-A time, consistent with reduced meltwater input into the GOM.

Our findings complement those of Toucanne et al. (2015), who reconstructed Channel River runoff during the deglaciation and found that it was a reliable indicator of European Ice Sheet margin evolution/recession (Denton et al., 2022; Toucanne et al., 2015). They found evidence for increased runoff and glacial melt during early HS-1, from 18.2 ± 0.2 to 16.7 ± 0.2 cal ky BP. However, the impacts of post-glacial relative sea-level rise near the Channel River truncated the meltwater discharge reconstruction at ~ 17 cal ky BP, leaving the later part of HS-1 unconstrained (Ménnot et al., 2006; Toucanne et al., 2015). Taken together, our records are consistent with quickening deglaciation occurring in the Northern Hemisphere throughout HS-1.

Our meltwater reconstruction is at odds with the interpretation that HS-1 was a year-round cold excursion that affected summer and winter temperatures equally in the Northern Hemisphere. Rather, our interpretation of the

GOM oxygen isotope record lends support to the ideas of Denton et al. (2005, 2022), which suggest that stadials were marked by amplified seasonality around the Northern Atlantic, driven by summer warming consistent with Southern Hemisphere climate records. Meltwater plays a central role in the mechanics of the Seasonality Hypothesis by promoting extensive winter sea ice (Denton et al., 2005). Our findings indicate a substantial release of meltwater into the Gulf of Maine during HS-1, most likely from intensified surface ablation on the LIS due to atmospheric warming. This warming is not clearly expressed in the Mg/Ca BWT record, which primarily reflects seafloor conditions rather than overlying atmospheric temperatures, and is therefore influenced by changes in circulation and RSL as well as atmospheric temperature.

4. Conclusions

1. A basal radiocarbon age of 17.5 ± 0.2 cal ky BP provides a minimum-limiting age for the deglaciation of Franklin Basin. However, deglaciation was likely underway prior to this age as the core (EN-669 GGC-40) did not reach the bottom of the deglacial sequence seen in seismic stratigraphy (Figure S4 in Supporting Information S1).
2. Paired $\delta^{18}\text{O}_{\text{bf}}$ and Mg/Ca-derived bottom water temperature (BWT) estimates from benthic foraminifera in the Gulf of Maine (GOM) were used to demonstrate changes in $\delta^{18}\text{O}_{\text{sw}}$ during the most recent deglaciation (17.5–11.7 cal ky BP). Sedimentological evidence paired with stable isotope records show a distinct two-phased structure to HS-1 in the GOM, with a geochemical and sedimentological shift indicating the beginning of a local ice recession event occurring ~ 15.7 cal ky BP, matching terrestrial estimates from eastern Maine (Hall et al., 2025).
3. HS-1b (17.5–15.9 cal ky BP) is characterized by high $\delta^{18}\text{O}_{\text{sw}}$ variability but little average shift in $\delta^{18}\text{O}_{\text{sw}}$, multiple peaks in MS, and a progressive $\delta^{13}\text{C}_{\text{bf}}$ depletion. We interpret this as episodic meltwater input into the GOM, interrupted by incursions of relatively $\delta^{18}\text{O}$ -enriched Atlantic slope water.
4. HS-1a (beginning at 15.9 ± 0.6 cal ky BP, and lasting until the onset of the B-A), is characterized by an isotopic depletion of -1.7‰ VSMOW in $\delta^{18}\text{O}_{\text{sw}}$, accompanied by depletion of $\delta^{13}\text{C}_{\text{bf}}$. We interpret this shift to indicate the onset of sustained input of meltwater into the GOM, unlike the variability of HS-1b. The HS-1a depletion trend is reversed in the early B-A, and $\delta^{18}\text{O}_{\text{sw}}$ values partially return HS-1b values, in tandem with a warming of BWTs. We interpret this as a waning of meltwater input into the GOM during the B-A likely because much of the ice in Maine was gone by this time.
5. The Franklin Basin meltwater reconstruction is consistent with a rapid recession of the southeastern margin of the LIS and terrestrial deglaciation of eastern Maine centered on $\sim 15.3 \pm 0.5$ cal ky BP (Hall et al., 2025). We also find broad agreement between the GOM meltwater reconstruction and comparable records from around the North Atlantic (e.g., Rasmussen et al., 2016; Toucanne et al., 2015) that show an increase in meltwater flux during HS-1.
6. Our record supports the hypothesis that North Atlantic stadials were highly seasonal episodes (Denton et al., 2005, 2022) characterized by warm summers and melting ice, juxtaposed against cold winters with extensive sea-ice cover.

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Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

Data are archived at the National Centers for Environmental Information (Herlinger et al., 2026). Plots for this paper were made using Matplotlib version 3.10.8 (Hunter, 2007). Maps were made using Cartopy version 0.24.0 (<http://scitools.org.uk/cartopy>). Code used to create figures is reposted at Zenodo (Herlinger, 2026).

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